Investigating the effect of a wave-dependent momentum flux in a process oriented ocean model

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New expressions of the drag coefficient were developed using measurements from the Östergarnsholm site in the Baltic Sea. The drag coefficient was significantly lower in the presence of waves travelling faster than the wind (swell). The expressions were implemented in an oceanographic process-oriented model in a 45-year simulation. Since no wave information was included we did an analysis of the potential impact of swell on an ocean model. Current velocity and surface stress were significantly altered during periods with low wind speed but the temperature and the mixing depth in the ocean were not significantly changed. The implementation of the swell effect in a process oriented ocean model is thus of limited importance. There is, however, an indication that for studies of current velocity it is crucial to have a correct description of the drag coefficient.

Introduction

When modelling the atmosphere as well as the ocean it is of crucial importance to correctly describe the boundary conditions. The atmosphere–ocean boundary is an important source of turbulence in the atmosphere as well as in the ocean and there is a significant exchange of momentum, heat and moisture. The turbulence in the ocean is mainly generated by surface waves and current shear governed by the surface stress, and controls the depth of the ocean mixed layer together with the heat fluxes.

Traditionally the atmospheric surface layer is described by the Monin-Obukhov Similarity Theory (MOST). It assumes stationary and homogeneous conditions and a solid surface. Numerous field measurements have confirmed MOST over land (e.g. Businger et al. 1971, Haugen et al. 1971, Högström 1990) as well as over the ocean (Smedman et al. 1994, Guo Larsén et al. 2003). However, several investigations in the marine surface layer have found important deviations in the applicability of MOST (Smedman et al. 1994, Smedman et al. 1999, Rutgersson et al. 2001). The surface is not solid but changing as a response to the atmospheric forcing. In the wave boundary layer (WBL) the waves directly influence the atmospheric turbulence. Above the viscous sub-layer the total wind stress or momentum flux \( \tau \) can be expressed as:

\[
\tau = \tau_i + \tau_w = \rho_a u_*^2
\]

Here \( \tau_i \) is the turbulent part, \( \tau_w \) the wave induced part and \( u_* = \left[ (\overline{u' w'})^2 + (\overline{v' w'})^2 \right]^{1/2} \) is the measured friction velocity, with \( u' \), \( v' \) and \( w' \) being...
the longitudinal, lateral and vertical wind fluctuations, respectively (bar denoting the Reynolds average); and \( \rho_a \) is the air density.

The wind stress is generally described by a bulk formulation, where the friction velocity is related to the wind speed by a bulk coefficient according to:

\[
\frac{\tau}{\rho_a} = u_r^2 = C_D U_{10}^2
\]

(2)

where \( C_D \) is the drag coefficient and \( U_{10} \) the mean wind speed at 10 m. The drag coefficient depends on the sea surface roughness and the atmospheric stability. The neutral drag coefficient \( C_{DN} \), valid for neutral stratification, is a function only of the roughness parameter \( z_0 \). Smoother surface yields a smaller \( C_{DN} \). Several parameterisation studies arrive at a wind speed dependent \( C_{DN} \) (e.g. Large and Pond 1981, Jansen et al. 1987, Fairall et al. 2003). An alternative way to calculate \( z_0 \) and thus \( C_{DN} \) is to use the Charnock equation

\[
z_0 = \alpha u_r^2 \frac{\rho_a}{\rho}
\]

(Charnock 1955), where \( \alpha \) is the Charnock parameter and \( g \) is the gravitational acceleration. Some studies show the Charnock parameter depending on the state of the waves (e.g. Drennan et al. 2003).

The state of the waves can be described by wave age \( \frac{c_p}{U_{10}} \), where \( c_p \) is the phase speed of the dominant waves. The definition of wave age sometimes includes a factor \( 1/\cos \beta \), where \( \beta \) is the angle between the local wind and the waves. This would give unrealistic results since, for swell waves, this factor can be very large. It would imply increasing atmospheric effects with increasing values of \( 1/\cos \beta \), predictions which disagree with our data for the approximate range \( \beta < 90^\circ \). As a result we choose not to include the \( \cos \beta \) term here.

