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Marine Meteorological Conditions and Air-Sea Exchange Processes Over the Northern Baltic Sea in 1990s

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Abstract

Marine meteorological conditions and air-sea exchange processes over the northern Baltic Sea were studied using data from three open-sea lighthouse meteorological stations and the R/V Aranda automatic weather station from the period 1991-1999. The ship data were analysed for the period from April to August from the region of 59 to 62°N, in the basic analyses only considering cases with the fetch from the coast exceeding 30 km. Compared to a previously published climatology of the Baltic Sea basin for the period 1961-1990, considerable differences in the air temperature and wind speed were found. They are partly related to the warm and windy conditions in 1990s and partly to the fact that the lighthouse station data from the open-sea regions were not available for the previous study. The turbulent surface fluxes, calculated using the bulk method from the R/V Aranda data, showed that the monthly median sensible heat flux was a few W m^{-2} *upwards in April, July, and August. In May and June, a stable stratification and a downward sensible heat flux prevailed. The monthly median latent heat fluxes were from sea to air from April to August, but downward latent heat fluxes occurred occasionally. In April-August, the relative humidity, sensible heat flux, and the Bowen ratio had their diurnal maximum values in the morning, while the latent heat flux had it in the afternoon. The fetch effect on the marine meteorological quantities was studied in stationary conditions. Depending on the meteorological conditions, the air temperature, specific and relative humidity and sensible heat flux either increased or decreased with fetch. The decrease of wind speed could be explained by probable mesoscale circulation systems, and the decrease of relative humidity with fetch was due to a dominating increase of air temperature compared to an increase of specific humidity. The changes in the sensible heat flux were related to an air-mass modification towards a neutral stratification. The latent heat flux had its largest values within 30 km off the coast.*

Key words: Baltic Sea, marine meteorology, air-sea interaction, surface fluxes, fetch from the coast

1. Introduction

Despite its shallowness and limited spatial extension, the Baltic Sea strongly interacts with the atmosphere and influences the weather and climate of northern Europe. Due to the large heat capacity of the sea, the sea surface temperature follows the air and land surface temperature with a considerable delay in the annual cycle. In late spring and autumn, the absolute difference between the land and sea surface temperature often exceeds 10°C, and in winter it can even reach 30°C if the sea is open. In addition to the surface temperature, the land-sea differences in surface moisture and aerodynamic roughness are often large. These differences result to large gradients in the turbulent surface fluxes in the coastal zone, and the air-mass advected over the Baltic Sea is therefore subject to a modification. The modification in the atmospheric boundary layer (ABL) mostly takes place via the turbulent fluxes of momentum, sensible heat, and latent heat, but the contribution of the net longwave radiation can also be large. The horizontal scales of the Baltic Sea are characterized by basin scales of the order of 10^2 to 10^3 km, while the straits have horizontal scales of the order of 10^1 to 10^2 km. These scales are important from the point of view of air-mass modification towards equilibrium with the underlying surface.

A comprehensive summary of the climate of the Baltic Sea basin in the normal period from 1961 to 1990 has been presented by *Mietus* (1998). It is based on data from coastal and archipelago stations as well as from ships. The Baltic Sea was also included in a study on the Atlantic Ocean climatology based on ship reports in the period from 1940 to 1979 (*Lindau*, 2001). Evaporation from the Baltic Sea has been studied by several authors, e.g., *Henning* (1986), *Omstedt et al.* (1997), and *Rutgersson et al.* (2001). Most studies on the structure of the ABL over the Baltic Sea have based on coastal and archipelago stations (*Smedman et al*., 1995; 1997; *Källstrand*, 1998) and airborne measurements (*Tjernström and Smedman*, 1993; *Smedman et al*., 1994; *Högström et al*., 1999). The ABL structure has also been studied on the basis of rawinsonde soundings from ships (*Uotila et al*., 1997). The coastal effects on wind speed and evaporation have been addressed by *Bumke et al*. (1998), and the wind conditions over the open sea far from the coasts by *Launiainen and Laurila* (1984) on the basis of observations at lighthouses. The ABL over the Baltic Sea ice cover has been studied by, e.g., *Joffre* (1982; 1983)*, Launiainen et al.* (2001), and *Vihma and Brümmer* (2002). The influence of the Baltic Sea on the ABL represents a challenge for the modelling community (e.g., *Gustafsson et al*. 1998; *Pettersson and Rontu*, 1999).

It is well known that the climate in Fenno-Scandia (and globally) has been warm in 1990s (e.g. *Tuomenvirta et al*., 2000; *Klein Tank et al*., 2002). This information is, however, mostly based on observations over land and coastal areas, and we are not aware of specific analyses on the marine climate far from the coasts over the northern Baltic Sea in 1990s. Such an analysis would be useful for oceanographers, marine meteorologists, and climatologists for at least the following reasons: (a) in providing basic marine meteorological information, (b) in understanding the differences in climate and weather between land, coastal areas, and the open Baltic Sea, and (c) in understanding the applicability of various atmospheric forcing fields for marine models. The analysis should therefore include specific studies on the air-sea exchange processes as well as on the horizontal gradients in the coastal regions. To achieve these goals we analyze a data set recorded by three marine meteorological stations in various regions in the northern Baltic Sea (Bogskär, Kalbådagrund, and Kemi I) and R/V Aranda from 1991 to 1999 (Fig. 1). We study the representativeness of the R/V Aranda data set, and then estimate the stratification parameters and turbulent fluxes of sensible and latent heat on the basis of the wind, temperature and humidity data from R/V Aranda. The statistics of the observed and calculated quantities are studied paying particular attention to the seasonal and diurnal variations. Considering the spatial variations, we study the effect of the fetch over the sea by analyzing cases of air-mass modification.

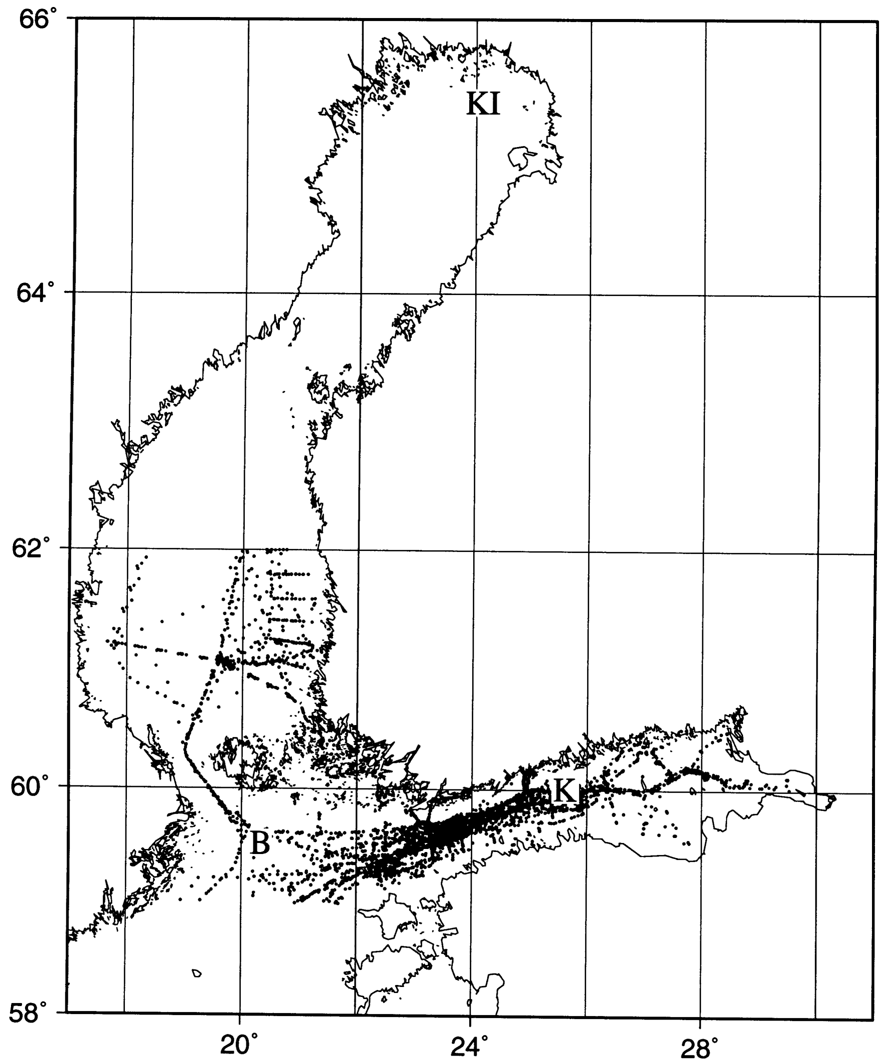


Fig. 1. R/V Aranda marine meteorological observations in the region of 59 to 62°N in the Baltic Sea in April–August in 1991-1999. The lighthouse stations of the Finnish Meteorological Institute are marked by letters: B stands for Bogskär, K for Kalbådagrund, and KI for Kemi I.

2. Data

2.1 Marine meteorological station data

We analyzed the data on air temperature, humidity, wind speed, and wind direction from three lighthouse stations operated by the Finnish Meteorological Institute (FMI) in the northern Baltic Sea (Fig. 1). They are the FMI stations least influenced by the coast and archipelago. Bogskär locates in the northern Baltic Proper 55 km from the nearest coast. Kalbådagrund locates in the Gulf of Finland 29 km from the Finnish and 37 km from the Estonian coast. If the wind is from the sector 235-265° or 65-120°, the fetch over the sea is longer than 100 km. The lighthouse Kemi I locates in the Bothnian Bay 42 km from the nearest coast, and winds from the sector 180-265° have over-sea fetches longer than 100 km. In Bogskär, the air temperature and humidity were measured onwards from 6 August, 1992, but there was an interruption in the humidity measurements from 14 June, 1996, to 30 September, 1997. In Kalbådagrund the air temperature was measured onwards from 31 March and air humidity onwards from 5 July, 1992. In Kemi I the air temperature measurements covered whole our study period, and the air humidity was measured onwards from 11 April, 1991. Kemi I is surrounded by sea ice typically from December to May, while the ice conditions in Kalbådagrund and Bogskär vary from year to year. In the mild 1990s, Bogskär was free of ice except in February – March 1994 and 1996, and the ice season in Kalbådagrund typically lasted from January to March or mid-April.

The data were recorded as instantaneous values with 3 h intervals, except of the wind speed that was given as 10-minute means. The observation heights ranged from 26 to 32 m above the sea level, except that at Kemi I the air temperature and humidity were measured at the height of 22 m. Since the mutual differences in the observation heights between the three stations were small and the vertical gradients are small at such heights above the sea surface, no height correction was made for the mutual comparison of the data.

2.2 R/V Aranda observations

The R/V Aranda is equipped with an automatic ship weather station (MILOS 200, Vaisala Ltd.) including sensors for wind speed (V) and direction (DD), air temperature (Ta), air relative humidity (RH), atmospheric pressure (p), and sea surface temperature (Ts). The observation height is 19 m for the wind speed and direction and 14 m for the other meteorological quantities. The sensors are regularly calibrated and maintained by Vaisala. In general, the following data are registered: the hourly mean value, the instantaneous value (averaged over 2 s for the wind speed) once an hour, and the maximum and minimum values during the hour. For the wind speed, also 1 minute means are registered, and during expeditions specifically addressing the air-sea interaction all data are registered more frequently. In this study we use the hourly means, if not otherwise mentioned.

The data set (1991-1999) includes about 21000 hourly mean values for each quantity (V, DD, Ta, RH, Ts, p) over the Baltic Sea. The data set is, however, not representative enough for climatological analyses for the autumn and winter seasons, and in no season there are enough data north of 62°N and south of 59°N. The highest data coverage is from the region of 59-62°N from April to August. We accordingly restrict our analyses to this region and season. Further, concentrating on the conditions over the open sea far from the coasts, we analysed only cases with the fetch over the sea exceeding 30 km. Using these criteria (season, region, and fetch) we had 6785 hourly means. In addition, we selected cases for studies of air-mass modification with fetch over the sea (Section 5).

