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Vertical coarse aerosol fluxes in the atmospheric surface layer over the North Polar Waters of the Atlantic

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[1] Sea spray emission fluxes were calculated on the basis of vertical gradients of the aerosol concentration data collected with a laser particle counter during four polar scientific cruises carried out in the summers of 2000, 2001, 2002, and 2003. The calculated flux values are the first obtained with a gradient method from data measured on the open ocean; such a method has a much better physical basis than any involving calculating fluxes from concentrations measured at one level only. The sea spray generation function we obtained was compared with other functions from the literature. We show that most literature sea spray generation functions are underestimated for the 1–8 μm radius range, some of them by a factor of 6.

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1. Introduction

[2] Measurements of marine aerosol production and transport are important for many branches of earth sciences such as cloud physics, atmospheric optics, environmental pollution studies and interaction between ocean and atmosphere. Without precise determination of the flux of drop emission from the sea surface, it is not possible to decide whether sea spray affects the air-sea fluxes of heat and latent heat. This problem has been the subject of many studies [*Bortkovskii*, 1972, 1987; *Borisenkov*, 1973; *Petelski*, 1986, 1996; *Mestayer et al.*, 1989; *Andreas*, 1992; *Andreas et al.*, 1995; *Edson et al.*, 1996].

[3] Mechanisms of sea salt particle transport from the sea surface to the atmosphere are well understood thanks to the efforts of many researchers, among others: *Blanchard* [1954, 1989] and *Cipriano et al.* [1983]. Drop emission processes from whitecaps and sea foam are quite well known thanks to laboratory studies of, e.g., *Wu* [1973], *Monahan* [1982], and *Monahan et al.* [1982]. However, the problem of parameterization of aerosol emission from the sea surface has not been solved yet.

[4] Sea spray generation function (SGF) is used for this parameterization. The function quantifies the rate at which sea spray droplets of initial radius r_0 are produced at the sea surface: for example as a function of wind speed. *Andreas* [1998] reviewed the published sea spray generation function models. His conclusion was that the authors of models do not agree even on the order of magnitude of aerosol generation flux from the sea surface. *Andreas* [2002] chose some of the previously published functions as the most reliable. The choice was performed using physical arguments but without direct measurements of aerosol fluxes in the marine atmospheric boundary layer. No one can be sure how close the actual source function would be to the chosen prototypes. However, direct measurements of aerosol emission fluxes from the sea surface are virtually impossible. Therefore, emission values are estimated from aerosol vertical fluxes in the near-water atmospheric layer. The most direct way one may perform such a measurement is using the eddy-covariance method [*Nilsson et al.*, 2001; *Nilsson and Rannik*, 2001]. Unfortunately, it can be used only to determine fluxes of the smallest aerosol particles, as it requires very fast measurements of instantaneous concentration values, which are possible only with average concentrations large enough to provide good statistics. Much lower average concentration values are required when fluxes are calculated from measured vertical gradients of aerosol concentration [*Petelski*, 2003] which makes it a method suitable for determining coarse aerosol fluxes.

[5] So far, SGF have been determined using two methods. Firstly, field studies of the relationship between whitecap coverage and wind speed have been coupled with measurements of the particulate productivity of a given whitecap [*Monahan et al.*, 1986]. Secondly, a variety of simple boundary layer models have been used to derive the generation function from field measurements of aerosol particle concentration [e.g., *Smith et al.*, 1993; *De Leeuw et al.*, 2000; *Reid et al.*, 2001]. In both these methods, vertical fluxes are determined only indirectly. The former uses laboratory measurement results which are difficult to utilize for open ocean processes without first defining an appropriate similarity scale. The latter usually uses a so-called 'dry deposition method' to determine SGF from aerosol concentration values assuming balance between deposition and emission fluxes. The inherent error of the method results from the fact that aerosol concentration near the sea surface is not solely determined by vertical aerosol fluxes. An advection component plays a very important role in creating aerosol concentration. Even under the conditions of horizontal homogeneity in the studied layer, as well as

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stationary process conditions, the aerosol vertical fluxes can be balanced only with an accuracy determined to within the addition of an arbitrary constant value (the transport equation provides the balance of flux gradients, not the fluxes themselves). One should remember that the spatial scales at which a given area may be seen as horizontally homogeneous for small aerosol particles are much bigger than synoptic scales for the wind field [*Petelski*, 2005].

[6] In this study, in order to estimate aerosol emission, we calculate vertical aerosol fluxes from vertical concentration gradients. The method is based on Monin-Obukhov theory which makes it possible to make a better approximation of aerosol turbulent fluxes than the dry deposition method. As far as we know, this study is the first using a gradient method for data measured on the open ocean. It is obvious that such a method has much better physical basis than any involving calculating fluxes from concentrations measured at one level only. We also compare the results obtained using this method with results from other studies as well as with SGF values calculated using the Monahan method [*Monahan*, 1988] and dry deposition methods using data collected in the same ocean area and the same season but in different years (gradient data from years 2000, 2001 2003 and 2003, one level aerosol concentrations from years 1995, 1997, 1998, 2000 and 2001, whitecap data from years 1998, 1999 and 2000).

