Wave-Induced Wind in the Marine Boundary Layer

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ABSTRACT

Recent field observations and large-eddy simulations have shown that the impact of fast swell on the marine atmospheric boundary layer (MABL) might be stronger than previously assumed. For low to moderate winds blowing in the same direction as the waves, swell propagates faster than the mean wind. The momentum flux above the sea surface will then have two major components: the turbulent shear stress, directed downward, and the swell-induced stress, directed upward. For sufficiently high wave age values, the wave-induced component becomes increasingly dominant, and the total momentum flux will be directed into the atmosphere. Recent field measurements have shown that this upward momentum transfer from the ocean into the atmosphere has a considerable impact on the surface layer flow dynamics and on the turbulence structure of the overall MABL. The vertical wind profile will no longer exhibit a logarithmic shape because an acceleration of the airflow near the surface will take place, generating a low-level wave-driven wind maximum (a wind jet). As waves propagate away from their generation area as swell, some of the wave momentum will be returned to the atmosphere in the form of wave-driven winds.

A model that qualitatively reproduces the wave-following atmospheric flow and the wave-generated wind maximum, as seen from measurements, is proposed. The model assumes a stationary momentum and turbulent kinetic energy balance and uses the dampening of the waves at the surface to describe the momentum flux from the waves to the atmosphere. In this study, simultaneous observations of wind profiles, turbulent fluxes, and wave spectra during swell events are presented and compared with the model. In the absence of an established model for the linear damping ratio during swell conditions, the model is combined with observations to estimate the wave damping. For the cases in which the observations showed a pronounced swell signal and almost no wind waves, the agreement between observed and modeled wind profiles is remarkably good. The resulting attenuation length is found to be relatively short, which suggests that the estimated damping ratios are too large. The authors attribute this, at least partly, to processes not accounted for by the model, such as the existence of an atmospheric background wind. In the model, this extra momentum must be supplied by the waves in terms of a larger damping ratio.

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Although it might seem intuitive that fast-running waves (swell) arriving on light wind areas will have an impact on the local wind field, this concept had not been given proper attention until the laboratory experiments by Harris (1966). During several experiments performed

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in an indoor wave tank using a mechanical wave generator, it was noticed that a weak wind immediately above the waves was always present. Harris (1966) named this phenomenon the "wave-driven wind."

Observations of the air-sea interaction regime in the presence of swell are relatively rare and sparse. Nevertheless, studies in the early 1970s from different Soviet ocean campaigns (Volkov 1970; Benilov et al. 1974) and later from Lake Michigan (Davidson and Frank (1973), from the Baltic Sea (Smedman et al. 1994, 1999; Rutgersson et al. 2001), and from several campaigns in the Atlantic and Pacific Oceans (Donelan et al. 1997; Grachev and Fairall 2001) have found evidence that the presence of fast-running waves during light winds induces an upward momentum flux, directed from the water surface to the atmosphere.

The study from Smedman et al. (1999) was based on observations collected in the aftermath of a gale from a tower located on the southern tip of the small island of Östergarnsholm, east of Gotland Island in the Baltic Sea. During periods of strong swell regime, upwarddirected momentum fluxes were recorded from turbulence sensors. In addition, wind measurements at several levels showed a well-defined negative wind gradient above the first measuring level (around 8 m high above the mean sea level). This negative gradient indicated the presence of a low-level wind maximum in the lower marine atmospheric boundary layer (MABL). These findings were in agreement with what Harris (1966) had already postulated, namely that it would be possible that a wave-driven wind might produce a perturbation on the velocity profile by increasing the wind velocity in the direction of wave propagation at low elevations above the water.

Until recently, the wave-driven wind has been looked upon as a peculiarity or, as Grachev and Fairall (2001) suggest, an exotic case. In spite of being an intriguing process, the dominant idea has been that it only occurs in a thin layer above the water surface and that it presumably has no impact on the dynamics of the atmosphere (Janssen 2004). Sullivan et al. (2000) and Rutgersson and Sullivan (2005), using direct numerical simulation (DNS), and Sullivan et al. (2008), using large-eddy simulation (LES), investigated the impact of swell on the MABL. Their findings indicate a stronger impact, in agreement with previous (Smedman et al. 1999) and more recent (Smedman et al. 2009) field measurements. The impact of swell was shown, both by measurements and simulations, not only to generate a wave-driven wind but also to influence the overall turbulence structure of the MABL.

The basic concept behind the wave-driven wind and momentum transfer from the waves into the MABL is

that swell waves perform work on the overlying atmosphere because they propagate faster than the wind, producing a forward thrust on the flow. Hence, swell loses momentum and energy to the atmosphere as it gradually decays, accelerating the airflow. Under swell influence, the wind profile exhibits a low-level wind maximum and a negative (or constant) gradient from there on, violating the logarithmic wind profile law. The Monin–Obukhov similarity theory cannot be claimed as valid in this situation (Miller et al. 1999; Smedman et al. 2009).

As Hristov et al. (2003) pointed out, the incomplete understanding of the atmosphere-ocean interchanging processes reduces the predictability not only of climate models but also of weather and wave forecasting models. Swell is known to propagate thousands of kilometers across entire oceans (Snodgrass et al. 1966), crossing the tropics and the equatorial regions where light wind regimes prevail. The picture that emerges from this feedback process is of momentum being transferred from the wind into the ocean at mid and high latitudes, where storms are more frequent. Part of this momentum is used in the wave generation process along storm tracks. As waves propagate away from their generation area as swell, some of this momentum is returned to the atmosphere, mainly at lower latitudes, in the form of wave-driven winds. Therefore, a better physical understanding of this process is of considerable interest from a global climatological point of view.

Although the attenuation might be small, there is some decay in the swell energy as it propagates, and the physical mechanisms responsible for this attenuation remain poorly understood (Komen et al. 1994; Ardhuin and Jenkins 2006; Kantha 2006). A major question is how (and to where) the wave energy is transferred and how to model this air-ocean exchange process. Kudryavtsev and Makin (2004) presented numerical solutions from a onedimensional stationary model for the MABL flow in the presence of swell. They have imposed an inner region and an outer region structure on the MABL, with the wave influence on the atmosphere being limited to the inner region. The model is conceptually based on the energy transfer from the waves to the atmosphere when the momentum flux is directed upward. Although in the situation in which the wind is aligned with the swell propagating direction the model reproduces a low-level wave-induced wind maximum, as found by Smedman et al. (1999) and Sullivan et al. (2008), the wind maximum is located at a lower height.