Approximating wind stress measurements during swell the scatter of \( C_{DN} \) increases significantly. When the wind direction follows the wind direction (following swell) the drag coefficient is reduced (Guo Larsén et al. 2003 and Drennan et al. 1999a). When the wind is cross or against the wind direction (cross/counter swell) \( C_{DN} \) has been seen to increase (Guo Larsén et al. 2003, Drennan et al. 1999a). This can be explained by \( \tau_u \), which gives an upward contribution of momentum during following swell, that is, wave energy is transported from the waves to the wind. The physical mechanism is, however, not yet fully understood. The turbulent contribution, \( \tau_t \), always gives a negative momentum transport and for further increase of the wave age there can be a sign reversal of \( \tau_t + \tau_u \), i.e. a net positive upward momentum transport (Smedman et al. 1999, Grachev and Fairall 2001). Other deviations from MOST observed in the presence of swell include a low level wind maximum (Holland et al. 1981, Donelan 1990, Rutgersson and Fairall 2001). In Guo Larsén et al. (2004) the non-dimensional wind gradient

\[
\phi_m = \frac{\partial U}{\partial z} \frac{\kappa z}{u_r},
\]

is modified for following swell. \( z \) is height and \( \kappa = 0.4 \) is von Karman’s constant (Högström 1996).

During swell the waves are not generated by the local wind. This is often the case during a decaying storm or when waves are transported from a distant storm. Since the swell can travel long distances without dissipating it has been assumed that there is no transport of energy from the waves to the wind. However, during swell the WBL is significantly deeper and the structure in the whole boundary layer could be affected (Smedman et al. 1994, Smedman et al. 1999).

Effects of swell on the turbulence structure in the overlying airflow are supported by both DNS (direct numerical simulation) (Sullivan et al. 2000, Rutgersson and Sullivan 2005) and LES (large eddy simulation) (Sullivan et al. 2008).

One of the effects of swell is seen in the drag coefficient. In most investigations analysing measurements during swell the scatter of \( C_{DN} \) is modified for following swell. \( \phi_m = \frac{\partial U}{\partial z} \frac{\kappa z}{u_r} \)

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In this study, new empirical parameterisations are developed for the drag coefficient, taking into account the state of the waves. The data are from a measurement site in the Baltic Sea. A filter-
ing method is developed to reduce the scatter in measured data seemingly connected to irregular low frequency variations. The parameterisations are then implemented and tested in the process-oriented ocean model PROBE-Baltic in a sensitivity study. Since there is no wave information in the model some assumptions are made and we restrict ourselves to compare two runs with different sea states: one run where the expression for growing sea is used, and the other where following swell is assumed for low wind speed. The effect of breaking waves is also implemented in the ocean model for a more reliable description of the ocean surface layer turbulence and surface currents.

**Theory**

By the MOST the wind speed profile is related to the friction velocity according to:

\[
U = \frac{u_*}{\kappa} \left[ \ln \frac{z}{z_0} - \psi_m \right]
\]

(3)

where \(\psi_m\) is the integrated non-dimensional wind gradient (Paulson 1970). The non-dimensional wind gradient \(\phi_m\) is an empirical function of the stability parameter \(\zeta\):

\[
\zeta = \frac{z}{L} = \frac{-g\kappa w'\theta}{u_* T_0}
\]

where \(L\) is the Obukhov length and \(w'\theta\) the buoyancy flux (kinematic flux of virtual temperature) and \(T_0\) is the mean temperature in the surface layer.