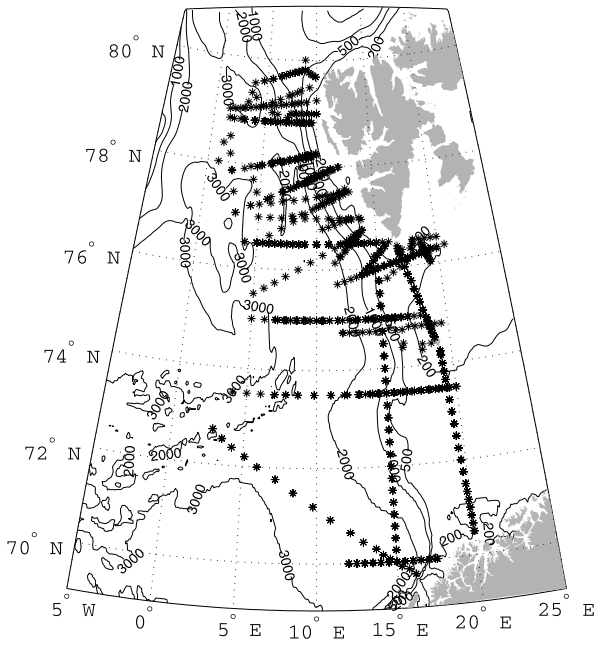

Figure 1. Locations of hydrographic measurement stations during cruises of the R/V *Oceania* in the Norwegian and Greenland Seas (with bathymetry data in meters) in summer periods in 1995–2003.

2. Area and Method of Measurements

[7] Data we used as input for aerosol gradient method were collected during cruises of R/V *Oceania* to Arctic Seas (Norwegian Sea, Greenland Sea, and Barents Sea) in 2000, 2001, 2002 and 2003. The aerosol concentrations were measured by a laser particle counter CSASP-100-HV (deployed on the mast of R/V *Oceania* in a way which allowed the instrument to be moved vertically). The instrument, made by PMS, is an optical spectrometer of aerosol using He-Ne laser.

[8] The measurements were taken at 5 levels: 8, 11, 14, 17, and 20 m above sea level. A single measurement at each level lasted 2 minutes. Every measuring cycle consisted of at least 4 measurements at each level. The instrument detected particles with diameters from 0.5 to 36 μm. The instrument bin width equals 1 μm. The counter inlet was always placed horizontally facing the wind. During the measurements, we used only the original inlet, installed and tested by the manufacturer. The 18 cm long inlet is a convex tube with 2.3:1 acceleration of the airflow. All measurements were conducted at a distance from the mast equal to at least 10 times its radius, upwind from the ship itself, directly over the sea surface (outside the actual ship hull outline).

[9] The measurements were made on board R/V *Oceania*, while the ship was drifting at hydrological stations. The cruise routes are shown in Figure 1. Measurements were performed only during favorable meteorological conditions: while no precipitation or fog was observed, relative humidity was less than 95% and wind speed was greater than 5.5 m/s.

[10] Air temperature (T_p) and humidity were determined from measurements of wet- and dry-bulb air temperatures with a sling (Assman) psychrometer. The recordings were made at the same elevation above sea level as the aerosol measurements. Figure 2a presents a histogram of relative humidity, while wind speed U_{10} values are presented in Figure 2b and a histogram of surface water temperature is presented in Figure 2c.

[11] The wind speed at 10m a.s.l. (U_{10}) was measured with a cup anemometer. Ship's course, speed, and position from the GPS system were also recorded. Measurements were carried out during summer periods: June, July and August. At this time of the year, storms are rare in this region. Hence the results presented below were collected at winds velocities ranging from 5 to 12 m/s only.

3. Vertical Aerosol Concentration Gradients

[12] Sea spray emitted from the sea surface is transported upwards by means of turbulent diffusion, once the particles have reached their immediate injection heights of 180mm or less [*Blanchard*, 1963]. In the near-surface atmospheric layer, the diffusion can be described using Monin-Obukhov layer self-similarity theory [*Petelski*, 2003]. According to the theory, the conditions of neutral stability of this layer are met, if only the following condition is met:

$$z/L \to 0 \tag{1}$$

where z is the altitude over the sea, $L = \frac{-u_*^3}{\kappa\beta Q}$ is the Monin-Obukhov length scale, u_* is the friction velocity, Q is the sensible heat flux, $\beta = g/T$ is the buoyancy parameter, $u_* = (\tau/\rho)^{1/2}$ is the friction velocity, τ is the momentum flux, and κ is the Karman constant. Then the vertical profile of marigenic aerosol concentration should have a logarithmic functional shape. Because the value of z/L is close to zero, not only for very large L values but also for very small