The ship tends, or at least is liable, to affect the measured wind, air temperature, and humidity. The ship weather station has therefore two sensors for each observation quantity, one on the port side and the other on the starboard side; the data from the up-wind side are registered. This practically eliminates the ship's interference with the air temperature and humidity.

The sea surface temperature was taken from a sensor installed on the ship's hull 3 m below the sea surface. In conditions of light winds this observation may naturally differ from the real sea surface temperature. To better understand this source of error, we investigated a set of 150 randomly selected CTD soundings from the northern Baltic Sea in April–August, paying attention to the temperature difference between the surface and the depth of 3 m (ΔT_3). In 77% of the cases $|\Delta T_3|$ was less than 0.2 K, in 87% less than 0.5 K, and in 96% less than 1.0 K. The maximum $|\Delta T_3|$ was, however, as large as 3.1 K. The cases of $|\Delta T_3|$ exceeding 0.5 K occurred mostly in calm and sunny conditions in June with the temperature always increasing towards the surface. The surface temperature measured by a CTD sonde represents, however, the bulk temperature of the sea surface. It often differs from the skin temperature (*Grassl*, 1976), which should be used for the calculation of the turbulent surface fluxes. The skin temperature is typically 0.1-0.4 K colder than the bulk temperature (*Wick et al.,* 1996). Hence, on average, the R/V Aranda data most likely do not have as large a cold bias as the CTD comparison would indicate. In individual cases with a large ΔT_3, the bias is, however, considerable. We cannot base our study on the more accurate CTD data, because soundings have not been made frequently enough. Also, satellite-based Ts data was found to be still too in-accurate over the Baltic Sea.

We studied the disturbance of the wind speed measurement caused by the ship on the basis of data gathered from a meteorological mast deployed on sea ice in the Gulf of Bothnia in March 1998. The 10-m-high mast was located some 300 m from R/V Aranda, which was anchored in the land-fast ice. The wind speed measured at the up-permost level (10 m) of the mast was compared to the R/V Aranda weather station measurements (at a height of 19 m). The effect of the different observation heights de-pended on the thermal stratification, which was determined on the basis of the mast data, and this effect was taken into account in the calculations. Due to the asymmetric superstructure of the ship, the difference between the measurements depended on the direction of the wind with respect to the ship. We see from Figure 2 that the disturbance caused by the ship is typically less than 7%, but reaches 14% for a certain wind direc-tion. The results indicate that with a typical marine wind speed of 10 m/s, the absolute error ranges from 0.2 to 1.4 m s^{-1}. During an hour-long observation period the wind di-rection relative to the ship typically varies, and therefore it was not possible to correct the data for this error. In the open sea during strong winds, the rolling of the ship may cause some additional error.

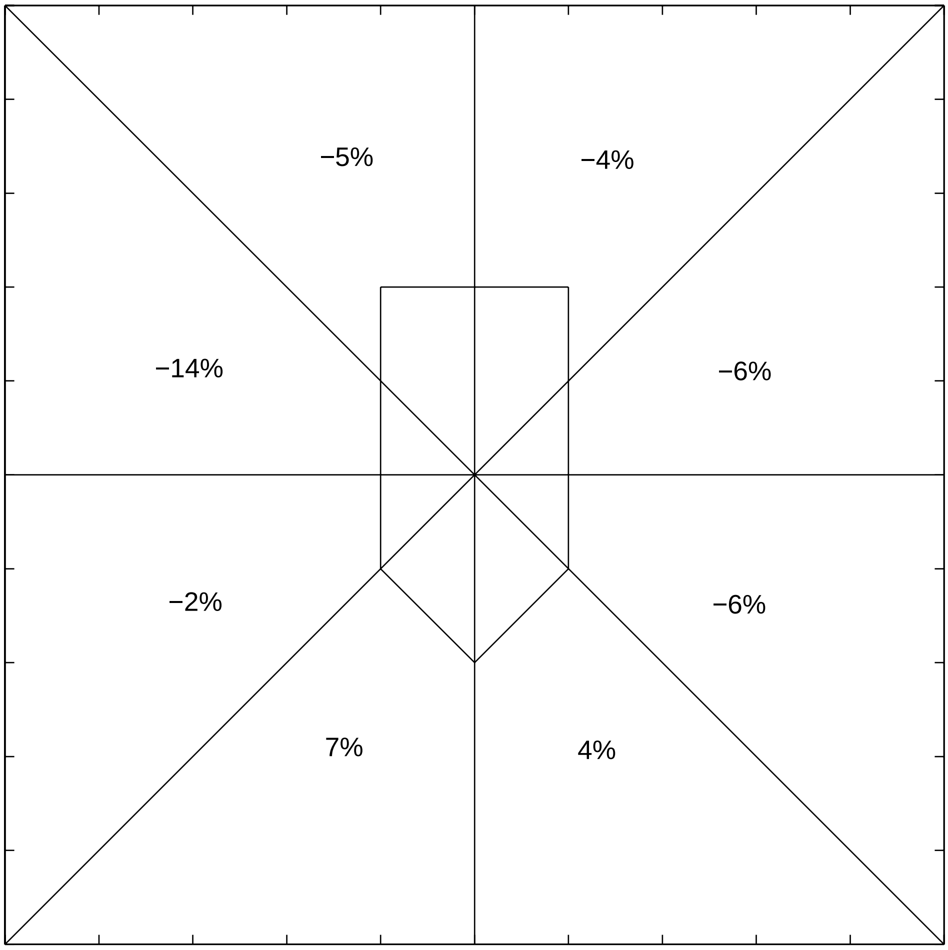


Fig. 2. Relative error in the wind speed measurement of the R/V Aranda weather station compared with a mast measurement over sea ice. The error is shown for various sectors of wind direction with respect to the ship. The bow of the ship is pointing towards the bottom of the page.

2.3 Calculated quantities based on R/V Aranda data

In addition to the direct observations, we calculated the turbulent surface fluxes and the stratification parameters. In the lack of observations on the sea surface temperature, these could not be calculated on the basis of the FMI station data.

The turbulent surface fluxes of sensible heat (H), latent heat (LE), and momentum (τ) were calculated applying the bulk formulae:

$$H = \rho c_p k^2 \left[\ln(z/z_0) - \Psi_M(z/L)\right]^{-1} \left[\ln(z/z_{tq}) - \Psi_{HE}(z/L)\right]^{-1} (\theta_S - \theta_Z) V_Z = \rho c_p C_{HEZ} (\theta_S - \theta_Z) V_Z \tag{1}$$

$$LE = \rho \lambda k^2 \left[\ln(z/z_0) - \Psi_M(z/L)\right]^{-1} \left[\ln(z/z_{tq}) - \Psi_{HE}(z/L)\right]^{-1} (q_S - q_Z) V_Z = \rho \lambda C_{HEZ} (q_S - q_Z) V_Z \tag{2}$$

$$\tau = \rho u^{*2} = \rho k^2 \left[\ln(z/z_0) - \Psi_M(z/L)\right]^{-2} V_Z^2 = \rho C_{DZ} V_Z^2 \qquad (3)$$

where V is the wind speed, θ is the potential temperature and q the specific humidity. The subscript s refers to the surface, and z to a reference height in the air. The ρ is the air density, c_p is the specific heat of air, k is the von Karman constant ($\cong 0.4$), λ is the latent heat of evaporation. z_0 and z_{tq} are the roughness lengths for momentum and heat/moisture, respectively, and u_* is the friction velocity. Functions ψ_M and ψ_{HE} describe the stability effects on the turbulent mixing, and depend on the parameter z/L, where L is the Obukhov length. The transfer coefficients for momentum (C_D) and heat/moisture (C_{HE}) are related to the roughness lengths, and to the Ψ-functions as shown in (1)-(3). Assuming an equality for the transfer of heat and moisture (e.g. *Smith*, 1989; *DeCosmo et al.*, 1996) we use the single variables z_{tq}, ψ_{HE}, and C_{HE}. The effect of the wind speed and thermal stratification on the transfer coefficients, as well as the complexity due to the different observation heights for the wind and air temperature and humidity, were taken into account by applying the algorithm of *Launiainen and Vihma* (1990). The neutral C_D at the height of 10 m was calculated according to the wind-dependent formula of *Wu* (1980): $C_{DN10m} = (0.80 + 0.065 \mathrm{x} V_{10m})/1000$, where V_{10m} was calculated from the observed V_{19m}. The equation should be valid for a basin with shorter fetches than over the ocean. The wind effect on the neutral 10-m transfer coefficient for heat/moisture reads: $C_{HEN10m} = (0.70 + 0.040 \mathrm{x} V_{10m})/1000$. We calculated the effect of thermal stratification on the Ψ - functions following *Holtslag and de Bruin* (1988) in stable conditions, and *Högström* (1988) in unstable conditions.

The stability parameter z/L is defined as:

$$\frac{z}{L} = -\frac{zgkH(1 + 0.61T_0 c_p E / H)}{u_*^3 T_0 c_p \rho} \qquad (4)$$

where T_0 is the mean temperature (in K) in the atmospheric surface layer, and g is the acceleration due to gravity. As z/L and the turbulent fluxes depend on each other (via u_*, H, and the evaporation E), they were obtained via the same iteration routine (*Launiainen and Vihma*, 1990). A bulk-Richardson number Ri was also applied in studies related to the thermal stratification; it was calculated for z = 14 m as follows:

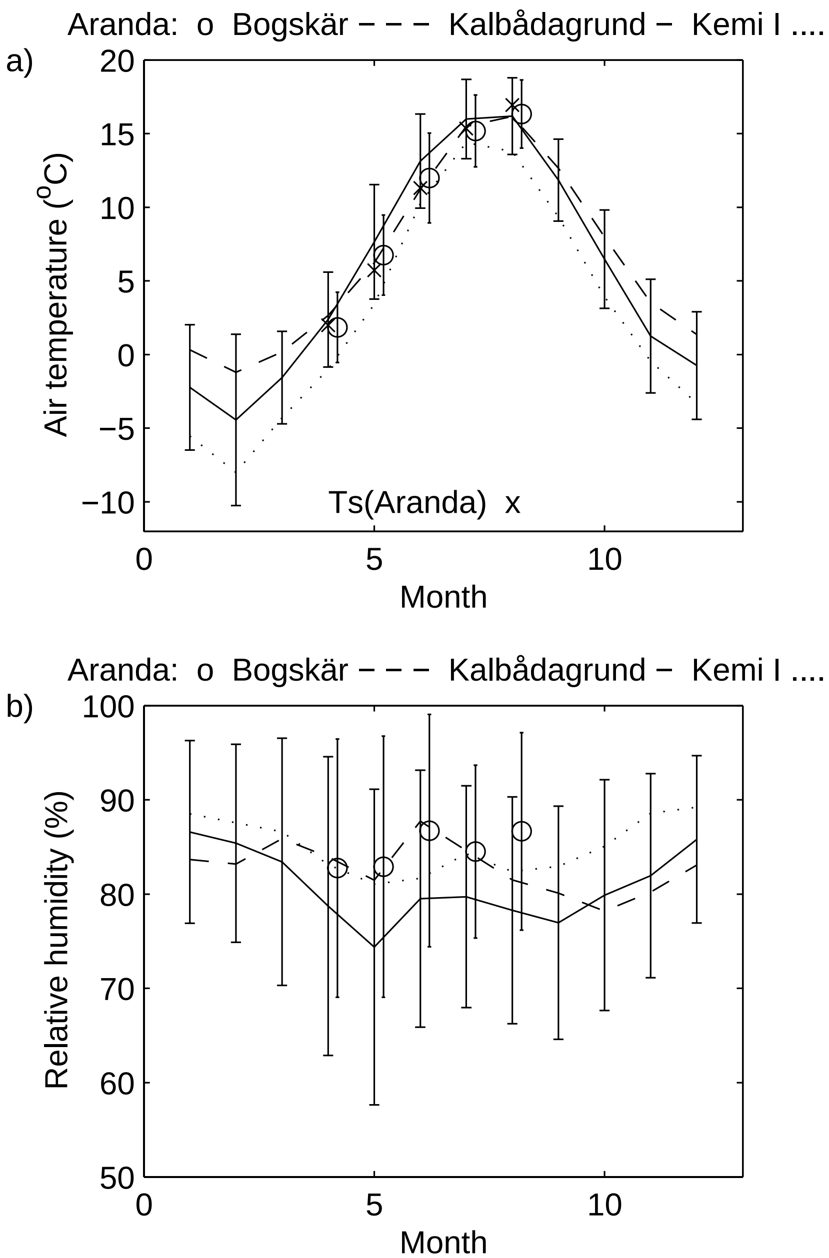
$$Ri = \frac{zg(\theta_z - \theta_s)}{\overline{T} V^2} \qquad (5)$$

where $\overline{T}$ is the mean temperature (in K) in the layer between the reference height z and the sea surface. The stability parameters were normally calculated only for wind speeds exceeding 1 m/s. Errors in the fluxes and stratification parameters are discussed in section 4.2.

