Turbulent aerosol fluxes over the Arctic Ocean
1. Dry deposition over sea and pack ice

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Abstract. An eddy-covariance flux measurement system was applied successfully to measure aerosol number dry deposition over open sea, ice floes, and over the leads in between the ice floes during the Arctic Ocean Expedition 1996. The aerosol number dry deposition velocity geometric mean was $1.9 \times 10^{-2} \text{ m s}^{-1}$ for the open sea. The dry deposition velocity to the ice floes averaged approximately $3 \times 10^{-4} \text{ m s}^{-1}$ with no significant difference between the melting summer ice and frozen, partly snow-covered ice in the early freeze-up. Over the leads the dry deposition velocity was $3.4 \times 10^{-4}$ and $9.1 \times 10^{-4} \text{ m s}^{-1}$ for the summer and the freeze-up period, respectively. For neutral and stable stratification the geometric averages were $6.5 \times 10^{-4}$ and $2.9 \times 10^{-4} \text{ m s}^{-1}$, respectively. The dry deposition changed mainly with friction velocity and particle size, as expected from theory, and with static stability, and we are able to recommend an empirical parameterization that works reasonable well over the Arctic Ocean pack ice. The resistance through the quasi-laminar sublayer dominated over the aerodynamic resistance and the ultrafine and Aitken mode particles dominated the dry deposition flux. The average turnover time for the submicrometer aerosol in the boundary layer was only 4.6 days, but for cases when the ultrafine and Aitken modes dominated the aerosol size spectra, the turnover times were 1.3 and 3.5 days, respectively. The small turnover times were largely caused by the shallow boundary layer over the Arctic Ocean pack ice.

1. Introduction

The aim of the interdisciplinary atmospheric program on the Arctic Ocean Expedition 1996 (AOE-96) with the icebreaker Oden was to study the sulfur aerosol and its effect on clouds and regional climate (C. Leck et al., unpublished manuscript, 2000). The data set includes the ice free Barents Sea to deep within the pack ice covered Arctic Ocean at 88°N in July and August 1996 during about 15 days distributed on a 5 day ice camp and several 12-24 hour long stations, see Figure 1. To understand the atmospheric aerosol, it is necessary to study its sources and sinks. For the submicrometer aerosol we would expect the sinks to be coagulation, detrainment, dry deposition, and wet deposition. Given the low aerosol number concentrations in the central Arctic, coagulation can be neglected. The high static stability and small turbulent mixing except near surface result in small turbulent transport sinks of aerosol particles (E. D. Nilsson et al., unpublished manuscript, 2000). Furthermore, the low frequency of rain and snow over the Arctic Ocean in the given season limits the wet deposition sink [Nilsson and Bigg, 1996; Nilsson and Barr, this issue]. Deposition in fogs are probably an efficient sink for the accumulation mode particles [Nilsson and Leck, 1997], but for the particles smaller than 100 nm, the ultrafine and Aitken modes, dry deposition may be the only efficient sink. The aim of this study is to quantify and characterize this sink: the submicrometer aerosol number dry deposition over the central Arctic Ocean in summer during the AOE-96. Our principal tool in this task is the eddy covariance flux measurements performed on two stations in open sea and six of the stations plus the ice camp within the pack ice, see Figure 1.

2. Experimental Methods

We used an eddy covariance (EC) system, radiosondes [E. D. Nilsson et al., unpublished manuscript, 2000], a differential mobility particle sizer (DMPS), and an optical particle counter (OPC) (P. P. Aalto, unpublished manuscript, 2000). We also used a condensation particle counter (CPC) mounted in a helicopter. The DMPS system measured aerosol number size distributions from 5 to 610 nm in dry diameter ($D_p$), see (P. P. Aalto, unpublished manuscript, 2000), and the OPC system measured aerosol number size distributions from 130 nm to 15 μm in dry $D_p$. The DMPS and OPC were sampling at 27 m height from a container on board Oden. During the ice camp the DMPS system sampled air at 4 m height from above a laboratory deployed on the ice floe, see Leck et al. [this issue]. The helicopter CPC was equipped with a diffusion battery with an alternative flow line bypassing the diffusion battery, so that the air could be sampled
either directly through about 1 m of 4 mm stainless steel tube in front of the CPC, or through the diffusion battery. Thus we could measure the gave the aerosol number for $D_p > 20$ and $D_p > 140$ nm respectively (E. D. Nilsson et al., unpublished manuscript, 2000).

Throughout the expedition, the ship speed and heading were logged from the ship's navigation system. Position, speed, and heading were also stored from a special high-accuracy geographical position satellite system (GPS) on Oden for the marine geological drilling program on the AOE-96. In addition, during the ice camp, two GPS receivers were located at a 300 m distance on the ice floe to allow us to follow the motions of the ice floe. These data have been used to calculate the true wind speed and wind direction. The correction in wind speed was small for measurements made within the pack ice because the drifting speed of ice floes (or of Oden moored to an ice floe as during the 12-24 hour stations) was small compared to the wind speed.