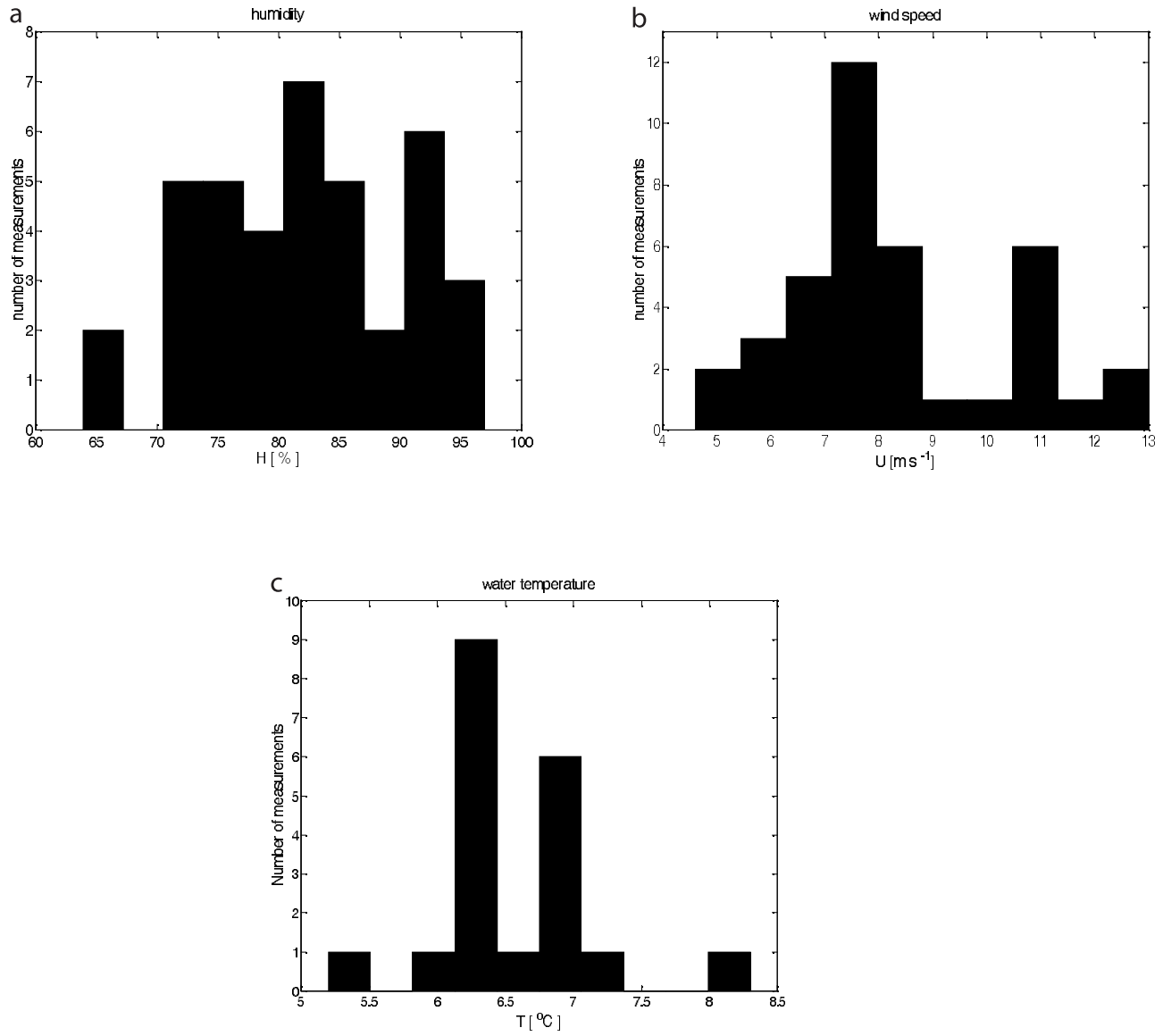

Figure 2. Histograms of (a) humidity at 8 m a.s.l [%], (b) wind speed at 10 m a.s.l. U_{10} [ms^{-1}], and (c) water temperature [°C] measured during the experiments. The bar heights represent data numbers.

altitudes z, in most cases, the following formula can be used for atmospheric surface layer over sea:

$$N(z) = N_* \ln(z) + C \quad (2)$$

Most profiles of marine-derived aerosol concentration profiles could be described using formula (2). Figures 3 and 4 show examples of measured aerosol vertical profiles and best fit functions derived using the formula, respectively.

[13] Measurements of aerosol vertical profiles were conducted only when whitecapping was present. Together, in 4 cruises, we conducted 61 profile measurement sets. Nondimensional stability parameter z/L estimated for those data sets ranged from −0.1 to + 0.01 which is close to neutral stability conditions described by *Zhilitinkievitch et al.* [1978] as $-0.07 \leq z/L \leq 0$. Exact determination of z/L values without direct measurements of heat and momentum fluxes is not possible. Therefore, we used a simpler criterion for applicability of formula (2), namely a linear correlation coefficient between measured functions $N(z)$ and $\ln(z)$. The highest correlation values between $N(z)$ and $\ln(z)$ were obtained for the smallest particle size bins. For big particles of radius r > 5 μm, the correlation coefficient decreased. This is due to poor number statistics for large particles. It can be clearly seen in Figures 3 and 4 that in the case of small particles (upper plots) the measurement points are scattered closer to the best fit line than for large particles (lower plots). The reason for this is the rapidly increasing relative error for increasing particle sizes caused by decreasing number of particles counted. Typical size distribution of aerosol particles in the area studied can be described as $\exp(-0.5r)$ [*Petelski*, 2005]. This size distribution implies that for 8 minutes collection time, we have only 1% relative error for 1 μm particle radius and as much as 20% for 10 μm. As the concentration decreases rapidly with

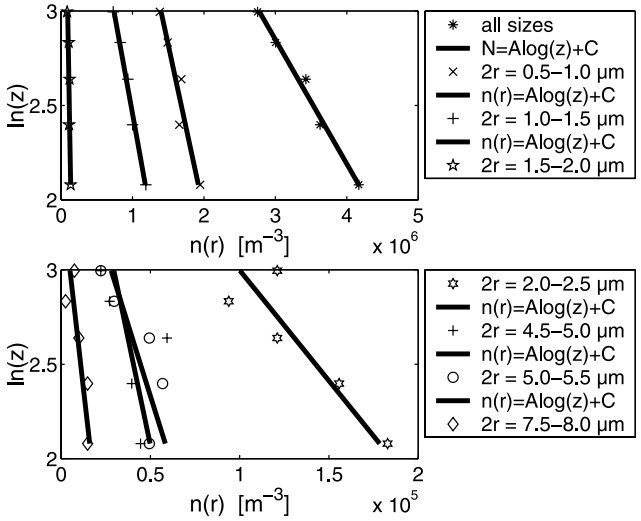

Figure 3. Typical example of measured aerosol concentration vertical profiles. The markers represents measured concentration values. The lines are extrapolated functions. Each line and kind of marker represent concentrations of particles of different size.

