J1.4 EVAPORATION AND SENSIBLE HEAT EXCHANGE FOR A SHALLOW LAKE

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1. INTRODUCTION

Due to the smaller scales of the presently used models, which are applied for weather forecast and climate research, it is necessary to parameterize the evaporation as well as the heat and mass exchange (this paper mainly concentrates on evaporation), even for small lakes. These are often shallow and covered with small waves so the usual bulk equations for air water exchange are not valid.

In principle, equations of the Penman or Priestley-Taylor type (Stull, 1988) can be used. However, these equations do not include the type of lake and, additionally, the parameterizations depend upon radiation input and include parameters of a more climatological scale, which do not allow exact calculations on a time scale of one hour (Foken, 2003).

Intensification of the energy-mass exchange of the shallow basin with the atmosphere is caused mainly by changes in both the thermal regime of the shallow water basin and the aerodynamic roughness of its surface. Thus, if the thermal regime changes are well represented by the actual surface temperature data used, the impact of the roughness changes is usually not examined.

Two different approaches for deep sea conditions will be used and validated with a data set from the LITFASS-98 experiment (Beyrich et al., 2002a) which was conducted over a shallow lake on a time scale of 30 minutes using eddy covariance data.

2. MODEL APPROCHES

2.1. Bulk equation for deep water

Assuming that the heat and humidity exchange in the near-the-water atmosphere layer can be described within the framework of dimensionless coefficients

$$\begin{aligned} C_U &= \overline{w'u'}/U_{10}^{\;2} \\ C_T &= \overline{w'T'}/\Delta T_{S,Z}\cdot U_{10} \\ C_e &= \overline{w'a'}/\Delta a_{S,Z}\cdot U_{10} \end{aligned} \qquad (1)$$

(C_D drag coefficient, C_T coefficient of heat exchange – Stanton number, and C_e coefficient of evaporation – Dalton number) it follows with L: Obukhov length, h_s: height of the roughness elements, z: height, δ_v: height of the molecular sublayer (Panin, 1985) that

$$\begin{aligned} C_T &= F_T\left(z/h_S;\ z/L;\ h_S/\delta_\nu;\ \mathrm{Pr}_T\right) \\ C_e &= F_e\left(z/h_S;\ z/L;\ h_S/\delta_\nu;\ \mathrm{Pr}_e\right) \end{aligned} \qquad (2)$$

Assuming that the Prandtl numbers are constant and equal among themselves Pr_T= Pr_e= Pr= 1.0, that the adjustable heights are fixed (for example, z=10m), and that the height of the molecular sublayer with precision up to a numeral multiplier could be represented in the equation $\delta_v \approx v/u_*$ and $h_s \approx z_0$ (z_0: roughness parameter), we can rewrite (1) in the form:

$$C_{T,e} = F_{T,e}\left(z/L;\ z_0 u_*/\nu\right). \qquad (3)$$

Using data of fluxes of impulse, heat, and moisture in the form for the determination of the coefficient of heat/moisture exchange in stratified conditions, we achieve a simple means to account for the dependence of stratification on the heat/moisture exchange (Panin and Krivitskii, 1992):

$$C_{T,e} = C_{T,e}{}^{N}\cdot\begin{cases} (1-z/L)\left[1+10^{-2}\left(z_0u_*/\nu\right)^{3/4}\right] \\ at\ \ z/L<0 \\ \left[\ 1/(1+3.5z/L)\ \right]\cdot\left[1+10^{-2}\left(z_0u_*/\nu\right)^{3/4}\right] \\ at\ \ z/L>0 \end{cases} \qquad (4)$$

