

Characterization of distinct Arctic Aerosol Accumulation Modes and their Sources

Lange, R.; Dall'Osto, Manuel; Skov, H.; Nielsen, I.E.; Beddows, David; Simo, R.; Harrison, Roy; Massling, A.

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3 **Characterization of distinct Arctic**

4 **aerosol accumulation modes and their**

5 **sources**

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8 R. Lange^{1*}, M. Dall'Osto², H. Skov¹, J. K. Nøjgaard¹, I. E.

9 Nielsen¹, D.C.S. Beddows³, R. Simo², Roy M. Harrison^{3†}

10 and A. Massling¹

11

12

13 *¹Department of Environmental Science, Aarhus University, Frederiksborgvej*

14 *399, 4000, Roskilde, Denmark*

15

16 *²Institute of Marine Sciences, ICM-CSIC, Passeig Marítim de la Barceloneta,*

17 *37-49. E-08003, Barcelona, Spain; * also E-mail: dallosto@icm.csic.es.*

18

19 *³National Centre for Atmospheric Science, University of Birmingham,*

20 *Edgbaston, Birmingham, B15 2TT, United Kingdom*

21

22 *†Also at: Department of Environmental Sciences / Center of Excellence in Environmental Studies, King*

23 *Abdulaziz University, PO Box 80203, Jeddah, 21589, Saudi Arabia*

24

25 ***Corresponding author: Robert.lange@envs.au.dk**

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Keywords

Arctic aerosol, accumulation mode, Arctic haze, biogenic Aerosol, CCN

Highlights

- Accumulation mode aerosol is measured in North East Greenland during a 7 year record, apportioning 56% of total aerosol size distributions.
- Three aerosol categories are found: accumulation *Haze* (32%), accumulation *Aged* (14%) and accumulation *Bimodal* (6%).
- Accumulation categories have very distinct chemical and physical properties across different seasons.
- Arctic accumulation mode aerosols during summer coexist with a smaller Aitken mode, likely biogenic.
- Cloud Condensation Nuclei (CCN) measurements suggest that ultrafine aerosol (~30-60nm) drives CCN concentrations in the Arctic during summer.

59 **Abstract**

60

61 Measurements of aerosol number size distributions (9-915 nm), as well as aerosol
62 chemistry and cloud condensation nuclei (CCN) activity, were undertaken at Villum
63 Research Station, Station Nord (VRS) in North Greenland during a 7 year record (2010-
64 2016). Clustering analysis on daily size distributions identified several k-means SMPS
65 clusters. K-means clusters of accumulation aerosols (with main size modes >100 nm)
66 accounted for 56% of the total aerosol time sampling period (89-91% during February-
67 April, 1-3% during June-August). By air trajectory association, diurnal variation patterns,
68 and relationship to meteorological and pollution variables, three typical accumulation-mode
69 aerosol categories were identified: *Haze* (32% of the time), *Bimodal* (14%) and *Aged* (6%).
70 In brief: (1) *Haze* accumulation aerosol shows a single mode at 150 nm, peaking in
71 February-April, with highest loadings of sulfate and black carbon concentrations; (2) *Aged*
72 accumulation aerosol shows a single mode at 213 nm, peaking in September-October and
73 is associated with cloudy and humid weather conditions during autumn; and (3)
74 Accumulation *Bimodal* aerosol shows two modes at 38 nm and 150 nm, peaking in June-
75 August, with the highest ratio of organics to sulfate concentrations. The three aerosol
76 categories were considered alongside Cloud Condensation Nuclei (CCN) concentrations.
77 We suggest that organic compounds - likely biogenic in nature and responsible for the
78 smaller mode in the *Bimodal* category - contribute significantly to the CCN activity. It is
79 concluded that - at least during summer - an Aitken mode, biogenic in origin always
80 coexists with an accumulation mode, stressing the importance of better characterizing the
81 marine ecosystem and the aerosol-mediated climate effects in the Arctic.

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1. Introduction

The Arctic, one of the most sensitive regions to climate change, is warming at a rate twice as rapid as the global average (AMAP, 2011). Currently, atmospheric aerosols are poorly characterized in Arctic climate models. Aerosol particles may perturb the radiative balance of the Arctic environment in numerous ways (Carslaw et al., 2013; Ramanathan et al., 2001). Overall, different aerosol chemical species, as well as particle size and abundance, may determine the magnitude of the aerosol induced direct forcing. Furthermore, aerosols also constitute the seeds upon which cloud droplets form (Ramanathan et al., 2001). Improved understanding of the spatial and temporal variability of the microphysical properties of the aerosol in the Arctic is required in order to determine the magnitude and direction of future climate change in this important region.