increasing particle radius, longer measurement times are required for bigger particles to obtain concentration values with reasonable relative error values. We used only one particle counter and therefore the acquisition time had to be relatively short and thus even though we measured particles from a wide size range, reliable results were obtained usually only for a much narrower radius range $0.25 < r < 7$ μm. As shown by one of us [*Petelski*, 2003], for all particles in this size range the dominant factor controlling the vertical gradient in near-water atmospheric layer is turbulent diffusion. Therefore, a logarithmic vertical gradient for small particles implies the same for larger ones, belonging to this size range. The increasing scatter of points (statistical error) with increasing particle radius in Figures 3 and 4 is caused directly by the smaller number of larger particles collected. Because we wanted to calculate the fluxes for as wide size range as possible, we decided to use also data of relatively lower quality. Therefore, we assumed a two-sided Pearson 80% confidence in logarithmic data fitting as the condition of data acceptance. For five measurement levels, this implies correlation coefficient greater than 0.67 ($r > 0.67$). There were 39 profiles that met this condition, meaning that over 60% of all measured profiles could be accepted as logarithmic. It is worth noting that for some profiles, the correlation coefficient was as high as $r = 0.99$. It clearly proves that logarithmic profiles can be found in near-water atmospheric layer, a finding confirming our results presented at International Sea Spray Workshop in Skipton, UK, in May 2004.

4. Sea Spray Fluxes

[14] Using all the measured vertical profiles which have logarithmic functional shapes, we calculated the scale N_* and subsequently the vertical aerosol fluxes using the following formula [*Petelski*, 2003]:

$$FN = u_* N_* = N_* \sqrt{C_{10}} U_{10} \tag{3}$$

where U_{10} is the wind velocity at 10m a.s.l. and C_{10} is the drag coefficient.

[15] Figure 5 presents the calculated flux FN as a function of wind speed. At the same time, the figure includes a line representing a function proportional to the third power of wind speed, for comparison. It is evident that the flux functions can be approximated with a function proportional to wind speed cubed. Only the measurements done in year 2000 are different in a significant way. A possible reason is that during the 2000 measurement campaign relative

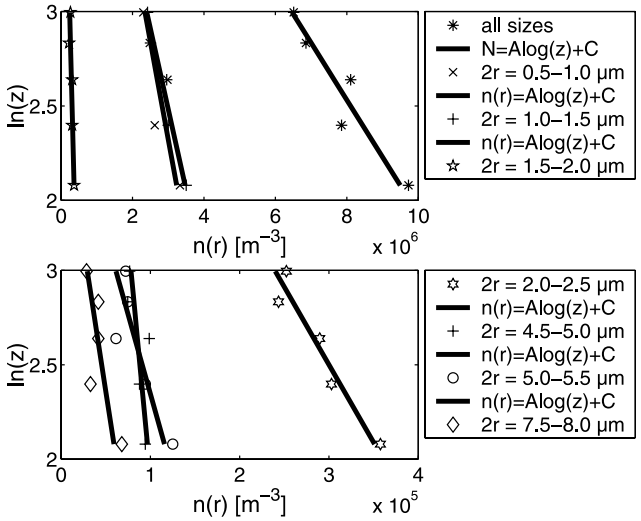

Figure 4. Typical example of measured aerosol concentration vertical profiles. The markers represent measured concentration values. The lines are extrapolated functions. Each line and kind of marker represent concentration of particles of different size.

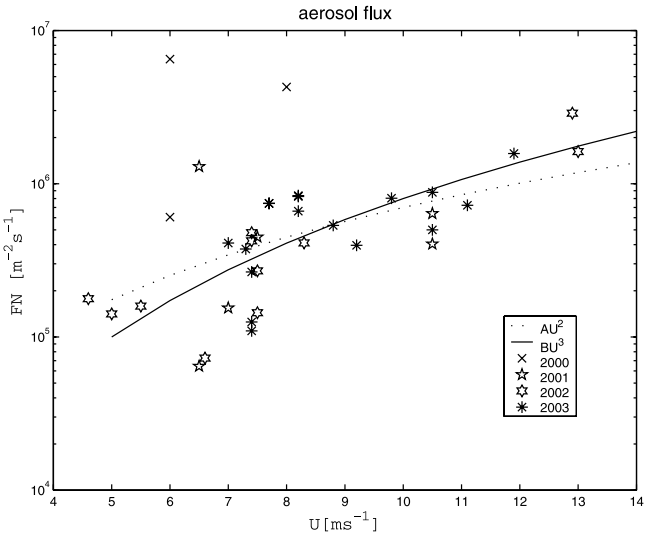

Figure 5. Aerosol flux calculated from vertical aerosol gradients as a function of wind speed. Best fit for U^2 and U^3 functional shapes for better illustration of our data scatter comparing to possible aerosol flux functions.

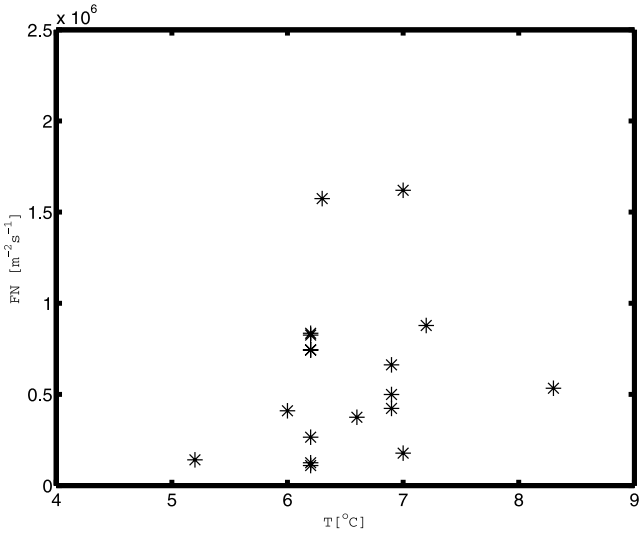

Figure 6. Aerosol flux as function of surface water temperature.

humidity values were generally higher than in other cruises. We calculated the fluxes for ambient conditions. On the other hand, most SGF function used by other authors are recalculated to 80% relative humidity. Therefore, in order to make our functions comparable, we decided to use only the functions that were measured when RH was less than 90%.