Guo Larsén et al. (2004) found that \(\phi_m\) is a function of the wave state of the waves. For growing sea \((c_p/U_{10} < 0.8)\) and unstable conditions:

\[
\phi_m = (1 - 19\zeta)^{-1/4}
\]

(4)

which agrees with the recommended expression by Högström (1996) over land. For following swell \((c_p/U_{10} < 1.2)\):

\[
\phi_m = 1 - (-3\zeta)^{-1/2} \quad -1 < \zeta < 0
\]

\[
\phi_m = -0.73 \quad \zeta < -1
\]

In the interval \(0.8 < c_p/U_{10} < 1.2\) we usually have a mixture of wave conditions. There are not yet any conclusive results for the \(\phi_m\) in this wave regime and we here use the \(\phi_m\) valid for growing sea.

Using Eqs. 2 and 3 we can express the roughness length as:

\[
z_0 = 10 \exp \left( \frac{\kappa}{\sqrt{C_D}} - \psi_m \right)
\]

(6)

which can be used to calculate the neutral drag coefficient:

\[
C_D = \left( \frac{10}{\ln \frac{z_0}{z_0}} \right)^2
\]

(7)

The drag coefficient \(C_D\) is calculated from Eqs. 6 and 7 with:

\[
C_{DN} = \left( \frac{\kappa}{\sqrt{C_D} - \psi_m} \right)^2
\]

**Measurements**

**Site and instrumentation**

The measurements used in this investigation were taken from the field station at the island of Östergarnsholm. Östergarnsholm is situated about 4 km east of Gotland in the Baltic Sea (at A in Fig. 1). The island has very few trees and is very flat. On the southernmost tip a 30-m tower is erected at about 1 m above the mean sea level. For the wind directions 80–220° the data have been shown to represent open ocean conditions. In a previous analysis of the disturbance in the wave field in combination with a footprint analysis Smedman et al. (1999) showed a very small impact of the limited water depth near the shore.

Mean profile data were taken from slow response instruments measuring wind speed and direction and temperature at 5 levels (8, 12.5, 15, 21 and 29 m above mean sea level) and humidity (8 m). Turbulence data were calculated from sonic anemometers Solent 1012R2 at three heights (10, 18 and 26 m) measuring virtual temperature and the three wind components and sampled at 20 Hz. To get 10 m values of wind and temperature the average of the 8 and 12.5 m values were taken.
A wave-rider buoy (run and owned by the Finnish Institute for Marine Research) is moored 3.5 km in the direction of 115° from the tower, giving the wave field and sea surface temperature (SST). For a more detailed description of the site and instrumentation, see Smedman et al. (1999).

Data were selected using the following criteria: wind from the sector 80–220°, wind speed above 2 m s\(^{-1}\), only upward sensible flux (unstable stratification) and with a dominant wave field from the sector 40–210°. Turbulent fluxes were calculated using the eddy-correlation method and all turbulence statistics were subject to a 10-min running average to remove trends. The data included hourly averages from the period 1995–2004.

**Filtering method**

The turbulent fluctuations can be illustrated by a spectrum achieved by a Fourier transform on the correlation function. It represents the energy distribution of the components of turbulence in frequency space \( n \). Corresponding fluxes can be described by the cospectrum (real part of the transform). The \( \text{uw-cospectrum} \ C_{\text{uw}} \) for instance, can be integrated between a selected cut-off frequency \( n_{\text{lim}} \) and the Nyquist frequency 10 Hz to get the longitudinal part of the kinematic momentum flux.

\[
\overline{u'w'} = \int_{n_{\text{lim}}}^{10} nC_{\text{uw}}dn \tag{8}
\]

The 10-min running average applied to the turbulence data corresponds to a high-pass filter cut-off frequency \( n_{\text{lim}} = 0.0017 \) Hz.

The phase shift between the \( u \) and \( w \) components is described by the phase angle \( \phi_{\text{uw}} \) as function of frequency.

\[
\phi_{\text{uw}}(n) = \tan^{-1} \frac{Q_{\text{uw}}(n)}{C_{\text{uw}}(n)}
\]
$Q_{uw}$ is the quadrature spectrum, i.e. the imaginary part of the Fourier transform (see Lumley and Panofsky 1964). For 0° the components are in phase and 180° there is an anti-phase. As the momentum flux is negative, i.e. downwards, we should find the phase angle to be close to 180°. Over land and for growing sea the momentum flux this is generally the case.