More recently, Hanley and Belcher (2008) investigated how ocean waves affect the dynamics of the whole MABL by proposing different models of momentum budget above the ocean surface. The models were further



FIG. 1. Map of the Baltic Sea, with a close-up of the measuring site. The wave buoy is moored at \sim 4 km east-southeast of the tower in the island, at 36-m-deep water.

used to assess the effect of swell on the wind profile and on the entire MABL dynamics. Their study was based on Ekman theory, modified by introducing a waveinduced component on the total stress and on the Ekman wind profile. In addition to qualitatively reproducing the LES experiment from Sullivan et al. (2008), they have proposed several criteria for the existence of swelldriven wave-induced jets using different parameterizations of the eddy viscosity coefficient and different turbulence closures; however, as with Kudryavtsev and Makin (2004), their study was not compared with field measurements.

In the present paper, a model that qualitatively reproduces the wave-induced stress in the surface layer of the MABL wind is proposed. New parameterizations for the wave-induced stress at the surface (expressed as a function of the swell energy decay rate and wave slope) and variation with height are included in the model. The model results are compared with observations from a tower at Östergarnsholm Island in the Baltic Sea.

In section 2, the measuring site and the data used in the comparisons are described. The selection criteria for the different cases used in the comparisons are also explained in this section. The model is derived in section 3, with two different parameterizations for the eddy viscosity (linearly varying with height and as a function of the turbulent kinetic energy). An extension of the model in which the assumption of constant total stress is relaxed is presented in the end of section 3. In section 4 some sensitivity tests and comparisons with wind speed profile observations are shown. In section 5 the results are discussed with reference to previous findings.

2. Observations

The measurements used in this study were taken at the Östergarnsholm site in the Baltic Sea. This air–sea interaction measuring site consists of an instrumented 30-m-high tower, situated at the southernmost tip of Östergarnsholm (57°27'N, 18°59'E; see Fig. 1), and a directional wave rider (DWR) buoy. The DWR is run and owned by the Finnish Institute of Marine Research (FIMR) and was moored about 4 km southwest of the tower, where the water depth is 36 m. The island is very low and flat, with virtually no trees and very scarce vegetation. The tower base is located at 1 m above the mean sea level, with a ± 0.5 -m sea level variation.

High-frequency Solent sonic 1012 anemometers (Gill Instruments, Lymington, United Kingdom), mounted at 9, 16.5, and 25 m above the tower base, recorded turbulence data of the three wind components (and also temperature) at 20 Hz. In addition, slow response

("profile") sensors recorded temperature and wind speed and direction at 6.9, 11.8, 14.3, 20, and 28.8 m above the tower base at 1 Hz. A high-pass filter based on a 10-min linear detrending was applied to the turbulence time series to remove trends. Both turbulence and slow response data are 60-min averages. The wave measurements were recorded once every hour for 26min-long periods. The significant wave height was calculated using the integration trapezoidal method, from spectral frequencies between 0.025 and 0.58 Hz. The spectral variance, mean direction, directional spreading, skewness, and kurtosis were calculated over a frequency range of 68 bins from 0.025 to 0.58 Hz. For longer period waves (swell), which are the focus of this study, a wave transformation correction was applied (see appendix of Smedman et al. 1999).

Both meteorological and wave data have been collected almost continuously since 1995 at this site. For the wind direction in the 80° to 210° sector, the data have been shown to represent open sea conditions in the sense that the wave field is mainly undisturbed and the atmospheric turbulence is not influenced by the low water depths very near the shore (Högström et al. 2008). Additional details about the measuring site—including the flux footprint analysis concept and a detailed analysis of the wave field, the fetch conditions, and the bottom topography in the vicinity of the buoy and around the southern shore of the island—can be found in Smedman et al. (1999, 2003) and Högström et al. (2009).

The measurements used for the model comparisons in section 4 were obtained from two different periods: 16–19 September 1995 and 22–23 September 1996. A total of six cases were selected: four from the first period and two from the second.

Relatively low wind speeds were measured during the selected cases (indicating a light wind regime), and the wind direction was roughly aligned with the swell propagation direction. A swell-dominated wave field was always present. In all cases the 60-min averaged total momentum flux was negative and therefore directed upward. Relatively small positive heat fluxes were present, giving a slightly unstable stratification in both selected periods.

3. Model

Over the sea surface, when waves are present, the wind velocity has an additional component. Besides the mean and the turbulent components, found over land or rigid surfaces like ice, a wave-induced term is now present. The total kinematic stress τ_{tot} is therefore partitioned into turbulent shear stress τ_{turb} , wave-induced stress τ_{wave} and viscous stress τ_{visc} (Phillips 1977):

$$\tau_{\rm tot} = \tau_{\rm turb} + \tau_{\rm wave} + \tau_{\rm visc},\tag{1}$$

where the viscous component is neglected because it is not important from a distance on the order of millimeters above the sea surface.

During the wave developing (or growing) process, the wave-induced stress (or wave-induced momentum flux) is directed downward and is positive ($\tau_{wave} > 0$; Komen et al. 1994). In this situation, energy and momentum are being supplied from the atmosphere to the sea surface. We define the wave age parameter as c_p/u_* .

As waves start propagating away from their generation area, the wave-induced momentum flux gradually decreases, reaches zero, and reverses sign, becoming negative ($\tau_{wave} < 0$) (Smedman et al. 1999; Grachev and Fairall 2001). When the wave-induced momentum flux becomes dominant over the turbulent stress (i.e., $|\tau_{wave}| > |\tau_{turb}|$), the total momentum flux will reverse sign, becoming negative ($\tau_{tot} = \tau_{wave} + \tau_{turb} < 0$) and upward directed. The negative total momentum flux indicates that energy and momentum are being transferred from the sea surface to the atmosphere.

a. Constant flux model

A neutrally stratified MABL is now considered. It is assumed that in the surface layer the effect of the Coriolis term is negligible and therefore the total stress and its turbulent and wave-induced components are confined to the x direction. For two-dimensional stationary flow with no horizontal gradients, it follows from the principle of conservation of momentum that the shear stress is constant in the turbulent boundary layer:

$$\frac{d\tau_{\rm tot}}{dz} = 0. \tag{2}$$

Here, z is the vertical coordinate, which is positive upward. The turbulent stress will be parameterized as

$$\tau_{\rm turb} = K_m \frac{dU}{dz},\tag{3}$$

where K_m is the turbulent eddy viscosity and U is the mean horizontal wind. Inserting (3) in Eq. (2) yields an equation for the mean horizontal wind:

$$\frac{dU}{dz} = \frac{\tau_{\rm tot} - \tau_{\rm wave}}{K_m}.$$
(4)

The wave stress is

$$\tau_{\rm wave} = -\langle \tilde{u}\tilde{w} \rangle, \tag{5}$$

where u and w are the longitudinal and vertical components of the flow. The angled brackets indicate time

averaging, and the tilde denotes wave-induced flow fluctuations.