3. *Temperature, humidity and wind*

3.1 *Annual cycles and distributions*

When discussing seasonal statistics we define spring as the period from March to May, summer as June to August, autumn as September to November, and winter as December to February. The annual cycles of the air temperature as well as air specific and relative humidity are shown in Figure 3. The monthly means are shown for the three FMI stations together with the standard deviations for one of the stations (Kalbådagrund). The R/V Aranda observations are shown for the period from April to August.



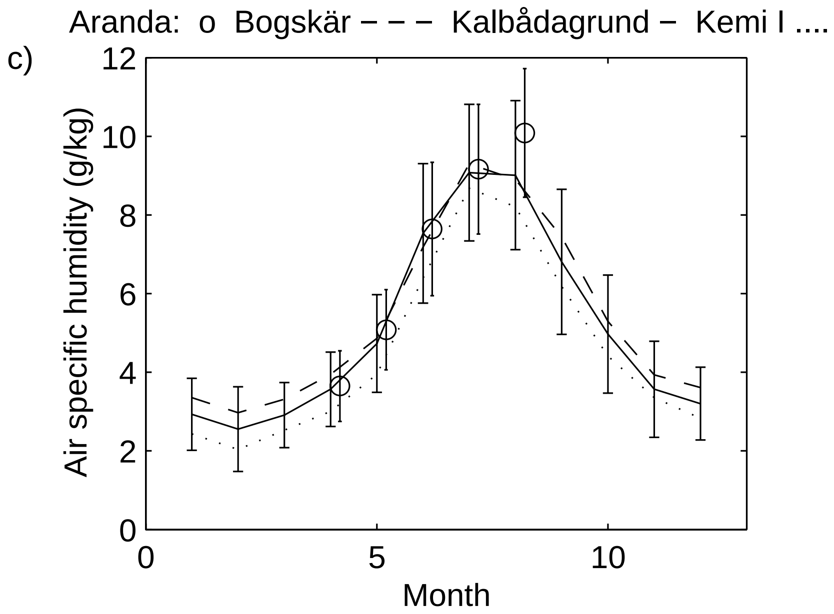


Fig. 3. Monthly means of the (a) air and sea surface temperature (the latter at R/V Aranda only), (b) relative humidity, and (c) specific humidity in 1991-1999 at the lighthouses of Bogskär, Kalbådagrund, and Kemi I, as well as at R/V Aranda in the region of 59-62°N with a fetch from the coast longer than 30 km. The vertical bars show the standard deviation at Kalbådagrund and R/V Aranda (bars of the latter shifted to the right).

We see that the maximum values for air temperature are reached in August, except for Kemi I in July. The maximum specific humidities are reached in July, except for R/V Aranda data in August. The relative humidity shows a less clear annual cycle, although the Kemi I and Kalbådagrund data suggest maximum values in winter. The air temperature has its largest standard deviation in February, the air relative humidity in March–June, and the air specific humidity in July–September. In winter, the air temperature in Kemi I is 5 to 7 K lower than in Bogskär, but in summer the difference is only 1 to 2.5 K. In Kalbådagrund the air temperature is most of the year between those of Bogskär and Kemi I, but from May to July Kalbådagrund is the warmest station, and in April and August equally warm to Bogskär.

We compared our air temperature data to the climatology over the Baltic Sea for the normal period 1961-1990 presented by *Mietus* (1998). The largest differences occurred in the Bothnian Bay in winter: the mean temperatures at Kemi I for 1991-1999 were -3.3 (December), -5.5 (January) and -8.0°C (February), while they were -9, -11.5, and -10.5°C, respectively, in *Mietus* (1998, the values were interpolated from his figures 1.42-1.53). In January, the data of 1991-1999 showed warmer temperatures also in Bogskär (1.3 K warmer than in *Mietus* (1998)) and Kalbådagrund (3.5 K warmer). On the other hand, in April and May the temperatures in *Mietus* (1998) were warmer, the largest differences of 1.5 K occurring at Kemi I. The annual mean temperatures were 6.4°C at Bogskär, 5.5°C at Kalbådagrund, and 2.7°C at Kemi I, indicating a south-north gradient of 0.56 K / 100 km between Bogskär and Kemi I. This is lower than the value of 0.7 to 1.2 K / 100 km calculated by *Mietus* (1998) for the coast of the Gulf of Bothnia. In agreement with Mietus, the spatial variability is much larger in winter than in summer (Fig. 3). The relative humidities agreed with *Mietus* (1998) within some 5%. The low spring and summer values at the Gulf of Finland (Kalbådagrund) are in agreement with *Mietus* (1998), and may be related to the advection of air from the continent.

An apparent annual cycle is found in the wind speed, and in particular in its variance (Fig. 4), as in *Launiainen and Laurila* (1984). The annual cycles are related to synoptic-scale activity, which affects the variation of the geostrophic wind. The annual cycles are additionally affected by the surface layer stratification, which is most stable during spring and early summer, reducing the near-surface wind speeds. The monthly mean wind is highest in November (Kemi I), December (Kalbådagrund), or in January (Bogskär). According to *Launiainen and Laurila* (1984), in 1977-1982 the maximum at Kemi I was reached in October and at Kalbådagrund in November. *Mietus* (1998) reported that the Finnish coastal stations along the Gulf of Bothnia reach monthly maximums in October. The differences between the stations are largest in winter, when the winds are strongest at Bogskär and weakest at Kemi I. In this data set from 1991 to 1999 at all the three stations, the monthly mean wind was throughout the year stronger than in the 1961-1990 climatology of *Mietus* (1998). The differences in the monthly mean wind were from 1 to 5 m s^{-1}.

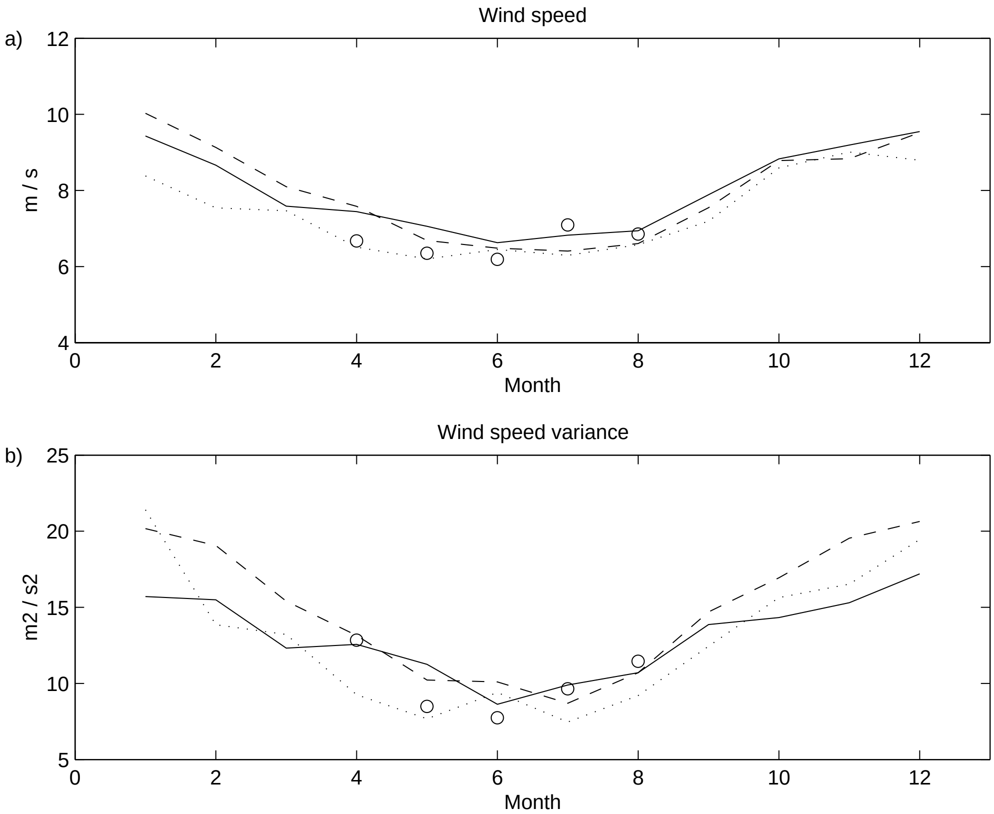


Fig. 4. Monthly means of (a) the wind speed and (b) the wind speed variance in Bogskär (dashed line), Kalbådagrund (solid line), and Kemi I (dotted line) in 1991-1999. The R/V Aranda data are shown for April to August (circles).

The distributions of air temperature, and specific and relative humidity at Bogskär and Kemi I are shown in Figure 5. The distributions at Kalbådagrund (not shown) resemble those at Bogskär, except that the temperature and relative humidity distributions are a bit wider indicating a less marine climate (compare to Fig. 3). At Bogskär the annual air temperature distribution has two peaks, the primary one at 1-3°C and the secondary one at 13-15°C. At Kemi I the peak is around 0°C and no clear warmer peak exists. The air specific and relative humidities have skewed distributions. The relative humidities exceeding 95% are more common at Kemi I than Bogskär. This result is supported by the charts of *Mietus* (1998), which show that the occurrence of fog over the Baltic Sea often has its maximums over the Bothnian Bay and the Danish Straits.

The shape of the wind speed distribution is close to the Weibull distribution. At Bogskär the mean wind speed exceeds 10 m s^{-1} for 25% and 20 m s^{-1} for 0.5% of the observations. The corresponding numbers for Kemi I are 19% and 0.1%, and for Kalbådagrund 24% and 0.2%. *Launiainen and Laurila* (1984) calculated much lower values for the period 1977-1982: 13.6% and 0.09% at Kemi I, and 15.5% and 0.008% at Kalbådagrund.

3.2 Diurnal cycles

To avoid confusion between the winter and summer time and between the local time of Finland and Sweden, all numbers are given with respect to the UTC time. Over the study region on average, the local solar time is 3.7 h in advance of the UTC time. When using terms 'morning' etc., we refer to the solar time of day.