2.1. The Eddy-Covariance System

The EC system was either located at about 12.5 m height in a sky lift directed upwind Oden or at 10 m height in a mast on an ice floe, see Leck et al. [this issue] for all logistical details. The system included a TSI 3010 condensation particle counter (CPC) which measured the number concentration ($N$) with $D_p > 10$ nm at 1 Hz. The vertical wind component ($w$) was sampled at 20.8 Hz from a Gill Solent 1012R research model ultrasonic anemometer, which also measures horizontal wind components and temperature. The anemometer was mounted on a boom extending 3 m upwind from the sky lift or at the top of the mast. The aerosol was sampled from below the sensing head of the anemometer (about 15 cm from the center) through an 11 m long copper tube (inner diameter 4 mm) with a flow rate of 1 L min$^{-1}$. The particle penetration curve through the sampling tube for given flow conditions is presented in Figure 2, estimated for horizontal and vertical tube assumptions according to Brockmann [1993]. Diffusional losses of particles larger than 20 nm in diameter were small, and sedimentation of accumulation mode particles was insignificant even for the worst case assumption of a horizontal tube. The optional analogue (0-10 V) output of the TSI 3010 in the 0-10000 cm$^{-3}$ range was used with a simple circuit limiting the voltage to 5 V. This allowed us to feed the signal into the analogue auxiliary input (0-5 V sampled at 10 Hz, converted to 20.8 Hz by a built in interpolation algorithm) of the Solent interface for digitalization together with wind and temperature data, for further
processing and storage on a computer. The resulting range of 0-5000 particles cm\(^{-3}\) for the EC system was never exceeded during the expedition. The CPC, a computer, the Solent interface etc were mounted in an isolated aluminum box with thermostat regulated temperature and heating system located either in the sky lift or at the foot of the 10 m mast. A LICOR 6262 gas analyzer was also included in the system to measure water vapor concentration (and latent heat) with first-order response time of 0.1 s; see Nilsson et al. [this issue (a)] for a more detailed description. The underestimation of particle number fluxes due to tube damping of particle concentration fluctuations [Lenchow and Raupach, 1991] and limited frequency response of the CPC was corrected according to Moore [1986]. The correction technique relies on calculation of flux underestimation by a measurement system, in which components are characterized by frequency transfer functions. In the calculation the model cospectra of atmospheric surface layer turbulence were used. Typical magnitude of corrections were from 20 to 50\%, depending on wind speed and stability conditions.

2.2. The Eddy-Covariance Method and Theory

These fast measurements allowed us to separate the aerosol number concentration \(N\) and the vertical wind velocity \(w\) into mean values \(\bar{N}\) and \(\bar{w}\) (resulting from the mean flow) and turbulent fluctuations \(N'\) and \(w'\)

\[
N = \bar{N} + N' \\
\bar{w} = w + w'
\]  

(1)

where linear detrending was used to extract the mean and fluctuation parts from turbulent records. A three-dimensional co-ordination of wind vector to the local streamlines was performed according to Kaimal and Finnigan [1994]. The \(N\) signal was delayed because of travelling time in the sampling tube by maximizing the fluxes. The vertical turbulent surface layer aerosol number particle fluxes \(N'w'\) were calculated over 30 min periods from the turbulent fluctuations \(N'\) and \(w'\). The uncertainty of the time average flux estimates arises due to two reasons: the stochastic nature of turbulence [see, e.g., Rannik and Vesala, 1999], and discrete counting of aerosol particles [e.g., Fairall, 1984]. These uncertainties are of random nature. The uncertainty due to turbulence is typically of the order of 10\%. The uncertainty due to counting can be estimated as [Fairall, 1984]

\[
\Delta w'N' = \sigma^2 N Q T 
\]

(2)

where \(\sigma\) is the standard deviation of vertical wind speed, \(Q\) is the volumetric flow rate through the counting device, and \(T\) is the averaging time. While the absolute error decreases with decreasing concentration, the error normalized with concentration (which is the measure of the error of dry deposition) increases correspondingly. Therefore the counting noise becomes significant for the low particle number concentrations typically observed in the Arctic. The error due to turbulence will be neglected in the analysis.

Turbulent fluxes of momentum, heat, and water vapor (latent heat) were calculated in a similar way to that of particle fluxes; see Nilsson et al. [this issue (a)] for the evaluation of these data. In total, the anemometer and the CPC were fully operational simultaneously for 175 hours, which gave 350 data points. This is less than for the other turbulent fluxes because the CPC sampling line froze on a few occasions.

2.3. Footprints of the EC System

Model estimations of the footprint of this EC system over the Arctic Ocean, following Leclerc and Thurtell [1990], show that the upwind location of peak contribution to the flux was about 200 m or less. The upwind distance that contributed 80\% of the flux was on the order of 1 km or less. At the ice camp the EC system mast was located on an ice floe about 300 m from the closest part of the ice floe edge. In other wind directions the distance was up to 1 km, see Leck et al. [this issue] for an ice camp overview map. The leads must have contributed to, but not dominated, the flux during the ice camp. During station time, Oden was moored to an ice floe, typically on the port side, the same side as the EC system. As a result, most of the EC system data from the stations represents fluxes over the ice floe, but some of the data are representative of a lead. With help of the log of ship position, speed and heading, photographs, video recordings, and careful documentation of the surroundings on each station the data have been divided into five categories according to the nature of the surface within the footprint of the EC system: (1) open sea, (2) open lead in the pack ice or along the edge of a lead and an ice floe, (3) smooth ice floe surfaces (no ridges), and (4) Rough ice floe surfaces with one or several ridges 0.5 to 3 m high.

3. Theory of Dry Deposition

It is useful to normalize number fluxes by the aerosol number concentration if the flux can be assumed to
depend on \( N \). Such is the case for fluxes caused by diffusion, interception or gravity sedimentation to the surface. The negative \( N'w' \) over the pack ice covered Arctic Ocean corresponds to deposition velocities

\[
v_d = \frac{N'w'}{N}
\]

of typically \( 10^{-5} \) to \( 10^{-2} \) m s\(^{-1}\).