A parameterization for τ_{wave} nevertheless is still needed. For irrotational waves, the orbital velocity components decay as e^{-kz} , where k is the wavenumber. In this case, the vertical and horizontal components are 90° out of phase. The wave stress is then of course zero. When a small amount of work is performed at the surface, the velocity components are slightly phase shifted, as observed in the LES of Sullivan et al. (2008). Their findings indicate a wave-induced stress that exponentially decays with height, in agreement with prior numerical calculations presented by Chalikov and Belevich (1993) and with the field measurements of Högström et al. (2009). In this study it will be assumed that the upward directed wave-induced stress will have the form

$$\tau_{\rm wave} = \tau_{\rm wave}^0 e^{-2kz},\tag{6}$$

where τ_{wave}^0 is the wave-induced stress at the surface.

This surface wave stress can be related to the energy damping through the rate of work performed at the surface. For one harmonic wave component, the energy per unit area is

$$E = \frac{1}{2}\rho_w g a^2, \tag{7}$$

where ρ_w is the water density, g is the acceleration of gravity, and a is the wave amplitude. The rate of change of wave energy caused by a surface stress is proportional to the phase velocity multiplied by the stress:

$$\frac{\partial E}{\partial t} = \rho_a c \tau_{\text{wave}}^0, \tag{8}$$

where c is the phase velocity. By linear theory, the rate of change of wave energy in decaying waves may be written as

$$\frac{\partial E}{\partial t} = \beta E,\tag{9}$$

where β is the growth rate or the wave damping coefficient depending on the sign. Using Eqs. (7), (8), and (9), the relation between τ_{wave}^0 and β is found to be

$$\tau_{\rm wave}^0 = \frac{1}{2} \frac{\beta g a^2}{sc},\tag{10}$$

where $s = \rho_a / \rho_w$. By substituting this in Eq. (6), the wave stress can be written as

$$\tau_{\rm wave} = \frac{1}{2} \frac{\beta g a^2}{sc} e^{-2kz}.$$
 (11)

For a wave spectrum, the wave stress is found by adding the contribution from all wave components:

$$\tau_{\text{wave}} = \int_0^\infty \frac{\beta g S(f)}{sc} e^{-2kz} df.$$
(12)

Here, S(f) is the wave spectrum and f is the frequency. In this case, β is positive in the high-frequency range and changes sign for long waves traveling faster than the wind. If we define β^g as the growth rate and β^d as the damping coefficient, the wave stress can be rewritten as

$$\tau_{\text{wave}} = \int_0^{f_c} \frac{\beta^d g S(f)}{sc} e^{-2kz} df + \int_{f_c}^{\infty} \frac{\beta^g g S(f)}{sc} e^{-2kz} df, \quad (13)$$

where f_c is the frequency corresponding to a wave phase speed equal to the 10-m wind speed. This frequency can be seen as a separation between the swell and the young sea parts of the wave spectra; it is

$$U_{10} = c = \frac{\omega}{k} = \frac{g}{\omega} = \frac{g}{2\pi f_c} \Rightarrow f_c = \frac{g}{2\pi U_{10}}, \qquad (14)$$

where $\omega = \sqrt{gk}$ is the angular frequency for deep water waves and U_{10} is the wind speed at 10 m.

The eddy viscosity can be related to the turbulent kinetic energy b and the mixing length l as

$$K_m^b = l\sqrt{b}.\tag{15}$$

In this study, a neutral stratification is assumed and the mixing length will be taken to be

$$l = \kappa z, \tag{16}$$

where $\kappa = 0.4$ is the von Kármán constant. Following Kudryavtsev and Makin (2004), we assume a balance among shear production, the vertical rate of change of energy flux, and dissipation of turbulent kinetic energy. The turbulent kinetic energy budget is then (Tennekes and Lumley 1972)

$$\tau_{\rm tot} \frac{dU}{dz} + F_w - \epsilon = 0, \tag{17}$$

where ϵ is the energy dissipation. The wave term in Eq. (17) is

$$F_w = -\frac{1}{\rho_a} \frac{d}{dz} (\langle \tilde{p} \tilde{w} \rangle).$$
(18)

Here, \tilde{p} is the fluctuating part of the pressure due to the waves and ρ_a is the density of the air. The energy dissipation term is often parameterized as

$$\epsilon = \frac{b^{3/2}}{l}.\tag{19}$$

Combining Eqs. (3), (15), (17), and (19) yields an equation for the turbulent kinetic energy:

$$b^{2} = |\tau_{\text{tot}}(\tau_{\text{tot}} - \tau_{\text{wave}}) + l\sqrt{b}F_{w}|.$$
 (20)

Here, the total stress can be negative (i.e., upward directed). This will be the case whenever the magnitude of the wave stress is larger than the turbulent stress. The shear production term in Eq. (17) may therefore become negative. The absolute value of the first term on the right-hand side is taken to avoid negative production of turbulent kinetic energy.

The energy flux in Eq. (20) is related to the rate of work:

$$-\langle \tilde{p}\tilde{w}\rangle = -\rho_a c \langle \tilde{u}\tilde{w}\rangle \Rightarrow -\langle \tilde{p}\tilde{w}\rangle = \rho_a c \tau_{\text{wave}}^0 e^{-2kz}.$$
 (21)

Accordingly, the pressure perturbation term for in (20) for one harmonic component becomes

$$F_w = -2kc\tau_{\text{wave}}^0 e^{-2kz}.$$
 (22)

For a wave spectrum, the expression can be written as

$$F_{w} = -\int_{0}^{f_{c}} \frac{2\beta^{d}gkS(f)}{s} e^{-2kz} df - \int_{f_{c}}^{\infty} \frac{2\beta^{g}gkS(f)}{s} e^{-2kz} df.$$
(23)

If the damping and growth rates are known, the wave stress and wave energy flux can be calculated from Eqs. (13) and (23). The wind profile is then found by solving Eqs. (4) and (20) numerically. Using a prescribed eddy viscosity instead of Eq. (20) yields a simplified solution that nicely illustrates the general behavior of the model. For this purpose we will use an eddy viscosity that varies linearly with height. Strictly, this is only valid in the absence of waves. However, because the eddy viscosity will have to increase with height near the surface, we will assume that the assumption of a linearly increasing eddy viscosity does not violate the general structure of the solution. It will be presented here to demonstrate how the traditional boundary layer model is modified in the presence of swell. The eddy viscosity is then taken to be

$$K_m^l = \kappa z u_*. \tag{24}$$

Here, $u_* = \sqrt{|\tau_{tot}|}$ is the friction velocity. Using Eq. (11) for the wave stress, the solution to Eq. (4) is

$$U(z) = \frac{\tau_{\text{tot}}}{\kappa u_*} \ln\left(\frac{z}{z_0}\right) - \frac{\tau_{\text{wave}}^0}{\kappa u_*} \int_{z_0}^z \frac{e^{-2kz}}{z} dz, \qquad (25)$$

where z_0 is the aerodynamic roughness length. In the absence of wave stress, Eq. (25) reduces to the well-known logarithmic profile. We see that the introduction of the wave stress yields a new term that modifies the

traditional logarithmic distribution. For cases in which the total stress is upward directed, the logarithmic term is negative and the wave term becomes a production term for mean wind in the MABL.