The mean diurnal cycles of the air temperature, specific and relative humidity, and wind speed are shown in Figure 6. The peak in the air temperature diurnal cycle occurs approximately five hours later than over land areas in Finland (*Heino*, 1977). In summer, the range of the lowest and highest daily air temperatures over the Baltic Sea is between those over the oceans and lakes. At Kalbådagrund and Kemi I the summertime range is 1.7 K, while it is only 0.8 K at Bogskär, which represents the most marine environment, and 1.1 K for the Aranda data. The corresponding range over the oceans is 0.5-0.6 K (*Roll*, 1965). *Elomaa* (1976) reported diurnal air temperature ranges of 7 to 12 K over a medium-size Finnish lake, and *Heikinheimo et al*. (1999) found ranges of 6 to 10 K over a Swedish lake. In Haparanda, at the coast of the Bay of Bothnia only 50 km north of Kemi I, the mean diurnal temperature range is 9 K in summer and 8 K in winter (*Tuomenvirta et al*., 2000; study period 1890-1990). At Bogskär, Kalbådagrund and Kemi I, the diurnal air temperature range in winter is only 23-28% of that in summer (Figures 6 a and b), i.e., at Kemi I the winter range is less than 0.6 K, only 7% of that in the nearby Haparanda.

The relative humidity usually reaches its highest values in the morning. In summer, the diurnal range is 8% at Kalbådagrund and Kemi I, 5% for Aranda data, but only 3% at Bogskär. In winter, the ranges are less than 2%, and the smallest one, 0.4%, occurs at Kemi I, where the relative humidity keeps through the day close to saturation with respect to ice (the numbers are shown with respect to water saturation, as conventional).

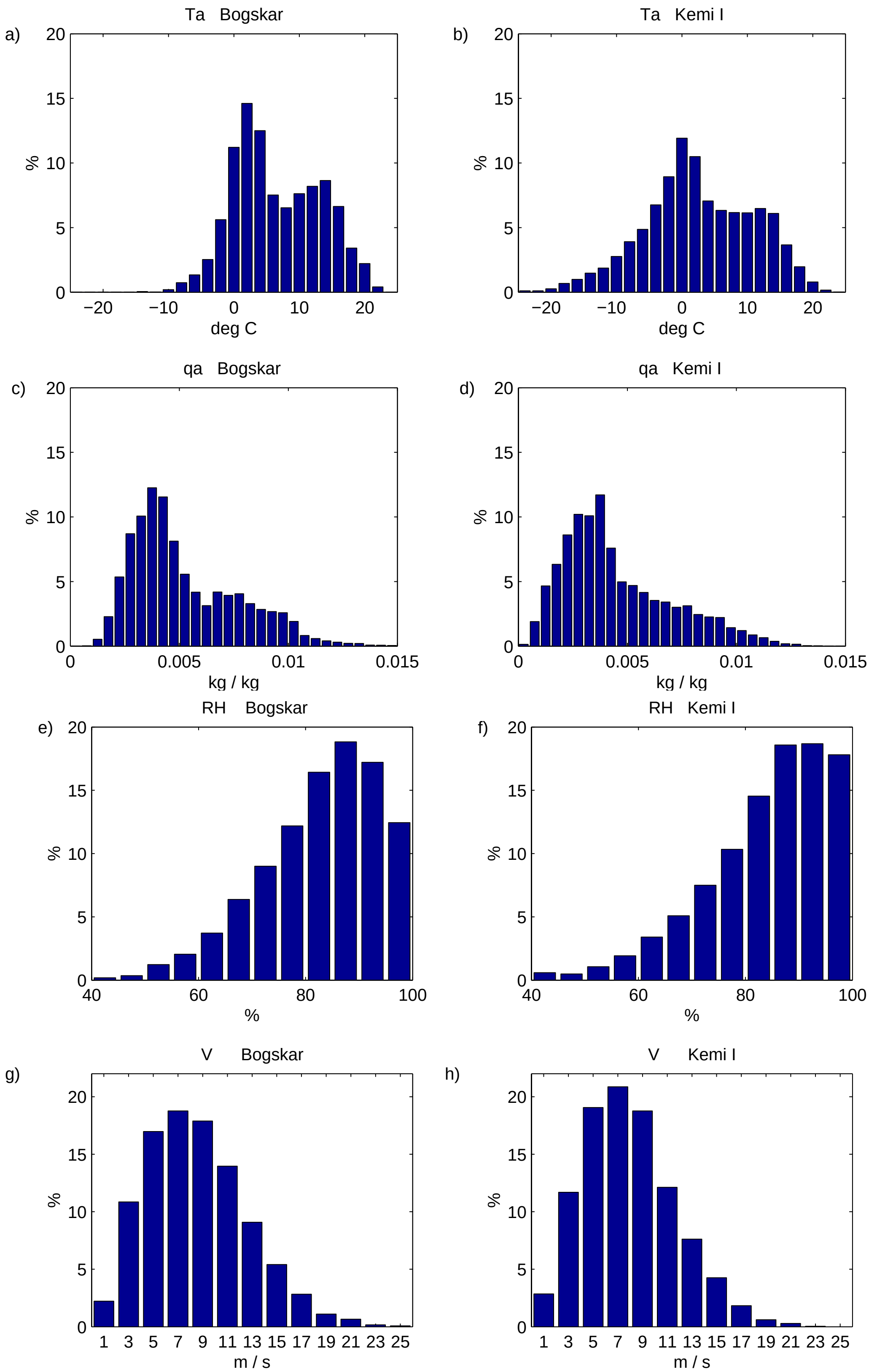


Fig. 5. Mean annual distributions of the air temperature (Ta), specific humidity (qa), relative humidity (RH), and wind speed (V) in Bogskär and Kemi I in 1991-1999.

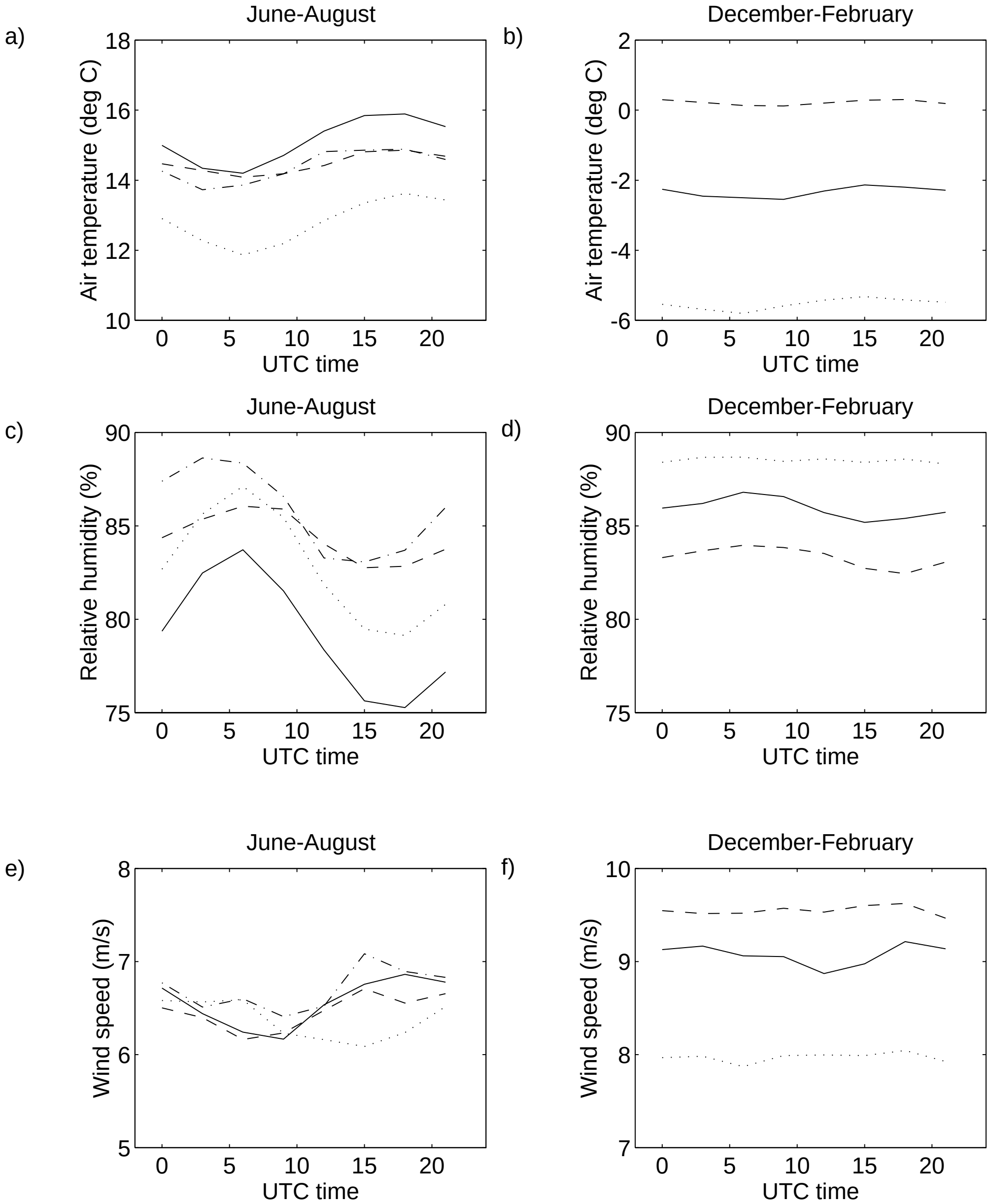


Fig. 6. Mean diurnal cycle of the air temperature, relative humidity, and wind speed in summer and winter in Bogskär (dashed line), Kalbådagrund (solid line), Kemi I (dotted line), and R/V Aranda (dot-dashed line, summer data only).

In summer, at Bogskär, Kalbådagrung, and R/V Aranda the wind speed has its diurnal minimum values in the morning or noon, while at Kemi I the minimum occurs in the afternoon. A sea-breeze circulation could generate an afternoon maximum. On the other hand, if air masses heated over a land surface are advected over the colder sea, the stability effect would reduce the afternoon wind speeds. The sea-breeze effect is well-known over the Gulf of Finland (*Savijärvi and Alestalo,* 1988), and may dominate for the Kalbådagrund and Aranda data, while the stability effect may dominate at Kemi I. The reason for the diurnal cycle at Bogskär remains unclear. In winter, a diurnal cycle can be identified only at Kalbådagrund, where the minimum occurs in the afternoon.

3.3 *Comparisons of lighthouse and R/V Aranda data*

Comparisons in Figures 3, 4, and 6 suggest that the R/V Aranda data set represents well the marine meteorological statistics for the period from April to August in the region of 59-62°N for the cases with fetch over the sea exceeding 30 km. This is evident because the monthly means, standard deviations, and diurnal cycles are reasonably well in agreement with the Bogskär and Kalbådagrund data. We therefore consider the R/V Aranda data set representative enough to be used for the calculation and statistical analyses of the surface fluxes. Another aspect is the inaccuracy related to calculation of the turbulent surface fluxes and stratification parameters in individual cases. Here we present some comparisons that can be used as a reasonable approach to the error estimates, and the flux errors will be discussed in section 4.2.