Dry deposition of aerosol particles takes place through: (1) aerodynamic turbulent eddy diffusion transport through the surface layer followed by (2a) Brownian diffusion transport or (2b) interception in a very thin quasi laminar sublayer just above the surface. (3) Sedimentation of particles may also contribute. It is useful to express \( v_d \) in terms of

\[
v_d = \frac{1}{r_a + r_p + r_a r_p v_s + v_s} \]

where \( r_a \) is the aerodynamic resistance in the surface layer, \( r_p \) is the resistance in the quasi-laminar sublayer and \( v_s \) is the terminal settling velocity due to gravitation from Stokes law [see, e.g., Hinds, 1982]. The contribution from \( v_s \) will be small for submicrometer aerosol particles since \( D_p^2 \). The aerodynamic resistance can be expressed in the form [Seinfeld et al., 1998]

\[
r_a = \frac{1}{k_{ua} \left[ \ln \left( \frac{z}{z_0} \right) + 4.7 \left( \zeta - \zeta_0 \right) \right]}
\]

\( \zeta \geq 0 \) \( \text{stable} \)

\[
r_a = \frac{1}{k_{ua} \left[ \ln \left( \frac{z}{z_0} \right) + \ln \left( \frac{(n_0^2 + 1)(n_0 + 1)}{(n^2 + 1)(n + 1)} \right)^2 
+ 2(\tan^{-1} n - \tan^{-1} n_0) \right]}
\]

\( \zeta \leq 0 \) \( \text{unstable} \)

where \( u_* \) is the friction velocity, \( k \) is von Karmans constant (=0.4), \( z \) is the measurement height (12.5 or 15 m in our case), \( z_0 \) is the surface roughness height, \( \zeta = z/L \) and \( \zeta = z_0/L \) are the dimensionless heights for \( z \) and \( z_0 \), respectively, where \( L \) is the Monin-Obukov length [Monin and Obukov, 1954], and \( n = (1-15\zeta)^{1/4} \) and \( n_0 = (1-15\zeta_0)^{1/4} \), respectively.

The resistance through the quasi laminar sublayer \( r_p \) can be expected to depend directly on \( D_p \), \( z_0 \), \( u_* \), diffusion coefficient (\( D \)), and kinematic viscosity (\( v \)). Various expressions are suggested in literature. For example, adapting the expressions given by Schack et al. [1985], we can write for neutral conditions and a completely rough surface

\[
r_p = \left( \frac{D_p}{\nu_0} \right)^2 \left( \frac{1}{z_0} \right)^2 + \left( \frac{BD_p}{\nu_0} \right)^2 \left( \frac{1}{z_0} \right)^2 + \left( \frac{1}{\nu} \right)^2 \right)^{-1}
\]

where the right-side parts of the equation are for the diffusion and interception ranges, respectively and \( A \) and \( B \) are empirical factors depending on the surface type, for example, the collection surface per unit cross-sectional area parallel to the flow. In current study the values \( A = 0.4 \) and \( B = 20 \) are used (see section 4.2 below). The aerodynamic deposition velocity should be indirectly dependent on \( U \), \( \zeta \), and \( z_0 \) through the \( u_* \) dependency since

\[
\left[ \frac{1}{k} \ln \left( \frac{z}{z_0} \right) + 4.7(\zeta - \zeta_0) \right]^{-1} \zeta \leq 0
\]

4. Results and Discussion

The vertical turbulent surface layer aerosol particle number flux \( N'w' \) ranges and several statistics are given in Table 1, separated into different surface types and season: summer or freeze-up. Typical summer conditions are wet, snow-free, and gradually melting ice floes and an increasing amount of open leads [cf. Nilsson and Barr, this issue; Nilsson et al., this issue (a)]. The freeze-up period is characterized by frozen ice surfaces, some new snow and the first formation of new thin ice on the leads. Negative \( N'w' \) is toward the surface and suggests that dry deposition to the surface was dominating, and positive \( N'w' \) suggests that there existed a particle source at the surface of larger magnitude than the dry deposition sink. For all surface types, when positive as well as negative fluxes are included, fluxes are not significantly different from zero, see Table 1. However, \( N'w' \) was not normal distributed. Instead it consisted of two roughly lognormal distributions, one at positive fluxes, the other at negative fluxes, with a minimum at zero. Therefore we will for now on treat the net fluxes dominated by upward and downward fluxes, respectively, separately. The upward as well as the downward fluxes were significantly different from zero, except for the rough ice, which only distinguishes itself by a large \( z_0 \), which does not influence the aerosol fluxes in any obvious way. Over the open sea the upward flux was an order of magnitude or more larger than over the pack ice, and significantly different from all other surfaces. It is worth noticing that the upward fluxes over the summer leads are significantly different from those over the ice floes and over the freeze-up leads. In contrast, the upward flux during the freeze-up period over open leads is not significantly different from the upward flux over the ice during the same period. In the reminder of this paper, we will investigate in more detail the downward fluxes in terms of aerosol dry deposition. The upward flux will be examined in an accompanying paper [Nilsson et al., this issue (b)], in which we suggest that the positive average fluxes over open sea and during the summer period was due to a marine source of primary aerosol particles.
Table 1. Average Aerosol Number Fluxes, Dry Deposition Velocities and Roughness Length Values for Various Surface Conditions During the Arctic Ocean Expedition 1996, First Leg