b. Nonconstant flux model

The assumption of a constant momentum flux in the surface MABL allows for relatively broad variations with height: up to 10% of its magnitude over the total surface layer (Stull 1988). Recent field campaigns have collected observations that do not confirm the constant flux assumption under a swell-dominated wave field. The total momentum flux magnitude has been observed to decrease with height (Smedman et al. 2009). This decrease in magnitude is related to the impact of swell on the turbulence structure in the MABL. The starting point is the fact that swell induces modifications in the turbulence production mechanism close to the surface. These modifications have an impact on the turbulence structure and dynamics of the entire MABL and are expected to drive the nonconstant momentum flux, breaking one of the assumptions of the Monin-Obukhov similarity theory. If we assume that τ_{tot} is no longer constant but varies linearly with height, tending asymptotically to values close to zero outside the surface MABL, then

$$\tau_{\rm tot} = \tau_{\rm tot}^0 + \alpha z, \qquad (26)$$

where τ_{tot}^0 is the value of the flux at the surface and α its vertical gradient, which is assumed to be constant and positive; thus, the magnitude of τ_{tot} is actually decreasing. Following the same steps that lead to Eq. (25), the wind speed profile then becomes

$$U(z) = \frac{\tau_{\text{tot}}}{\kappa u_*} \ln\left(\frac{z}{z_0}\right) - \frac{\tau_{\text{wave}}^0}{\kappa u_*} \int_{z_0}^z \frac{e^{-2kz}}{z} dz + \frac{\alpha}{\kappa u_*} (z - z_0).$$
(27)

An additional term is present on the right-hand side. This term will determine the departure from the original wind speed profiles due to the vertical variation of the total momentum flux. The impact of the nonconstant momentum flux MABL will be explored in greater detail in the following section.

4. Results

a. General behavior of the model

In this section general model results, along with some model sensitivity tests, are presented. The behavior of the wind speed profiles from Eq. (25) and the numerical solution of (4) and (20), corresponding to the linearly varying with height eddy viscosity K_m^l and the turbulent kinetic energy (TKE)-dependent eddy viscosity K_m^b , respectively, are investigated. The input parameters to the

model are an upward-directed constant momentum flux $\tau_{tot} = -10^{-2} \text{ m}^2 \text{ s}^{-2}$, a wave damping parameter $\beta = -5 \times 10^{-5} \text{ s}^{-1}$, and a constant roughness length $z_0 = 10^{-5} \text{ m}$. We stress that these values have been chosen only to demonstrate the overall behavior. Realistic values will be discussed in more detail later. The monochromatic wave field is assumed to have a wave amplitude a = 1 m and a wavenumber $k = 0.1 \text{ m}^{-1}$, corresponding to a wave phase speed $c = 9.9 \text{ m s}^{-1}$, wave period $T \approx 6.3 \text{ s}$, and wavelength $L \approx 63 \text{ m}$.

The two wind speed profiles, corresponding to the two eddy viscosity formulations, are shown in Fig. 2. Note that in most of the figures presented here, the x axis starts below zero to highlight gradients close to the surface. Additional thin lines have been added to the figures to mark zero. A distinctive wave-induced lowlevel wind jet (or wind maximum) in the surface MABL, for both eddy viscosities, can be seen in Fig. 2a. The wind profile calculated using the K_m^l eddy viscosity (dashed line) has a more pronounced bulge and a higher wind speed at the wind maximum (jet strength) compared with the profile evaluated with the K_m^b eddy viscosity (dotted-dashed line). The heights of the wind maxima are the same for both cases ($z \approx 3$ m). The wind speeds at the jet are 3.6 and 2.5 m s⁻¹ for the K_m^l and K_m^b eddy viscosities, respectively. Figure 2b shows the two wind speed profiles for the entire MABL, normalized by the background flow U_B . The height of the boundary layer is assumed to be 200 m, and the background flow is the wind speed at that height. The wave-induced wind speed departures from U_B at low levels can be seen for both eddy viscosity formulations. Above the jet the wind speed reduces smoothly to the background flow in both profiles. This feature is in qualitative agreement with the LES predictions from Sullivan et al. (2008).

The vertical profiles for the two eddy viscosities are shown in Fig. 3. The inclusion of the TKE on the K_m^b formulation gives rise to an increase of the eddy viscosity up to about 25 m, compared with the linearly varying K_m^l formulation. The immediate result is a vertical diffusion of momentum. This turbulence diffusion leads to a less pronounced wind jet, as seen in Fig. 2a. The lower jet strength is also related to the momentum diffusion.

An additional horizontal wind speed profile from Eq. (27), with the K_m^l eddy viscosity, is shown in Fig. 4. The input parameters remained unchanged, with the exception of the total stress that is now decaying (in magnitude) with height according to Eq. (26), where $\tau_{tot}^0 = -0.01 \text{ m}^2 \text{ s}^2$ and $\alpha = 3 \times 10^{-4} \text{ m s}^{-2}$. The surface MABL height, where the total stress eventually becomes zero and the TKE vanishes, is assumed to be 30 m. The decrease in magnitude of τ_{tot} resulted in an increase of the wind speed at higher levels and a less enhanced jet



FIG. 2. (a) Wind speed profiles for both eddy viscosity formulations, K_m^l (dashed) and K_m^b (dotted–dashed), at the surface MABL. The triangle and the diamond represent the wind speed maxima for the K_m^l and K_m^b eddy viscosities, respectively. (b) Normalized wind speed profiles for both eddy viscosity formulations, with the boundary layer depth being 200 m. The thin horizontal line represents the x axis.

(dashed line) when compared with the original wind speed profile computed with a constant total momentum flux (solid line).