We compared the Kalbådagrund and Aranda data in cases when the ship was within 1 km from the lighthouse (Fig. 7). The mean absolute differences were 0.2 K for the air temperature, 3% for the relative humidity, 10° for the wind direction, and 0.4 m/s

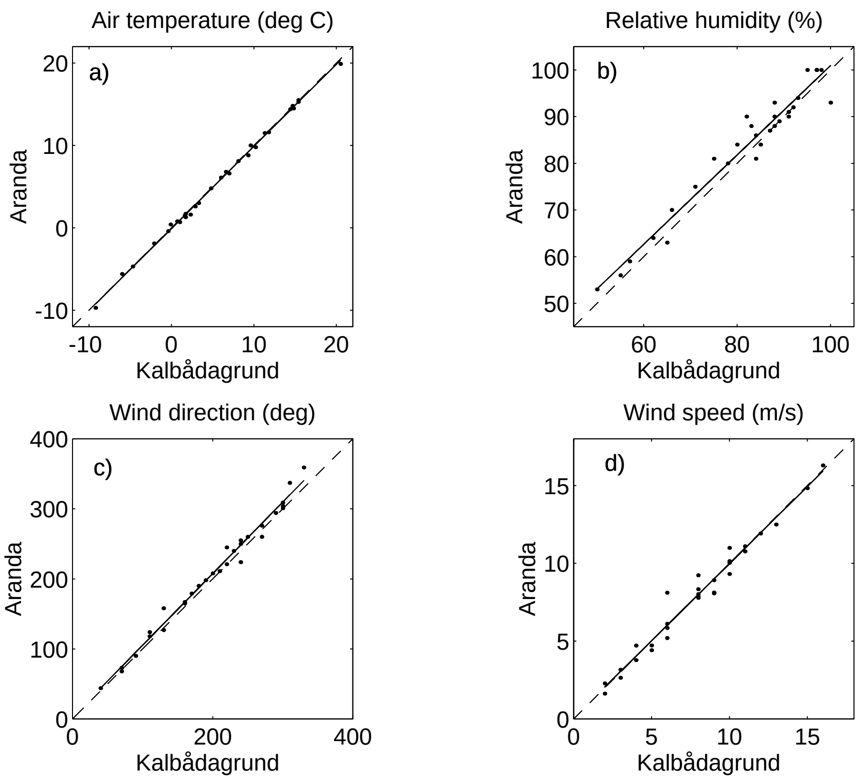


Fig. 7. Comparison of observations of (a) air temperature, (b) relative humidity, (c) wind direction, and (d) wind speed in cases when R/V Aranda was within 1 km of Kalbådagrund. The linear regression is shown as a solid line and the 1:1 line as a dashed one.

for the wind speed. A stability-dependent height correction was made before comparing the wind data. We stress that it is not clear whether the lighthouse observations are more accurate than the ship observations or vice versa (because of a difficult maintenance of lighthouse stations). Anyway, this comparison gives an overview of the uncertainty related to marine meteorological observations.

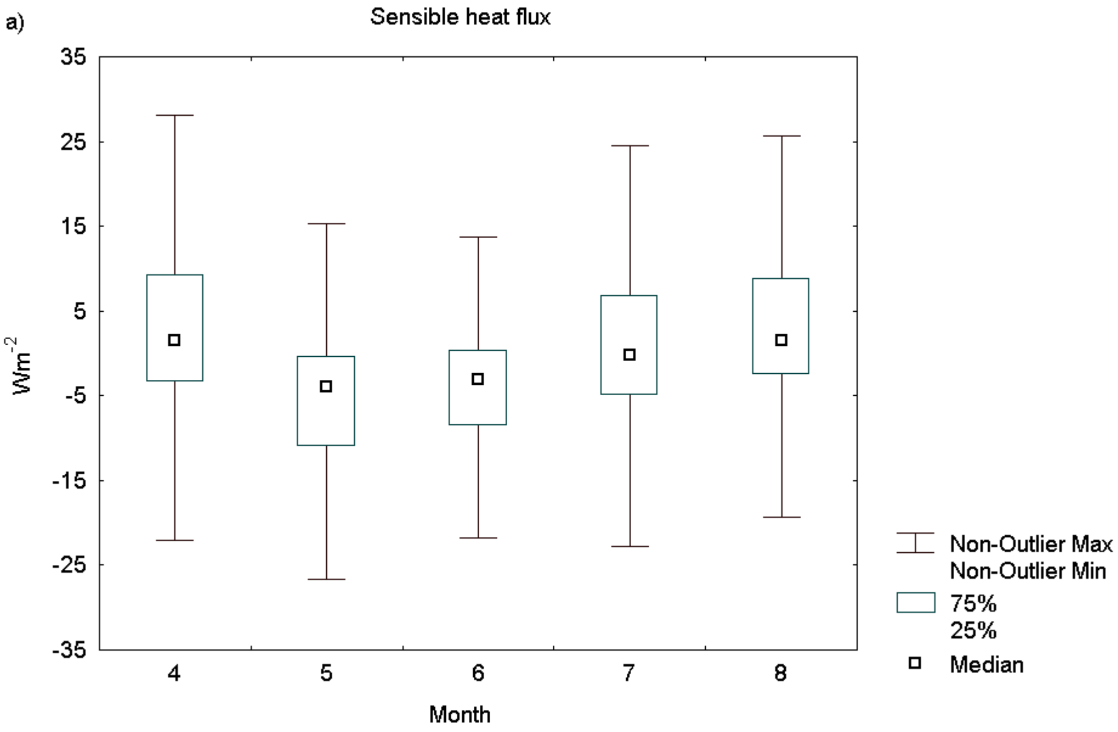
4. Turbulent surface fluxes

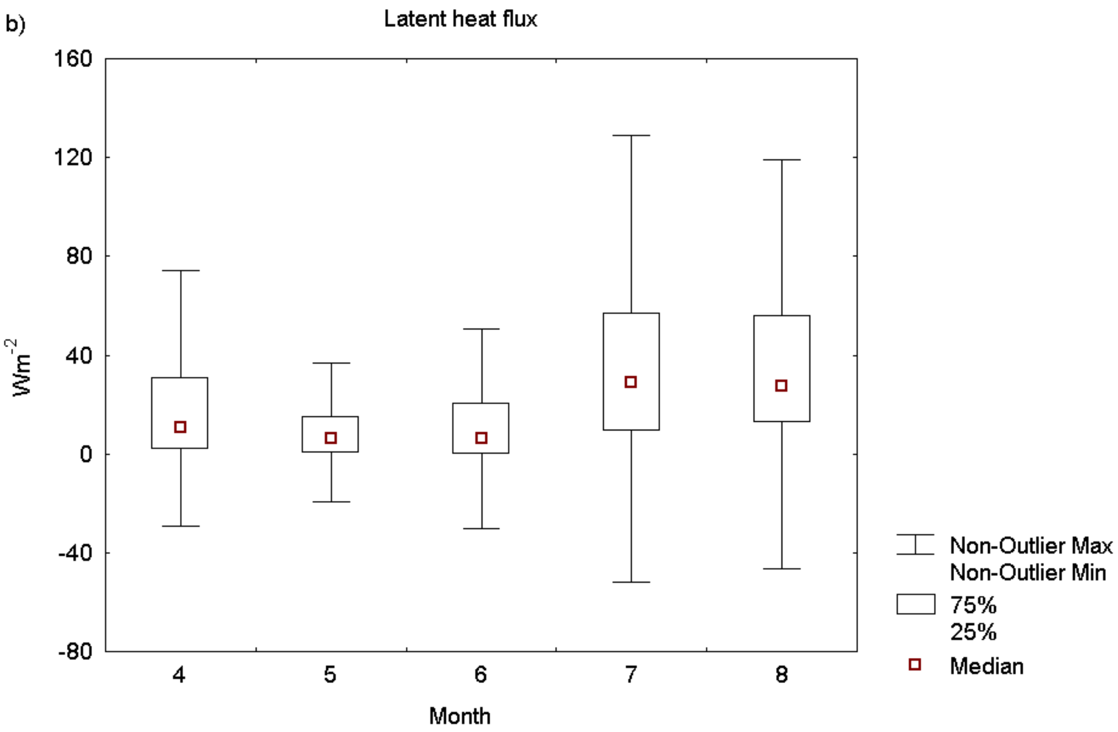
4.1 Results

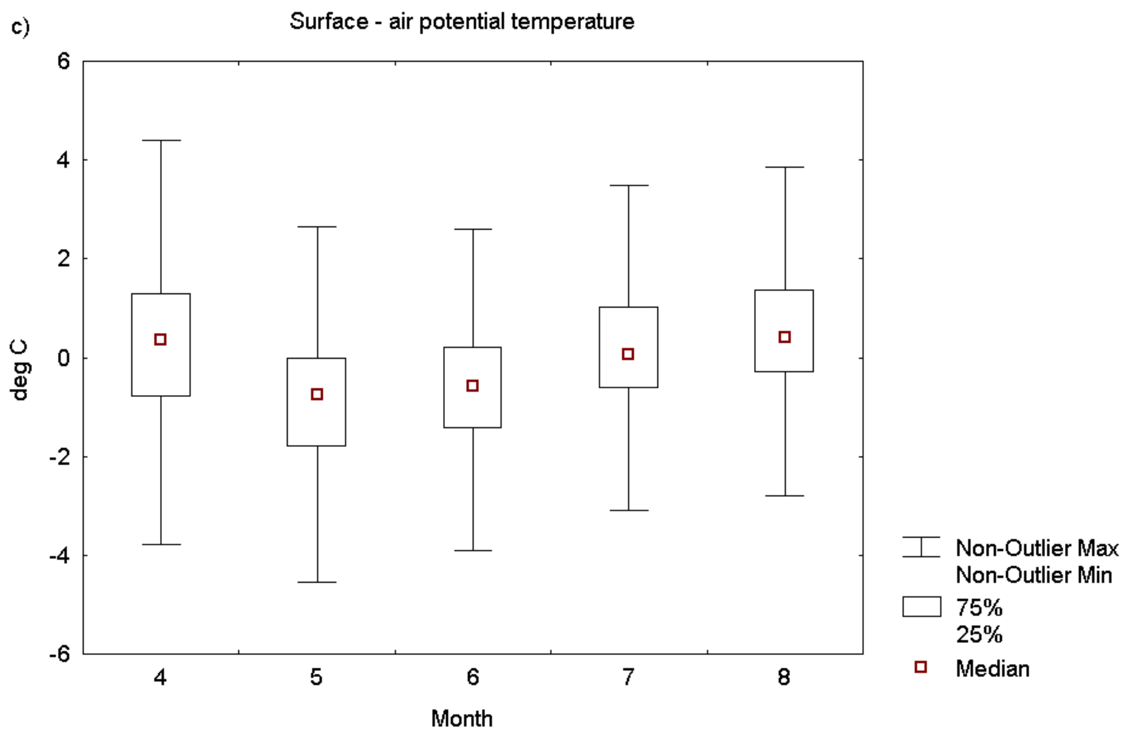
For the sensible and latent heat fluxes, the monthly median values, the range of values containing the middle 50% of the data, and the non-outlier maximum and minimum values are shown in Figure 8. (By an outlier we mean a value that differs from the median by more than two times the standard deviation.) We see that the magnitude of the monthly median sensible heat flux is very small, less than 5 W m^{-2}, in every month from April to August. The monthly median flux is from air to sea in May and June, practically zero in July, and from sea to air in April and August. The latent heat flux is from sea to air in each month for at least 75% of the time, and even more in July and August, when the largest fluxes occur. The reliability of the downward fluxes will be discussed in Section 4.2. In May and June the sea is colder than the air (Figure 8c), which reduces the latent heat flux. In at least 50% of the data, the surface-air potential temperature difference is less than 2 K (cf. Section 2.2 for the Ts errors). The distribution of LE is much wider than that of H (note the different scales in Figures 8a and b), and in 72% of the data $|H/LE| = |Bo| < 1$. The Bowen ratio (Bo) has a negative value in 40% of the cases, indicating a different direction of H and LE. The stability parameter 10/L suggests typically stable stratification in May and June.

The mean diurnal cycles of the sensible heat flux, latent heat flux, Bowen ratio, and the stability parameter 10/L in April-August are shown in Figure 9. (In the case of 10/L, the diurnal cycle is based on the median values of each hour.) It should be noted that the temporal variations are dominated by synoptic scale disturbances, and the mean diurnal cycles can be drawn only by averaging a large amount of data. The sensible heat flux has its largest values in the morning and the lowest ones in the evening. This is due to the fact that the sea surface temperature had a smaller diurnal cycle than the air temperature. Our data are in agreement with the observations of *Venäläinen et al*. (1999) over Swedish lakes. The latent heat flux had a diurnal cycle opposite to that of the sensible heat flux. This is because the sea surface temperature and the wind speed reached their peak values in the evening, while the air specific humidity was practically constant throughout the day.