One example of a spectrum from 6 Nov. 1997 (Fig. 2a) is shown. The momentum flux was constituted of turbulent fluctuations roughly in the range 0.005 Hz to 10 Hz. The corresponding phase angle between the $u$ and $w$ components was also near 180° in this range (Fig. 2c). For lower frequencies the spectrum values were close to zero. The small fluctuations at the low frequencies are irregular fluctuations and should not be included when analysing the turbulence. At the transition frequency the phase angle was also seen to change (Fig. 2c). In this case it is rather clear that the cut-off frequency $n_{lim} = 0.005$ Hz gives a good estimate of the momentum flux. However, 0.0017 Hz (10 min) will work as well as the contribution is small between 0.0017 and 0.005 Hz.

When $u_*$ is small, which is the case at low wind speed and during swell, irregular low frequency variations become prominent and influence the cospectra at higher frequencies than 0.0017 Hz (Fig. 2b shows a typical example). These fluctuations do not contribute to the turbulent flux and should not be included when analysing turbulence. These patterns contribute to the large scatter that often is seen in the drag coefficient estimated from measurements during swell. Further, a systematic increase of phase angle with decreasing frequency is seen for higher frequencies (Fig 2d). This type of behaviour is present in many swell cospectra and was identified in Smedman et al. (1999) in their analysed period with swell present. This is possibly connected to an indirect swell contribution on the atmospheric turbulence in a broad spectrum band below the frequency of the swell itself.
Drennan et al. (1999b) also found a swell effect on momentum spectra for frequencies lower than that of the dominant swell waves.

Since we wanted the total surface stress during swell we needed to differentiate between the irregular low frequency variations and the turbulence fluctuations generating stress during swell conditions. This can be done using an Ogive-curve method, by integrating the cospectrum from high to low frequencies (e.g., Guo Larsén et al. 2003). In the method presented here the phase angle acted as a steering separator instead.

New cut-off frequencies (replacing the 10-min cut-off) were used if: The phase angle $\phi_{uw}$ (for each frequency in the range $0.0017 < n < 0.05$ Hz) differed more from $180^\circ$ than the limiting phase angle ($\phi_{lim} = 30 + 3000n$, dashed-dotted curves in Fig. 2c and d). We chose $\phi_{lim}$ to be rather large for $n > 0.01$ Hz since this gives the possibility to include the wave effects corresponding to the systematic deviation of the phase angle, mentioned above. The irregular low frequency motions can be filtered out at lower frequencies as $\phi_{lim}$ is smaller in that region. The cut-off frequency $n_{lim}$ (dashed lines in Fig. 2) was found where $|\phi_{uw} - 180^\circ| > \phi_{lim}$, going from higher $n$. The same procedure was used for $C_{vw}$ using the frequency limits determined for $C_{uw}$.

The method was applied to all swell data (222 hours). Outliers were removed for non-stationary conditions and if there where single spikes for the phase angle or in spectra, or if the momentum flux was upward (only one hour). Upward momentum flux is an indication of swell, but present theory cannot include positive momentum flux as $u$, then becomes imaginary. Further, only data with angle $\beta < 90^\circ$ between the dominant wave and wind direction were considered, due to limited amount of data with larger angles. The resulting data with swell were 183 hours. The scatter of the drag coefficient $C_D$ (not reduced to neutral) decreased significantly using the filtering method, in particular for $U_{10} < 4$ m s$^{-1}$ (Fig. 3). The mean value of $C_D$ was reduced as well, most significantly for the lower wind speeds.

This method to reduce the irregular low frequency variations gives a similar result to using the Ogive curve, with the advantage that relatively low frequencies affected by the swell can be included. Applying the method for data during growing/mixed sea gave no difference since the irregular variations do not dominate in the same way during these situations (e.g., Fig. 2a and c).

