JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 103, NO. C1, PAGES 1137–1152, JANUARY 15, 1998

Air-sea exchange of heat in the presence of wind waves and spray

V. K. Makin

Royal Netherlands Meteorological Institute, De Bilt, The Netherlands

Abstract. Waves extract a considerable part of the surface stress. While breaking, they eject spray into the atmosphere. Spray evaporates and influences a balance of heat and moisture above the waves. A one-dimensional model of the stratified marine surface boundary layer (MSBL) accounts directly for the impact of waves on the momentum flux and the impact of sea spray on fluxes of heat and moisture. The model is viewed as a higher order parameterization of the MSBL compared to the bulk parameterization. The model is based on the balance equations of momentum, the turbulent kinetic energy and the dissipation rate, heat, and moisture. A general experimental knowledge is used to parameterize the jet droplet concentration above the sea. That is, the surface droplet concentration is proportional to the cube of the friction velocity of the air, and the fast decay of droplet concentration with elevation above waves is parameterized by exponential decay. The exchange coefficients for heat, moisture, and momentum are computed from the wind speed and the sea state. Consistency of the dynamical part of the model is checked against measurements of the drag coefficient. Consistency of the thermodynamical part is checked against measurements of the sensible heat flux for light to moderate winds. The impact of spray is then assessed for stronger winds. It is shown that for a wind speed of about 25 $\mathrm{m\,s^{-1}}$ and above the impact of sea spray on heat and moisture fluxes becomes significant. The magnitude and the sign of the spray mediated heat and moisture fluxes depend on stratification of the atmosphere. To settle the issue whether or not sea spray plays an important role in exchange of heat and moisture above the sea, simultaneous direct measurements of sensible heat and moisture flux under different stratification conditions at wind speeds of about 25 $\mathrm{m\,s^{-1}}$ are needed.

1. Introduction

Wind waves, being the visual manifestation of the air-sea interaction, play an active role in this process. In fact, waves are responsible for the formation and regulation of the surface transfers of momentum, sensible heat, and moisture. The role of waves in air-sea interaction increases with increase of the wind speed. Waves enhance transfer rates even by fairly moderate winds when the sea surface is fairly smooth and continuous. They further enhance them when they begin to break and form sea sprays. The impact of waves on the dynamics of air flow in the marine surface boundary layer (MSBL) is now understood rather well [*Janssen*, 1989; *Makin*, 1990; *Chalikov and Makin*, 1991; *Caudal*, 1993; *Jenkins*, 1993; *Chalikov and Belevich*, 1993; *Makin et al.*, 1995; *Makin and Mastenbroek*, 1996]. Waves extract a considerable part of the surface stress (by the form drag mechanism, that is, the correlation of the surface wave-induced pressure with the wave slope) and determine the vertical structure of turbulence in the MSBL [*Makin and Mastenbroek*, 1996]. This results in considerable increase of sea drag (the exchange coefficient for momentum) with increase of the wind speed. This increase is described rather well by the Charnock relation for sea roughness [*Charnock*, 1955].

It is well established by field measurements that the sensible heat and humidity exchange coefficients over the sea are much less dependent on wind speed than the drag coefficient [*Anderson*, 1993; *DeCosmo et al.*, 1996; *Friehe and Schmitt*, 1976; *Geernaert*, 1990; *Katsaros et al.*, 1987, 1994; *Large and Pond*, 1982; *Smith*, 1980, 1988, 1989]. The wind speed dependence of the drag coefficient and sensible heat exchange coefficient can be explained by the difference in the exchange mechanism of momentum and heat at the sea surface [*Makin and Mastenbroek*, 1996]. Momentum to a large extent is transported by the organized wave-induced motions correlated with the waves (form drag). Heat is transported only by viscosity. Form drag dominates the surface stress and determines the vertical structure of tur-

Paper number 97JC02908.
0148-0227/98/97JC-02908$09.00

bulence in the MSBL. Sensible heat and humidity flux above the waves is determined by the diffusivity of turbulence, which is affected by waves. In this case, waves have only indirect impact on heat (sensible and latent) fluxes. Sensible heat and humidity exchange coefficients are taken usually as constants, i.e., independent of wind speed [*Anderson*, 1993; *DeCosmo et al.*, 1996; *Friehe and Schmitt*, 1976; *Large and Pond*, 1982; *Smith*, 1980, 1988, 1989].

However, waves can directly influence sensible heat and humidity fluxes. Waves break. They eject spray into the atmosphere. The spray droplets evaporate and can change the balance of sensible heat and moisture in the MSBL. An additional term, related to the evaporation of droplets, appears in the balance equations for heat and humidity. This term is "wave-induced" and is analogous to the form drag term in the balance equation for momentum above waves. If this wave-induced, or better expressed as the spray mediated sensible heat and humidity flux, can become comparable to a direct turbulent flux, then sea spray will play an important role in the exchange processes of sensible heat and moisture.

The first attempts to determine the effect of sea spray on evaporation from the sea surface, and thus on heat and moisture fluxes, were made in the early seventies [*Bortkovskii*, 1973; *Wu*, 1974]. Though considerable progress was made in the study and understanding of this problem during the last two decades (see a comprehensive review by *Andreas et al.* [1995]), many uncertainties remain. As is stated by *Andreas et al.* [1995, p.3], "the field is ... still rife with controversy because experimental work is difficult, the processes are complex and interactive, and theories of sea spray generation and heat and moisture transfer by spray are still rudimentary."

In the literature it is argued that for wind speeds above about 15 $\mathrm{m\,s^{-1}}$ the effect of the sea spray evaporation becomes important and influences sensible heat and humidity fluxes. *Andreas* [1992] has suggested a theory and found for a wind speed of $U_{10} = 20$ $\mathrm{m\,s^{-1}}$ a droplet mediated latent heat flux of 45% of a direct latent heat flux. *Fairall et al.* [1994] used a simplified and generalized derivative of an approach developed by *Andreas* [1992] and have shown that the contributions of oceanic spray to air-sea fluxes become comparable to direct turbulent fluxes at wind speeds above about 20 $\mathrm{m\,s^{-1}}$. On the contrary, *Hasse* [1992], using some general theoretical arguments, argued that there could hardly be any effect of spray evaporation on heat fluxes (perhaps except for hurricane wind strength).

The hostile environment of rough seas makes direct measurements of sensible and humidity fluxes over the open sea extremely difficult, and only few measurements exist. The main goal of the Humidity Exchange Over the Sea Main Experiment (HEXMAX) program [*DeCosmo et al.*, 1996; *Katsaros et al.*, 1987, 1994; *Smith et al.*, 1996] was to find accurate empirical heat and moisture flux parameterization formulas for high wind conditions over the sea. It had been postulated that breaking waves and sea spray, which dominate the air-sea interface at high wind speeds, would significantly affect the air-sea heat and moisture exchange for wind speeds above 15 $\mathrm{m\,s^{-1}}$. The outcome of the experiment was rather disappointing; no significant variation with wind speed has been found for wind speeds up to 18 $\mathrm{m\,s^{-1}}$ for the humidity exchange coefficient and up to 23 $\mathrm{m\,s^{-1}}$ for the sensible heat exchange coefficient. *Large and Pond* [1982] and *Anderson* [1993] reported measurements of sensible heat flux up to 23 $\mathrm{m\,s^{-1}}$ under unstable conditions. No significant variations with wind speed were found either. Thus far, there is no direct experimental evidence that the evaporation of droplets plays a significant role in heat and moisture transfers. Measurements beyond a wind speed of about 23 $\mathrm{m\,s^{-1}}$ are needed to settle the issue.

The effect of the evaporating droplets on heat and humidity transfers in the MSBL is determined by the evaporation function. The evaporation function is calculated from the droplet concentration above the sea waves. Many experiments were dedicated to establishing droplet distributions over the sea [*Blanchard and Woodcock*, 1980; *Bortkovskii*, 1987; *DeLeew*, 1986a, 1986b, 1987, 1990, 1993; *Monahan*, 1968; *Monahan et al.*, 1983; *Miller and Fairall*, 1988; *Preobrazhenskii*, 1973; *Smith et al.*, 1993; *Woolf et al.*, 1987; *Wu*, 1992; *Wu et al.*, 1984]. The rate of droplet generation at the sea surface is expressed usually in terms of a sea spray generation function, which can be related to the droplet concentration. The generation functions suggested by different authors are compiled by *Andreas* [1994] and *Andreas et al.* [1995], where detailed analyses are given. Though the spectral form of the generation functions seems to be reasonably established, the magnitude can vary significantly. It is generally known, however, that the generation of droplets is proportional to the energy flux from the wind, i.e., to the cube of the friction velocity of the air [*Andreas et al.*, 1995].

The vertical distribution of droplets is also not well established, and only a few measurements exist. Data of *Preobrazhenskii* [1973] gathered in the ocean suggest that the droplet concentration decreases exponentially with the distance from the sea surface. *Blanchard and Woodcock* [1980] came to the same conclusion. The data reported by *DeLeew* [1986a,b, 1987] show deviations from the exponential decay at heights close to the surface, while above 2 m his concentrations have a negative gradient, in agreement with the study of *Preobrazhenskii* [1973]. In his review, *Wu* [1990] argues that there are not enough data to quantify the vertical distribution of droplet concentrations in a consisting fashion. What is generally known, however, is that the concentration decreases very rapidly with elevation above the sea surface [*Wu*, 1990], which is well described by exponential decay. Laboratory measurements and their numerical simulation [*Edson and Fairall*, 1994; *Edson et* *al.*, 1996] also show that the concentration of droplets decays exponentially with height.

The present paper will try to contribute to understanding of this complicated problem by answering the following questions. If the contribution of the spray mediated flux in the balance equation for sensible heat and humidity is not negligible, how does this mediated flux influence the vertical distribution of temperature and humidity, and the vertical distribution of the direct turbulent flux of sensible heat and humidity above waves? What is the impact of the spray mediated flux on sensible heat and humidity exchange coefficients, and can this impact be measured?