<table>
<thead>
<tr>
<th>Surface Type</th>
<th>Open sea</th>
<th>Open Leads</th>
<th>Smooth Ice Floes</th>
<th>Rough Ice Floes</th>
<th>Open/Frozen Lead</th>
<th>Smooth Ice Floes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean ±σ N ( w \times 10^4 ) m(^2) s(^{-1})</td>
<td>26±297</td>
<td>1.5±5.8</td>
<td>0.78±19</td>
<td>1.2±21</td>
<td>-1.9±13</td>
<td>-1.3±13</td>
</tr>
<tr>
<td>Median</td>
<td>-25</td>
<td>1.3</td>
<td>0.94</td>
<td>2.3</td>
<td>-3.2</td>
<td>-1.1</td>
</tr>
<tr>
<td>90% range</td>
<td>-133 to 229</td>
<td>-3.7 to 7.6</td>
<td>-9.4 to 14</td>
<td>-21 to 27</td>
<td>-12 to 11</td>
<td>-14 to 16</td>
</tr>
<tr>
<td>Number</td>
<td>60</td>
<td>67</td>
<td>62</td>
<td>32</td>
<td>80</td>
<td>42</td>
</tr>
<tr>
<td>Probability of median = 0</td>
<td>0.91</td>
<td>0.74</td>
<td>0.88</td>
<td>1.00</td>
<td>0.81</td>
<td>0.87</td>
</tr>
<tr>
<td>Log geometric mean ±σ</td>
<td>1.94±0.57</td>
<td>0.49±0.39</td>
<td>0.71±0.47</td>
<td>0.99±0.41</td>
<td>0.86±0.33</td>
<td>0.68±0.52</td>
</tr>
<tr>
<td>Median</td>
<td>79</td>
<td>3.2</td>
<td>4.6</td>
<td>9.9</td>
<td>5.7</td>
<td>4.0</td>
</tr>
<tr>
<td>90% range</td>
<td>15 to 638</td>
<td>1.1 to 11.5</td>
<td>1.2-25.5</td>
<td>2.5-43.9</td>
<td>2.2-18.2</td>
<td>1.1-19.4</td>
</tr>
<tr>
<td>Number</td>
<td>28</td>
<td>37</td>
<td>37</td>
<td>17</td>
<td>33</td>
<td>19</td>
</tr>
<tr>
<td>Probability of median = 0</td>
<td>0.09</td>
<td>0.09</td>
<td>0.09</td>
<td>0.25</td>
<td>0.09</td>
<td>0.10</td>
</tr>
<tr>
<td>Log geometric mean ±σ</td>
<td>1.83±0.42</td>
<td>0.40±0.33</td>
<td>0.70±0.50</td>
<td>0.95±0.45</td>
<td>0.87±0.36</td>
<td>0.76±0.46</td>
</tr>
<tr>
<td>Median</td>
<td>53</td>
<td>-2.4</td>
<td>-5.1</td>
<td>-14</td>
<td>-6.6</td>
<td>-7.5</td>
</tr>
<tr>
<td>90% range</td>
<td>-394 to -27</td>
<td>-6.9 to -1.0</td>
<td>-26 to -1.1</td>
<td>-28 to -2.5</td>
<td>-21 to -3.1</td>
<td>-18 to -1.1</td>
</tr>
<tr>
<td>Number</td>
<td>28</td>
<td>37</td>
<td>37</td>
<td>15</td>
<td>43</td>
<td>23</td>
</tr>
<tr>
<td>Probability of median = 0</td>
<td>0.10</td>
<td>0.10</td>
<td>0.09</td>
<td>1.0</td>
<td>0.09</td>
<td>0.10</td>
</tr>
<tr>
<td>Log geometric mean ±σ</td>
<td>0.28±0.37</td>
<td>-0.47±0.33</td>
<td>-0.51±0.46</td>
<td>-0.44±0.45</td>
<td>-0.04±0.36</td>
<td>-0.46±0.46</td>
</tr>
<tr>
<td>Median</td>
<td>1.8</td>
<td>0.40</td>
<td>0.26</td>
<td>0.56</td>
<td>0.73</td>
<td>0.26</td>
</tr>
<tr>
<td>90% range</td>
<td>0.68 to 6.5</td>
<td>0.17 to 0.74</td>
<td>0.089 to 1.4</td>
<td>0.087 to 2.3</td>
<td>0.36 to 3.0</td>
<td>0.11 to 1.4</td>
</tr>
<tr>
<td>Number</td>
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<td>25</td>
<td>31</td>
<td>15</td>
<td>44</td>
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<tr>
<td>Probability of median = 0</td>
<td>0.09</td>
<td>0.10</td>
<td>0.09</td>
<td>1.0</td>
<td>0.09</td>
<td>0.10</td>
</tr>
<tr>
<td>Log geometric mean ±σ</td>
<td>-6.06±2.21</td>
<td>-3.10±0.93</td>
<td>-2.94±0.92</td>
<td>-1.87±0.52</td>
<td>-3.26±0.54</td>
<td>-4.00±1.33</td>
</tr>
<tr>
<td>Median</td>
<td>4 (10^6)</td>
<td>9 (10^6)</td>
<td>2 (10^5)</td>
<td>2 (10^2)</td>
<td>6 (10^4)</td>
<td>2 (10^4)</td>
</tr>
<tr>
<td>90% range</td>
<td>3 (10^3) to 9 (10^7)</td>
<td>10 (4^2) to 10 (2^2)</td>
<td>3 (10^3) to 3 (10^3)</td>
<td>10 (4^2) to 2 (10^3)</td>
<td>7 (10^7) to 1 (10^3)</td>
<td></td>
</tr>
<tr>
<td>Number</td>
<td>31</td>
<td>67</td>
<td>83</td>
<td>34</td>
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</tr>
<tr>
<td>Probability of median = 0</td>
<td>0.09</td>
<td>0.09</td>
<td>0.09</td>
<td>0.09</td>
<td>0.09</td>
<td>0.09</td>
</tr>
</tbody>
</table>

