J 1.7 TURBULENT FLUXES AND AIR-ICE COUPLING IN THE BALTIC AIR-SEA-ICE STUDY

Jouko Launiainen *
Finnish Institute of Marine Research, Helsinki

INTRODUCTION

The overall goal of the European Baltic Air-Sea-Ice-Study (BASIS) was to create and analyse an experimental data set for verification and optimization of coupled air-ice-ocean models. Accurate determination of the turbulent fluxes and air-ice-ocean coupling and modelling were important goals of BASIS. The project was conducted by various Finnish, Swedish and German institutes during 1997-2000. The main field campaign was carried out in February-March, 1998, in the northern Baltic Sea. The location was in the boundary zone between the sea ice-covered and open sea. The experiment and the data are described in the BALTEX-BASIS Data Report (Launiainen, Ed., 1999) and the compilation of results in the final report (Launiainen and Vihma, Eds., 2001). BASIS is a subproject of the Baltic Sea program BALTEX, a study of WCRP/GEWEX, and it was financially supported by the EU.

In the air-ice coupling, the primary quantities to be studied include fluxes of momentum, heat and water vapour latent heat, radiative fluxes and, air-ice interfacial (surface) temperature. Below, results of a study of local air-ice coupling are introduced.

OBSERVATIONS AND METHODS

Momentum and sensible heat surface fluxes were measured as eddy-fluxes by a sonic anemometer (Metek USA-1). Wind and air temperature profile measurements were gathered on a 10 m high profile mast on the sea ice. In addition to the sonic and mast measurements, various air-ice and ice-water interaction quantities were measured.

The turbulent fluxes of momentum (τ) and sensible heat (H) were defined directly as covariances from the sonic measurements (method I in the Appendix). For comparison, the profile mast measurements allowed us to calculate the momentum and heat flux from profile gradients (LDM) using the Monin-Obukhov similarity theory (method II. 1). For the latter, the primary candidates for the universal functions were: the Businger (1971) - Dyer (1974) type forms with Högström´s (1988) coefficients for the unstable region and, for the stable region those of Webb (1970) and Holtslag and De Bruin (1988) were used. The results of the two methods above, and the flux-profile relationships allows us further to calculate the transfer coefficients of the bulk formulae (A1-A3 in the Appendix). The latent heat was estimated by the bulk aerodynamic method, with the aid one level moisture observation and the surface temperature estimate (yielding the surface saturation humidity) derived from the heat flux-temperature profile relationships (A4).

As the third method (III in the Appendix) to determine the turbulent surface fluxes, an air-ice coupled one-dimensional multilayer thermodynamic model (Launiainen and Cheng, 1998) was used. In the model, air and snow/ice are coupled by the heat fluxes and the interface temperature at each time step (10 min). For bulk calculation of the fluxes and the modelled surface temperature separate meteorological input data were used, so that the method is practically independent of the eddy flux and gradient methods described.

RESULTS

Fluxes, drag coefficient

During the three week BASIS period in winter 1997/1998 the weather was first warm and windy and later cool and calm. Accordingly, the momentum flux was first moderate to large and then rather low . As to the sensible heat, upward and downward fluxes up 50 to 80 Wm^{-2} alternated in the first part and later the fluxes were low.

The turbulent fluxes, both the momentum and sensible heat, derived from the sonic anemometer and the profile gradients (LDM) agreed mutually notably well (cf. Launiainen et al. 2001; Launiainen and Vihma, 2001). This methodological good concordance can be realized from Figures 1 and 2. Figure 1 gives the time serie of the neutral drag coefficient $C_{DN} = k^2/(\ln z/z_0)^2$ derived from the sonic and LDM method. The results are rather comparable. In Figure 2 is given the time serie of stability parameter and the both methods yield almost identical results. The consistent stability results, given in terms of $z/L = f(z, \tau, H, E)$, reveal the agreement of the flux results. As to the drag coefficient, neither methods did show apparent wind speed dependence. The mean $C_{DN}(10)$ = 1.28 $\times$ 10^{-3} corresponds to a mean aerodynamic roughness length z_0 = 1.2 $\times$ 10^{-4} m. On the contrary, the drag coefficient was distinctly dependent on the wind direction. This was because of the variable upwind roughness of the sea ice with the direction, and, because the measurement site was located in the archipelago. Our results were $C_{DN}(10)$ =1.0 x 10^{-3} (i.e. aerodynamic roughness $z_0 \cong$ 3 x 10^{-5}m) for smooth snow-covered ice, and 1.5 x 10^{-3} ($z_0 \cong$ 3 x 10^{-4} m) for deformed ice. By deformed ice we here mean thin (0.3 - 0.4 m) Baltic Sea coastal ice of 100% concentration including no ridges and roughness elements higher than 0.3 to 0.6 m. Even as to the dependence of the drag coefficient with wind direction, the both