A theoretical approach to describe the atmospheric boundary layer above the sea waves is based on balance equations of momentum, turbulent kinetic energy and dissipation, sensible heat, and humidity. The dynamical part of the model was developed by *Makin et al.* [1995], and *Makin and Mastenbroek* [1996], and in the latter paper the balance of sensible heat in near-neutral conditions was accounted for. The properties of the MSBL are related directly to the sea state described in terms of a wave spectrum. The lower boundary conditions are imposed on the instantaneous water surface, which allows one to relate the local roughness lengths to the scale of the molecular sublayer. In that way, no "Charnock type" relation for sea roughness is needed to be prescribed in the model. The roughness of the sea surface (or the drag coefficient) as well as the sensible heat exchange coefficient are obtained thus as a solution of the model. The open ocean data set of recent simultaneous measurements of momentum and sensible heat fluxes [*Anderson*, 1993] had been chosen to test the model results on the wind speed dependence of the drag coefficient and the sensible heat exchange coefficient. It was shown that, while the heat exchange coefficient remains virtually independent of wind speed in the range of 5 - 20 m s^{-1}, the calculated drag coefficient increases up to 100% in the same range of wind speed in excellent agreement with field measurements.

In the present paper the model is developed further to account for the impact of evaporating droplets on fluxes of sensible heat and humidity in a stratified atmosphere. To evaluate the "wave-induced" evaporation term in the balance equations of sensible heat and humidity, a simple model of droplet concentration above the waves, based on general experimental knowledge, is used. That is, the surface droplet concentration is proportional to the cube of the friction velocity of the air, and the fast decay of droplet concentration with elevation above waves is parameterized by exponential decay. The length of decay is related to the significant wave height [*Andreas*, 1992; *Andreas et al.*, 1995; *Fairall et al.*, 1994; *Edson et al.*, 1996]. The problem is simplified further: no spume produced droplets, resulting from the mechanical tearing of the sharpened wave crests by the wind, are considered. (As remarked by *Fairall et al.* [1994], this portion of the droplet flux spectrum is still speculative.) Only bubble-generated droplets are accounted for. The following conditions are required on the solution of the model. The calculated drag coefficient should agree with measurements. The calculated sensible heat exchange coefficients should agree with measurements, and further, no significant variations due to the impact of sea spray should be obtained for wind speeds below 18 m s^{-1}. The solution follows from the balance equations, which requires the finite and positive values for the exchange coefficients.

As a matter of fact, the droplet concentration function of *Wu* [1992] satisfies the above-mentioned conditions for wind speeds below 18 m s^{-1}. The exchange coefficients are then evaluated for wind speeds up to 30 m s^{-1}, and their wind speed dependence in the presence of sea spray is assessed. It is shown that for a wind speed of about 25 m s^{-1} and above the impact of sea spray becomes significant. Sensible heat and humidity exchange coefficients are increased (or decreased) up (or down) to 100% compared to their values at a wind speed of 10 m s^{-1}, when the spray effect is not important. It is shown that the magnitude and the sign of the sea spray mediated fluxes depend on stratification of the atmosphere. It is also concluded that to answer the question whether or not sea spray plays an important role in the exchanges of heat and humidity over the sea, direct measurements of sensible heat and humidity fluxes at about 25 m s^{-1} are needed.

2. Model Equations

The lowest part of the marine surface boundary layer, the so-called wave boundary layer (WBL), is considered. The upper boundary of the WBL is defined at the height where all wave-induced fluxes in the atmosphere are negligible. The $h = 10$ m height is a good estimate for the upper boundary of the WBL, defined by the decay of the wave-induced momentum flux [*Makin et al.*, 1995; *Makin and Mastenbroek*, 1996]. The fact that the droplet concentration decays exponentially with height above waves secures, as will be shown later, the decay of the wave-induced (spray mediated) fluxes of sensible heat and moisture at 10 m height.

Stationarity and spatial homogeneous conditions are assumed. The wind direction coincides with the mean direction of wave propagation, and the wave field (wave spectrum) is symmetrical relative to that direction.

The wave-induced fluxes are concentrated in the vicinity of the water surface. Explicit description of this region dictates the use of a vertical coordinate system which follows the instantaneous wave surface $z = \eta(\vec{x}, t)$, e.g.,

$$\hat{z} = h\frac{z-\eta}{h-\eta}. \tag{1}$$

The mean space averaged variable is defined in $(\vec{x}, \hat{z})$ coordinate system by the following procedure

$$f(\hat{z}) = \frac{1}{L}\int_0^L \bar{f}(\vec{x}, \hat{z})d\vec{x}, \quad (2)$$

where $\bar{f}$ is any ensemble phase averaged variable [see *Mastenbroek et al.,* 1996], and L is the averaging distance. The problem of the averaging process in geophysics is described, e.g., by *Monin and Yaglom* [1971], and *Donelan* [1990]. The averaging process, in our case the waves, must be contained within the space scale of the average, and this scale must be much smaller than the scale of the airflows that drive the waves. The estimated averaging timescale for air-sea studies appears to be about 1000 s [*Donelan*, 1990]. The characteristic distance/time of the wave process is the wave length/period of the wave component in the peak of a wave spectrum. For a fully developed sea, which will be considered in the present study, that is about 100 m and about 10 s. So, the averaging distance L is about 10^4 m.

The procedure (2) is defined in the domain $0 \leq \hat{z} \leq h$ $(\eta \leq z \leq h)$. Note that $\hat{z} = 0$ corresponds to the instantaneous water surface $z = \eta$. Hereafter the circumflex above z is dropped.

2.1. Balance Equations for Momentum, Heat, and Moisture

The spatially averaged balance equations of momentum, sensible heat, and moisture above waves read

$$\frac{\partial u}{\partial t} = -\frac{\partial \overline{u'w'}}{\partial z} + \frac{\partial \tau_m^w}{\partial z}, \quad (3)$$

$$\frac{\partial \theta}{\partial t} = -\frac{\partial \overline{\theta' w'}}{\partial z} + \mathcal{E}(z), \quad (4)$$

$$\frac{\partial q}{\partial t} = -\frac{\partial \overline{q' w'}}{\partial z} + E(z). \quad (5)$$

In equations (3)-(5), u is the mean horizontal wind speed, θ is the mean potential temperature in K, q is the mean specific humidity in $\mathrm{kg_w\,kg_a^{-1}}$. The $\tau_m^t = -\overline{u'w'}$, $\tau_\theta^t = -\overline{\theta' w'}$, $\tau_q^t = -\overline{q'w'}$ are the mean turbulent kinematic fluxes of momentum, sensible heat and moisture. In equation (3), τ_m^w is the wave-induced stress [e.g., *Janssen*, 1989; *Makin et al.*, 1995]:

$$\tau_m^w(z) = \int_0^\infty \int_{-\pi/2}^{\pi/2} \omega^2 S(k,\phi)\beta(k,\phi)\cos\phi d\phi e^{-z/z_d(k)} kdk. \quad (6)$$

Equation (6) presumes that (1) all undulations of the sea surface are considered as waves, which can then be statistically described by a directional wave spectrum $S(k,\phi)$, where the wave number k satisfies the dispersion relation

$$\omega^2 = gk + \mathcal{T}k^3 \quad (7)$$

($\mathcal{T}$ is the dynamical surface water tension and ϕ is the propagation direction of the k wave component, and ω is the wave frequency); (2) the energy input to waves from the atmosphere is known and described in terms of the growth rate parameter $\beta(k,\phi)$; (3) the decay length of the wave-induced flux of the individual wave component $z_d(k)$ is known. The decay length for the second wave-induced moment should be at least $z_d(k) = (2k)^{-1}$, but could be smaller [*Mastenbroek et al.*, 1996] as a result of the smearing of the wave fluctuations of turbulent stresses in the outer region [*Belcher and Hunt*, 1993]. The value of $z_d(k) = (5k)^{-1}$ was adopted by *Makin and Mastenbroek* [1996]; the value of $z_d(k) = (2k)^{-1}$ was used by *Kitaigorodskii and Donelan* [1984] and *Makin et al.* [1995]. *Makin and Mastenbroek* [1996] showed that the drag coefficient reduces with $(5k)^{-1}$ decay length, compared to $(2k)^{-1}$ decay, but only for 5%, which is not significant.

In equations (4) and (5) the E and $\mathcal{E}$ are the source (sink) functions resulting from the evaporation of spray. The function $\mathcal{E}$ relates to the evaporation function E:

$$\mathcal{E} = -\frac{L_v}{C_p}E \quad (8)$$

where $L_v \cong 2.45 \times 10^6\,\mathrm{m\,s^{-2}}$ is the latent heat of vaporization of water, $C_p = C_{pd}(1 + 0.84q)$ is the specific heat of moist air at constant pressure, and $C_{pd} = 1005\,\mathrm{m\,s^{-2}\,K^{-1}}$ is that of the dry air. The dimension of E is $\mathrm{kg_w\,kg_a^{-1}\,s^{-1}}$.

The evaporation of droplets increases humidity of the air and thus $E > 0$, and decrease the air temperature.

It is convenient to rewrite equations (4) and (5) in the flux form

$$\frac{\partial \theta}{\partial t} = -\frac{\partial \overline{\theta' w'}}{\partial z} + \frac{\partial \tau_\theta^s}{\partial z}, \quad (9)$$

$$\frac{\partial q}{\partial t} = -\frac{\partial \overline{q' w'}}{\partial z} + \frac{\partial \tau_q^s}{\partial z}, \quad (10)$$

where

$$\tau_\theta^s(z) = -\int_z^\infty \mathcal{E}(z)dz, \quad (11)$$

$$\tau_q^s(z) = -\int_z^\infty E(z)dz \quad (12)$$

are the spray mediated fluxes of sensible heat, and moisture.