Note: The data are averaged for measurements with the footprint covering open sea or dominated by an ice floe or an open lead during the late summer period of the expedition and during the early freeze-up period. For the freeze-up period, it is not clear that leads really dominated the flux in the lead column. Finally, statistics are given for periods when the footprint included rougher ice, with ridges on the height scale 0.5 to 3 m. The statistics are given as arithmetic or geometric (due to lognormal distributions) means plus or minus one standard deviation, the median, the upper and lower 90% ranges of the data, and the number of data points (half hour averages) for each group of data. Finally, the probability that the median is not significantly different from zero is given according to a Wilcoxon rank sum test (which is independent on the distribution).

4.1. Surface Type and Roughness

It can be seen from Table 1 that the open sea, leads, and smooth ice had small \( z_o \) (always less than \( 10^{-2} \) m). The surface category with the highest \( z_o \) was the rough ice, where ice ridges on the scale of 0.5 to 3 m were observed. With a \( z_o \) average of about \( 1.4 \times 10^{-1} \) m, this category had more than an order of magnitude larger \( z_o \) than the other surface types. A few peak values of \( z_o \) reached above 1 m. Measurements made with both an ice floe and a lead within the footprint had larger \( z_o \) than just over leads or...
## Table 2. Significant Differences Between Different Surface Types

<table>
<thead>
<tr>
<th></th>
<th>Open Sea (Summer)</th>
<th>Open Leads (Summer)</th>
<th>Smooth Ice Floes (Mainly Summer)</th>
<th>Rough Ice (Mainly Summer)</th>
<th>Open Leads (Freeze-up)</th>
<th>Smooth Ice Floes (Freeze-up)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$z_0$</td>
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<td>&gt;0</td>
<td>&gt;0</td>
<td>&gt;0</td>
<td>&gt;0</td>
<td>&gt;0</td>
</tr>
<tr>
<td>$N'w'$</td>
<td></td>
<td>&gt;0</td>
<td>&gt;0</td>
<td>&gt;0</td>
<td>&gt;0</td>
<td>&gt;0</td>
</tr>
<tr>
<td>$v_d$</td>
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<td></td>
<td>&lt;0.01</td>
<td>&lt;0.01</td>
<td>&lt;0.01</td>
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<td>$N'w'$</td>
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<td>$N'w'$</td>
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<td>$v_d$</td>
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</tr>
</tbody>
</table>

* Dashes indicate that there is no significant difference between the two groups for the specific parameter given in the second column. Here (<0.01) indicates that there is a significant difference and that the probability that the two populations are identical is <0.01. In cases with a probability between 0.01 and 0.2, the probability is given in the table for the reader to judge. The significance tests were made with a Wilcoxon rank sum test, which is distribution-independent.

The dry deposition was significantly different between the leads and the ice during the freeze-up period, see Tables 1 and 2, but not during the summer. The dry deposition over open sea was significantly larger than over the other surfaces. Because of the lack of earlier field measurements of submicrometer aerosol $v_d$ over pack ice and leads, we will compare our $v_d$ to previous measurements of $v_d$ to water and to snow, see Table 3. Indeed, the surface of the pack ice floes is more similar to aged compressed snow than of ice and may also be covered with a layer of snow. Ibrahim et al. [1983] reported values of 3.9 and 9.6x10$^{-4}$ m s$^{-1}$ for stable and unstable stratification, respectively, using ammonium sulfate particles tagged with radioisotope $^{35}$S. Bergin et al. [1995] reported values of 2.3 to 6.2x10$^{-4}$ m s$^{-1}$ from aerodynamic surrogate surfaces (airfoils). This is within the range of our measurements, and in between our geometric mean values for ice floes of 1.1x10$^{-3}$ and 1.0x10$^{-4}$ m s$^{-1}$ for summer and early freeze-up, respectively, see Table 1 and 3.

Dolske and Sievering [1979] reported values of 5x10$^{-3}$ m s$^{-1}$ on average and 2x10$^{-3}$ m s$^{-1}$ for stable stratification from the diabatic drag coefficient over water, which are...
larger than our measurements both over open sea and over open leads. In the later case, this may partly be because ice floes influence measurement over leads since no leads were large enough to fill the entire footprint. It may also be because the references sited in Table 3 used such methods that the given aerosol dry deposition velocities are representative of the aerosol mass deposition, and hence more influenced by the accumulation mode. Our measurements are representative of the aerosol number deposition, and the largest number of particles are found at sizes smaller than the accumulation mode.