[16] *Martensson et al.* [2003], in a laboratory experiment, found a dependence of aerosol emission on water surface temperature. Our data do not show such a relationship (Figure 6), possibly because they were measured in the same ocean area and always in the same season and therefore we did not record much water temperature variability.

5. Sea Spray Generation Function

[17] The general form of the sea spray generation function is a function of two variables: wind speed and radius of the emitted particle. In order to make our results comparable with SGF functions of other authors, we present the calculated flux values as a function of the two parameters. The data set was divided into wind speed classes, 1 m/s wide, to determine the functional dependence on particle radius for each of them. All flux values calculated for measurements done with wind speed between u − 0.5 and u + 0.5 [m/s] were prescribed to wind speed u (midpoint of the class bin). In Figure 10, we present results for classes 7, 10 and 13 m/s. For each class, we looked for a function FN(r) which was the best fit of calculated flux values. It turned out that the best approximation can be made using a linear function on a ln(*FN*) vs. *r* plane. It means that for all wind speed classes, we could find a function with the following form:

$$\log(FN) = -ar + b \tag{4}$$

where *r* is the particle radius and *FN* is the flux of particles of r in range $r \pm dr$; $dr = 0.5$ μm.

[18] We tried several other function formulas but we could not get any significant improvement of the approximation quality. Of course, this does not mean that the SGF cannot be approximated better with a function different from (4). However, our data do not support the idea that any other candidate function is better than the one we decided to use.

[19] Figure 7 presents coefficients *a* and *b* from (4) as functions of wind speed. Both increase with increasing wind speed. The increase is more pronounced for *b* (take note that the vertical axis in *b* plot is logarithmic). Using the simplest linear best fit for coefficients *a* and *b* from equation (4), we obtained the following source function formula:

$$\frac{dF}{dr} = \exp(-(0.05U_{10} + 0.64)r + 7.17 + 0.52U_{10}) \tag{5}$$

presented at the solid lines in Figure 7.

[20] The most typical form of a SGF is a product of two functions, one dependent on wind speed and the other on particle size. However our data suggest that not only is the emission value is dependent on wind speed but also the size distribution itself. The changes in *FN* function slope are visible in Figure 8 which presents flux as a function of particle radius for 7, 10 and 13 m/s wind speed bins. The left plot of Figure 8 presents linear best fit functions while the right one presents polynomial (cubic) best fit functions. One can easily see that for both approximation methods, *FN* functional dependence on particle radius is different for different wind speeds. One possible reason for the increase in aerosol flux for winds stronger than 9 m/s could be an additional aerosol source process. *Monahan et al.* [1986] suggest that at above 9 m/s droplets from the direct disruption of wave crests ("Spume Drop" production) became an important contribution to the sea surface droplet flux.

[21] Figures 9 and 10 present a comparison of the flux values calculated from our data, as well as the approximation functions *FN*(*r*) based on them, with flux values obtained by other authors. Star symbols denote values calculated from our data, broad solid lines denote the best fit function, lines with circle denote the Smith [*Smith et al.*, 1993] function, lines with six-point stars denote the *Andreas* [1998] function and lines with squares denote the Andreas

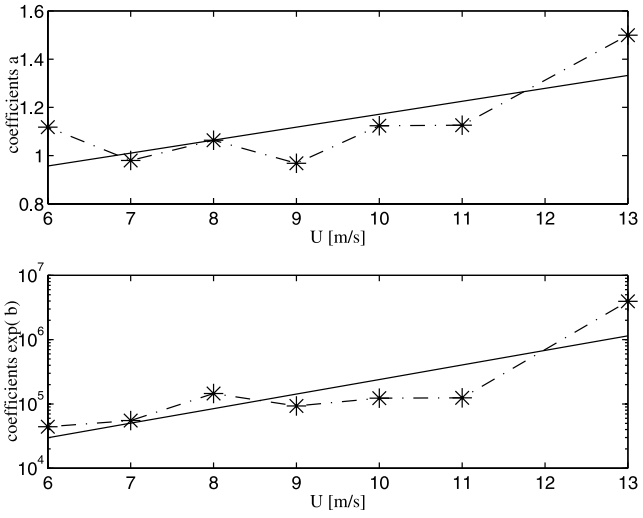

Figure 7. Coefficients *a* and exp(*b*) of equation (4) calculated from the flux approximation as a function of wind speed.