In the stationary WBL the total flux of momentum, sensible heat and moisture is constant over height. The wave-induced fluxes τ_m^w, τ_θ^s, and τ_q^s are negligible at the top of the WBL $z = h$, which is secured by the exponential decay of the wave-induced momentum flux, equation (6), and by the exponential decay of the droplet concentration, equation (51). Thus constants are defined by the turbulent fluxes at the top of the WBL, i.e., $-\overline{u'w'}_h = u_*^2$, $-\overline{\theta'w'}_h = \theta_* u_*$, and $-\overline{q'w'}_h = q_* u_*$. The total fluxes are

$$\tau_m = \tau_m^t(z) + \tau_m^w(z) = \text{const} = u_*^2 = -\overline{u'w'}_h, \quad (13)$$

$$\tau_\theta = \tau_\theta^t(z) + \tau_\theta^s(z) = \text{const} = \theta_* u_* = -\overline{\theta'w'}_h, \quad (14)$$

$$\tau_q = \tau_q^t(z) + \tau_q^s(z) = \text{const} = q_* u_* = -\overline{q'w'}_h. \quad (15)$$

2.2. Balance Equation of the TKE

The balance equations of the turbulent kinetic energy (TKE) e and the dissipation rate ε in a stratified atmosphere read

$$-\frac{\partial}{\partial z}\overline{(e'+p')w'} + P - \varepsilon = 0 \quad (16)$$

and

$$-\frac{\partial \overline{\varepsilon' w'}}{\partial z} + \frac{\varepsilon}{e}(c_{1\varepsilon}P - c_{2\varepsilon}\varepsilon) = 0. \quad (17)$$

In equations (16) and (17) the first terms on the left-hand side are the diffusive transport of the TKE and ε, P is the production of the TKE, $c_{1\varepsilon}$ and $c_{2\varepsilon}$ are empirical constants to be defined in the next section. Production of the TKE above waves has three contributions: production by interaction of the turbulent stress and mean velocity shear P^t, by interaction of the mean wave-induced stress τ_m^w and mean velocity shear P^w, and by buoyancy B:

$$P = P^t + P^w + B. \quad (18)$$

The production terms P^t and P^w represent the transfer of kinetic energy from the mean and mean wave-induced motions to the turbulent motion. Above the waves, they have the following form [*Makin and Mastenbroek*, 1996]:

$$P^t = \tau_m^t \frac{\partial u}{\partial z}, \quad (19)$$

$$P^w = \tau_m^w \frac{\partial u}{\partial z} + \frac{\partial \Pi}{\partial z} \quad (20)$$

where $\Pi(z)$ is the wave-induced flux of energy

$$\Pi(z) = \int_0^\infty \int_{-\pi/2}^{\pi/2} \omega^2 c S \beta k d\theta e^{-z/z_d(k)} dk \quad (21)$$

and $c = \omega/k$ is the phase speed of component k.

The buoyancy production term is [e.g., *Stull*, 1991]

$$B = \frac{g}{\theta_0}\overline{\theta'w'} + 0.61 g \overline{q'w'}. \quad (22)$$

2.3. Turbulent Closure Scheme

The local closure, eddy-viscosity K theory, which relates the turbulent flux to the gradient of the associated mean variable is used here. The turbulent fluxes which appear in the balance equations of momentum, heat, humidity, TKE and dissipation rate have the following form.

Turbulent stress is

$$\tau_m^t(z) = -\overline{u'w'} = K\frac{\partial u}{\partial z}. \quad (23)$$

Turbulent diffusive flux of the TKE is

$$-\overline{(e'+p')w'} = \frac{K}{\sigma_e}\frac{\partial e}{\partial z}. \quad (24)$$

Turbulent diffusive flux of the dissipation rate is

$$-\overline{\varepsilon' w'} = \frac{K}{\sigma_\varepsilon}\frac{\partial \varepsilon}{\partial z}. \quad (25)$$

Turbulent flux of sensible heat is

$$\tau_\theta^t(z) = -\overline{\theta'w'} = K_\theta \frac{\partial \theta}{\partial z}. \quad (26)$$

Turbulent flux of moisture is

$$\tau_q^t(z) = -\overline{q'w'} = K_q \frac{\partial q}{\partial z}. \quad (27)$$

The eddy-viscosity K is calculated from the TKE and the dissipation rate:

$$K = c_\mu \frac{e^2}{\varepsilon}. \quad (28)$$

The turbulent diffusivity of heat K_θ and moisture K_q depends on the state of turbulence and is closely related to the eddy-viscosity K via the turbulent Prandtl and Schmidt numbers:

$$K_\theta = \frac{K}{\mathrm{Pr}_t}, \quad (29)$$

$$K_q = \frac{K}{\mathrm{Sc}_t}. \quad (30)$$

Values of empirical constants which appear in (23)-(28) and in the balance equation for the dissipation rate (17) are $c_\mu = 0.09$, $c_{1\varepsilon} = 1.44$, $c_{2\varepsilon} = 1.92$, $\sigma_\varepsilon = 1.30$, $\sigma_e = 1$. These values are commonly used in simulations of atmospheric turbulent flows, [e.g., *Stull*, 1991] and of turbulent flows in different engineering applications [e.g., *Rodi*, 1984]. As is mentioned, e.g., by *Donelan* [1990] and *Rodi* [1984], the buoyancy forces produced by temperature and humidity differences affect the values of the turbulent Prandtl and Schmidt numbers. Empirical data compiled, e.g., by *Zilitinkevich* [1970, Figure 1.22], show that the Prandtl number for stable stratification is clearly larger than that for unstable stratification. Based on these data, the value of $\mathrm{Pr}_t = 0.9$, for unstable and the value of $\Pr_t = 1.25$ for stable stratification is adopted here. Further, there is commonly assumed that for the atmospheric flows $\mathrm{Sc}_t = \Pr_t$ [*Large and Pond*, 1982; *Donelan*, 1990].

2.4. Wave Spectrum and the Growth Rate Parameter

Makin et al. [1995], and *Makin and Mastenbroek* [1996] used the wave spectrum of *Donelan et al.* [1985], and *Donelan and Pierson* [1987]. However, this wave spectrum model is criticized [*Elfouhaily et al.*, 1997] for overestimating the total mean squared slope parameters. Moreover, it fails to reproduce the capillary-gravity spectrum range, as measured from controlled laboratory measurements [*Jähne and Riemer*, 1990]. *Elfouhaily et al.* [1997] have suggested an empirical wave spectrum model which does not suffer from these deficiencies. In this study this wave spectrum model is used. The model is described briefly below.

The omni-directional spectrum has the general form

$$S(k) = k^{-4}(B_l + B_h). \tag{31}$$

In (31) the long wave curvature spectrum B_l is

$$B_l = \frac{1}{2}\alpha_p \frac{c_p}{c} F_p J_p \tag{32}$$

where

$$\alpha_p = 0.006\Omega^{0.5} \tag{33}$$

is the equilibrium range parameter, dependent on the inverse wave-age parameter $\Omega = U_{10}/c_p$; $c_p = c(k_p)$ is the phase speed at the spectral peak and $c = c(k)$ is the phase speed. The inverse wave-age parameter used in this study is $\Omega = 0.83$, which corresponds to a fully developed sea. The side-effect function F_p is

$$F_p = \exp\left\{-\frac{\Omega}{\sqrt{10}}\left(\sqrt{\frac{k}{k_p}} - 1\right)\right\}. \tag{34}$$

J_p is the JONSWAP peak enhancement function

$$J_p = \gamma^\Gamma \exp\left\{-\frac{5}{4}\left(\frac{k_p}{k}^2\right)\right\}, \tag{35}$$

$$\gamma = 1.7 + 6\log(\Omega),$$
$$\Gamma = \exp(-(\sqrt{k/k_p} - 1)^2/2\sigma^2),$$
$$\sigma = 0.08(1 + 4\Omega^{-3}). \tag{36}$$

The short wave curvature spectrum B_h is

$$B_h = \frac{1}{2}\alpha_m \frac{c_m}{c} F_m J_p \tag{37}$$

where $c_m = 0.23\,\mathrm{m\,s^{-1}}$ is the phase speed at the secondary gravity-capillary peak, and $\alpha_m = 0.0625u_*$ is the equilibrium range parameter for short waves. The short wave side-effect function F_m is

$$F_m = \exp\left\{-\frac{1}{4}\left(\frac{k_m}{k} - 1\right)^2\right\}. \tag{38}$$

The directional spectrum $S(k, \phi)$ is written by *Elfouhaily et al.* [1997] in the form

$$S(k, \phi) = S(k)\Phi(k, \phi), \tag{39}$$

and the angular spreading function is

$$\Phi(k, \phi) = \frac{1}{\pi}[1 + \Delta(k)\cos(2\phi)], \tag{40}$$

$$\Delta(k) = \tanh\left\{a_0 + a_p(\frac{c}{c_p})^{2.5} + a_m(\frac{c_m}{c})^{2.5}\right\}.$$

In (40), $a_0 = 0.173$, $a_p = 4$, and $a_m = 0.565u_*$. The mean squared elevation is defined by

$$\overline{\eta^2} = \int_0^\infty \int_{-\pi/2}^{\pi/2} S(k, \phi) k dk d\phi. \tag{41}$$

The wave growth rate parameter β is defined as

$$\frac{\rho_a}{\rho_w}\beta = \frac{1}{\omega S}\frac{\partial S}{\partial t}. \tag{42}$$

The β parameterization adopted by *Makin et al.* [1995] is used in this study:

$$\beta = 1.25\frac{u_{*\lambda}}{c}\cos\phi(\frac{u_\lambda}{c}\cos\phi - 1.15), \tag{43}$$

where u_λ is the wind speed and $u_{*\lambda}$ is the friction velocity at height $\lambda(k) = 2\pi/k$.

2.5. Evaporation Function

After *Pruppacher and Klett* [1978] (see also *Fairall et al.* [1994]), the evaporation function $E(z)$ can be calculated from the known distribution of the droplet concentration $n(r, z)$. The rate of loss of mass by evaporation $\dot{m}$ by a single droplet of radius r (in m) is [*Pruppacher and Klett*, 1978, pp. 414 and 440]

$$\dot{m} = -4\pi F_p D_v \rho_a r(q_s - q) \tag{44}$$

where $\rho_a = 1.23\,\mathrm{kg_a\,m^{-3}}$ is the density of the air, q is the ambient specific humidity, and q_s is the saturation specific humidity at the droplet temperature. The diffusivity of water vapor D_v [*Pruppacher and Klett*, 1978, p. 413] is

$$D_v = 0.21 \times 10^{-4}\left(\frac{T}{T_0}\right)^{1.94}\left(\frac{p_0}{p}\right) \tag{45}$$

with $T_0 = 273°\mathrm{K}$, $p_0 = 101.3\,\mathrm{kPa}$. Dimension of D_v is in $\mathrm{m^2\,s^{-1}}$. For the present study we shall use for pressure p its value at sea level for a standard atmosphere, i.e., $p = p_0$. The difference between the absolute temperature T and the potential temperature θ in the surface boundary layer can be neglected.

The droplet ventilation factor [*Pruppacher and Klett*, 1978, p. 440; *Fairall et al.*, 1994] is

$$F_p = 1 + 0.25\left(\frac{2V_f r}{\nu}\right)^{0.5} \tag{46}$$

where $\nu = 14 \times 10^{-6}\,\mathrm{m^2\,s}$ is the kinematic viscosity of the air and $V_f = 8000r$ is the terminal fall velocity [*Pruppacher and Klett*, 1978, pp. 323-324], in $\mathrm{m\,s^{-1}}$.