2.2 Three-layer model for deep water

Approximately 30-40 years ago, several publications described the integration of the profile coefficient over three layers (Bjutner, 1974; Kitajgorodskij and Volkov, 1965; Mangarella et al., 1972; Mangarella et al., 1973). The basic approach was the bulk equation where the product from the bulk coefficient with the velocity was replaced by the so-called profile coefficient Γ (H_T sensible heat flux; T_S surface temperature, T_Z temperature at the height z)

$$H_T = \Gamma\left[T_S - T_z\right], \qquad (5)$$

where Γ can be determined by integration over the three layers

$$\Gamma = \left(\int_0^z \frac{dz}{K_T+\nu_{Tt}+\nu_T}\right)^{-1}. \qquad (6)$$

where K_T is the turbulent exchange coefficient for temperature, where v_T is the molecular temperature diffusion coefficient, and where v_{Tt} is the molecular-turbulent temperature diffusion coefficient..

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Direct measurements of the molecular temperature sublayer revealed a dimensionless thickness of this layer which depends upon the wave structure at the water surface (Foken et al., 1978). The same measurements revealed parameterization for the dimensionless temperature difference (normalized by the dynamic temperature scale) of the buffer layer given by $\Delta T^+ \approx 4$ (Foken, 1984; Foken et al., 1978). For friction velocities $u_* < 0.23$ m s^{-1} it follows that

$$\Gamma = \frac{\kappa \cdot u_*}{\left(\kappa \cdot \mathrm{Pr} - \frac{1}{6}\right) \cdot \delta_T^+ + 5 + \ln \frac{u_* \cdot z}{30\,\nu}} \qquad (7)$$

For heights > 1 m Equation (7) can be completed by a stability dependent term in accordance with the similarity theory of Monin and Obukhov (1954). Assuming similar temperature and humidity structure near the water surface and replacing the Prandtl number with the Schmidt number, Equations (5) and (7) can also be used for the calculation of the evaporation.

2.3 Shallow lake correction

We have accomplished special experimental research regarding the impact of the energy-mass exchange process on shallow waters, which has allowed us to account for the influence of the basin depth on evaporation and heat exchange intensities in a final form (Panin et al., 1996):

$$H_T^{SW} = H_T + H_T \cdot k_T^{SW} \cdot \frac{h}{H} \approx H_T(1 + 2h/H)$$
$$E^{SW} = E + E \cdot k_E^{SW} \cdot \frac{h}{H} \approx E(1 + 2h/H) \qquad (8)$$

In Equation (8) are: the empirical coefficient $k_T^{SW} \approx k_E^{SW} \approx 2.0$, h: the mean waves height of the lake at the measuring point, H: the depth of the lake, and H_T, E: the evaporation and sensible heat exchange of deep water, and, finally, H_T^{SW}, E^{SW}: the evaporation and sensible heat exchange of shallow water.

If data on h are absent, then one could use the empirical relationship

$$h \approx \frac{0.07 \cdot U_Z^2 \cdot \left(gH/U_Z^2\right)^{3/5}}{g} \qquad (9)$$

(Davidan et al., 1985), where U_Z: wind speed on the height 10m, and g: gravity acceleration.

While the formulation of this dependence still needs to be refined, today it allows us to account for the influence of the basin depth on the evaporation and heat exchange intensities with an error rate of about 25% off from the calculated value.

A new parameterization of evaporation from shallow water surfaces under different wind velocities (Fig.1) has been recognized on the basis of the theoretical and experimental analyses performed on the intensive heat-mass exchange between deep and shallow sea/lake surfaces and the atmosphere.

According to Fig. 1, it can be stated that in natural conditions (within the range of actual wind velocities) evaporation from shallow waters theoretically might exceed more than 1.5 times the usual evaporation magnitudes for deep waters. In reality, the recurrence of high wind velocities of 25-30 m s^{-1} is not very likely, therefore their input into the final amount of evaporation should not be that important.