The vertical profiles of the total stress and its components—the turbulent and wave-induced stresses, used as input parameters—are shown in Fig. 5. In Fig. 5a, τ_{tot} is constant with height (vertical solid line), τ_{wave} (dashed line) is evaluated from Eq. (11), and the



FIG. 3. Eddy viscosity profiles: K_m^l (dashed) and K_m^b (dotted-dashed). The thin horizontal and vertical lines represent the *x* and *y* axes, respectively.

turbulent stress (dotted-dashed line) is computed from $\tau_{\rm turb} = \tau_{\rm tot} - \tau_{\rm wave}$. The vertical profiles of the stresses resulting from the nonconstant stress model are shown in Fig. 5b, where $\tau_{tot} \approx \tau_{turb} \approx 0$ at z = 30 m. The less pronounced bulge is due to the decrease (in magnitude) of τ_{turb} after a certain height, tending to a decreasing τ_{tot} . The characteristics of the wind profiles in Fig. 2 are related to the stress profiles shown in Fig. 5a and to the dynamical processes behind them. Below the jet the wind speed gradient is positive. In this layer the wave-induced stress accelerates the wind because there is a significant amount of momentum being transferred from the waves to the atmosphere; the turbulent stress is positive (and downward). At the height of the jet, τ_{turb} is zero and reverses sign and direction from there up (becoming negative and upward), corresponding to a smoother negative wind speed gradient. The physical sense of this process is in agreement with the LES results from Sullivan et al. (2008) and with the field observations and findings of Högström et al. (2009).

Model sensitivity tests to the variability of the wave damping parameter and the roughness length, using the model formulation from Eq. (25), are performed. The model response to the variations of β^d is shown in Fig. 6. The original profile with $\beta^d = -5 \times 10^{-5} \text{ s}^{-1}$ is shown as a full line. The wave damping parameter is then slightly varied ($\Delta\beta^d = \pm 3 \times 10^{-6} \text{ s}^{-1}$), keeping all the remaining parameters unchanged. Increased values of β^d , corresponding to a larger loss of energy from the



FIG. 4. Wind speed profiles for constant and nonconstant total stress models. The solid line is the wind speed profile for K_m^l eddy viscosity, with a constant total stress, as in Fig. 2. The dashed (dotted–dashed) line indicates the departure from the wind speed profile (solid line), owing to a total momentum flux that decreases with height. The triangle represents the wind speed maximum. The thin horizontal and vertical lines represent the *x* and *y* axes, respectively.

waves into the atmosphere, lead to a stronger and higher jet (wind profiles in dashed lines in Fig. 6). The opposite effect, with lower values of β^d and consequently less energy transfer from the waves, leads to a weaker and lower jet. The horizontal shift of the wind profiles in Fig. 6 occurs because only β^d is varied and all the other parameters are kept unchanged.

The fact that the wind profile over the ocean, under swell conditions, is no longer logarithmic makes the correct evaluation of the roughness length a cumbersome problem (see Figs. 3 and 7 in Smedman et al. 2003). For the model sensitivity tests to roughness length variations, several formulations are used (Table 1). Once again the original profile evaluated with $z_0 = 10^{-5}$ m is kept as a reference and is shown as a solid line in Fig. 7. The profiles corresponding to several roughness lengths from the different formulations are shown as dashed lines. The effect of roughness length variations is different from the one obtained with the variations of the damping parameter β^d . The impact is now caused not only by the wave-induced part of Eq. (25) but also by the logarithmic component of the profile. Variations of z_0 have an impact on the wind speed only and not on the height of the wind speed maxima, as can be seen in Fig. 6. Larger values of z_0 give rise to lower wind speed values and vice versa. The sensitivity of the model to the roughness length, however,



FIG. 5. Profiles of the total momentum flux, wave-induced stress, and turbulent stress, for the (a) constant and (b) nonconstant stress model. The full thick line is the total momentum, which is negative (upward directed) and constant with height. The dashed line is the wave-induced stress, which is negative (upward directed) and tends asymptotically with height to zero ($\tau_{wave} \rightarrow 0$). The dotted–dashed line is the turbulent stress, which can be either positive (downward directed) or negative (upward directed) and tends asymptotically with height toward the total stress ($\tau_{turb} \rightarrow \tau_{tot}$). The thin horizontal line represents the *x* axis.

is relatively small. The wave age-dependent relation roughness length from Smith (1992)— $z_0 = 0.48u_*^2/g(u_*/c_p)^{1.0}$ —will be used in the model comparisons with observations in the following section.



FIG. 6. Model sensitivity tests to the variability of the wave damping parameter β^d . The full line is the wind speed profile for the K_m^l eddy viscosity, as in Fig. 2. The dashed line wind speed profiles are the result of variations in the wave damping parameter ($\Delta\beta^d = \pm 3 \times 10^{-6} \text{ s}^{-1}$). The black triangle symbols represent the wind speed maxima. The height of the wind speed maxima decreases for decreasing values of β^d . The thin horizontal and vertical lines represent the x and y axes, respectively.

b. Comparisons with field measurements

In the present section the modeled wind profiles from Eq. (25) and the numerical solution of Eqs. (4) and (20), corresponding to the K_m^l and K_m^b eddy viscosities, are compared with wind speed measurements from the 30-m-high tower at Östergarnsholm Island. The observed input parameters to the modeled wind profiles are the total stress form highest measuring level, the roughness length, the wave amplitude, and the wave phase speed. The wave amplitude is approximated by one half of the observed significant wave height ($a \approx H_s/2$). Despite the impact that convective processes can have in the vertical distribution of momentum in the MABL, a neutrally stratified MABL is assumed because of the relatively small heat fluxes measured in the tower.

TABLE 1. Aerodynamic roughness length (z_0) formulations.

Reference	Formulation			
Kudryavtsev and Makin (2004)	$z_0 = 0.1v/u_* + 0.012u_*^2/g$			
Smooth flow	$z_0 = 0.11 v/u_*$			
Charnock (1955)	$z_0 = 0.012 u_*^2/g$			
Donelan (1990)	$z_0 = 0.184 \sigma (u_*/c_p)^{2.53}$			
Smith (1992)	$z_0 = 0.48 u_*^2 (u_*/c_p)$			



FIG. 7. Model sensitivity tests to the variability of the aerodynamic roughness length. The full line is the wind speed profile for the eddy viscosity, as in Fig. 2, computed from the Charnock (1955) relation. The dashed line wind speed profiles are the result of variations in the roughness length due to different formulations. The symbols represent the wind speed maxima corresponding to each formulation: the circle indicates Kudryavtsev and Makin (2004); the triangle, smooth flow; the pentagram, Donelan (1990); and the square, Smith (1992). Lower (higher) values of roughness lead to higher (lower) wind speed maxima. The thin horizontal and vertical lines represent the *x* and *y* axes, respectively.

Of all the needed input parameters, the one that has not been observed is the wave attenuation/growth parameter β . A clear conclusion about the parameterizations of the parameter β is still lacking for both the growth and the decaying regimes (but particularly the latter), as Hanley and Belcher (2008) discuss. As far as we know, there is no study or measurements available in the open literature describing specifically the damping rate of swell aligned with the wind. There are some laboratory studies on wave energy attenuation, but they are mainly focused on the effect of an opposing aligned wind on wave energy decay (Donelan 1999; Pierson et al. 2003). The results from these laboratory experiments are not consensual and most of the time are either not in agreement with the theory or cannot be directly applied in the open ocean (Kudryavtsev and Makin 2004).