To account for the fact that droplets are maintained at the wet bulb temperature rather than the air temperature, we follow *Fairall et al.* [1994] and write

$$q_s - q = \gamma\left[q_s(\theta) - q\right] \tag{47}$$

where γ is given by

$$\gamma = \left[1 + \frac{\epsilon L_v^2}{RC_pT^2}q_s(\theta)\right]^{-1}, \tag{48}$$

$\epsilon = 0.622$ is the ratio of gas constants for air and water vapor, and $R = 287\,\mathrm{m^2\,s^{-2}\,K^{-1}}$ is the gas constant for air.

It is convenient to express the saturation specific humidity q_s as a function of temperature. According to *Stull* [1991, p. 276]

$$q_s = \epsilon\frac{0.6112}{p_0}\exp\left[\frac{17.67(T - 273.16)}{T - 29.66}\right]. \tag{49}$$

The total loss rate of liquid water by evaporation at some level z is obtained by integrating equation (44) over the number concentration of droplets, $n(r, z)$:

$$-E(z) = \int \dot{m}n(r,z)dr. \tag{50}$$

As $q_s(\theta_z) - q(\theta_z) > 0$, $\dot{m} < 0$, and the evaporation function $E(z)$ is positively defined.

The vertical profiles of temperature θ_z and humidity q_z are found from equations (14)-(15) and (26)-(27), after the boundary conditions at the surface and at the top of the WBL are specified, and the evaporation function E is known. To evaluate the evaporation function E, the vertical distribution of the droplet concentration $n(r, z)$ should be known.

2.6. Droplet Concentration

The approach of *Smith et al.* [1993] is followed here, and it is assumed that the average droplet concentration, at a given wind speed, represents the equilibrium distribution. That is, the loss of droplets due to turbulent deposition and gravitational sedimentation is balanced by the new droplet production.

To evaluate the average droplet concentration $n(r, z)$ above the waves, a simple model based on general experimental knowledge is used. The model captures the main known features of the droplet distribution in the MSBL, namely, the proportionality of the droplet concentration to the cube of the friction velocity (wind speed) of the air, and a fast decay of droplet concentration with height above the waves.

The sea spray generated by bursting bubbles, produced by the air entrainment by breaking waves, is discussed here. Two types of droplets are produced by bursting bubbles. Film droplets are produced through the fragmentation of bubble caps, and jet droplets are produced through the breakup of a water jet formed by the collapse of bubble cavity [*Wu*, 1992]. As was shown by *Wu* [1989], the sea spray under various wind conditions consists mainly of jet droplets, as the dominance of jet droplets over film droplets is just too overwhelming. The contribution of film droplets to the average concentration thus can be neglected.

Field measurements [*Preobrazhenskii,* 1973], laboratory measurements, and their numerical simulation [*Edson and Fairall*, 1994; *Edson et al.*, 1996] show that the concentration of jet droplets decays exponentially with height

$$n(r,z) = N(r)_0\exp(-z/z_e) \tag{51}$$

where z_e is a characteristic decay length and $N(r)_0$ is the surface droplet concentration. It is usually assumed that the decay length is proportional to the significant wave amplitude [*Andreas*, 1992; *Andreas et al.*, 1995; *Fairall et al.*, 1994; *Edson et al.*, 1996], which is a half of the significant wave height $z_e = 1/2H_s$. The latter can be directly calculated from the directional wave spectra (39), (31), (40), and (41):

$$H_s = 4\sqrt{\overline{\eta^2}}. \tag{52}$$

(For the fetch limited conditions, relation (52) with the wind wave spectra (39) corresponds to the relation $z_e = 1/2H_s = 0.015U_{10}^2$, which is actually used by the above-mentioned authors.)

Wu [1992] suggested calculating the concentration of jet droplets from the bubble population at the sea surface $n_b(D)$

$$n_b(D) = N_0p(D) \tag{53}$$

where D is a bubble diameter in μm and $p(D)$ is a probability density

$$\begin{array}{ll} p(D) = 0 & D < 67\,\mu\mathrm{m} \\ p(D) = 0.015 & 67\,\mu\mathrm{m} < D < 100\,\mu\mathrm{m} \\ p(D) = 1.5\times10^6D^{-4} & D > 100\,\mu\mathrm{m}. \end{array} \tag{54}$$

N_0 is the total bubble concentration per cubic meter at the surface, given by *Wu* [1992] as

$$N_0 = 2.9\times10^6u_*^3. \tag{55}$$

The production of jet droplets by bubbles is expressed by *Wu* [1992]:

$$n_j(D) = 7\exp(-D/3000), \tag{56}$$

and the concentration of droplets as a function of the bubble diameter is thus [*Wu*, 1992]

$$N(D)_0 = N_0p(D)n_j(D). \tag{57}$$

Relating the bubble diameter to the droplet radius [*Wu*, 1992] $D = 16r$ the vertical droplet concentration $n(r, z)$, equation (51), and the evaporation function $E(z)$, equation (50) can be finally calculated. The range of the jet droplets contributing mainly to the total droplet concentration is $5\,\mu$m $< r < 100\,\mu$m [*Andreas*, 1994; *Edson and Fairall*, 1994; *Fairall et al.*, 1994].

2.7. Boundary Conditions

Each of the balance equations (3)-(5), (16), (17) requires two boundary conditions. The upper boundary condition at the top of the WBL $h = 10$ m is

$$z = h : u = U_{10}, \quad \frac{de}{dz} = 0, \quad \frac{d\varepsilon z}{dz} = 0, \qquad \theta = \theta_h, \quad q = q_h. \tag{58}$$

The surface boundary condition is related to the local roughness length

$$z = z_0^l : u = u_0, \quad \frac{de}{dz} = 0, \quad \frac{d\varepsilon z}{dz} = 0; \qquad z = z_{0\theta}^l : \theta = \theta_0; \quad z = z_{0q}^l : q = q_0 \tag{59}$$

where u_0 is the surface current of any origin, here taken to be zero. The surface value of humidity equals its saturation value over pure water $q_0 = q_s(\theta_0)$ and $q_0 = 0.98 q_s(\theta_0)$ over saltwater [*Large and Pond*, 1982]. With q_r being the relative humidity of the air at height h, the ambient humidity at that height is $q_h = q_r q_s(\theta_h)$.

The local roughness reflects the local dynamical properties of the instantaneous water surface, which if not actively involved in breaking, is a smooth surface. Except at low wind speeds, breaking always occurs in the ocean. It is a highly nonlinear local process which disrupts the water surface. However, if the process is weak-in-the mean, as in the formalism widely used in the description of sea waves [*Hasselmann*, 1968; *Komen et al.*, 1994], the instantaneous water surface can be regarded as a smooth one. Using for the instantaneous fraction of the sea surface covered by whitecaps relation $W(U) = 3.84 \times 10^{-6} U^{3.41}$ [e.g., *Monahan et al.*, 1986], one can find the whitecap coverage of 1%, 4%, 10%, and 22% of the sea surface for the corresponding wind speeds of 10 m s^{-1}, 15 m s^{-1}, 20 m s^{-1}, and 25 m s^{-1}, which supports the weak-in-the mean assumption for wave breaking.

Notice that the instantaneous water surface is distinguished from the sea surface. The latter, due to the presence of waves, is a rough surface.

As all undulations of the sea surface are considered to be waves which are covered by the molecular sublayer, the instantaneous water surface can be treated as smooth, as was shown by *Makin et al.* [1995]. The local roughness lengths for momentum, temperature, and humidity are thus related to the scale of the molecular sublayer:

$$z_0^l = c_0 \frac{\nu}{u_*^l}, \quad z_{0\theta}^l = c_{0\theta} \frac{\nu}{u_*^l}, \quad z_{0q}^l = c_{0q} \frac{\nu}{u_*^l} \tag{60}$$

where the local friction velocity u_*^l is

$$u_*^l = u_* \sqrt{1 - \frac{\tau_m^w(0)}{u_*^2}}. \tag{61}$$

The surface turbulent stress $\tau_m^t(0) = (u_*^l)^2$ in (61) is only a part of the total stress $\tau_m = u_*^2$, as the form drag $\tau_m^w(0)$ (calculated from (6) at the instantaneous water surface $z = 0$) dominates the surface momentum flux for wind speeds above 6 m s^{-1} [*Janssen*, 1989; *Makin et al.*, 1995].

For the momentum roughness length z_0^l, $c_0 = 0.1$ is adopted [e.g., *Monin and Yaglom*, 1971]. *Kader and Yaglom* [1972] suggested that

$$c_{0\theta} = \exp\left(-\frac{\mathcal{F}(\mathrm{Pr})\kappa}{\mathrm{Pr}_t}\right), \tag{62}$$

where $\mathcal{F}$ is a function of the Prandtl number $\mathrm{Pr} = \nu/\chi_H \simeq 0.70$ (χ_H is the diffusivity of heat):

$$\mathcal{F}(\mathrm{Pr}) = 12.5\mathrm{Pr}^{2/3} + 2.12 \ln \mathrm{Pr} - 5.3. \tag{63}$$

From (62), (63) the value of $c_{0\theta}$ is 0.21 for the turbulent Prandtl number $\mathrm{Pr}_t = 1$, and is 0.29 for $\mathrm{Pr}_t = 1.25$. *Liu et al.* [1979] recommend a value of $c_{0\theta} = 0.18$ for temperature and a value of $c_{0q} = 0.29$ for humidity. *Makin and Mastenbroek* [1996] showed that the solution is not sensible to these constants. A value of 0.21 is adopted here for both temperature and humidity roughness lengths.

Note that the local roughness lengths follow from the solution of the problem, as local friction velocity (61) depends on the form drag $\tau_m^w(0)$ and friction velocity u_*^2.

A relaxation method [*Press et al.*, 1992, p. 753], is used to solve the balance equations (3)-(5), (16), (17) with boundary conditions (58) and (59).