4.2. Measurements Versus Parameterization

We can use measured values of $u_*, z_{0r}$, $\zeta$, $N(D_p)$, $T$, and $p$ and equations (4) to (6), integrating over $D_p$, to

<table>
<thead>
<tr>
<th>Conditions</th>
<th>$v_{de}$, m s$^{-1}$</th>
<th>Comment</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow, stable stratification</td>
<td>$3.9 \times 10^{-4}$</td>
<td>ammonium sulfate tagged with $^{35}$S</td>
<td>Ibrahim et al. [1983]</td>
</tr>
<tr>
<td>Snow, unstable stratification</td>
<td>$9.6 \times 10^{-4}$</td>
<td>ammonium sulfate tagged with $^{35}$S</td>
<td>Ibrahim et al. [1983]</td>
</tr>
<tr>
<td>Snow</td>
<td>$2.3 - 6.2 \times 10^{-4}$</td>
<td>aerodynamic surrogate surfaces (airfoils)</td>
<td>Bergin et al. [1995]</td>
</tr>
<tr>
<td>Water, average</td>
<td>$5 \times 10^{-3}$</td>
<td>diabatic drag coefficient</td>
<td>Dolske and Sievering [1979]</td>
</tr>
<tr>
<td>water, stable stratification</td>
<td>$2 \times 10^{-3}$</td>
<td>diabatic drag coefficient</td>
<td>Dolske and Sievering [1979]</td>
</tr>
</tbody>
</table>

Figure 3. One day of measurements from the ice camp. (a) Observed surface roughness (grey triangles), calculated dry deposition velocities (empty circles) based on measured surface roughness, friction velocity, stability, temperature, and aerosol number size distributions. Observed dry deposition velocities are marked with solid circles together with the uncertainty (vertical bars) from equation (2). (b) Observed friction velocity (black triangles) and stability (grey circles). (c) Observed total number of aerosol particles with diameters larger than 10 nm (grey pentagrams) and calculated aerodynamic resistances (black circles) and quasi-laminar sublayer resistance (empty circles). (d) Measured aerosol number size distribution as $\text{d}N/\text{d}\log D_p$ in isopleth lines and grey scale from 0 (white) to 400 (black) cm$^{-1}$ in steps of 40.
calculate $r_u$, $r_h$, $v_a$, and finally $v_d$ for each half hour data point. By doing so, we can compare the calculated $v_d$ with the directly measured $v_d$ and examine the total quality of the measurements and the overall agreement between theory and measurements. Equation (6) was derived by Schack et al. [1985] from a number of wind tunnel experiments over various surfaces, weighting the diffusion and interception effects with the empirical parameters $A$ and $B$. However, lacking wind tunnel data, Schack et al. [1985] reported no $A$ and $B$ for pack ice, or even for any ice or snow surface. It is not possible to derive values for $A$ and $B$ by fitting Equation (6) to our field experimental data owing to its larger variability and errors compared to the wind tunnel data with its controlled environment. If we use $A=0.4$ and $B=20$, the average ratio of measured and calculated $v_d$ are unity. These values of $A$ and $B$ are closest to those reported for water surfaces at $u=0.44$ m s$^{-1}$ by Schack et al. [1985], but $A$ in the Brownian diffusion component of equation (6) are a factor of 2 larger. With this choice of $A$ and $B$, equations (4) to (6) could be used to parameterize the dry deposition over the Arctic Ocean in models.

Figure 3 shows an example of measured $z_m$, measured and calculated $v_d$, measured $u_*$ and $\zeta$, measured $N$, calculated $r_u$ and $r_h$, and number size distribution as $dn/d\log D_p$ for 1 day during the ice camp. Both measured and calculated $v_d$ decreased gradually, see Figure 3a, owing to a gradually increasing $\zeta$ and decreasing $u_*$, see Figure 3b. The good agreement indicates high-quality measurements and good agreement between theory and experiment (i.e., calculated and measured $v_d$ are correlated). However, the magnitude of calculated $v_d$ depend on our choice of $A$ and $B$, see above. Even with the best choice of $A$ and $B$, individual data points disagree up to 1 order of magnitude. The most important factor for the disagreement between the calculations and measurements were probably large errors in the determination of $z_m$ for very small $z_m$ (plus or minus 1 order of magnitude). This could result in a factor 6 error in $v_d$. Diffusion losses in the EC system CPC sampling line of <50% of the 10 nm $D_p$ particles may have caused a 50% error during nucleation events with high numbers of small particles. It is unclear if equation (6) fully takes into account the effect of stability, or if the theory is incomplete because the wind tunnel experiments, on which equation (6) is based, were probably made in neutral stratification. There are also other possible explanations for the discrepancy. At strong stability there could be large vertical gradients in aerosol number concentration and distribution between the measurement level of the EC and the DMPS systems. Leads within the footprint may have caused a source flux of particles from the sea to counteract the deposition fluxes.

It can also be seen from Figure 3c that $r_u$ is 1 order of magnitude smaller than $r_h$, suggesting that the resistance through the quasi-laminar sublayer dominated. This was typical throughout the expedition, although $r_u$ approached $r_h$ during periods of high stability. Clearly, in most cases $r_h$ will dominate over $r_u$, that is the resistance through the quasi-laminar sublayer will be the limiting factor for the dry deposition, not the resistance through the turbulent surface layer. At stable stratification ($\zeta > 0$) $r_u$ can occasionally be as large as $r_h$ Not only $r_u$ but also $r_h$ increased with increasing $\zeta$ (corresponding to decreasing
Table 4. Aerosol Number Dry Deposition Velocity $v_d$ Over Pack Ice and Leads During the Arctic Ocean Expedition 1996, First leg

<table>
<thead>
<tr>
<th>$\xi$&lt;0.1, Unstable and Near Neutral</th>
<th>$\xi$&gt;0.1, Stable</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of cases</td>
<td>75 (8 unstable)</td>
</tr>
<tr>
<td>Log geometric mean+s</td>
<td>-3.19±0.41</td>
</tr>
<tr>
<td>Median</td>
<td>6.0x10^{-4}</td>
</tr>
<tr>
<td>90% range</td>
<td>2.1x10^{-4} to 2.0x10^{-3}</td>
</tr>
</tbody>
</table>

*Units are in m s$^{-1}$.