Ardhuin and Jenkins (2006) concluded that in addition to losing energy to the atmosphere, swell also transfers energy to the ocean mixed layer. The loss of energy from swell into the ocean occurs when waves and upper ocean turbulence coexist and the interaction between the two leads to a transfer of energy from waves to TKE. They have also concluded that for swell propagating in the wind direction, the transfer of wave energy into the ocean is at least one order of magnitude smaller than the loss of energy to the atmosphere.

The lack of a proper parameterization for the parameter β makes the comparisons of the model wind profiles proposed in the present study with field observations rather challenging. Instead of using one of the proposed parameterizations available in the literature, the wind speed observations from the tower are used as an input to the model. By using the formulation for the total (net) wave-induced stress in Eq. (13), the damping parameter β^d is then taken as a residual parameter by forcing both modeled profiles to go through the observed wind speed at the lowest level. For that purpose we introduce the weighted average of the damping parameter over the low-frequency range of the observed wave energy spectra β^d , such that

$$\tau_{\text{wave}} = \widehat{\beta^d} \int_{f_{\min}}^{f_c} \frac{gS(f)}{sc} e^{-2kz} df + \int_{f_c}^{f_{\max}} \frac{\beta^g gS(f)}{sc} e^{-2kz} df,$$
(28)

where f_{\min} and f_{\max} are the minimum and maximum frequencies in the observed spectra. The frequency f_c is computed by using an interpolated U_{10} between the two lowest wind speed observation levels. In a similar way, the wave term in Eq. (20) is calculated as

$$F_w = -\widehat{\beta^d} \int_{f_{\min}}^{f_c} \frac{2gkS(f)}{s} e^{-2kz} df - \int_{f_c}^{f_{\max}} \frac{2\beta^g gkS(f)}{s} e^{-2kz} df.$$
(29)

For the young sea part of the wave spectra, the growth of the waves remains frequency dependent and is assumed to be explained by the parameterization from Belcher and Hunt (1993):

$$\beta^g = C_\beta \omega s \left(\frac{u_*}{c}\right)^2,\tag{30}$$

where C_{β} is a constant and ω is the wave frequency in radians. Following Hanley and Belcher (2008) the wave growth rate coefficient is taken as $C_{\beta} = 0.32$. The observed 1D wave energy density spectra are used as an input for computing the total (net) wave-induced stress and the wave energy flux from Eqs. (28) and (29).

As mentioned in section 2, the selection of the six cases had as a main a priori condition an upward-directed momentum flux (60-min averaged), followed by the conditions established in Högström et al. (2008) for the so-called swell cases at the footprint area of the Östergarnsholm measuring site. The six cases are ordered chronologically and are designated A to F. Despite the limitation that the lowest wind speed measuring level is as high as 7.9 m, the fact that from this level up the wind profiles in all cases exhibit a well-defined negative gradient allows us to plausibly speculate that a wind speed maximum will be somewhere between this level and the surface (Smedman et al. 1999).

Figures 8 and 9 show the comparisons between observed wind speeds (black squares joined by solid line) and modeled wind speed profiles computed for the six cases (dashed and dotted–dashed lines for the K_m^l and K_m^b eddy viscosities, respectively). The observed wave energy spectra for each case are also shown in Figs. 8 and 9. The wave spectra are divided in two parts by a dashed vertical line. This line—the separation frequency f_c —separates the swell energy part from the young sea part.

There is a very good qualitative and quantitative agreement between the observed and the modeled wind speed profiles in cases A, B, C, and F, with not so good agreement in cases D and E. The quality of the agreement between the observed and modeled profiles seems to be directly related to the wave age parameter, here defined as c_p/u_* . All cases are clearly swell dominated because for all of them $c_p/u_* > 20$. Nevertheless, the lowest wave age values, indicating a less swell-dominated wave field, are the ones measured during cases D and E, whereas for cases A, B, C, and F the wave age values are higher. The values of the total stress, damping ratios, amplitudes, wave lengths, and attenuation lengths are listed in Table 2.

In cases A, B, C and F, the wave damping parameter $\hat{\beta}^d$ is consistently of the order of -10^{-4} s^{-1} for both modeled profiles. The exceptions are cases A and B (only for the K_m^l eddy viscosity), for which the values (magnitudes) are slightly smaller $(-\hat{\beta}^d = 0.89 \times 10^{-4} \text{ s}^{-1}$ and $-\hat{\beta}^d = 0.69 \times 10^{-4} \text{ s}^{-1}$ for cases A and B, respectively). The values for cases D and E are one order of magnitude higher for both eddy viscosities. The values of the damping parameter $\hat{\beta}^d$ for all cases are listed in Table 2. The inverse wave age plotted against $\hat{\beta}^d$ is shown in Fig. 10. There is a clear decrease in $|\hat{\beta}^d|$ as the inverse wave age (or as the swell dominance) decreases, in agreement with Kudryavtsev and Makin (2004).

The height of the jet was evaluated for all cases for both eddy viscosities. The height is higher in the profiles evaluated with the K_m^b eddy viscosity, when compared with the ones evaluated with the K_m^l eddy viscosity. The reason is related to momentum diffusion, as before. The difference between the height on the modeled profiles is relatively more pronounced (1–2 m) for cases A, B, C, and F than for cases D and E (~0.5 m). For these two last cases, the height of the jet was also lower. The justification for this lower height might be the possible disruption of the wind-driven jet by the presence of a background flow or just the poor agreement these two cases show when compared with the observations. On the other hand, assuming that the effect of an unaccounted background flow is very small for cases A, B, C, and F, the wind speed maxima at heights of 4–6 m can be assumed to be driven (mostly) by the upward-directed momentum flux from the waves and are therefore closer to reality.

The spatial scale of attenuation can be calculated as

$$L_a = c_g / \left| \beta^d \right|, \tag{31}$$

where c_g is the group velocity, approximated for the peak wave speed as $c_g \approx c_p/2$. The results are listed in Table 2. The values seems to be rather small even when considering the relatively short wave lengths in these observations (also in Table 2), indicating that the damping ratios are too large. The wave lengths range from 20 to 48 m, with attenuation lengths from 98 km down to almost 6 km for some cases. We see that the longest (and probably closest to realistic) attenuation lengths are found for cases A to C. These are also the cases with best agreement in the comparison between the modeled and the observed wind distribution. We believe that the too large damping ratios can be explained by background wind not accounted for in the model. If a background wind exists during the observations, the model must adjust for this by an artificially large momentum flux and hence a too large damping ratio. The model is very sensitive to the values of τ_{tot} . Uncertainties in the measured total momentum flux might also have contributed to such high tuned values of the damping ratios. An additional source of error might be the fact that a neutrally stratified MABL is assumed. As seen in the LES by Sullivan et al. (2008), the effect of buoyancy in the dynamics of a swell-dominated MABL can be considerable.