Corresponding author address: Jouko Launiainen, Finnish Institute of Marine Research (FIMR), Box 33, FIN-00931 Helsinki, jouko.launi@fimr.fi

methods of determination agree mutually well. As to the third flux determination method, the coupled air-ice model estimated fluxes (and surface temperature) compare well with those derived from the two other methods.

Temperature roughness length

The BASIS data allowed us to study the temperature roughness (z_T) and bulk heat transfer coefficient $C_{HN} = k^2 / (\ln z/z_0)(\ln z/z_T)$. The temperature roughness was calculated from (A4). To be most precise, we used for z_T derivation the cases in which we were assured the surface temperature was defined most accurately i.e. for cases T_s = 0 °C, corresponding to the melting temperature of the fresh snow-ice. Those cases were defined both by inspecting the coupled air-ice model results, surface heat balance excess, air temperature and visual observations.

Unlike the aerodynamic roughness length and drag coefficient, the z_T results indicated no distinct dependency on the wind direction but on the wind speed. The results indicated z_T to be slightly larger than z_0 for low wind velocities but $z_T < z_0$ for winds higher than 4 - 5 ms^{-1}. In terms of a roughness Reynolds numer $Re = (z_0 \cdot V)\nu^{-1}$, where ν is the kinematic viscosity, the roughness length ratio found (Launiainen et al., 2001) was

$$z_0/z_T = 0.035\, Re^{0..98} \qquad 20 < Re < 300$$

As parameterized linear with respect to wind speed the results read

$$\ln(z_0/z_T) = s + a\,V \qquad \text{or}$$

$$C_{HN} = k C_{DN}^{1/2} (\ln z / z_T)^{-1} \cong C_{DN} (1 + C_{DN}^{1/2} k^{-1} (s + aV))^{-1}$$

where s = - 0.80 and a = 0.15 for wind speed at a height of 2 m. For a wind referred to 10 m, a = 0.13. For a site with C_{DN} or z_0 known or estimated, the above should give reasonable estimate of z_T *and* C_{HN}. The region of our z_0 observations in the analysis was limited of from 3×10^{-5} to 9×10^{-4} m, i.e. C_{DN} (10) = 1.0×10^{-3} to 1.9×10^{-3} and wind speeds from 3 to 15 ms^{-1}. The analysis with respect to Re suggested that for higher aerodynamic roughnesses and wind speeds an asymptotic value $\ln(z_0/z_T) \cong 2$ is attained. This value is in good agreement with the findings of Garratt (1992) for rough surfaces with z_0/z_T = 7.3. Accordingly, our results might serve as proper first estimates for even higher wind speeds and roughness lengths. A comparison of our results with those given with respect to the frequently used $Re = (z_0 \cdot u_*)\nu^{-1}$ (Owen and Thomson, 1960; Andreas, 1987) is not straighword, because of a different definition of Re we prefer. However, for the region of z_0 and wind speed corresponding to ours, our simple linear form gives comparable results (Figure 3) with those estimated by the semi-empirical theory by Andreas (1987), the mutual difference in the bulk heat transfer coefficients being ≅ 5%.