**Model**

PROBE-Baltic is a process oriented ocean model developed for the Baltic Sea. It is described in detail in Omstedt and Nyberg (1996) and Omstedt and Axell (2003). In the model the Baltic Sea is divided into 13 sub-basins (Fig. 1). Each sub-basin is coupled with surrounding sub-basins.
through in- and outflows. The calculated properties, temperature, salinity and current velocity, are horizontally averaged over each basin. The basins are resolved in the vertical by an expanding grid ranging from 1 m at the boundaries up to 2–10 m, depending on the depth of the basin. Description of the turbulence and the implementation of a wave breaking effect are described in the Appendix. This implementation had a significant impact on both surface currents and the temperature profile.

Input to the model is meteorological data and river runoff together with water-levels in the North Sea. The meteorological forcing, including temperature, geostrophic wind, cloud cover and relative humidity, is from ERA40 data (1958–2001) and the SMHI (1 × 1°) gridded data set (2002–2004). See Omstedt et al. (2005) for an evaluation of the difference between the two data sets. The focus in this study is on the eastern Gotland Basin, but the other basins and the Baltic Sea as whole are expected to follow the same pattern. The model was run for the period November 1958 until December 2004.

Results

Parameterisation of $C_{DN}$

For the parameterisation of the drag coefficient $C_{DN}$ we focused on the unstable data since influence of swell is not clear for stable data. According to Rutgersson et al. (2001) the stability parameter $\zeta$ ceases to be a relevant scaling parameter during swell and no systematic variation of $C_{DN}$ with stability can be seen. For the swell data we limited $\zeta$ to be equal to $-0.4$ even when the measured $\zeta$ was smaller than $-0.4$. The neutral drag coefficient was calculated using Eqs. 2, 6 and 7. Equation 4 was used to calculate $f_m$ for both growing and mixed sea ($c_p/\bar{U}_{10} < 1.2$) and Eq. 5 was used to calculate $f_m$ for all swell data.

The available number of hours with growing or mixed sea conditions was 535. For the growing sea data at $\bar{U}_{10} > 8 \text{ m s}^{-1}$ there was an indication of higher $C_{DN}$ for the youngest sea as compared with that for the older growing sea data (Fig. 4). This agreed with Drennan et al. (2003) who showed higher values for younger seas. The mixed sea, however, gave even higher $C_{DN}$. Because of the weak wave age dependence at growing/mixed sea $C_{DN}$ was chosen to be a function only of wind speed (Fig. 5):

\[
C_{DN} = \max(1.03, 0.60 + 0.060\bar{U}_{10}) \times 10^{-3} \quad (9)
\]

The correlation coefficient $r$ for the parameterisation was 0.96.

The relatively limited wave age dependence of $C_{DN}$ was explained in Smedman et al. (2003), where it was shown that $C_{DN}$ is not only governed
by wave age, but by two parameters representing the wave state, wave age and $E_1/E_2$. There, $E_1$ and $E_2$ were defined as the energy of long waves and short waves, respectively. Thus the relative energy of the long (fast) and short (slow) waves is an important parameter when we want to investigate wave effects in combination with wave age for growing/mixed sea. When lacking information about this parameter we suggest a simple wind speed dependence.

For the swell data with $\beta < 90^\circ$ no variations of $C_{DN}$ with angle could be seen. There is a tendency of higher $C_{DN}$ for $\beta > 90^\circ$, but we have too few data in this regime to draw any further conclusions about the angle dependence of $C_{DN}$. We thus use the term following swell for $\beta < 90^\circ$ and developed a parameterisation only for that sea state. The neutral drag coefficient during following swell can be described as a function of wind speed only (Fig. 5) according to:

$$C_{DN} = (0.34 + 0.045 U_{10}) \times 10^{-3}$$

(10)

The correlation coefficient $r$ for this relation to the data was 0.83. As there were few data for $U_{10} < 4.5$ m s$^{-1}$ it should be remembered that the equation is rather tentative for these conditions. Guo Larsén et al. (2003) used partly the same data set (using data from the period 1995–1998) and their result for swell cases is close to our result.