5. Conclusions

The effect of fast-running waves on the overlaying MABL is stronger than has been assumed until recently, as seen from the LES predictions in Sullivan et al. (2008) and in the recent field measurements in Smedman et al. (2009) and Högström et al. (2009). A swell-dominated wave field is frequently in a state of disequilibrium in light wind conditions and loses momentum and energy to the overlaying atmospheric flow as it propagates. In this situation, the net momentum and energy exchange over the air–sea interface change direction and are directed upward to the atmosphere. The most



FIG. 8. Comparisons between observed and modeled wind speed profiles, for both (left) eddy viscosities and (right) wave spectra for cases A, B, and C. The dashed line profiles correspond to the K_m^l eddy viscosity, and the dotteddashed line profiles indicate the K_m^b eddy viscosity. The triangles and the diamonds represent the wind speed maxima. The squares represent the observed wind speeds in the tower. The dotted vertical line in the wave spectra represents the separation between the swell and young sea components of the spectra. The thin horizontal and vertical lines represent the x and y axes, respectively.



FIG. 9. Comparisons between observed and modeled wind speed profiles, for both (left) eddy viscosities and (right) wave spectra for cases D, E, and F. (Symbols are as in Fig. 8.)

Case	$ au_{\rm tot} \ (10^{-3} \ { m m}^2 \ { m s}^{-2})$	<i>a</i> (m)	λ (m)	c_p/u_*	K_m^l		K_m^b	
					L_a^l (km)	$\widehat{eta^d} \ (10^{-4} \ { m s}^{-1})$	L_a^b (km)	$\widehat{\beta^d} \ (10^{-4} \ \mathrm{s}^{-1})$
А	-1.22	0.43	48	248.1	81	-0.90	44	-1.54
В	-0.38	0.28	39	400.1	98	-0.70	50	-1.29
С	-0.93	0.30	40	259.6	48	-1.87	26	-3.26
D	-11.28	0.31	38	72.7	7.4	-13.5	5.9	-16.3
E	-4.88	0.19	21	82.4	6.3	-10.9	5.0	-13.4
F	-11.08	0.21	20	168.9	19	-3.52	11	-6.20

TABLE 2. Observed input parameters and tuned damping parameter for the two eddy viscosity formulations.

striking effect of this process is the flow acceleration in the lower surface MABL, leading to the formation of a low-level wind jet. Nevertheless, the effect of swell is not constrained to the surface because it can be extended throughout the MABL by changing its vertical turbulence structure up to the top.

In this paper a model for the effect of swell on the wind profile in the surface layer of the MABL is described. The model is based on the swell loss of energy into the atmosphere, quantified by the wave damping parameter. The governing parameter behind the model is the form drag, here proposed as a function of the wave energy loss, and the wave field characteristics. The wave-induced stress is assumed to decay exponentially with height, in agreement with Sullivan et al.'s (2008) LES results and with Högström et al.'s (2009) findings for swell-dominated wave fields. Two mixing length



FIG. 10. Observed swell damping rate as a function of inverse wave age. The triangles and the diamonds represent the wave age vs the damping ratio for the K_m^l and K_m^b eddy viscosities, respectively. The dashed and dotted–dashed lines show the fits for the two eddy viscosities.

closure schemes are used for the eddy viscosity: one assuming a linear variation with height and the second assuming a dependence on TKE, governed by the wave energy flux. The model assumes a swell-dominated wave field and a net upward-directed transport of momentum, so that at the surface the wave-induced stress dominates the turbulent stress. Although measurements of such situations are scarce and difficult to obtain, they have been observed over the ocean and are one of the key assumptions behind the LES in Sullivan et al. (2008). The observations used in the model comparisons with field measurements had to meet this criterion.

The proposed model reproduces the characteristics and the dynamics of the surface layer of a swell-dominated MABL for wind-following waves, in agreement with the findings of Sullivan et al. (2008) and Högström et al. (2009) and also with prior field observations, mainly from Smedman et al. (1994, 1999). The comparisons with field measurements show a good agreement between modeled and measured wind speed profiles. This agreement is especially good for the cases with a high wave age parameter, indicating a predominant wave effect in the surface MABL. The wave damping parameter is treated as a residual term by forcing the modeled profiles to fit the observed wind speed at the lowest level in the tower. Furthermore, the weighted average of the damping parameter over the swell part of the spectrum is calculated to allow for the wave spectra balance between the swell and the wind waves.

The attenuation length is calculated from the damping ratio and the group velocity. We obtained attenuation lengths shorter than 100 km. Swell is known to propagate over considerable distances, so these attenuation lengths seem to be small. Accordingly, the damping ratios must be too large. One possible explanation is the existence of a background wind not accounted for in the model. In the model, such additional momentum must be supplied by the waves, yielding artificially large damping ratios. The fact that the cases with clearest swell signals gave the largest attenuation lengths supports this hypothesis. This argument also explains the poor qualitative fitting of the modeled profiles in cases D and F. In all six cases, the observed damping decreases with decreasing inverse wave age. This behavior also suggests that the proposed parameterization of the form drag reflects the reality.

Because the lowest observation level was 8.6 m, the height of the jet was not measured but was inferred to be below that level because of the observed negative wind gradient from that level up. Therefore, comparisons between modeled and observed jet heights and strengths are not possible. Nevertheless the results obtained with the model with both eddy viscosity formulations for cases A, B, C, and F, can be viewed as close to reality when compared with the observations (Smedman et al. 2009; Högström et al. (2009) where a similar swell-dominated wave field was present. The rather extreme wave field characteristics imposed in the LES in Sullivan et al. (2008) do not allow for a quantitative comparison as far as the jet strength and height are concerned.

Both a linearly varying eddy viscosity and an eddy viscosity calculated from the TKE budget lead the model to reproduce the effect of swell on the surface MABL. In view of the values obtained for wave-damped parameter and for the swell attenuation length, the results based on a linearly varying eddy viscosity produced smaller (and hence more realistic) values of the wave damping parameter. Because the effect of the wave energy flux in the TKE budget in the lowest meters, below the height of the jet, cannot be ignored, this issue should be further explored in future work.

The driving parameter that ultimately determines the effect of fast-running waves in the MABL is the swell energy rate of change. Only extensive field measurements of this parameter, together with the effect of swell waves in the lower atmosphere, preferably in open ocean, can help improve our knowledge of this effect.