3. Results

3.1. Quantitative Analysis

In the bulk parameterization, the fluxes are related to the measured variables at the surface and at a certain height via the exchange coefficients. The kinematic sensible heat flux $\theta_* u_*$ and moisture flux $q_* u_*$ can be rewritten

$$\theta_* u_* = -\overline{\theta' w'} = C_H \Delta u \Delta \theta, \tag{64}$$

$$q_* u_* = -\overline{q' w'} = C_E \Delta u \Delta q \tag{65}$$

where C_H and C_E are the sensible heat and humidity exchange coefficients and Δ denotes the difference between a variable f at a certain height h and its surface value, i.e.,

$$\Delta f = f_{z=h} - f_{z=0}. \tag{66}$$

The total fluxes of heat τ_θ, equation (14), and moisture τ_q, equation (15), are constant over height in the stationary WBL which immediately follows from equations (9) and (10). Using relations (26) and (27), equations (14) and (15) can be integrated from the surface to the top of the WBL, and the following relations can be easily obtained for the sensible heat and humidity exchange coefficients. For sensible heat it reads

$$C_H = C_H^0 + C_H^s, \tag{67}$$

where

$$C_H^0 = \left[\Delta u \int_{z_{0\theta}}^{h} K_\theta^{-1} dz\right]^{-1} \tag{68}$$

is the sensible heat exchange coefficient in the absence of spray (or in the presence of spray, a part which is supported by the turbulent flux τ_θ^t), and

$$C_H^s = \frac{1}{\Delta\theta} \int_{z_{0\theta}}^{h} \tau_\theta^s K_\theta^{-1} dz \left[\Delta u \int_{z_{0\theta}}^{h} K_\theta^{-1} dz\right]^{-1} \tag{69}$$

is the spray mediated exchange coefficient.

For moisture the equation is

$$C_E = C_E^0 + C_E^s, \tag{70}$$

where

$$C_E^0 = \left[\Delta u \int_{z_{0q}}^{h} K_q^{-1} dz\right]^{-1} \tag{71}$$

is the humidity exchange coefficient in the absence of spray (or in the presence of spray, a part which is supported by the turbulent flux τ_q^t), and

$$C_E^s = \frac{1}{\Delta q} \int_{z_{0q}}^{h} \tau_q^s K_q^{-1} dz \left[\Delta u \int_{z_{0q}}^{h} K_q^{-1} dz\right]^{-1} \tag{72}$$

is the spray mediated exchange coefficient.

As the evaporation function E is positively defined by (50), i.e., $E(z) > 0$, and thus from equation (8), $\mathcal{E}(z) < 0$, the spray mediated flux of moisture, equation (12), is negative $\tau_q^s < 0$ while that of heat, equation (11), is positive $\tau_\theta^s > 0$.

It follows from (67)-(72) that for unstable conditions ($\Delta\theta < 0$ and $\Delta q < 0$) the spray mediated coefficient $C_H^s < 0$, while $C_E^s > 0$, and thus

$$\begin{aligned} C_H &< C_H^0 \\ C_E &> C_E^0. \end{aligned} \tag{73}$$

For stable conditions ($\Delta\theta > 0$ and $\Delta q > 0$) the opposite is true. The spray mediated coefficient $C_H^s > 0$ while $C_E^s < 0$, and thus

$$\begin{aligned} C_H &> C_H^0 \\ C_E &< C_E^0. \end{aligned} \tag{74}$$

For the small positive value of $\Delta\theta \simeq 1^o - 3^o$, the difference $\Delta q = q_r q_s(\theta_h) - q_z(\theta_0)$ can be negative, depending on the value of the relative humidity q_r. In this case

$$\begin{aligned} C_H &> C_H^0 \\ C_E &> C_E^0. \end{aligned} \tag{75}$$

If we now integrate equations (9) and (10) over height, we obtain

$$\frac{\partial}{\partial t} \int_{z0\theta}^{h} \theta dz = \tau_\theta^t(h) - \tau_\theta^t(0) - \tau_\theta^s(0), \tag{76}$$

$$\frac{\partial}{\partial t} \int_{z0q}^{h} q dz = \tau_q^t(h) - \tau_q^t(0) - \tau_q^s(0), \tag{77}$$

as the spray mediated fluxes are negligible at the top of the WBL. In the absence of spray, the turbulent fluxes of sensible heat and moisture are constant over height and the mean temperature and humidity profiles follow the diabatic distribution. When droplets evaporate, they generate the spray mediated fluxes. Terms $\tau_\theta^s(0)$ and $\tau_q^s(0)$ get the right-hand side of equations (76) and (77) out of balance. For heat $\partial/\partial t \int_{z0q}^{h} q dz < 0$ as $\tau_\theta^s(0) > 0$; for moisture $\partial/\partial t \int_{z0q}^{h} q dz < 0$ as $\tau_q^s(0) < 0$. To satisfy the constant total flux conditions (14)-(15), the temperature has to decrease, while humidity has to increase. This originates temperature and humidity profiles that differs from the diabatic ones.

The typical temperature and humidity profiles in the presence of evaporating droplets (under the condition that the spray mediated flux is comparable to the turbulent flux) as well as the total flux balance are shown in Figure 1 for the case of unstable, and in Figures 2 and 3 for the case of stable stratification. All three figures show (upper panels) that in the presence of evaporating droplets temperature is indeed decreased while humidity is increased, for both unstable and stable stratification. However, deviation from the diabatic profiles is so small that it cannot be measured.

Under unstable conditions (Figure 1) the spray mediated sensible heat flux has a negative sign. It counteracts the turbulent flux, which results in decrease of the total flux $\tau_\theta = \theta_* u_*$. The initial (no spray effect) total flux, which is supported only by the turbulence flux, in this case is 2.5 times larger. In contrast, the spray mediated humidity flux is added to the turbulent flux. The total humidity flux $\tau_q = q_* u_*$ increases compared to the initial total flux.

In strong stable conditions (in this case, $\Delta\theta = +5^o$ and the humidity difference Δq is positive) (Figure 2) the situation is just the opposite. While the heat spray mediated flux adds to the turbulent heat flux and increases the total flux, the humidity mediated flux has a negative sign and considerably reduces the total flux. When stability is not strong and is defined by the pos-

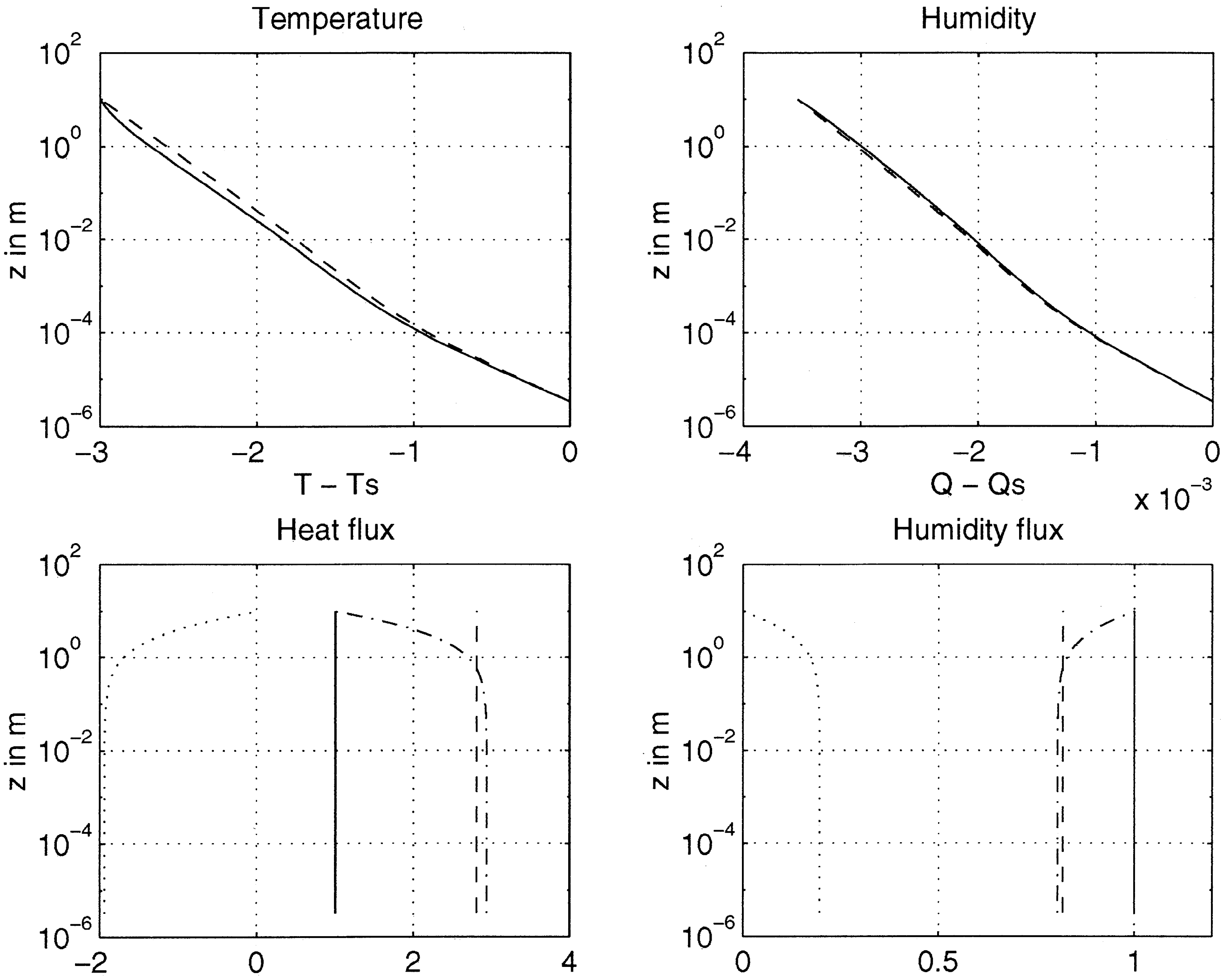

Figure 1. (top) Temperature and humidity profiles. The surface values are substracted. Dashed line indicates no spray effect; solid line, spray effect is accounted for. (bottom) Sensible heat and humidity flux profiles. Fluxes are normalized on their surface values $\theta_* u_*$ and $q_* u_*$ when spray effect is accounted for. Solid line shows the total flux, spray effect is accounted for (always equals 1); dashed-dotted line shows the turbulent flux; dotted line, spray mediated flux; dashed line, the total flux, no spray effect. Wind speed $U_{10} = 25\,\mathrm{m\,s^{-1}}$; unstable stratification $\Delta\theta = -3^o$.

itive temperature difference (Figure 3, $\Delta\theta = +1.5^o$) while the humidity difference Δq is negative, both sensible heat and humidity mediated fluxes add to the turbulent counterparts and increase the total fluxes of heat and moisture.

The sharp changes in turbulent flux of sensible heat and moisture occur in the layer above about 1 m over the instantaneous water surface and can, in principle, be measured with the use of a wave following device. Though, for high wind conditions when the impact of spray on the turbulent fluxes is anticipated to be pronounced, that would be a challenging technical operation.