Model results

In the model the equations for the neutral drag coefficient developed from the data in the previous section (Eq. 9 for growing/mixed sea and Eq. 10 for swell) were used together with Eqs. 4 and 5, respectively for the $\phi_m$-functions. For all stable conditions Eq. 9 was used. Since no wave information was available in the model we did a sensitivity test of the swell effects. We assumed that swell is present for wind speeds below 8 m s$^{-1}$ and that the wave field was homogeneous in each basin and aligned with the wind. There was no detectable effect in the model of the singularity at 8 m s$^{-1}$, separating growing/mixed sea from swell. Probably this is more important in a 3-dimensional model than in this in a way 1-dimensional model.

Two runs were simulated and denoted as:

- $C_{DU}$ (reference): .No swell effect included, using Eq. 9 for all wind speeds.
- $C_{DW}$: Following swell ($\beta = 0^\circ$), using Eq. 10 for $U_{10} \leq 8$ m s$^{-1}$ and Eq. 9 for $U_{10} > 8$ m s$^{-1}$.

When analysing the effect of the new parameterisation in the model only ice-free conditions were used. The 45-year period was used to investigate the long-term impact in the ocean model.

The mean surface stress $\tau$ for the eastern Gotland basin was 0.1 N m$^{-2}$ (run $C_{DU}$). If assum-
ing following swell (run CD\textsubscript{w0}) the total stress at the surface was reduced by 6%. For situations with winds below 8 m s\(^{-1}\) the stress was reduced by 20%. In some situations the assumption of following swell led to a reduction of the stress by more than 50%. Absolute maximum differences to the reference run were about 0.03 N m\(^{-2}\).

The sea surface temperature (SST) was somewhat changed due to a changed mixing of the ocean surface layer. The overall mean SST was, however, not significantly altered, an average decrease of 0.08 °C for following swell as compared with the reference run. For some periods, though, the difference could be large, exceeding 1 °C, both in terms of cooler and warmer.

For both \(\tau\) and SST there was a seasonal change in the differences between the following swell and the reference runs. The monthly averaged difference in surface stress was smallest in May (Fig. 6a). At this time the MABL is mostly
stably stratified. The largest difference (on average 0.01 N m$^{-2}$) was seen in early autumn when it is still relatively weak winds and a starting of longer periods with unstable atmospheric stratification.

The seasonal pattern in SST difference to the reference run was smaller than 0.2 °C (Fig. 6b). There was a small increase in temperature during June–July and a decrease in early autumn. The reason behind the seasonal change in SST difference was the yearly change in the temperature profile in the water. The mean effects on the mixing depth in the sea were limited.

The current velocity was on average only slightly influenced. The surface current speed was in mean reduced by 1% when assuming following swell. For situations with wind speed below 8 m s$^{-1}$ the reduction was 4%. At 5-m depth the reduction was 7% for the same wind speed interval. In absolute values this means differences measured in mm s$^{-1}$. However, there were occasions where both speed (Fig. 7a and c) and direction (defined as where the current comes from, Fig. 7b and d) differed considerably. For extreme cases the difference in speed could be of the order dm s$^{-1}$ (not shown).

**Discussion**

**The parameterisation**

There have been a number of studies focusing on the air–sea exchange of momentum. Most of them agree on a wind dependence of $C_{DN}$ (Fig. 8). The widely used COARE 3.0 algorithm (Fairall et al. 2003) can be considered present state of the art. Large and Pond’s (1981) classical curve is constant below 10 m s$^{-1}$ and increases linearly with wind speed above. The neutral drag
Wave-dependent momentum flux coefficient by the WAMDI Group (Janssen et al. 1987, WAMDI Group 1988) is often used for wave modelling. This expression is used in the original version of PROBE-Baltic. It is higher than the other curves discussed. Drennan et al. (2003) presented a wave age dependent neutral drag coefficient.

The result of the present study is also shown in Fig. 8. Our growing/mixed sea parameterisation follows the 0.8-wave age line of Drennan et al. (2003) for $U_{10} > 7 \text{ m s}^{-1}$ and near the curve of Large and Pond (1981) for $U_{10} > 8 \text{ m s}^{-1}$.