Our results for the sensible and latent heat flux are approximately in agreement with the climatology of *Lindau* (2001), although his focus on the Atlantic Ocean prevents detailed comparisons. Our sensible heat fluxes are some 10 W m^{-2} larger than those of *Hankimo* (1964) for the Bothnian Sea, and in the latent heat flux the difference is 10-20 W m^{-2}, except that the agreement is good in August. Our latent heat fluxes


a)
Sensible heat flux
35
25
15
5
-5
-15
-25
-35
Wm^{-2}
4
5
6
7
8
Month
Non-Outlier Max
Non-Outlier Min
75%
25%
Median


b)
Latent heat flux
160
120
80
40
0
-40
-80
Wm^{-2}
4
5
6
7
8
Month
Non-Outlier Max
Non-Outlier Min
75%
25%
Median



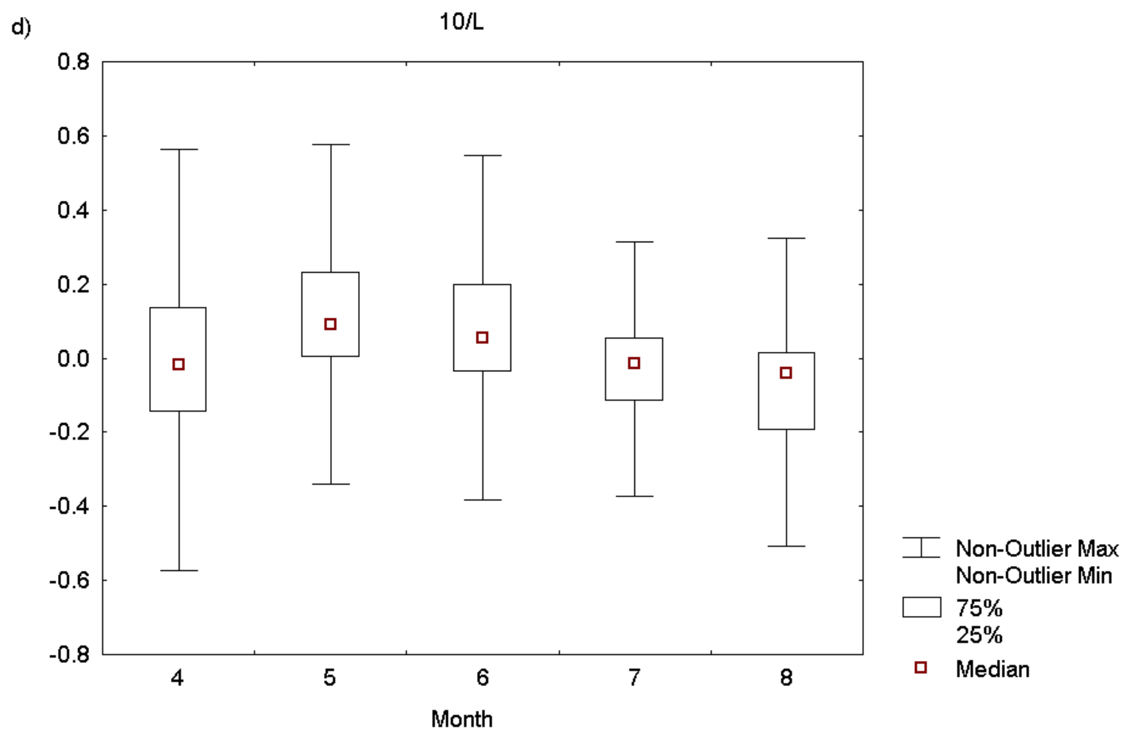


Fig. 8. The monthly median values, the range of values containing the middle 50% of the data, and the non-outlier maximum and minimum values of (a) sensible heat flux, (b) latent heat flux, (c) surface-air potential temperature difference, and (d) the stability parameter 10/L. An outlier is defined as a value that differs from the median by more than two times the standard deviation. The results are based on the R/V Aranda observations in the region of 59-62°N including only cases with a fetch from the coast longer than 30 km.

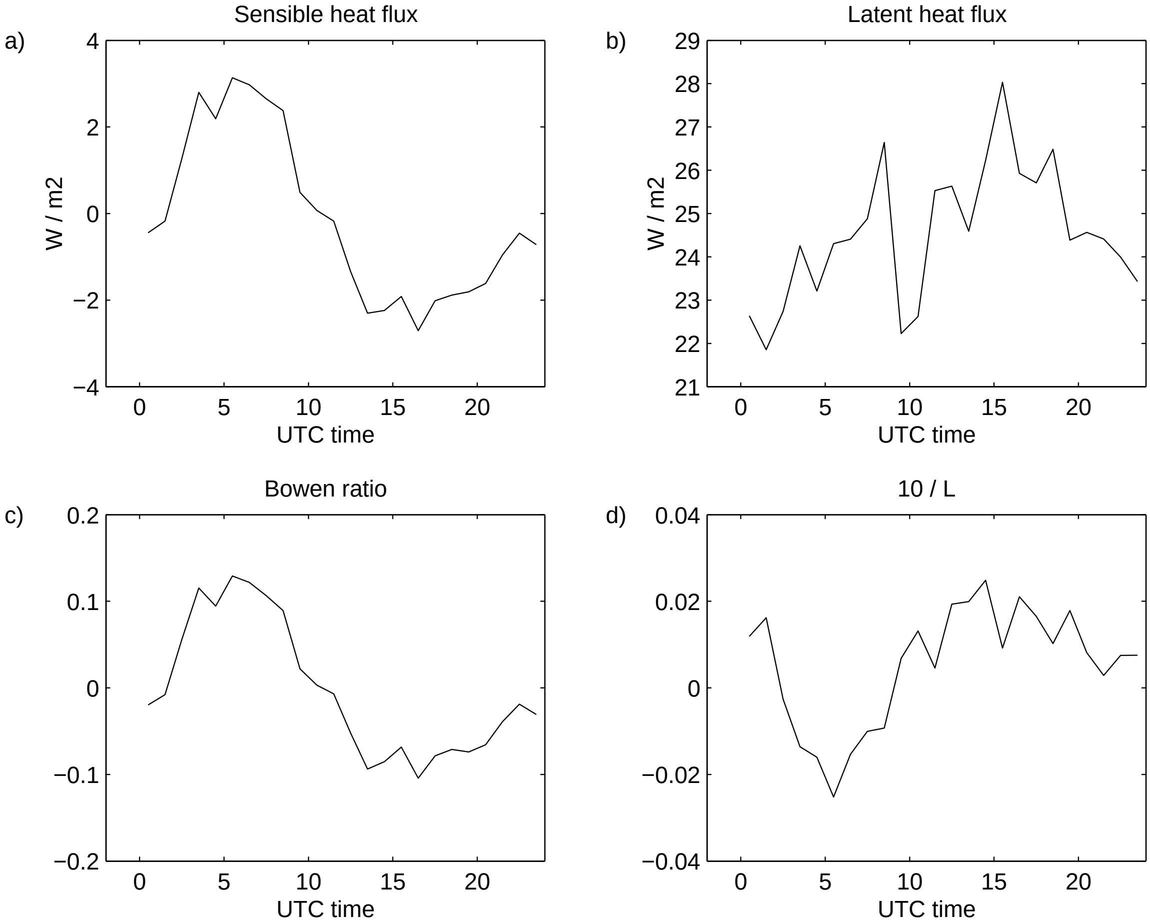


Fig. 9. Mean diurnal cycles of the (a) sensible heat flux, (b) latent heat flux, (c) Bowen ratio, and (d) 10/L (median values), based on the R/V Aranda observations from April to August in the region of 59-62°N including only cases with a fetch longer than 30 km.

agree well with those of *Henning* (1986) for the Bothnian Sea and the Gulf of Finland in April to August. The above-mentioned studies were all based on the bulk-aerodynamic method. *Omstedt et al.* (1997) estimated evaporation using an ocean model. They give the monthly values for the period 1981-1994 for the Bothnian Bay and East Gotland Basin (his Figure 3), which does not allow exact comparisons with our results. In general the agreement looks good, but in July their latent heat fluxes are only approximately 15 W m^{-2} for both basins, while in our results most of the July values are higher (Fig. 8) and the mean is 38 W m^{-2}. In July the sea surface temperature is, however, some higher in our study region than in the Bothnian Bay. *Omstedt et al.* (1997) stressed the large interannual and seasonal variations, which is also true in our data set: the mean LE for the period from April to August had its minimum of 8 W m^{-2} in 1994 and a maximum of 46 W m^{-2} in 1995.

We made some calculations to understand how large the turbulent fluxes can be in extreme conditions over the Baltic Sea. We applied the lighthouse station data to calculate the winter fluxes assuming the surface temperature at the freezing point of -0.3°C. The results accordingly represent fluxes over leads within the sea-ice zone. The maximum sensible heat flux was 520 W m^{-2}, and the simultaneous latent heat flux was 240 W m^{-2}. These occurred over leads at Kemi 1 (some leads were open even in this region) on 25 January, 1999, in conditions of Ta = -17.2°C, V = 18.0 m/s, and RH = 64%. Also the maximum sum of H and LE (760 W m^{-2}) occurred in this case. The maximum LEs do not, however, occur in winter but in autumn. Due to lack of Ts data, we can give reasonable estimates for the fluxes based on the lighthouse station data only for the sea-ice season, but the fluxes calculated on the basis of the Aranda data revealed a case of LE = 310 W m^{-2}. It occurred close to the Latvian coast on 14 November, 1994, in conditions of Ts = 8.3°C, Ta = -0.7°C, V = 17.9 m/s, and RH = 64%. Due to our limited data sets, these numbers do not represent any absolute maximums that can occur over the Baltic Sea, but they give an estimate for the maximum magnitudes.

4.2 Error estimates

Errors in the sensible and latent heat fluxes were estimated on the basis of the measurement inaccuracy of the surface and air temperature, relative humidity and wind speed. The inaccuracy of the surface temperature was assumed to be 0.3 K (Section 2.2; cases with a larger error are addressed separately below), and that of the wind speed as 5% (see Fig. 2). The corresponding inaccuracies for the air temperature (0.2 K) and relative humidity (3%) were based on the comparisons with Kalbådagrund data (Section 3.3). In cases of θs > θa, the maximum estimate for H was obtained by adding 0.3 K to θs and 5% to V, and simultaneously decreasing θa by 0.2 K. The minimum estimate was obtained by decreasing θs by 0.3 K and V by 5%, and simultaneously adding 0.2 K to θa. In cases of θa ≥ θs, the process was made similarly with respect to V, but vice versa with respect to θs and θa. The maximum and minimum estimates for LE, Ri, and 10/L were calculated analogously. For example, for LE the calculation was based on the observed qs – qz (Eq. 2) and on the inaccuracy of RH, θs, θa, and V. The errors in H, LE, Ri, and 10/L were calculated on the basis of the maximum and minimum estimates. The absolute and relative errors of hourly values in H, LE, 10/L, and Ri are shown in Table 1.

Table 1. Mean and median values for the absolute and relative errors in the turbulent fluxes and stratification parameters calculated on the basis of R/V Aranda observations.