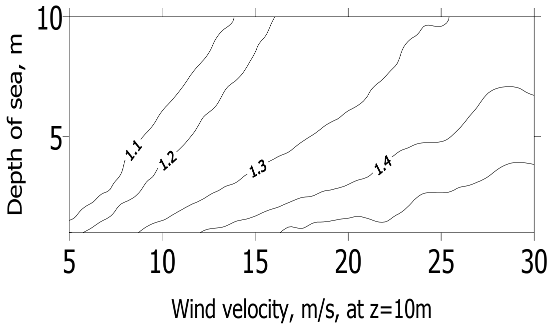

Fig. 1: Intensification of the evaporation and sensible heat exchange of the shallow sea/lake.

3. EXPERIMENTAL DATA BASIS

The data set for this investigation was measured during the LITFASS-98 experiment (Beyrich et al., 2002a) in a research area of the German Meteorological Service near the Meteorological Observatory Lindenberg, approximately 50 km southeast of Berlin. During this experiment, which took place in May and June 1998, turbulent fluxes were measured over different land use types (Beyrich et al., 2002b), one of them the lake 'Großer Kossenblatter See'. The turbulent fluxes were measured with a sonic anemometer Campbell CSAT3, a thin platinum wire sensor AIR 150, and a Campbell Krypton hygrometer KH20 located on a tower (52° 08' 17" N, 14° 06', 37" E, 43 m a.s.l.) at a height of 3.5 m. Furthermore, wind measurements (Vector A100R at 4.3 m), temperature and humidity measurements (Vaisala HMP35AC at 3.0 m), net radiation measurements (REBS Q6 at 2.0 m), and water temperature measurements (Campbell T-probe 107) were taken. For more details, like comparisons and quality checks, see Beyrich et al. (2002b). During the LITFASS-98 experiment only data from the wind sector 180°-330° were used because of the short fetches in the eastern sector, a village in the northeast, and the forest east of the site.

4. RESULTS

The data set from the LITFASS-98 experiment was first used in a comparison of the models by Panin (bulk approach, Chapter 2.1) and Foken (tree-layer approach, Chapter 2.2), both with the shallow lake correction by Panin (Chapter 2.3). With R^2 values of 0.99 and 0.98 for the latent heat flux (Fig. 2) and the sum of the latent and sensible heat fluxes (Fig. 3), respectively, a good agreement, especially for the latent heat flux with a regression coefficient of 0.98, was found. From Fig. 3 it follows that in some cases either the model by Foken underestimates or the model by Panin overestimates the sensible heat flux.

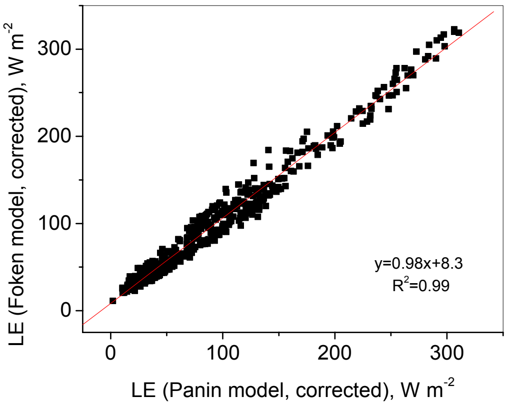

Fig. 2 Comparison of latent heat flux calculations according to the bulk model by Panin (Chapter 2.1) and the three-layer model by Foken (Chapter 2.2), both with the shallow water correction by Panin (Chapter 2.3), for the ‘Großer Kossenblatter See’ during LITFASS-98

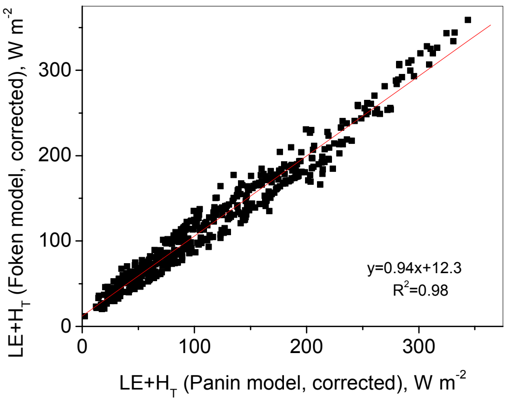

Fig. 3: Comparison of the sum of latent and sensible heat flux calculations according to the bulk model by Panin (Chapter 2.1) and the three-layer model by Foken (Chapter 2.2), both with the shallow water correction by Panin (Chapter 2.3), for the ‘Großer Kossenblatter See’ during LITFASS-98)