Our result on the following swell conditions is close to the result of Drennan et al. (1999a) who got $C_{DN} \approx 0.7 \times 10^{-3}$ at $U_{10} \approx 7 \text{ m s}^{-1}$ during a following swell case (not shown). The slightly smaller values in this study can be explained by the use of the $\phi_m$-expression valid for swell, which has a significant influence on $C_{DN}$ calculated from measurements.

**The model simulations**

Including the swell effect on the stress description in the PROBE-Baltic model did not alter the ocean mixed layer depth significantly as a long term average. This depth is mainly governed by the turbulent mixing of the ocean surface layer and the heat fluxes at the surface, together with the halocline (during winter). It is probably more influenced by high-wind situations than the low-wind situations when swell is present. In addition the turbulence in the ocean is mainly considered to be generated as well as dissipated near the surface.

For ocean current velocity the importance of swell is somewhat larger. In the modelling of dispersion of oil or biomass transport, for instance, it could be crucial to model the currents correctly. One should bear in mind here that this model is not suitable for current modelling as this parameter is unrealistically averaged over whole basins. It is primarily used as a parameter determining the turbulence. To do a realistic quantification of the impact of swell on the current velocity one should instead use a horizontally well-resolved model.

The use of the new expression for growing sea (Eq. 9) gave on average 18% lower stress as compared with that for the original setting in the model with the WAMDI formulation (see the previous sub-section). Large seasonal differences in the SST were also seen and the ocean mixing depth was changed due to this. Currents were almost completely different. This shows that the model is not insensitive to a change in the drag coefficient.

We included the wave breaking effect in the PROBE-Baltic model with the presumption that this effect was important to have in the model to be able to do a correct analysis of the impact of

![Fig. 8. The neutral drag coefficient from different studies. The numbers next to the lines (Drennan et al. 2003) represent wave age.](image-url)
swell. Yet, the wave breaking effect did not significantly change the impact of swell on the SST in the model. The surface currents, however, were very much dependent on a correct turbulence description.

Conclusions

During swell, i.e. when the dominant sea waves travel faster than the wind, the turbulence structure in the atmosphere is changed, as compared with growing sea conditions and over land. Among other processes the wind stress is altered. For wind following swell the stress is lowered and for counter swell the stress is enhanced as compared with those for growing sea conditions. A method to reduce the scatter in swell data of the neutral drag coefficient was developed and filters out irregular low frequency variations. Using the filtered data two new expressions of the drag coefficient were developed discriminating between growing/mixed sea and wind following swell (Eq. 10 for swell). The new expressions were implemented in the process oriented ocean model PROBE-Baltic to investigate the effect of swell on the oceanic surface layer. Since no wave model was included we did a sensitivity test with two runs, one reference with only growing/mixed sea, and one assuming following swell, when the wind speed was lower than 8 m s$^{-1}$. The model was run for 45 years. The results give a measure of the maximum influence from swell in the Baltic Sea but information about waves is needed to compare the results to observations.

The main results are:

- For the 45-year average the surface stress was reduced by 6% for wind following swell. Below 8 m s$^{-1}$ the stress was reduced by 20%.
- Current velocity was altered during periods with low wind speed, occasionally of the order dm s$^{-1}$. However, to realistically quantify this impact one needs a horizontally well-resolved ocean model.
- The temperature and the mixing depth in the ocean were not significantly changed.

To conclude, the implementation of the swell effect is of limited importance for a process oriented ocean model. There is an indication that current velocity is the most sensitive parameter. It is likely that the impact of swell is larger in the atmosphere.

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References


Ocean System. 9: 97–120.
Appendix

The turbulence in the PROBE-Baltic model is described by a $k$-$\varepsilon$ model. The transport equations for the turbulent kinetic energy (TKE) $k = \sum_{i=1}^{n} 0.5 u_i^3$, where $u_i$ are the velocity components, and dissipation rate $\varepsilon$ are explicitly solved (Axell and Liungman 2001).