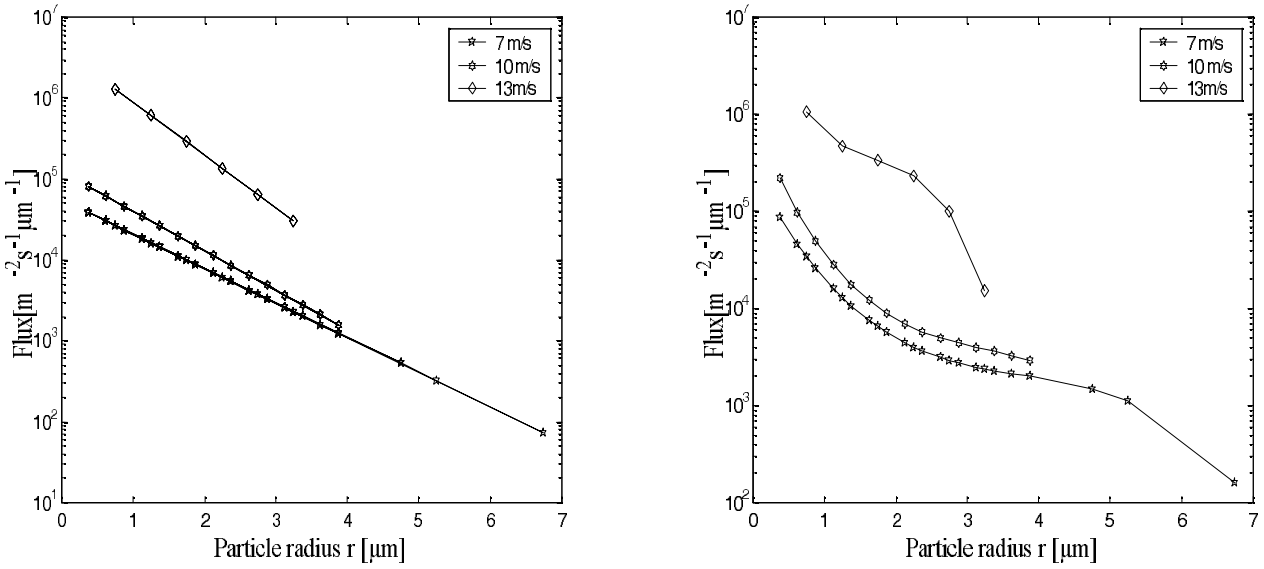

Figure 8. Aerosol flux as a function of particle radius for 7, 10, and 13 m/s wind speed bins. (left) Linear best fit functions and (right) cubic best fit functions.

functions multiplied by factor 6. Additionally, the figures present functions derived by us using two methods of flux calculations derived by other authors. The first method is based on a simple balance equation based on an assumption of horizontal uniformity and stationary conditions:

$$\frac{dF}{dr} = V_d \frac{dN}{dr} \tag{6}$$

where $\frac{dF}{dr}$ is the emission source function (the number of particles of radii from r to $r + dr$, emitted in a unit time from a unit sea surface area), V_d is the deposition velocity, and r is the particle radius. The second method was proposed by *Monahan* [1988] and it involves an estimation of the emission using the whitecap coverage W_C as the main parameter [*Monahan et al.*, 1986; *Monahan*, 1988]. He derived aerosol production from whitecaps in laboratory experiments multiplying the W_c parameter by the whitecap area, which yielded the source function.

[22] *Monahan* [1988] applied the following formula:

$$\frac{\partial F}{\partial r} = (\dot{W}) \frac{\partial E}{\partial r} \tag{7}$$

where $(\dot{W}) = W_C \tau^{-1}$ is the fraction of the sea surface from which whitecap bubble cluster disappear per second, $\frac{\partial E}{\partial r}$ is the number of particles per micrometer radius increment, produced during the decay of a unit area of whitecap, and τ is the time constant that defines the exponential rate of decay of the area of an individual whitecap. *Monahan* [1988] reported that a typical value of τ is 3.5 seconds. The most recent version of this function reported by *Monahan* [1988] is

$$\frac{\partial E}{\partial r} = A_0^{-1} * 4.40 * 10^5 * r^{-3} \left(1 + 0.057 * r^{1.05}\right) * 10^{1.19 \exp\left(-B^2\right)} \tag{8}$$

where $B = \frac{(0.380 - \log r)}{0.650}$.

[23] Using the deposition-production approach formula (6), the authors constructed a source function using the following deposition velocity V_d formula derived by Carruthers and Choularton [after *Smith et al.*, 1993]:

$$V_d = \frac{V_t}{1 - \exp\left\{-\left(\frac{V_t}{C_D U}\right)\right\}} \tag{9}$$

where V_t is a gravitational sedimentation velocity for given particle size, C_D is the drag coefficient, and U is the wind velocity at the measurement height. The function was obtained using the approximated size distribution, which was described in a previous paper [*Petelski*, 2005]. In that paper, the following formula was derived for coarse aerosol size distributions in the near water layer:

$$n(r) = \exp(0.21 U_{10} + 12.3 \pm 2) * \exp(-0.58 r) \tag{10}$$

Formulas (6), (9), and (10) lead to the following result:

$$\begin{aligned} \frac{\partial F}{\partial r} &= V_d n(r) \\ &= \frac{V_t}{1 - \exp\left\{-\left(\frac{V_t}{C_{10} U_{10}}\right)\right\}} \exp(0.21 U_{10} + 12.3 \pm 2) \\ &\quad * \exp(-0.29 * 2r) \end{aligned} \tag{11}$$

This function is denoted in Figures 9 and 10 as a line with rhombs.

[24] We show in Figures 9 and 10 functions based on the *Monahan* [1988] method with data collected in the same geographical region and the same season but in different years than data used above. We used the whitecap coverage dependence on wind speed appropriate for fully developed waves obtained by *Stramska and Petelski* [2003]