To improve the quality of the models and of the shallow water correction, the model outputs were compared with the eddy covariance data. In Fig. 4 this comparison is done for the latent heat flux using the model by Panin with a shallow water correction. According to this analysis, the model overestimates the flux by approximately 7 % for R^2=0.90. The sum of both the latent and sensible heat fluxes (Fig. 5) is in excellent agreement with the measurements (1 % but R^2=0.82). The differences found in the comparison are within an approximately 10% range of accuracy for flux measurements (Foken, 2003). However, the impact of the shallow water correction is slightly greater.

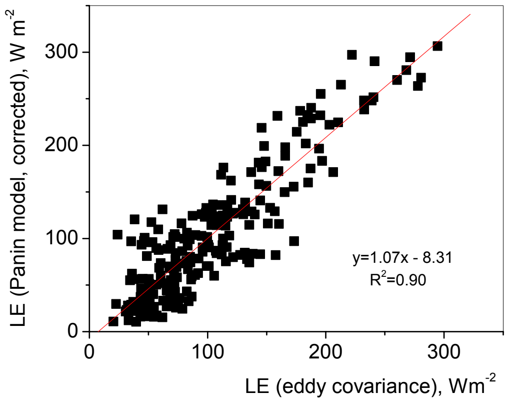

Fig. 4 Comparison of latent heat flux calculations according to the bulk model by Panin (Chapter 2.1) with the shallow water correction by Panin (Chapter 2.3) and eddy-covariance measurements for the ‘Großer Kossenblatter See’ during LITFASS-98

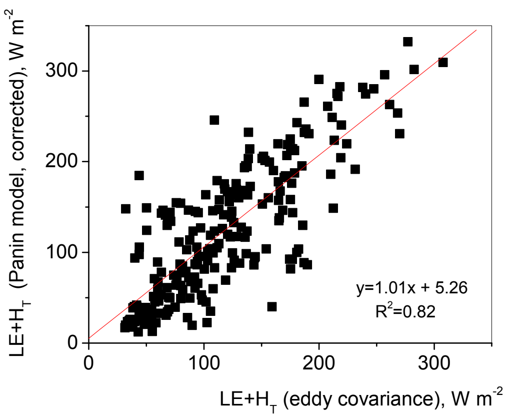

Fig. 5 Comparison of the sum of the latent and sensible heat flux calculations according to the bulk model by Panin (Chapter 2.1) with the shallow water correction by Panin (Chapter 2.3) and eddy-covariance measurements for the ‘Großer Kossenblatter See’ during LITFASS-98

5. CONCLUSIONS

It was shown that a comparison of the different approaches by Panin (bulk model, Chapter 2.1) and Foken (tree-layer model, Chapter 2.2) for the calculation of the sensible and latent heat fluxes for the open ocean under low wind conditions (only these were relevant for our investigations of shallow lakes) show very similar results. Such models can be very easily transferred to shallow water conditions and lakes including a depth dependent function. Even for low wind velocities, the effect of the increase of fluxes for shallow lakes is on the order of 10-20 %, which is higher than the differences between modelled and measured fluxes.

The validation of this model with the eddy covariance measurements of the LITFASS-98 experiment showed good results. Therefore, one may use the model presented for calculating the evaporation of lakes, where a standard data set of wind velocity, air and water temperature, and air moisture is available.

6. ACKNOWLEDGMENT

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