The upper boundary conditions for the TKE and dissipation are:

\begin{align*}
    k &= \left(1 + c_w \right) \frac{u_i^3}{\left(c_p^0\right)^3} + \frac{\max(B,0)}{l} \left(\frac{\left(c_p^0\right)^3}{l} \right)^{3/2} \\
    \varepsilon &= \left(c_p^0\right)^{3/2} \frac{\kappa^3}{l} 
\end{align*}

(A1) (A2)

Here

\[ u_t = \sqrt{\frac{\tau}{\rho_0}} \]

is the ocean friction velocity, where $\rho_0 = 1000 \text{ kg m}^{-3}$ is the water density. $B$ is the buoyancy flux at the boundary and $l = -\kappa z$ is the turbulent macro scale. $c_p^0 = 0.55$ is a constant and in the original version of the model the wave breaking parameter $c_w = 0$.

The boundary conditions for the current speed are:

\begin{align*}
    \nu_t \frac{\partial u}{\partial z} &= u_t^2 \\
    \nu_t \frac{\partial v}{\partial z} &= 0 
\end{align*}

(A3) (A4)

where $u$ and $v$ are the current velocity components in the mean wind direction and lateral direction, respectively, and $\nu_t$ is the eddy diffusivity, depending on $k$, $l$ and stability in the water. The surface stress $\tau$ is determined from the bulk relation in Eq. 2.

If including wave breaking one introduces a surface roughness length $z_w$ for the water side in $l = \kappa(-z + z_w)$ (Craig and Banner, 1994). It is parameterised in the model as $z_w = H_s = 114 u_t$ (Stacey 1999 and L. B. Axell pers. comm.) where $H_s$ is the significant wave height. The parameter $z_w$ should not be seen as a physical roughness length but rather as a length scale of the injected turbulence by the breaking waves (Umblauf et al. 2003).

One further assumes a balance between dissipation and diffusion (flux of TKE from the surface). For the turbulence input by the wave breaking we then have

\[ \frac{\nu_t}{\sigma_k} \frac{\partial \kappa}{\partial z} = m u_t^i \]

(A5)

The constant $m = 100$ is called the “wave energy factor” (Craig and Banner 1994). We can then determine $c_w$ in Eq. A1 ($\sigma_k = 1$):

\[ c_w = \sqrt{\frac{3\sigma_k}{2} c_p^0 m} = 67 \]

(A6)

according to Burchard (2001) where our $c_w$ refers to the whole second term inside the bracket at $z = 0$ in his eq. 12 and that his $c_p^0$ is defined as our $(c_p^0)^4$.

There are some uncertainties in this approach. The parameterisation for $z_w$ according to significant wave height, for instance, ranges from at least $0.5H_s$ to $1.6H_s$ (Terray et al. 1999, Burchard 2001). As the model does not include a wave model the uncertainty increases as $H_s$ itself must be parameterised. Further, $z_w$ can also be described by the Charnock relation with a much larger Charnock parameter.
$\alpha$ than used for the atmosphere. This $\alpha$ varies as well between different studies (e.g. Stacey 1999, Mellor and Blumberg 2004).

The model was run both excluding and including the wave breaking effect. The surface values and the gradient of the current speed close to the surface decreased due to the enhanced mixing from the injected turbulence by the breaking waves. A near logarithmic profile was achieved when no wave breaking was implemented. There was a change from the surface to the bottom giving larger temperature differences between bottom water and surface water when wave breaking was included and in addition a shallower ocean mixed layer was seen. This means that the turbulence intensity decreases rapidly near the surface. One of the sources of turbulence vanishes as the gradient decreases. Bothnian Bay showed an even more distinct change of profile and mixed layer depth. On a 45-year basis there was a mean increase of SST by nearly $0.1^\circ$ and damped extreme temperatures in the summer ($0.5 \, ^\circ C$) in the eastern Gotland basin.