$u_*$, given a constant wind speed), even if equation (6) fail to take the stability into account explicitly.

It was impossible to match measured and modeled $v_d$ by changing $B$, while changes in $A$ had large effects. The Brownian diffusion term of equation (6) dominated over the impaction term ($A$ over $B$). Comparisons above between measured and theoretical $v_d$ for various $z_0$, $u_*$, $\xi$ and $N$ ($D_p$) demonstrate how difficult it is to study the effect of only one parameter in field data where variations are so much larger than in controlled wind field experiments.

4.3. Stability

Figure 4 shows how $u_*$ depends on $\xi$, that is how friction velocity and the stability were correlated for the measurements during AOE-96. The curves in Figure 3 are based on equation (7) and the median and 90% intervals of observed $\bar{U}$ (4.8 m s$^{-1}$ and 2.5 to 9.4 m s$^{-1}$, respectively) and $z_0$ (7.8x10^{-4} m and 1.4x10^{-3} to 1.3x10^{-2}, respectively) to indicate in what range the measurements of $u_*$ should be found. The observations are not expected to follow the shapes of the curves since in reality the average wind speed and stability are correlated. The smallest values of $u_*$ appeared to follow the theoretical curve throughout the entirely observed range of $\xi$. For larger $u_*$, all samples are less stable, closer to neutral conditions. The high solar zenith angle in the central Arctic summer, and the high surface albedo often results in negative net radiation below clear skies [Nilsson et al., this issue (a)]. However, it appears that this radiative cooling is not sufficient to maintain the strongest stability (largest $\xi$) except at the lowest wind speeds (small $u_*$).

Figure 5. Dry deposition as a function of friction velocity. Measured values for unstable (upward triangles), near neutral (circles), and stable (downward triangles) conditions. Calculated dry deposition velocity for median and 90% variation intervals of surface roughness and stability for the ultrafine mode ($D_p$=14 nm), the Aitken mode ($D_p$=45 nm), and accumulation mode ($D_p$=170 nm).
The measured dry deposition velocity \( v_d \) was also a function of stability, but with much more scatter than \( u_* \). This is at least partly due to the strong influence of \( D_p \) on \( v_d \). During unstable or near neutral conditions, \( v_d \) was significantly larger than during stable conditions, see Table 4, still of the order of a \( 10^{-3} \) to \( 10^{-4} \) ms\(^{-1}\), and the variability within each group was large. The difference is comparable to the difference between unstable and stable conditions reported by Ibrahim et al. [1983], see Table 3. It appears that stability is not the most important factor to determine \( v_d \). According to the theory of dry deposition in section 3, \( v_d \) should be dependent primarily on particle sizes, friction velocity, and surface roughness. The dependence on atmospheric stability through the aerodynamic resistance should be marginal, since usually the laminar sublayer resistance dominates over the aerodynamic resistance, as seen in Figure 3.

### 4.4. Friction Velocity

Figure 5 includes the observed deposition velocities, and the deposition velocity curves based on median and 90% intervals of observed \( z_0 \) (see Table 1) and \( \zeta \) (0.035 and -0.11 to 0.74, respectively), for \( D_p = 14, 45 \) and 170 nm representing the ultrafine, Aitken and accumulation mode particles [Covert et al., 1996; P. P. Aalto, unpublished manuscript 2000]. The purpose of the calculated curves is to represent the intervals where most of the observations should be found.

The data as well as the theoretical curves show \( v_d \) to increase with increasing \( u_* \). The trend has a tendency to decrease as \( u_* \) increases. This is consistent with a \( v_d \) being dependent on a combination of \( u_*^{1/6} \) and \( u_*^{2/3} \) (i.e., equation (6) and Figure 5). The large effect, an order of magnitude, of \( D_p \) on \( v_d \) through \( r_d \) is evident from Figure 5. The variability of \( U \) and \( z_0 \) as well as in \( \zeta \) through \( r_d \) in equation (5) accounts for the wide range of observed \( v_d \). Measured \( v_d \) falls almost entirely within the range of \( v_d \) predicted by the curves, and the scatter in Figure 5 is reasonable for field measurements.

It appears that the measured \( v_d \) dropped off faster with decreasing \( u_* \) than the theoretical curves predict. This may be because most measurements at small \( u_* \) are made at stable conditions (the curves have been calculated for constant \( \zeta \)). Also, the laminar sublayer resistance \( r_s \) has been observed experimentally to depend on stability for vegetative surfaces [e.g., Gallagher et al., 1997]. To our knowledge, there is no corresponding experimental data for snow, ice, or water surfaces. When separating \( v_d \) into \( r_s, r_d \) and \( v_s \), as in Figure 3, \( r_s \) turns out to be an order of magnitude larger than \( r_s \), except for very small surface.
Figure 7. Aerosol number scale height versus aerosol number dry deposition velocity. The horizontal bars represent the uncertainty according to equation (2), and the vertical bars represent the range of observed scale heights in cases when more than one vertical aerosol profile were available. Pentagrams mark two nucleation episodes with high numbers of ultrafine particles. The solid and dotted lines marks 1, 5, and 10 days turnover time for the submicrometer aerosol.