Hourly values	**H**	**LE**	**10 / L**	**Ri**
Mean absolute error	4 W m^{-2}	8 W m^{-2}	0.08	0.006
Median absolute error	3 W m^{-2}	5 W m^{-2}	0.04	0.004
Mean relative error	60%	70%	100%	60%
Median relative error	50%	30%	50%	50%

We note that the above kind of an error estimate, where the maximum (minimum) estimate is calculated so that the errors in all three or four observation quantities are cumulative to simultaneously increase (decrease) the flux, results to large error margins, which are most likely overestimating the effects of the errors in direct observations. Another aspect is that error sources are not only related to the inaccuracy of the observations. An inaccuracy of approximately 10% can be caused by the selection of the empirical stability functions in Equations (1) and (2) (*Bumke et al.*, 1998). Also the application of the Monin-Obukhov theory itself may cause errors, because the theory is strictly valid only in the constant-flux layer in horizontally-homogeneous and semi-stationary conditions. According to *Geernaert* (2002), the horizontal inhomogeneity is, in general, the most important source of error for the momentum flux, and it can exert significant influence on the flux-profile relationships for fetches up to 5 km from the coast. The influence is less clear for scalar quantities. In our data set the fetch always exceeded 30 km, and therefore we believe that the application of Monin-Obukhov theory did not cause significant errors.

On the basis of the comparisons against Kalbådagrund data and the fact that the wind attack angle with respect to the ship varies, the calculated errors in hourly fluxes (Table 1) can be regarded as random errors. Hence, due to the large number of observations, the random errors in the monthly means are negligible. Systematic errors are, however, possible. Simultaneous systematic errors of 0.2 K in Ta and 0.5 m/s in V would cause a systematic error of 15-20% in the monthly mean sensible heat flux. A systematic error of 3% in RH simultaneously with 0.5 m/s in V would cause a systematic error of 20-25% in the monthly mean latent heat flux.

The calculations above did not, however, account for the individual cases with a large error (exceeding approximately 1 K) in the sea surface temperature (see section 2.2). In such cases, which we estimate to represent only a few percent of the data, the relative errors in H and LE can be very large. Such cases occurred, however, in calm conditions when the absolute magnitude and, accordingly, also the absolute error of H and LE were both small. The calculated results suggested that LE was from air to sea (indicating condensation) in 17% if the data. These cases occurred, however, in conditions typical for a large Ts error, and in many of them the negative bias of Ts may have caused a wrong direction for LE. We cannot therefore rely on the above-mentioned number (17%). It is, however, noteworthy that there occurred also definite cases of a downward LE. The one with the largest magnitude was observed over the Gulf of Finland on 15 June, 1998. Warm and moist air was flowing with a high speed from Estonia onto a cold sea surface (Ta = 16.3°C, RH = 88%, V = 17.8 m/s, and Ts = 11.6°C), and the calculated LE was –90 W m^{-2}. Under such a wind speed, the surface layer of the sea is expected to be well mixed and the error in Ts negligible. If we assume that in cases with V > 10 m/s Ts cannot exceed T(-3 m) by more than 0.5 K, and that Aranda RH observations have a positive bias of 3%, we can calculate a lower limit for the occurrence of a downward LE (by adding 0.5 K to the observed Ts and subtracting 3% from the observed RH). Then we in any case detect 76 cases (hourly means) with a downward LE. These represent 1% of the whole data set from April to August.

5. *Effect of the fetch*

We calculated the wind fetch over the sea for each hourly observation of R/V *Aranda*. In this analysis, all the ship data from the Baltic Sea were included (not only from the restricted periods and regions, as in Sections 3 and 4). Digital coastline data of the Baltic Sea were applied together with the data on wind direction and ship location. The coastline data had a resolution of 2 km for the complex regions (e.g., the south-western and southern coasts and archipelago of Finland) and 5 km for the other regions. The calculation method includes the assumption that the wind direction remains the same for the whole fetch. This can naturally cause some errors, in particular for larger fetches.

Regression analyses of the fetch effect on the air temperature, humidity, wind speed, and the turbulent fluxes on the basis of the whole year-round data set of 21000 hourly observations did not reveal many apparent signals. This was because the fetch effect was usually dominated by the stronger effects of seasonal and synoptic scale variability and south-north gradients. In any case, with a longer fetch the magnitude of the sensible heat flux was usually smaller: for fetches shorter than 30 km, mean |H| was 21 W m^{-2}, while for fetches longer than 100 km mean |H| was only 7 m^{-2}. The latent heat flux depended less on the fetch, and the corresponding numbers were 34 and 28 W m^{-2}. The difference in the mean values is significant in 99.9% confidence level for both H and LE (3225 cases with fetch < 30 km and 8532 cases with fetch > 100 km, standard deviations of H and LE were 30-40 Wm^{-2}).

We selected individual cases for studies of air-mass modification with fetch over the sea. In an ideal case the ship cruises perpendicularly to or from the coast, and the wind direction simultaneously remains stationary with its main component perpendicularly from the coastline. We identified 74 cases, in which the conditions were not far from the ideal ones: the ship cruise track had an angle of 88±22° (mean and standard deviation (std) among the 74 cases) with respect to the coastline. The corresponding angle for the wind direction was 90±30°, and the std of the wind direction was small (the mean and median values of the std were 20° and 11°, respectively).

For these 74 cases we used two further criteria: (1) the correlation coefficient (r) between the studied variable and fetch must be significant in 95% confidence level, and (2) the conditions must be stationary. The latter was studied on the basis of data from the nearest lighthouse or coastal station or, if the case occurred far (>100 km) from the nearest FMI station, the NCEP/NCAR reanalyses (*Kalnay et al.,* 1996). The maximum allowable temporal change during the observation period (which varied form 2 to 25 h) was set to 0.5 K for Ta, 5% for RH, and 1.0 m/s for V. Temporal changes in Ts are usually smaller than in Ta, and the large ones were assumed to be excluded by the criteria for Ta. For calculated quantities that depended on two or more observed variable, the criteria had to be met in each variable. For example, the fetch dependence of sensible heat flux was studied only in cases were the temporal changes were < 0.5 K for Ta and < 1 m/s for V. Finally, the cases with the change with fetch smaller than the temporal change in the particular case were omitted from the analyses. The following number of cases passed the all criteria: 16 for Ta, 10 for RH, 9 for V, 3 for qa, and 2 for H (0 for LE and Ri).

The regression lines for the air-sea interaction variables in the selected cases are shown in Figures 10 and 11. The plots of air temperature and specific humidity were made more readable by subtracting the coastal values, which were based on extrapolation of the regression lines.

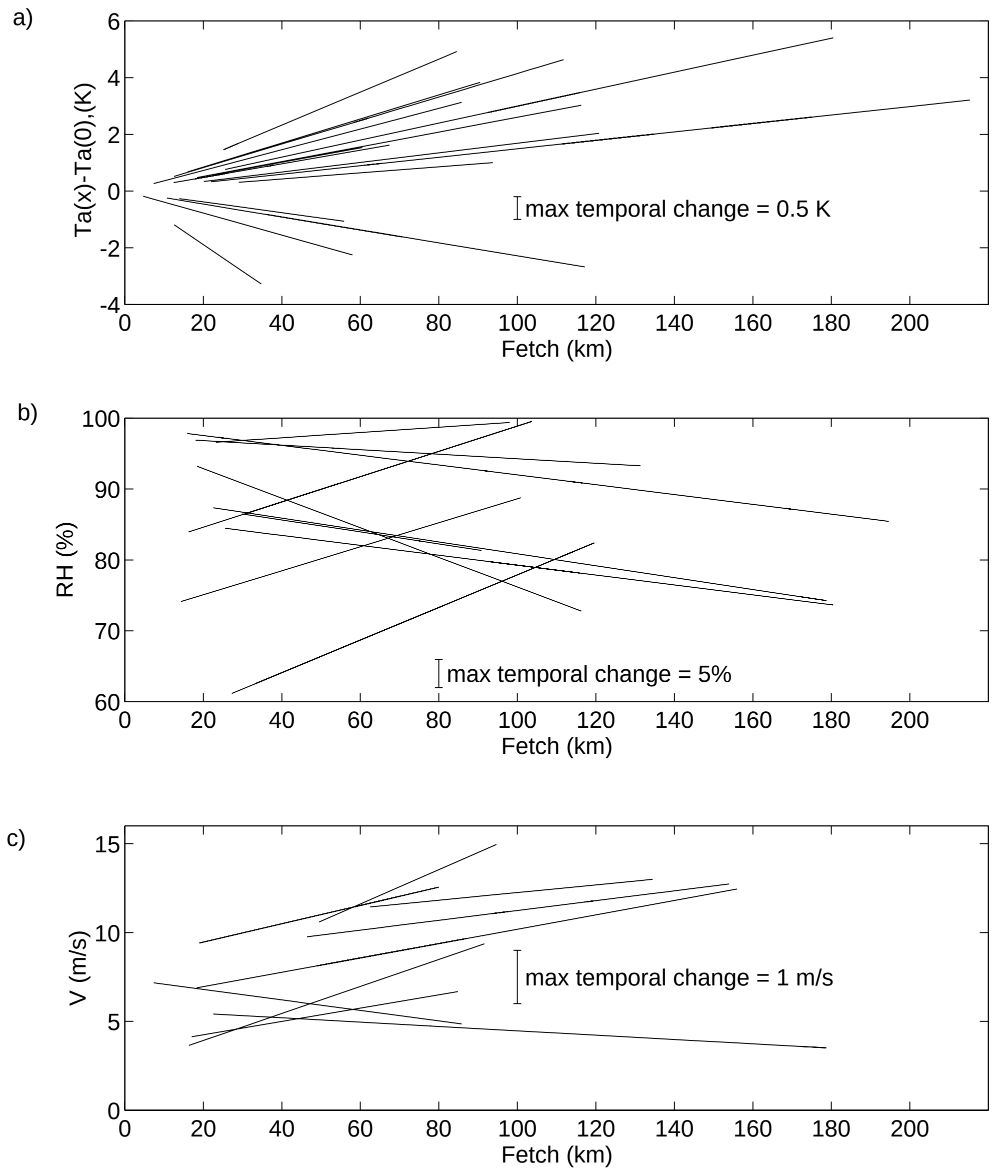


Fig. 10. Regression lines showing the dependence on the fetch in stationary cases for (a) air temperature minus its coastal value, (b) relative humidity, and (c) wind speed.

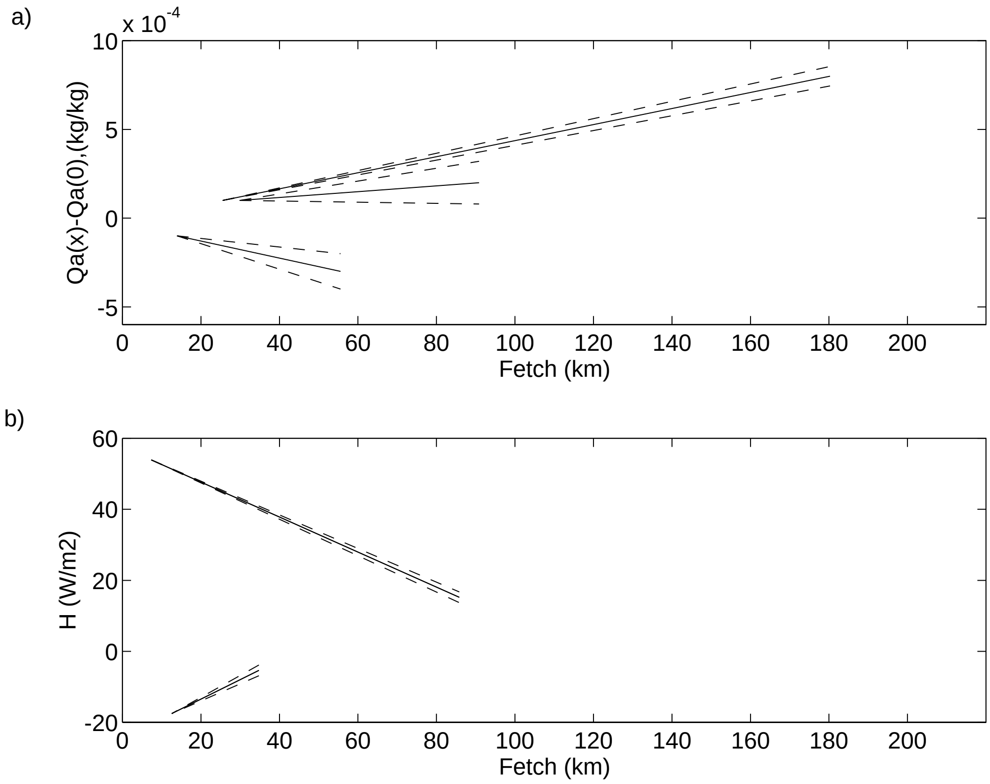


Fig. 11. Regression lines showing the dependence on the fetch in stationary cases for (a) air specific humidity minus its coastal value and (b) sensible heat flux. The dashed lines show the error margins related to temporal changes.