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REFERENCES

Ardhuin, F., and A. D. Jenkins, 2006: On the interaction of surface waves and upper ocean turbulence. J. Phys. Oceanogr., 36, 551–557.

- Belcher, S. E., and J. C. R. Hunt, 1993: Turbulent shear flow over slowly moving waves. J. Fluid Mech., 251, 109–148.
- Benilov, A. Y., O. A. Kuznetsov, and G. N. Panin, 1974: On the analysis of wind wave-induced disturbances in the atmospheric turbulent surface layer. *Bound.-Layer Meteor.*, 6, 269–285.
- Chalikov, D. V., and M. Y. Belevich, 1993: One-dimensional theory of the wave boundary layer. *Bound.-Layer Meteor.*, **63**, 65–96.
- Charnock, H., 1955: Wind stress on a water surface. *Quart. J. Roy. Meteor. Soc.*, **81**, 639–640.
- Davidson, K. L., and A. J. Frank, 1973: Wave-related fluctuations in the airflow above natural waves. J. Phys. Oceanogr., 3, 102–119.
- Donelan, M., 1990: Air-sea interaction. Ocean Engineering Science, B. LeMehaute and D. M. Hanes, Eds., Vol. 9, The Sea, Wiley Interscience, 239–292.
- —, 1999: Wind-induced growth and attenuation of laboratory waves. Wind-Over-Wave Couplings, S. G. Sajjadi, N. H. Thomas, and J. C. R. Hunt, Eds., Oxford University Press, 183–194.
- —, W. M. Drennan, and K. B. Katsaros, 1997: The air-sea momentum flux in conditions of wind sea and swell. J. Phys. Oceanogr., 27, 2087–2099.
- Grachev, A. A., and C. W. Fairall, 2001: Upward momentum transfer in the marine boundary layer. J. Phys. Oceanogr., 31, 1698–1711.
- Hanley, K. E., and S. E. Belcher, 2008: Wave-driven wind jets in the marine atmospheric boundary layer. J. Atmos. Sci., 65, 2646–2660.
- Harris, D. L., 1966: The wave-driven wind. J. Atmos. Sci., 23, 688–693.
- Högström, U., and Coauthors, 2008: Momentum fluxes and wind gradients in the marine boundary layer—A multi-platform study. *Boreal Environ. Res.*, **13**, 475–502.
- —, A. Smedman, E. Sahlée, W. M. Drennan, K. K. Kahma, H. Pettersson, and F. Zhang, 2009: The atmospheric boundary layer during swell—A field study and interpretation of the turbulent kinetic energy budget for high wave ages. J. Atmos. Sci., in press.
- Hristov, T. S., S. D. Miller, and C. A. Friehe, 2003: Dynamical coupling of wind and ocean waves through wave-induced air flow. *Nature*, **422**, 55–58.
- Janssen, P., 2004: The Interaction of Ocean Waves and Wind. Cambridge University Press, 300 pp.
- Kantha, L., 2006: A note on the decay rate of swell. Ocean Modell., 11, 167–173.
- Komen, G. J., L. Cavaleri, M. Doneland, K. Hasselmann, S. Hasselmann, and P. A. E. M. Janssen, Eds., 1994: *Dynamics* and Modelling of Ocean Waves. Cambridge University Press, 554 pp.
- Kudryavtsev, V. N., and V. K. Makin, 2004: Impact of swell on the marine atmospheric boundary layer. J. Phys. Oceanogr., 34, 934–949.
- Miller, S., C. Friehe, T. Hristov, J. Edson, and S. Wetzel, 1999: Wind and turbulent profiles in the surface layer over ocean waves. *Wind-Over-Wave Couplings*, S. G. Sajjadi, N. H. Thomas, and J. C. R. Hunt, Eds., Oxford University Press, 91–98.
- Phillips, O. M., Ed., 1966: The Dynamics of the Upper Ocean. Cambridge University Press, 269 pp.
- Pierson, W. L., A. W. Garcia, and S. E. Pells, 2003: Water wave attenuation due to opposing wind. J. Fluid Mech., 487, 345–365.

- Rutgersson, A., and P. P. Sullivan, 2005: The effect of idealized water waves on the turbulence structure and kinetic energy budgets in the overlying airflow. *Dyn. Atmos. Oceans*, **38**, 147–171.
- —, A.-S. Smedman, and U. Högström, 2001: Use of conventional stability parameters during swell. J. Geophys. Res., 106, 27 117–27 134.
- Smedman, A., M. Tjernström, and U. Högström, 1994: The nearneutral marine atmospheric boundary layer with no surface shearing stress: A case study. J. Atmos. Sci., 51, 3399–3411.
- —, U. Högström, H. Bergström, A. Rutgersson, K. K. Kahma, and H. Petterson, 1999: A case study of air-sea interaction during swell conditions. J. Geophys. Res., 104, 25 833–25 852.
- —, X. G. Larsén, U. Högström, K. K. Kahma, and H. Pettersson, 2003: Effect of sea state on the momentum exchange over the sea during neutral conditions. *J. Geophys. Res.*, **108**, 3367, doi:10.1029/2002JC001526.
- —, U. Högström, E. Sahleé, W. M. Drennan, K. K. Kahma, H. Pettersson, and F. Zhang, 2009: Observational study of marine atmospheric boundary layer characteristics during swell. J. Atmos. Sci., in press.

- Smith, S. D., 1992: Sea surface wind stress and drag coefficients: The HEXOS results. *Bound.-Layer Meteor.*, **60**, 109–142.
- Snodgrass, F. E., G. W. Groves, K. F. Hasselmann, G. R. Miller, W. H. Munk, and W. H. Powers, 1966: Propagation of ocean swell across the Pacific. *Philos. Trans. Roy. Soc. London*, 259A, 431–497.
- Stull, R. B., 1988: An Introduction to Boundary Layer Meteorology. Academic Press, 670 pp.
- Sullivan, P. P., J. C. McWilliams, and C.-H. Moeng, 2000: Simulation of turbulent flow over idealized water waves. J. Fluid Mech., 404, 47–85.
- —, J. B. Edson, T. Hristov, and J. C. McWilliams, 2008: Largeeddy simulations and observations of atmospheric marine boundary layers above nonequilibrium surface waves. J. Atmos. Sci., 65, 1225–1245.
- Tennekes, H., and J. L. Lumley, 1972: A First Course in Turbulence. MIT Press, 300 pp.
- Volkov, Y. A., 1970: Turbulent flux of momentum and heat in the atmospheric surface layer over a disturbed sea surface. *Izv. Atmos. Oceanic Phys.*, 6, 734–770.