An important conclusion can be drawn from the above analysis. If the spray mediated fluxes play a role in the balance of sensible heat and moisture over the sea, the simultaneously measured fluxes of sensible heat and humidity should show a tendency to decrease the sensible heat exchange coefficients while a tendency to increase the humidity exchange coefficients with increasing wind speed in unstable conditions, compared to their values under light and moderate winds, when spray is believed to play no role in exchanges. The situation is different for stable conditions over the sea. The sensible heat exchange coefficients have to show a tendency to increase while the humidity exchange coefficients have to show a tendency to decrease, if both $\Delta\theta$ and Δq are positive (relations (74), Figure 2). If Δq is negative, both coefficients C_H and C_E have a tendency to increase, (relations (75), Figure 3). It is also clear that the analysis of the measured fluxes should be done separately for each of the cases described above.

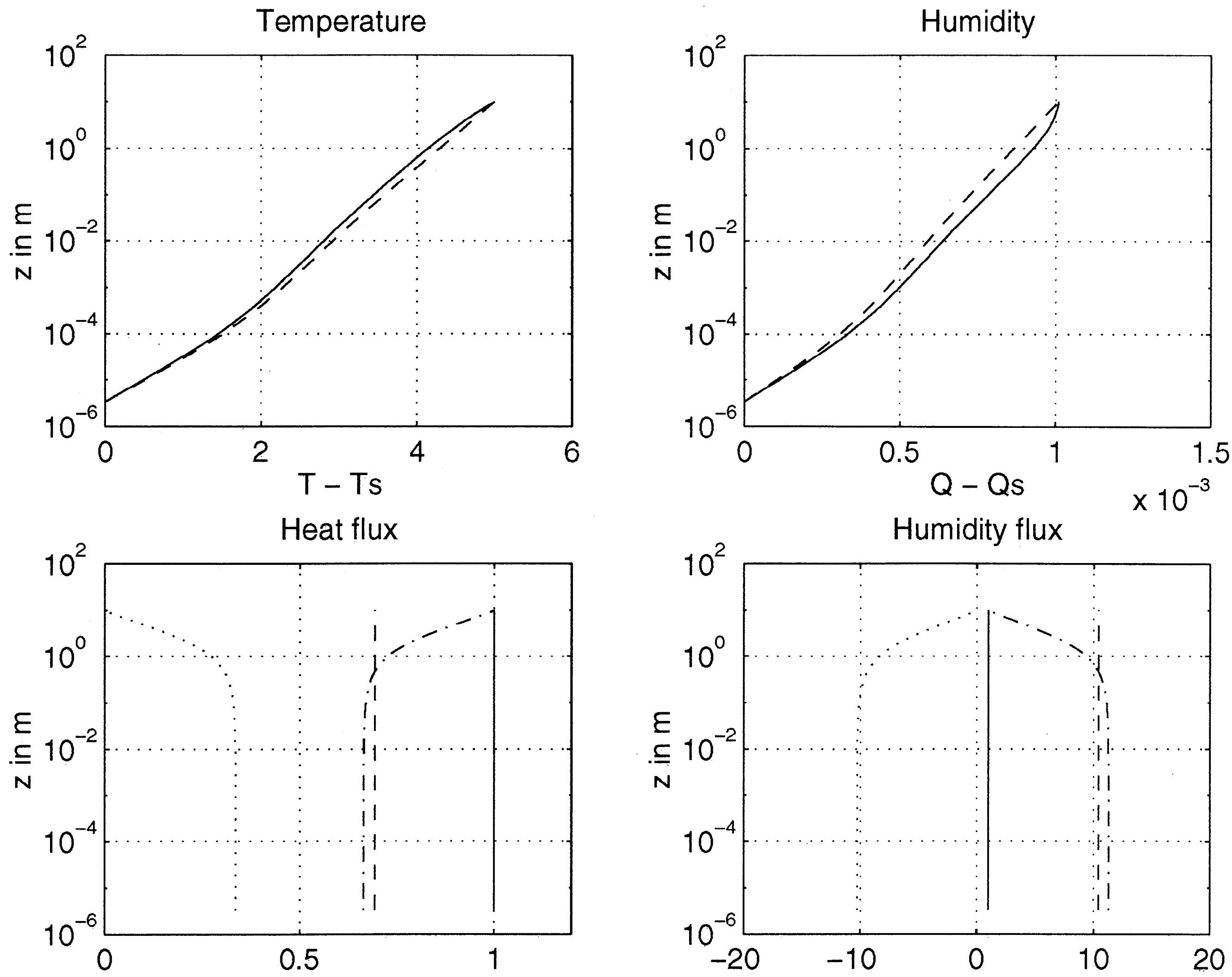

Figure 2. Same as Figure 1 but for wind speed $U_{10} = 25\,\mathrm{m\,s^{-1}}$, and stable stratification $\Delta\theta = +5^o$.

3.2. Exchange Coefficients

The dynamical part of the model is checked by comparing results for the drag coefficient with data of *Anderson* [1993] and *Large and Pond* [1982] (Figure 4). The regression line of *Large and Pond* [1982] is $10^3 C_D = 0.49 + 0.065 U_{10}$ in the wind speed range $10 < U_{10} < 20\,\mathrm{m\,s^{-1}}$, and $10^3 C_D = 1.14 \pm 0.20$ in the wind speed range $5 < U_{10} < 10\,\mathrm{m\,s^{-1}}$. Model results are for unstable ($\Delta\theta = -3^o$) and stable ($\Delta\theta = +3^o$) stratification. Model results are not corrected for stability effect. This correction is noticeable only for a wind speed of 5 m s^{-1}. It increases the drag coefficient for stable stratification and decreases for unstable stratification for about 10% and is not shown in the figure. The model results are found to be in reasonable agreement with data.

The sensible heat and humidity exchange coefficients for unstable conditions are shown in Figure 5. Data of *Large and Pond* [1982] and of *Anderson* [1993] of the sensible heat exchange coefficient are compiled in the same figure. Two runs for the heat exchange coefficient are shown. In the first run the commonly accepted value $z_e = 1/2 H_s$ for the decay length of droplet concentration (equation (51)) is used; in the second the length is reduced twice $z_e = 1/4 H_s$. The latter choice gives a better agreement with both data sets up to about 23 m s^{-1} and will be used hereafter. Unfortunately, the scatter of the data is so large that the model run with no spray effect (which shows an increase of C_H with an increase of wind speed, contrary to the case when the spray effect is accounted for and shows a decrease in C_H) also falls well in the data cloud, and one can actually not decide which wind speed dependence of C_H is true. Data of *Large and Pond* [1982] for the highest wind measurement seem to offer a possibility in decreasing C_H with wind speed, which will then support a point of view that evaporation of droplets plays a role in the heat balance above the waves. However, the authors themselves approximate their data by a constant value of C_H and are not very confident in data for the highest wind, as it was

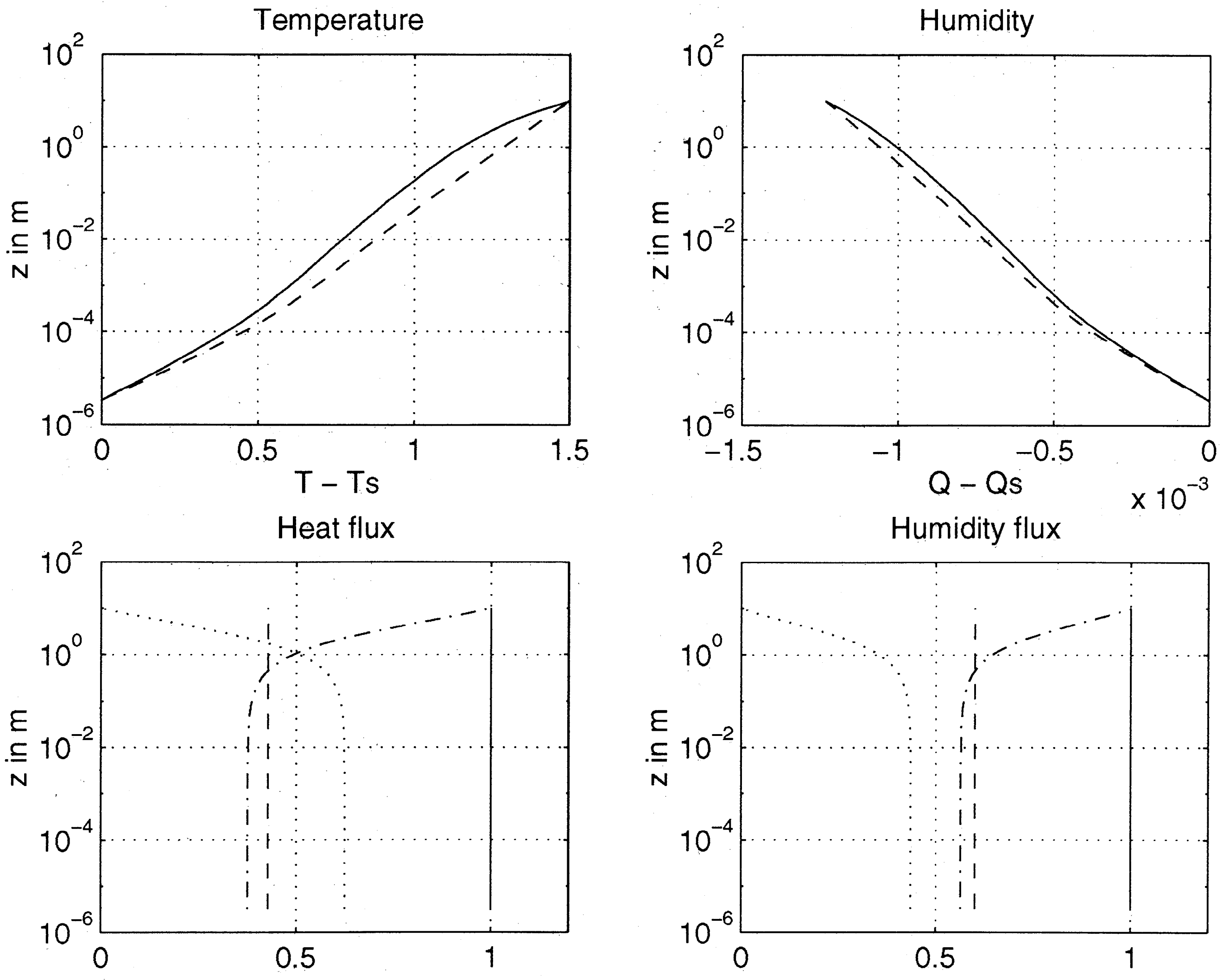

Figure 3. Same as Figure 1 but for wind speed $U_{10} = 25\,\mathrm{m\,s^{-1}}$, and stable stratification $\Delta\theta = +1.5^{o}$.

a storm event. Any further speculation on comparison model results and data is dangerous before direct measurements of heat fluxes up to about 25 $\mathrm{m\,s^{-1}}$ become available, which should then reveal whether or not the impact of sea spray is important for the balance of heat and moisture.