The observed quantities include inaccuracies related to the measurement inaccuracy, and the sensible heat fluxes include inaccuracy related to the error sources discussed in Section 4.2. We think, however, that these inaccuracies do not much affect the observed change with fetch. This is because the measurement conditions, such as the wind direction with respect to the ship, remained constant during the study cases. We therefore calculated the error estimates for the fetch dependence on the basis of the temporal change. For the air temperature, relative humidity, and wind speed, the errors in the fetch dependence are accordingly comparable to the criteria for stationarity: 0.5 K for Ta, 5% for RH, and 1 m/s for V. We see from Figure 10 that the change in fetch always exceeds these limits. For the specific humidity and sensible heat flux, which depend on more than one directly observed quantity, the error estimates had to be calculated separately for each case using the observed temporal changes. The error margins are shown in Figure 11.

We see from Figures 10 and 11 that the slopes of the regression lines vary a lot. The relative humidity shows a variable fetch dependence (Fig. 10b). The increasing cases can be regarded as expected ones, but there are also six cases with RH decreasing with fetch over the open sea. All these cases occurred in a cold season (October–February, Ts ≤ 7°C), but the sea surface was much (6 to 12 K) warmer than the air. Hence the increase of air temperature with fetch dominated over the increase of specific humidity, i.e. the saturation water vapour pressure increased faster than the actual water vapour pressure resulting to a decreasing relative humidity. This mechanism works in cold conditions with H comparable to or larger than LE. In warmer conditions LE usually dominates over H, and RH increases with fetch.

The wind speed typically increases with fetch, which is natural due to the reduced surface roughness compared to the land. There were, however, two cases with decreasing wind speed. The first one, with a longer maximum fetch in Figure 10c, occurred over the Bothnian Sea on 26–27 October, 1992. The wind was only 2 m/s at the coastal stations, and the sea surface was 6 K warmer than the air. We therefore believe that a mesoscale circulation cell may have developed over the sea, and the higher wind speeds at the shorter fetches were related to it. The second case on 21 March, 1994, over the Bothnian Sea in a low-concentration sea ice zone, was a comparable one. The sea surface was 4 K warmer than the air (Ta = -4.1°C and Ts = -0.2°C), and the coastal winds were only ~ 1 m/s.

There is a case of air specific humidity decreasing with fetch (Fig. 11a). It was observed in winter in saturation conditions with the air temperature decreasing with fetch, and therefore we believe that the decrease of specific humidity was related to condensation of fog. The two cases with qa increasing with fetch were observed in October (the one with a longer maximum fetch in Figure 11a) and February over the Bothnian Sea. The sea surface was warmer than the air by 7 K (October case) and 12 K (February, no ice in the study region), and the latent heat fluxes of 70 and 140 W m^{-2}, respectively, resulted to an increase in qa.

For the sensible heat flux, only two cases passed the strict stationarity criteria. In both of them, H approached zero with increasing fetch (Fig. 11b). In the above-mentioned case of 21 March, 1994, H decreased with fetch, while in a case with a stable stratification with Ta = 7.7°C and Ts = 6.2°C (27 May, 1994, over the Bothnian Sea) |H| decreased with fetch.

The slope of the regression equations for the air temperature showed a dependence on the fetch-averaged Richardson number (Fig. 12). The increase of Ta with fetch was strongest in the most unstable conditions, and the decrease was strongest in the most stable conditions. A large |H| with a weak wind is an ideal condition for a Lagrangian change in Ta: a weak wind allows more time for the air mass modification under a certain fetch, and therefore the effect of a certain flux is larger. It should be noted that we calculated the Richardson number on the basis of air temperature and humidity data from the height of 14 m and wind data from the height of 19 m. Using different reference heights, the quantitative relations between Ri and the slope naturally change.

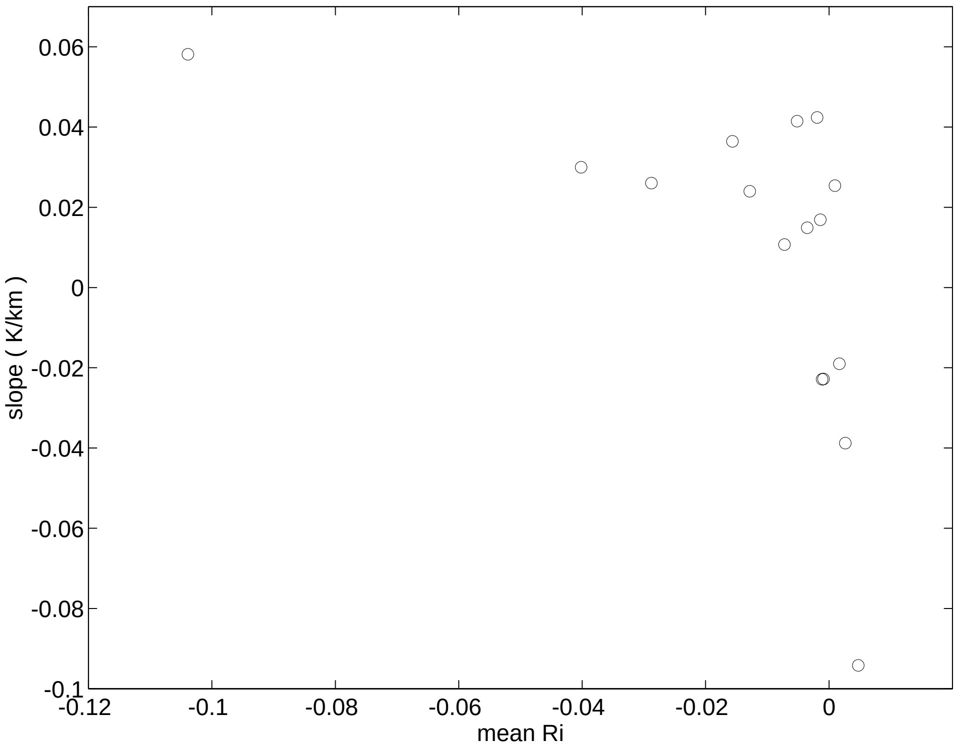


Fig. 12. Slope of the regression line for air temperature as a function of the fetch-averaged Richardson number (referred to the height of 14-19 m).

6. Conclusions

We analysed marine meteorological data from the northern Baltic Sea in the period of 1991 to 1999. The analysis was based on data from three open sea lighthouse stations (Bogskär, Kalbådagrund, and Kemi I) and the R/V Aranda. The ship data were analysed for the period from April to August from the region of 59 to 62°N, only considering cases with the fetch from the coast exceeding 30 km.

Compared to the climatology of the Baltic Sea basin published by *Mietus* (1998), we noticed considerable differences in the air temperature and wind speed. They may be partly related to the warm and windy conditions in 1990s compared to the normal period of 1961-1990. In addition, the data from the FMI stations Bogskär, Kalbådagrund, and Kemi I were not available for the analyses of *Mietus* (1998), which may have affected his results particularly in the data-sparse Gulf of Bothnia. The following finding makes us believe that this is probable. In April and May, the data of *Mietus* (1998) show warmer temperatures than the FMI station data from 1991-1999, the largest differences of 1.5 K occurring at the region of Kemi I. According to *Tuomenvirta et al.* (2000), however, the springs were warmer in 1990s than in 1961-1990.

In our analyses for the period 1991-1999, the monthly mean wind speed was throughout the year stronger than in the 1961-1990 climatology of *Mietus* (1998). The difference was up to 5 m/s. In Kemi I and Kalbådagrund, strong winds and storms were much more common in the period of 1991-1999 than in the period of 1977-1982 analysed by *Launianen and Laurila* (1984) on the basis of data from the same stations. In 1990s, the annual maximum wind speeds occurred a month later than in the previous studies of *Mietus* (1998) and *Launianen and Laurila* (1984).

We calculated the turbulent surface fluxes by the bulk method from the R/V Aranda data. We believe that the monthly means and medians were estimated within an accuracy of ±15-20% for H and ±20-25% for LE. Due to random errors in the observations, the inaccuracy in the hourly values can be larger (Table 1), particularly in cases of light winds and a warm surface layer of the sea. The monthly mean of the surface latent heat flux was upwards (from sea to air) from April to August, but downward fluxes occurred for at least 1% of the time. The monthly mean sensible heat flux was a few W m^{-2} upwards in April, July, and August. In May and June, a stable stratification and a downward sensible heat flux prevailed. Our results for the turbulent surface fluxes had a reasonable agreement with previous studies.

The Baltic marine weather stations showed diurnal cycles for the air temperature that were between those observed over oceans and Fenno-Scandian lakes. The relative humidity, sensible heat flux, and the Bowen ratio had their diurnal maximum values in the morning, while the latent heat flux had it in the afternoon. These were affected by the different diurnal cycles of the air temperature and specific humidity: the former had a larger diurnal amplitude than the sea surface temperature, while the latter had practically no diurnal cycle. The diurnal cycle of the wind speed varied between the three stations. The large differences between the coastal and marine conditions were apparent when comparing the diurnal air temperature cycles at Haparanda (*Tuomenvirta et al.*, 2000) and Kemi I: in winter the mean diurnal range exceeded 8 K at Haparanda, but was less than 0.6 K at Kemi I.

We studied the R/V Aranda cruise track and selected cases, in which the fetch effect on the marine meteorological quantities could be reliably distinguished. Depending on the meteorological conditions, the air temperature, specific and relative humidity and sensible heat flux either increased or decreased with fetch. The decrease of wind speed could be explained by probable mesoscale circulation systems, and the decrease of specific humidity by condensation to fog. The decrease of relative humidity with fetch was due to a dominating increase of air temperature compared to an increase of specific humidity. The changes in the sensible heat flux were related to an air-mass modification towards a neutral stratification. The change of Ta with fetch depended on the Richardson number, the increase (decrease) being strongest in the most unstable (stable) conditions.

Bumke et al. (1998) paid attention to the fact that the reduced wind speed reduces the latent heat flux in the coastal zone. Our analyses suggest that during off-shore winds the larger (qs – qa) in the coastal zone is usually more important than the reduced wind speed. Although we did not find good individual cases to study the fetch dependence of LE, the statistics of all observations indicated ~20% larger mean value for the cases with fetch from the coast less than 30 km than for the cases with fetch exceeding 100 km. This topic requires further studies during on-shore and shore-parallel winds.

Although we showed reasonably strong linear dependencies between the fetch and several quantities, we cannot assume that the dependence is always linear. We had few data for fetches smaller than 10 km, and a non-linear dependence is expected close to the coast. Finally we note that in many cases the modification of the air-mass properties extended to fetches of the order of 100 km. This suggests that in a large part of the Baltic Sea the atmospheric surface layer is not in balance with the local sea surface. The very variable and often strong fetch dependence of various quantities, above all the wind speed and air temperature, indicates that a use of coastal observations or coarse resolution model data as atmospheric forcing for marine models can lead to significant errors.

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