Effect of stability, universal functions

The universal functions tested for the determination of turbulent fluxes and transfer coefficients by the LDM method and by the ice model yielded results of reasonable agreement with those measured by the sonic anemometer. Strictly however, a detailed analysis of the diabatic eddy-flux-based transfer coefficients with the observed fluxes and the *M-O* similarity theory-based flux-profile relationships suggests the current universal functions for the stable region to suppress the turbulence and transfer coefficients too much. This is to bee seen in Figure 4. For stability up to $10/L$ = 0.5, the well-known universal functions of Webb (1970) and Holtslag and De Bruin (1988) yield 10 to 12 % lower bulk transfer coefficients than those defined from our data. For $10/L$ = 1 the above difference was 15 to 17 %. Actually, for less stable region up to $10/L$ the both mentioned formulae yield comparable results i.e. $\Psi_M \cong \Psi_H \cong -5z/L$. For strong stability, our data suggests $\Psi_M = -3.5z/L$ and stability thus to suppress turbulence and the bulk transfer coefficients less than estimated by those universal functions. The search for an overall improved universal function form is still in progress. (Note: Figure 4 is given in terms of $2/L$).

CONCLUSION

The agreement of the gradient-method and coupled air-ice model results with respect to the eddy-flux results supports the validity of the Monin-Obukhov similarity theory, in conditions satisfying the preconditions for the theory. The local drag coefficient did not reveal a wind-speed dependence. Analysis of the roughness lengths indicated the temperature roughness to be comparable to or slightly larger than the aerodynamic roughness for low wind speeds, but $z_0 > z_T$ for moderate to strong winds. The temperature roughness length and bulk heat transfer coefficients can be reasonably approximated for winds up to 20 ms^{-1} based on the simple formulae suggested. Methods to determine the relevant surface temperature allowed us to estimate the latent heat flux using the bulk method. For strong stability, our data suggest stability to suppress turbulence and the bulk transfer coefficients less than estimated by the most common current universal functions.

Finally, we cannot overemphasise the need for accurate and strict calibration of sensors (temperature, wind speed), especially for the gradient method. In addition, to the sensor calibration, we may note e.g. that an inaccuracy in the measuring heights, say 5 cm when defining gradients in the lowest few metres, causes a significant error (~10%) in fluxes and bulk transfer coefficients determined by the LDM method.

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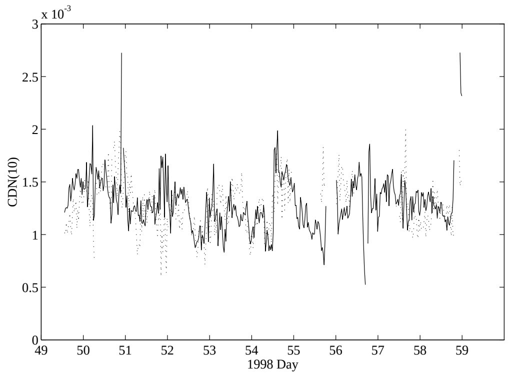

Figure 1. Neutral drag coefficient C_{DN} (10) time series as determined from the eddy flux (dotted) and profile-gradient (continuous line) method during BASIS.

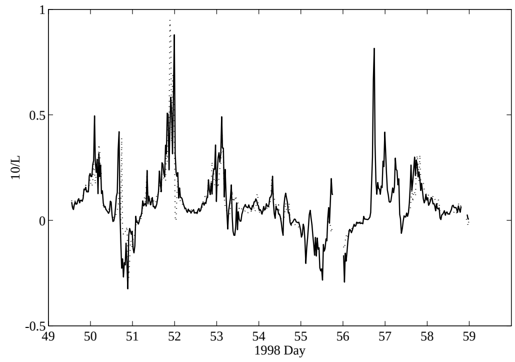

Figure 2. Time series of the stability parameter 10/*L* as calculated from the eddy flux results (broken) and profile gradient LDM method (continuous).

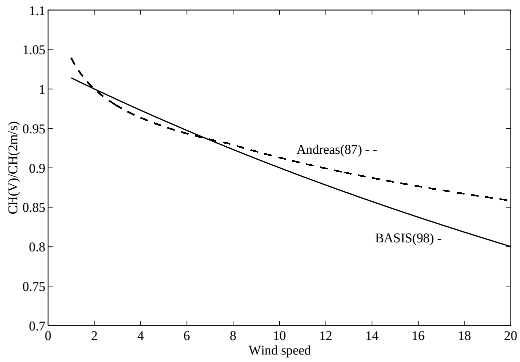

Figure 3. Neutral bulk heat transfer coefficient versus wind speed given as ratio of $C_{HN}(V)/C_{HN}(2)$ where $C_{HN}(2)$ corresponds to the transfer coefficient for 2 m/s wind speed, according to BASIS results (continuous) and the formulation by Andreas (1987; broken). The aerodynamic roughness $z_0 \cong 1.2 \times 10^{-4}$ m as observed.