$$W_c = 5.0 * 10^{-5} * (U_{10} - 4.47)^3 \tag{12}$$

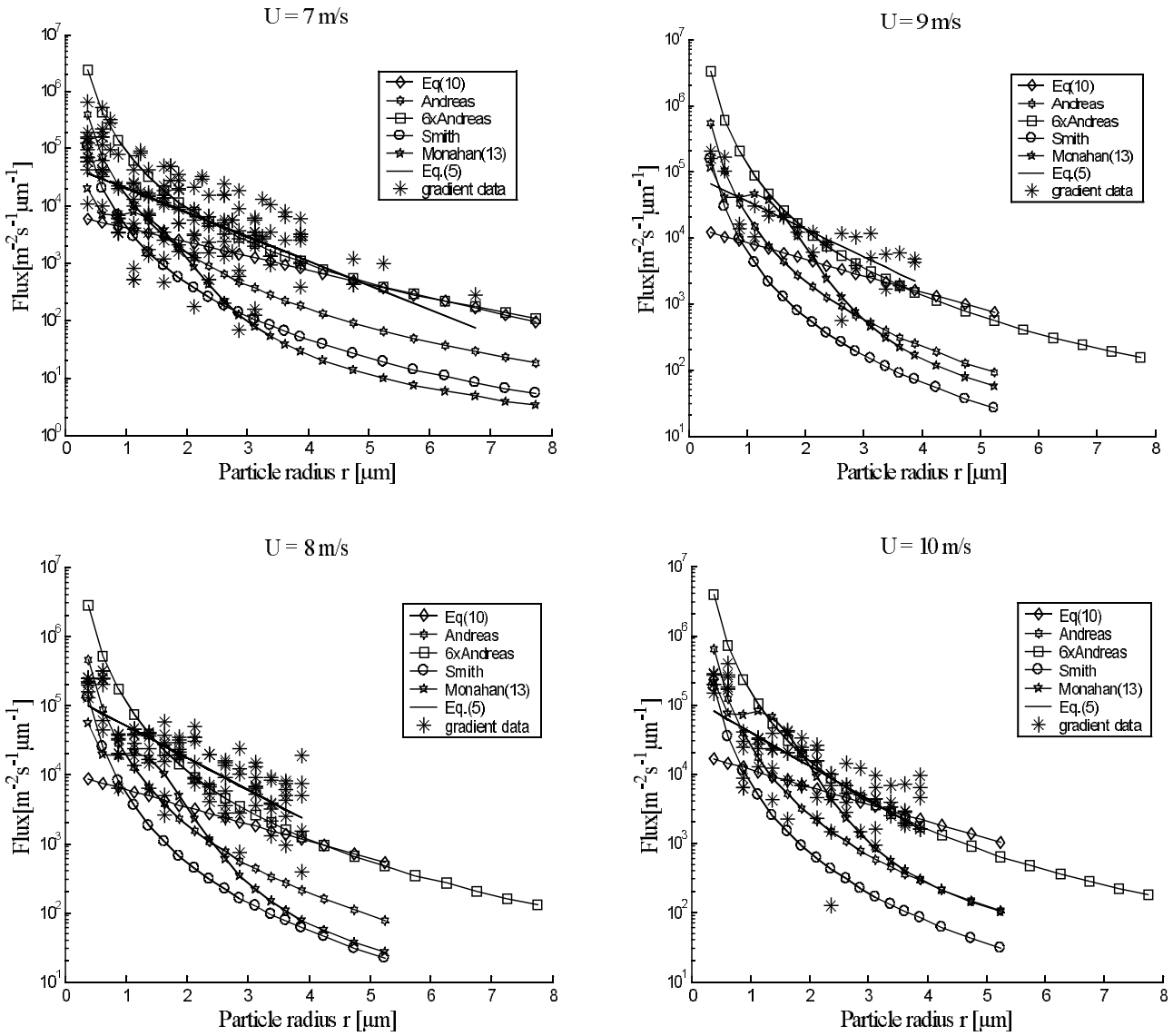

Figure 9. Sea spray generation functions versus particle radius for 7, 8, 9, and 10 m/s wind speed bins. The lines represent SGF, and large stars represent calculated values from measured values of aerosol fluxes. See text for details.

and $\frac{\partial E}{\partial r}$ (8) by *Monahan* [1988]. This yielded the following:

$$\frac{\partial F}{\partial r} = A_0^{-1} * 4.40 * 10^5 * r^{-3}\left(1 + 0.057 * r^{1.05}\right) * 10^{1.19\exp\left(-B^2\right)} * 5 * (U_{10} - 4.47)^3 \quad (13)$$

where $B = \frac{(0.380 - \log r)}{0.650}$.

[25] Function (13) is denoted on Figures 9 and 10 as a line with five-pointed stars. Figure 11 presents a comparison of the best fit (4) function with the functions shown in previous Figures 9 and 10 as well as with functions of *Gong* [2003] and *Martensson et al.* [2003]. Both the latter functions were constructed for submicron aerosols but their domains cover at least part of our particle size range. In Figure 11, for clarity, we do not show measurement values but only best fit functions.

[26] Figures 9, 10, and 11 show that all the literature functions underestimate our experimental data for most particle sizes. Only for radii smaller than 1 μm, does our experimental data overlap with the functions. This statement is true for all wind speeds we have data for. The exception is wind speed 13 m/s (Figure 10, left) where all experimental data are above functions by other authors. We managed to conduct only two measurement series for winds in this speed bin. Unfortunately, we have no measurements for 12 m/s or for winds stronger than 13 m/s. Most of the measurements were done with winds in the range of 6–11 m/s. In this range, the pattern for all wind speed values is similar. Namely, the original Andreas and Smith functions have smaller values than our experimental data, coinciding with the data "cloud" only for the smallest particle sizes. On the other hand, the Andreas function multiplied by a factor of 6 runs through our data space and overshoots our results only for the smallest particles. The original function