What we can well say is that model results for wind speeds below 18 $\mathrm{m\,s^{-1}}$ support the HEXMAX conclusion that there is no drastic impact of spray on heat and moisture fluxes in this range of wind speed. We also notice that the mean value of the modeled heat exchange coefficient $10^3 C_H = 1.1$ (in the same wind speed range) agrees well with the mean estimate of *Anderson* [1993] of $10^3 C_H = 1.1$ (no standard error is reported); *Large and Pond* [1982] of $10^3 C_H = 1.13 \pm 0.32$; and HEXMAX [*DeCosmo et al.*, 1996] of $10^3 C_H = 1.14 \pm 0.35$.

The calculated values of the humidity exchange coefficient are the same as the values of the heat exchange coefficient when no spray is present. However, in the presence of spray, it increases faster with wind speed, contrary to decrease in the sensible heat flux. The increase is about 20% compared to the value of C_E at a wind speed of 10 $\mathrm{m\,s^{-1}}$. This model result seems to support a result of the HEXMAX experiment that an increase of C_E by as much as 20% is not ruled out, but a sudden and drastic increase of C_E due to evaporation of spray droplets at wind speeds beyond about 15 $\mathrm{m\,s^{-1}}$ is not found (measurements were made up to 18 $\mathrm{m\,s^{-1}}$ in unstable conditions). Again, that remark remains pure speculation unless reliable data at wind speeds of about 25 $\mathrm{m\,s^{-1}}$ become available.

Results of modeling the exchange coefficients in stable conditions (Figure 6) give the same conclusions. The mean value of calculated sensible heat exchange coefficient $10^3 C_H = 0.78$ agrees well with measurement: *Large and Pond* [1982] give $10^3 C_H = 0.66 \pm 0.14$; *Anderson* [1993] gives $10^3 C_H = 0.79$. The impact of evaporating droplets is not noticeable below 18 $\mathrm{m\,s^{-1}}$. No-

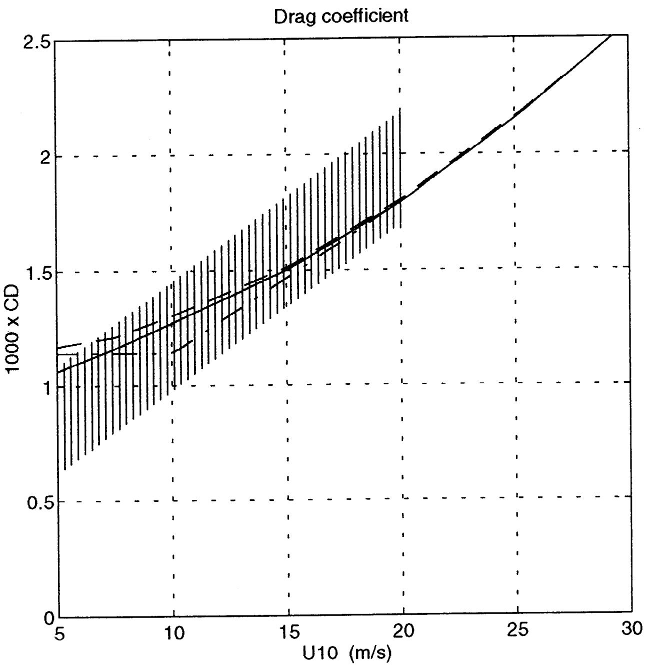

Figure 4. Drag coefficient versus U_{10}. Model results are shown as follows: dashed line, unstable stratification ($\Delta\theta = -3^o$); solid line, stable stratification ($\Delta\theta = +3^o$). Data of *Anderson* [1993] fall inside the shaded area. Regression line from *Large and Pond* [1982] is shown by dashed-dotted line.

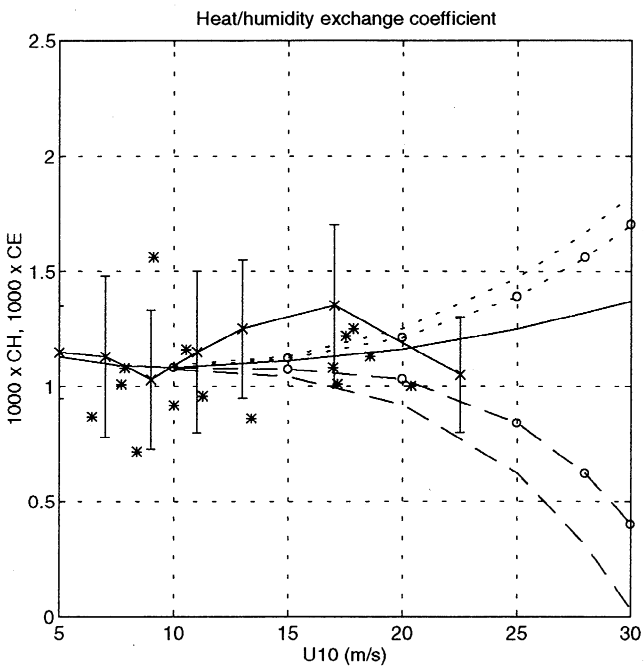

Figure 5. Heat and humidity exchange coefficient versus U_{10}, unstable stratification ($\Delta\theta = -3^o$). For the heat exchange coefficient, stars show data of *Anderson* [1993], unstable conditions, compiled from his Figure 8a. Solid line with crosses and error bars are data of *Large and Pond* [1982, Figure 10]. Model results are shown as follows: solid line, no spray effect; dashed line, spray effect is accounted for; dashed line with circles, the decay length in (51) is $z_e = 1/4H_s$. For the humidity exchange coefficient, model results are shown as follows: solid line, no spray effect; dotted line, spray effect is accounted for; dotted line with circles, the decay length in (51) is $z_e = 1/4H_s$.

tice that while in strong stable conditions the impact of spray will lead to an increase of the heat exchange coefficient and decrease of the humidity exchange coefficient, both will increase in weak stable conditions. These trends (if they exist) should be measurable at wind speeds of about 25 m s^{-1}.

Direct measurements of sensible heat and humidity flux at wind speeds of about 25 m s^{-1} should give a clear answer as to whether the sea spray plays an important role in exchanges of heat and humidity or not.

In unstable conditions the sensible heat spray mediated surface flux is about 75% of the direct turbulent flux at wind speed 25 m s^{-1}. At the same wind speed the humidity spray mediated flux is about 25% of the direct turbulent one, which is smaller than estimates by *Andreas* [1992] and *Fairall et al.* [1994]. However, their models do not account for the proper balance between direct turbulent and spray mediated fluxes. In strong stable conditions the heat spray mediated flux is about 40%, and the humidity spray mediated flux is about 100% of the turbulent flux. In weak stable conditions the heat spray mediated flux is about 60%, and the humidity spray mediated flux is about 45% of direct turbulent fluxes.

The sensitivity of the model to the magnitude of the droplet concentration is assessed by increasing and decreasing the surface droplet concentration $N(r)_0$, equation (51), to 10 times. The decrease of the concentration results in no spray impact on fluxes. The increase leads to unrealistic results. In unstable stratification the sensible heat exchange coefficient and in stable stratification the humidity exchange coefficient tend to zero already at wind speed 20 m s^{-1}, a trend which is not supported by measurements. This test shows that the droplet concentration function of *Wu* [1992] is rather realistic. The use of other functions which considerably exceed Wu's function in magnitude will lead to unrealistic results. However, this statement is only true in the scope of the present model. There could exist a physical feedback mechanism which will quench the spray mediated fluxes with increase of droplet concentration.

4. Conclusions

The sensible heat, humidity, and momentum exchange coefficients can be directly calculated from the wind speed and the sea state using an approach which is based on the conservation of momentum, heat, and humidity in the turbulent boundary layer above sea waves in the presence of sea spray.

In unstable conditions the spray mediated fluxes decrease the sensible heat exchange coefficient and in-

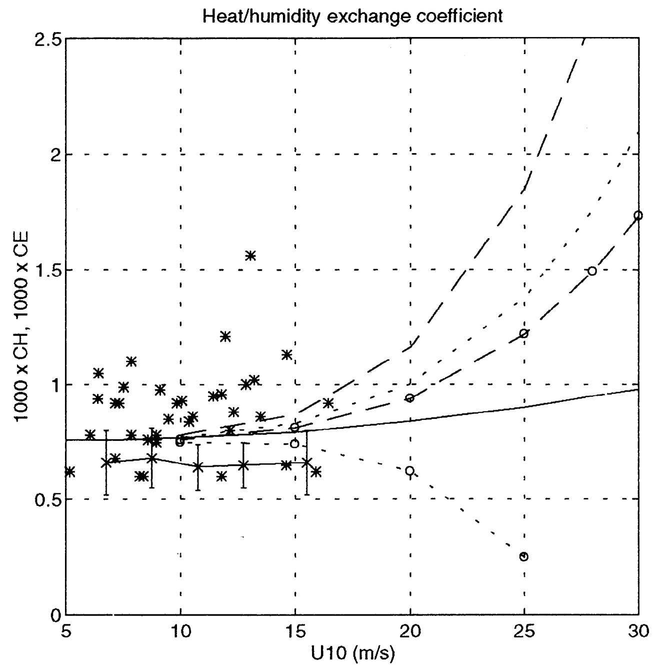

Figure 6. Heat and humidity exchange coefficient versus U_{10}, stable stratification ($\Delta\theta = +1.5^o$). For the heat exchange coefficient, stars show data of *Anderson* [1993], stable conditions, compiled from his Figure 8a. Solid line with crosses and error bars are data of *Large and Pond* [1982, Figure 10]. Model results are shown as follows: solid line, no spray effect; dashed line, spray effect is accounted for ($\Delta\theta = +1.5^o$); dashed line with circles, $\Delta\theta = +5^o$. For the humidity exchange coefficient, model results are shown as follows: solid line, no spray effect; dotted line, spray effect is accounted for ($\Delta\theta = +1.5^o$); dotted line with circles, $\Delta\theta = +5^o$. All model results are for the decay length in (51) $z_e = 1/4H_s$.

crease the humidity exchange coefficient with increasing wind speed, compared to exchange coefficients at 10 $\mathrm{m\,s^{-1}}$ wind speed, when the spray effect is negligible.