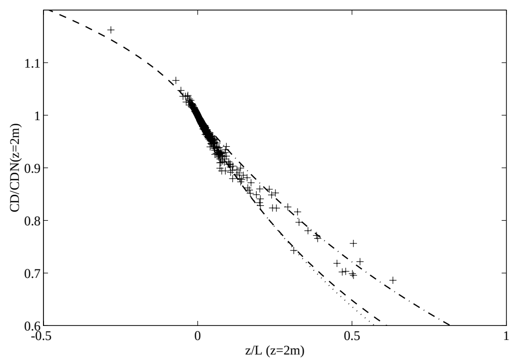

Figure 4. Observed stability dependence of the drag coefficient (crosses; sonic measurements) compared with the estimations calculated by the universal functions of Webb (1972; dotted) and Holtslag and De Bruin (1988; broken). The dash-dot line corresponds to Ψ_M = -3.5 *z*/*L*.

Appendix

1. Forms and methods of determination of the turbulent surface fluxes:

I. As *covariances* (from the sonic)

$\tau = -\rho \overline{u'w'}$ (Pa)

$H = \rho c_p \overline{\Theta' w'}$ (Wm^{-2})

$E = \rho \overline{q' w'}$ ($kgm^{-2}s^{-1}$)

II. From profile gradients (M-O Similarity Theory):

M-O universal profile gradients below read:

Those yield =>

$$\frac{\partial U}{\partial z}\frac{kz}{u_*} = \Phi_M(z/L)\,; \qquad u_*^2 = \frac{\tau}{\rho}$$

$$\frac{\partial \Theta}{\partial z}\frac{z}{\theta_*} = \Phi_H(z/L)\,; \qquad \theta_* = \frac{-H}{\rho \cdot c_p \cdot k \cdot u_*}$$

$$\frac{\partial q}{\partial z}\frac{z}{q_*} = \Phi_E(z/L)\,; \qquad q_* = \frac{-E}{\rho \cdot k \cdot u_*}$$

II. 1° *Level difference method (LDM)*

From mast; by u, z, q differences (levels may be arbitrary)

$$\frac{\Delta U}{\Delta z_u} \Rightarrow \tau, u_*, z_0, z/L$$

$$\frac{\Delta U}{\Delta z_u} and \frac{\Delta \Theta}{\Delta z_\Theta} \Rightarrow H$$

$$\frac{\Delta U}{\Delta z_u} and \frac{\Delta q}{\Delta z_q} \Rightarrow E$$

Because Φ_M and Φ_H and Φ_E = $f_{M,H,E}(z, \tau, H, E)$
=> iterative solution.

II.2° As integrated, M-O gradients yield

The bulk formulae:

$$\tau = \rho C_{Dz} U_z^2 \qquad \text{A (1)}$$

$$H = \rho c_p C_{Hz}(\Theta_z - \Theta_s)U_z \qquad \text{A (2)}$$

$$E = \rho C_{Ez}(q_z - q_s)U_z \qquad \text{A (3)}$$

Because $C_{Dz}, C_{Hz}, C_{Ez} = f_{M,H,E}(z_0, z_T, z_q, z/L, \Psi_{M,H,E}(z/L))$
and z/L = $f_{M,H,E}(z, \tau, H, E)$, solution is iterative.
For snow/ice, surface temperature Θ_s (and $q_s(\Theta_s)$) is frequently difficult and inaccurate to define.

III. *Thermodynamic ice model* solves for the surface temperature Θ_s and fluxes.
Fluxes are calculated according to II.1° or II.2°, using the modelled Θ_s and $q_s(\Theta_s)$.
For II.2° and III, the roughnesses Z_0, Z_T and Z_q are to be known/estimated.

2. From the flux-profile relationships, the surface temperature (T_s = Θ_s) is obtained

$$T_s = \Theta_z + \frac{H(\ln z/z_0 - \Psi_M(z/L))(\ln z/z_T - \Psi_H(z/L))}{\rho c_p k^2 V_z} \qquad \text{A (4)}$$