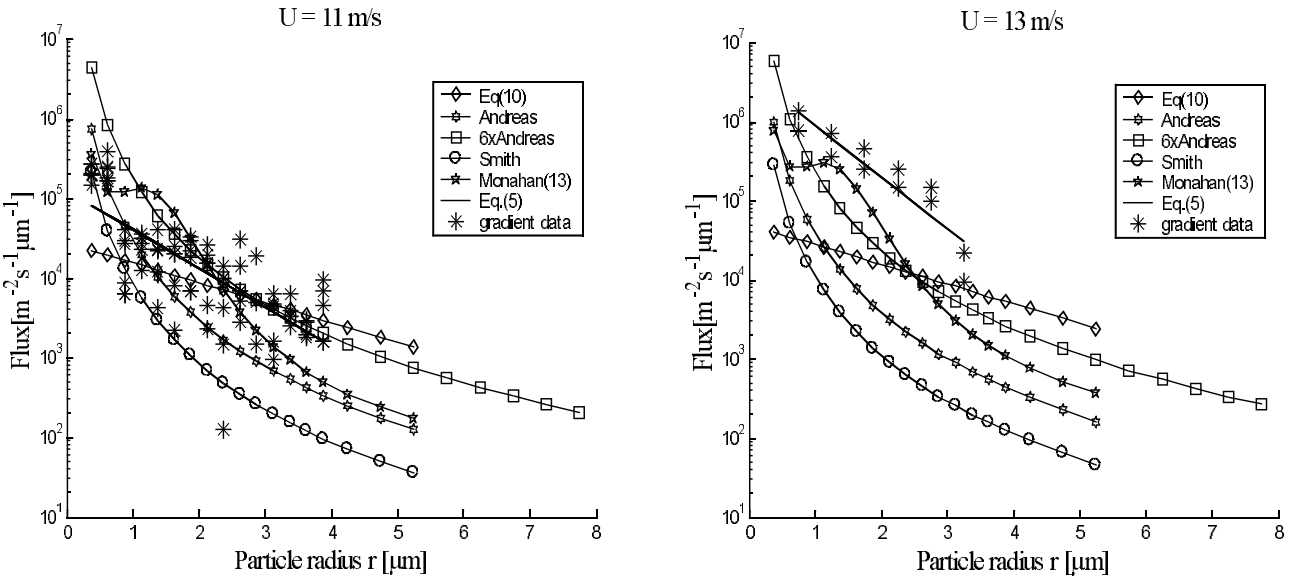

Figure 10. Sea spray generation functions versus particle radius for 11 and 13 m/s wind speed bins. The lines represent SGF, and large stars represent calculated values from measured values of aerosol fluxes. See text for details.

by *Monahan et al.* [1986] (in Figure 11 denoted as a line with triangles with their bases perpendicular to the line), similarly to functions by Andreas and Smith, enters into the range of our experimental data only for the smallest particles. The function, derived using the *Monahan* [1988] method variant of *Stramska and Petelski* [2003], overlaps with the experimental data space for a much wider range of particle sizes. However, it can be seen in Figures 9 and 10 that the function decreases quickly with increasing particle size. Function (11) (line with rhombs), estimated using dry deposition method with formula (10), has the smallest decrease rate of all the discussed functions. It overlaps with the data space in most of their range but has a lower slope value (is flatter) than the best fit function (line with dots). The functions are closest to each other in the range 3–5 μm for all the wind speeds we have data for. A possible reason for this behavior is that for this very size range, we had the highest correlation coefficients between aerosol concentration in the near-water layer and the wind speed [*Petelski*, 2005]. Formula (10) for aerosol concentration as a function of wind speed which was derived from data collected during many summer seasons should be free from advection. It means that the significant discrepancy between flux values derived from formula (11) and gradient measurements for small particles ($r < 3$ μm) can be attributed to erroneous values of the deposition velocity.

[27] The large scatter of measurement points in Figures 9 and 10 was caused in a large part by the fact that aerosol emission is controlled not only by wind speed but also by other factors, for example wave parameters [*Petelski et al.*, 2005]. The data would have less statistical error if we were able to collect them for a longer time on each measurement level. Unfortunately we were able to measure with one particle counter, only. A large improvement in the data quality would be possible with simultaneous measurements of aerosol gradients and turbulent fluxes, fluxes of sensitive heat and latent heat using, for example, the eddy correlation method. We are fully aware that the bulk formulas we used out of necessity are only a rough approximation.

6. Conclusion

[28] Measurements which were conducted over several summer seasons in the North Polar Waters of the Atlantic showed that logarithmic vertical profiles of aerosol concentration are common for whitecap covered ocean areas. Our calculated flux values are the first obtained with a gradient method (as opposed to methods based on measuring aerosol concentrations at one level only) for data measured over the open ocean.

[29] Aerosol flux values calculated from measured vertical concentration profiles may indicate an underestimation

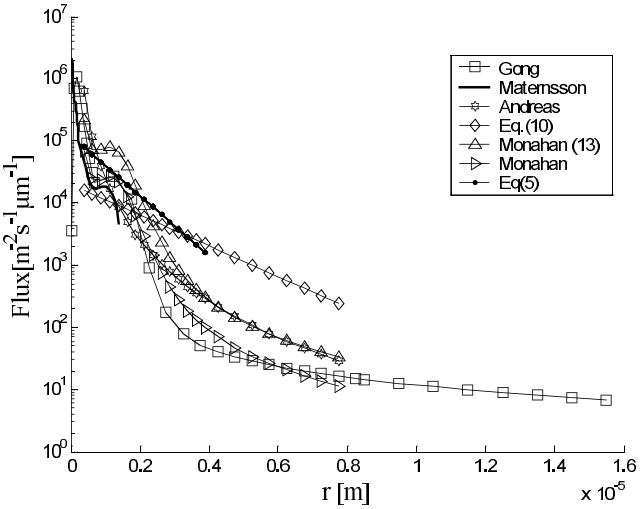

Figure 11. Comparison of different SGF for 10 m/s wind speed. See text for details.

of emission values by literature SGFs for particles with radii between 1 and 8 μm.

[30] The results of this study show also that the assumption used often in the research of marine aerosol fluxes, that the SGF can be divided into two independent parts: the size spectrum and factor dependent on environmental parameters, may not be true. However, we need more measurements for strong winds ($U > 11$ m/s) in order to clarify this topic in a satisfactory manner.

[31] The method of calculating aerosol fluxes from measured vertical concentration gradients, based on Monin-Obukhov scaling theory, gives a better approximation of turbulent aerosol fluxes than the dry deposition method. If used with data acquired from multiple particle counters, located at several altitudes with simultaneous measurement of additional parameters (especially wave parameters as well as momentum and heat turbulent fluxes), this method should make it possible to definitely solve the problem of coarse aerosol mode emission parameterization.

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