4.5. Particle Size

Since we measured the total $\overline{N'w'}$ or $v_d$ of all particles larger than 10 nm $D_p$, and since the submicrometer aerosol number size distribution typically includes particles in a wide size range, it is difficult to relate $v_d$ to $D_p$. For instance, even a small number of particles at $D_p$ around 20 nm can contribute as much to the average $v_d$ as a much larger number of particles at $D_p$ around 200 nm.

However, some order emerges in the data if we look only at those flux measurements made when the number size distribution was dominated by one of either the ultrafine or Aitken modes [Covert et al., 1996; P. P. Aalto, unpublished manuscript 2000], see Figure 6a. In Figure 6b we present the corresponding averages as well as ranges of observed $v_d$. The dry deposition velocity of each mode fall within the theoretically reasonable range, as represented by curves based on median and 90% intervals of observed $z_0$ (see Table 1). ζ (see previous section) and $u_*$ (0.18 and 0.08 to 0.36 m s⁻¹, respectively). The average observed $v_d$ for the nucleation and Aitken modes, $1.4 \times 10^{-3}$ and $5.1 \times 10^{-4}$ m s⁻¹ (geometric means of $5.0 \times 10^{-4}$ and $2.5 \times 10^{-4}$ m s⁻¹), respectively, are close to the median condition theoretical curve. Furthermore, there is a significant difference between the ultrafine and Aitken modes. The relative decrease in measured average $v_d$ when we go from the ultrafine mode to the Aitken mode agrees well with the theoretical $D_p^{-2}$ decline in the size range dominated by Brownian diffusion. There were a few periods when accumulation mode dominated the aerosol number, but there were always enough of the smaller particles for these to dominate the aerosol dry deposition. Therefore we could not separate a specific dry deposition velocity for the accumulation mode. This may still be possible within a larger data set than the present.

4.6. Turnover Time for the Submicrometer Aerosol by the Dry Deposition Sink

During AOE-96 a helicopter was used to profile the aerosol number in two size classes, $D_p>20$ nm and $D_p>140$ nm, from surface up to about 1200 m (see E. D. Nilsson et al., unpublished manuscript 2000). The
boundary layer scale height $h$ for aerosol particles was derived from these profiles. The scale height was determined according to Nilsson [1996] as the sum of the mixed layer height (if there were any) and the integral depth scale of the stable layer [Stull, 1988]. The combination of measurements of dry deposition velocities and boundary layer scale heights, see Figure 7, allow us to estimate the average turn over time $\tau$ of submicrometer aerosol particles in the boundary layer in nonprecipitating, cloud-free conditions from

$$\tau = \frac{h}{v_d}.$$  

The average $\tau$ was 4.6 days, but as can be seen from Figure 7, where $\tau=1, 5$ and 10 days are indicated with lines, individual estimates of $\tau$ ranged from 0.5 to 13 days. Considering that our observed $v_d$ are smaller than over many other surface types, for example land or vegetation, it may be surprising that $\tau$ was not larger. This is entirely due to the shallow nature of the boundary layer over the Arctic Ocean, for example $h$ in Figure 7 averaged only about 150 m and ranged from less than 50 m to just below 500 m. Even a small dry deposition velocity will be able to turn over the aerosol number in a relatively short time when the boundary layer is that shallow.

Two of the cases with the shortest turnover time, marked with stars in Figure 7, were associated with nucleation events, described by E. D. Nilsson et al. (unpublished manuscript, 2000) and Bigg and Leck [this issue]. Obviously, the presence of many small and very young particles with a large Brownian diffusion rate to the surface will result in much shorter $\tau$ than an aerosol size distribution dominated by Aitken or accumulation mode particles. Using the average $\tau$ for the ultrafine and Aitken modes from Figure 6 and the average $h$ results in an average $\tau$ of 1.3 and 3.5 days for the ultrafine and Aitken modes, respectively. We would expect the accumulation mode to have a larger $\tau$ (of the order of 2 weeks) than the Aitken mode. Such a long turnover time for accumulation mode particles by dry deposition would only be applicable for those accumulation mode particles not activated as condensation nuclei in clouds and fogs because the turnover time for wet deposition will be smaller [Nilsson and Leck, 1997].

5. Summary and Conclusions

We have established averages and ranges of dry deposition velocities for submicrometer aerosol particles for various conditions over the pack ice and open Arctic waters. The dry deposition primarily on friction velocity and particle size, as expected from theory, and to some degree on static stability. Resistance through the quasi laminar sublayer appeared to dominate over aerodynamic resistance, except in the most stable conditions. Our measurements and dry deposition theory agree well qualitatively. We recommend the parameterization of the quasi-laminar sublayer by Shack et al. [1985], which agrees quantitatively with our measurements if used with certain empirical parameters, as reported in 4.2 above. The dry deposition fluxes were obviously dominated by the ultrafine and Aitken modes since these have the largest number of particles.

Because the boundary layer over the Arctic ocean pack ice is so shallow, even the small aerosol number dry deposition velocities reported here resulted in submicrometer aerosol turnover times of a few days. Therefore in situ production of particles, surface sources, or entrainment from the free troposphere are needed to maintain the observed aerosol concentration even in the absence of wet deposition, or to cause the episodic peaks in the number of particles we observed in the ultrafine and Aitken modes. Indeed, very young particles or aerosol production, so called nucleation, was observed frequently (P. P. Aalto, unpublished manuscript, 2000). Since we expect the accumulation mode to have a longer turnover time, this argument does not hold for the accumulation mode particles, even though we suspect there exists an in situ production of primary aerosol particles in this size range from the open leads within the pack ice [Nilsson et al., this issue (b)].

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