In strong stable conditions the spray mediated fluxes increase the sensible heat exchange coefficient and decrease the humidity exchange coefficient with increasing wind speed. In weak stable conditions defined by a positive temperature difference $(\theta_h - \theta_0)$ and a small negative humidity difference, the spray mediated fluxes increase both the sensible heat and humidity exchange coefficients.

Model results obtained with the chosen surface droplet concentration function [*Wu*, 1992] and the vertical decay length of droplet concentration, proportional to the significant wave height, satisfy the experimental fact that in the range of wind speed $< 18\,\mathrm{m\,s^{-1}}$ no impact of sea spray on heat and moisture fluxes is detected.

Model results show that the impact of sea spray becomes significant at wind speeds of about 25 $\mathrm{m\,s^{-1}}$ and above.

To settle the issue whether or not sea spray plays an important role in exchange of heat and moisture above the sea, simultaneous direct measurements of sensible heat and moisture flux under different stratification conditions at wind speeds of about 25 $\mathrm{m\,s^{-1}}$ and above are needed.

Acknowledgment. To Vladimir Kudryavtsev for fruitful discussions.

References

Anderson, R.J., A study of wind stress and heat flux over the open ocean by the inertial-dissipation method, *J. Phys. Oceanogr.*, *23*, 2153-2161, 1993.

Andreas, E.L., Sea spray and the turbulent air- sea heat fluxes, *J. Geophys. Res.*, *97*, 11,429- 11,441, 1992.

Andreas, E.L., Reply, *J. Geophys. Res.*, *99*, 14,345-14,350, 1994.

Andreas, E.L., J.B. Edson, E.C. Monahan, M.P. Rouault, and S.D. Smith, The spray contribution to net evaporation from the sea: a review of recent progress, *Boundary Layer Meteorol.*, *72*, 3-52, 1995.

Belcher, S.E., and J.C.R. Hunt, Turbulent shear flow over slowly moving waves, *J. Fluid Mech.*, *251*, 109-148, 1993.

Blanchard, D.C., and A.H. Woodcock, The production, concentration, and vertical distribution of the sea-salt aerosol, *Ann. N.Y. Acad. Sci.*, *338*, 330- 347, 1980.

Bortkovskii, R.S., On the mechanism of interaction between the ocean and the atmosphere during a storm, *Fluid Mech. Sov. Res.*, *2*, 87-94, 1973.

Bortkovskii, R.S., *Air-Sea Exchange of Heat and Moisture During Storms*, 194 pp., D. Reidel, Norwell, Mass., 1987.

Caudal, G., Self-consistency between wind stress, wave spectrum, and wind-induced wave growth for fully rough air-sea interface, *J. Geophys. Res.*, *98*, 22,743- 22,752, 1993.

Chalikov, D.V., and M.Yu. Belevich, One- dimensional theory of the wave boundary layer, *Boundary Layer Meteorol.*, *63*, 65-96, 1993.

Chalikov, D.V., and V.K. Makin, Models of the wave boundary layer, *Boundary Layer Meteorol.*, *56*, 83-99, 1991.

Charnock, H., Wind stress on a water surface, *Q.J.R. Meteorol. Soc.*, *81*, 639-640, 1955.

DeCosmo, J., K.B. Katsaros, S.D. Smith, R.J. Anderson, W.A. Oost, K. Bumke, and H. Chadwick, Air-sea exchange of water vapor and sensible heat: The Humidity Exchange Over the Sea (HEXOS) results, *J. Geophys. Res.*, *101*, 12,001-12,016, 1996.

DeLeeuw, G., Size distribution of giant aerosol particles close above sea level, *J. Aerosol. Sci.*, *17*, 293-296, 1986a.

DeLeeuw, G., Vertical profiles of giant particles close above the sea surface, *Tellus*, *38B*, 51-61, 1986b.

DeLeeuw, G., Near-surface particle size distribution profiles over the North Sea, *J. Geophys. Res.*, *92*, 14,631-14,635, 1987.

DeLeeuw, G., Profiling of aerosol concentrations, particle size distributions and relative humidity in the atmospheric surface layer over the North Sea, *Tellus*, *42B*, 342-354, 1990.

DeLeeuw, G., Aerosols near the air-sea interface, *Trends Geophys. Res.*, *2*, 55-70, 1993.

Donelan, M.A., Air-sea interaction, *The Sea Ocean Eng. Sci.*, *9*, 239-292, 1990.

Donelan, M.A., and W.J. Pierson, Radar scattering and equilibrium ranges in wind-generated waves with application to scatterometry, *J. Geophys. Res.*, *92*, 4971-5029, 1987.

Donelan, M.A., J. Hamilton, and W.H. Hui, Directional spectra of wind generated waves, *Philos. Trans. R. Soc. London, Ser. A*, *315*, 509-562, 1985.

Edson, J.B., and C.W. Fairall, Spray droplet modeling, 1, Lagrangian model simulation of the turbulent transport of evaporating droplets, *J. Geophys. Res.*, *99*, 25,295-25,311, 1994.

Edson, J.B., S. Anquetin, P.G. Mestayer, and J.F. Sini, Spray droplet modeling, 2, An interactive Eulerian- Lagrangian model of evaporating spray droplets, *J. Geophys. Res.*, *101*, 1279-1293, 1996.

Elfouhaily, T., B. Chapron, K.B. Katsaros, and D. Vandemark, A unified directional spectrum for long and short wind-driven waves, *J. Geophys. Res.*, *102*, 15,781- 15,796, 1997.

Fairall, C.W., J.D. Kepert, and G.J. Holland, The effect of sea spray on surface energy transports over the ocean, *Global Atmos. Ocean Syst.*, *2*, 121-142, 1994.

Friehe, C.A., and K.F. Schmitt, Parameterizations of air-sea interface fluxes of sensible heat and moisture by the bulk aerodynamic formulas, *J. Phys. Oceanogr.*, *6*, 801-809, 1976.

Geernaert, G.L., Bulk parameterizations for the wind stress and heat fluxes, in *Surface Waves and Fluxes, vol.1*, edited by G.L. Geernaert and W.J. Plant, 336 pp., Kluwer Acad., Norwell, Mass., 1990.

Hasse, L., On the contribution of spray droplets to evaporation, *Boundary Layer Meteorol.*, *61*, 309- 313, 1992.

Hasselmann, K., Weak-interaction theory of ocean waves, *Basic Dev. Fluid Dyn.*, *2*, 117-182, 1968.

Jähne, B., and K.S. Riemer, Two-dimensional wave number spectra of small-scale water surface waves, *J. Geophys. Res.*, *95*, 11,531-11,546, 1990.

Janssen, P.A.E.M., Wave-induced stress and the drag of air flow over sea waves, *J. Phys.Oceanogr.*, *19*, 745-754, 1989.

Jenkins, A.D., A simplified quasi-linear model for wave generation and air-sea momentum flux, *J. Phys. Oceanogr.*, *23*, 2001-2018, 1993.

Kader, B.A., and A.M. Yaglom, Heat and mass transfer laws for fully turbulent wall flows, *Int. J. Heat Mass Transfer*, *15*, 2329-2353, 1972.

Katsaros, K.B., S.D. Smith, and W.A. Oost, HEXOS - Humidity Exchange Over the Sea: A program for research [redacted]

Makin, V.K., and C. Mastenbroek, Impact of waves on air-sea exchange of sensible heat and momentum, *Boundary Layer Meteorol.*, *79*, 279-300, 1996.

Makin, V.K., V.N. Kudryavtsev, and C. Mastenbroek, Drag of the sea surface, *Boundary Layer Meteorol.*, *73*, 159-182, 1995.

Mastenbroek, C., V.K. Makin, M.H. Garat, and J.P. Giovanangeli, Experimental evidence of the rapid distortion of turbulence in the air flow over water waves, *J. Fluid Mech.*, *318*, 273-302, 1996.

Miller, M.A., and C.W. Fairall, A new parameterization of spray droplet production by oceanic whitecaps, in *Proceedings of the 7th Conference on Ocean- Atmosphere Interaction*, pp. 174-177, Am. Meteorol. Soc., Boston, Mass., 1988.

Monahan, E.C., Sea spray as a function of low- elevation wind speed, *J. Geophys. Res.*, *73*, 1127- 1137, 1968.

Monahan, E.C., C.W. Fairall, K.L. Davidson, and P. Jones-Boyer, Observed inter-relations between 10 m winds, ocean whitecaps and marine aerosols, *Q. J. R. Meteorol. Soc.*, *109*, 379-392, 1983.

Monahan, E.C., D.E. Spiel, and K.L. Davidson, A model of marine aerosol generation via whitecaps and wave disruption, in *Oceanic Whitecaps and Their Role in Air- Sea Exchange Processes* , edited by E.C. Monahan and G. MacNiocaill, pp. 167-174, D. Reidel, Norwell, Mass., 1986.

Monin, A.S., and A.M. Yaglom, *Statistical Fluid Mechanics: Mechanics of Turbulence*, 769 pp., MIT Press, Cambridge, 1971.

Preobrazhenskii, L.Y., Estimate of the content of spray-drops in the near-water layer of the atmosphere, *Fluid Mech. Sov. Res.*, *2*, 95-100, 1973.

Press, W.H., S.A. Teukolsky, W.T. Vetterling, and B.P. Flannery, *Numerical Recipes in Fortran: The Art of Scientific Computing*, 2nd ed., 963 pp., Cambridge Univ. Press, New York, 1992.

Pruppacher, H.R. and J.D. Klett, *Microphysics of Clouds and Precipitation*, 714 pp., D. Reidel, Norwell, Mass., 1978.

Rodi, W., *Turbulence Models and Their Application in Hydraulics - A State of the Art Review*, 2nd ed., 104 pp., Inst. fur Hydromech. Univ. of Karlsruhe, Int. Assoc. for Hydraul. Res., Karlsruhe 1984.

[redacted]

Wu, J., Bubble flux and marine aerosol spectra under various wind velocities, *J. Geophys. Res.*, *97*, 2327-2333, 1992.

Wu, J., J.J. Murray, and R.J. Lai, Production and distributions of sea spray, *J. Geophys. Res.*, *89*, 8163-8169, 1984.

Zilitinkevich, S.S., *The Dynamics of the Atmospheric Boundary Layer*, 292 pp., Gidrometeoizdat, Leningrad, 1970.

V. K. Makin, Royal Netherlands Meteorological Institute, P.O. Box 201, 3730 AE De Bilt, The Netherlands. (e-mail: makin@knmi.nl)

(Received March 20, 1997; revised October 7, 1997; accepted October 10, 1997.)