The Arctic aerosol has been shown to be highly variable. Broadly, over the Arctic region the aerosol mass, surface area and number size distribution properties are a strong function of season, and this seasonality is repeated from year to year (Freud et al., 2017; Nguyen et al., 2016; Tunved et al., 2013). It is well established that the Arctic winter and spring atmosphere is more heavily impacted by transport of air pollution from lower latitudes compared to summer (Heidam et al., 2004; Law and Stohl, 2007). The continent-derived winter and spring aerosols, known as Arctic haze, reach their maximum number concentration during late spring, approximately in April. The transition from the Arctic haze conditions to the lower aerosol loadings over the summer period is driven by increasing wet scavenging due to increasing temperatures over a period of about two weeks (Browse et al., 2012; Croft et al., 2016; Engvall et al., 2008). It is becoming increasingly evident that biogenic ultrafine (including freshly nucleated) particles dominate ambient aerosols in Arctic areas during summer (Dall'Osto et al., 2017a; Dall'Osto et al., 2017c). Occasionally long-range pollution transport events also occur during summer (Iziomon et al., 2006; O'Neill et al., 2008). Towards the end of summer the intensity of sunlight decreases and, despite a low concentration of large particles, new particle formation comes to a halt as the production of nucleating vapors is too slow.

122 Aerosol number size distributions from multi-year measurements have been reported from
123 different arctic research stations: Zeppelin (Tunved et al., 2013), Tiksi (Asmi et al., 2016),
124 Alert (Croft et al., 2016), Barrow (Latham et al., 2013) and Villum Research Station (VRS),
125 Station Nord (Nguyen et al., 2016). All studies broadly converge in a similar scenario: the
126 haze period characterized by a dominating accumulation mode aerosol (March–May), is
127 followed by the sunlit summer with high concentration of small particles (June–August).
128 The remaining year is characterized by low concentration of accumulation mode particles
129 and negligible abundance of ultrafine (<100 nm) particles (September–February).

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131 Despite this, information on different types of aerosol accumulation modes is scarce.
132 Tunved et al. (2013) reported the occurrence of a typical accumulation mode geometric
133 mean particle diameter (D_p) of 161-185 nm during winter months (November to March)
134 and 130-163 nm during April to October, at Zeppelin, Svalbard. Nguyen et al. (2016)
135 reported that the larger D_p accumulation mode persists further into the summer at VRS,
136 Greenland, than at Zeppelin, with a typical mode geometric mean diameter D_p of 167-179
137 nm for months November to May and 107-119 nm for months June to September. A recent
138 inter-comparison of particle number size distributions observed at several Arctic stations
139 by Freud et al. (2017) suggests variations between the different stations throughout the
140 year. The most prominent differences are observed between the stations at Barrow and
141 Zeppelin. Barrow features a wider accumulation mode, with higher concentrations and
142 smaller D_p than Zeppelin in months September to May.

143

144 To the best of our knowledge, no long-term studies on Arctic aerosol have identified
145 several distinctively different accumulation mode aerosols. In this study we provide further
146 evidence that multiple accumulation mode aerosol clusters exist in the Greenlandic high
147 Arctic, and that these are present at different proportions throughout the year. These
148 accumulation mode aerosol clusters are characterized both physically and chemically, and
149 statistically significant differences are highlighted. In conclusion, the aim of the present
150 paper is to improve the understanding of Arctic aerosol accumulation modes, and to
151 describe them in tandem with meteorological parameters, gaseous concentrations, aerosol
152 chemical species and cloud condensation nuclei properties.

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2 Methodology

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2.1 Location

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2.2 Scanning Mobility Particle Sizer (SMPS)

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We analyzed continuous Scanning Mobility Particle Sizer (SMPS) data collected in the period 2010-2016 in the size range of 9-915 nm in diameter. The sampling setup has been described in detail by Nguyen et al. (2016), with the difference that since summer 2015 the SMPS has been situated in the newly constructed air observatory measurement hut, described above. The instrument is custom-built with a Vienna-type medium column, similar to SMPS instruments described in Wiedensohler et al. (2012). Our SMPS used either a condensation particle counter (CPC) model TSI 3010 or model TSI 7220. To ensure correct functioning, volumetric flow rates, temperatures and relative humidity (RH) of the aerosol- and sheath flow were monitored, as well as inlet ambient pressure. No additional drying was performed, as the transition from the low ambient temperatures outside of the huts (-45 to +15 °C, yearly average -15 °C) to the heated inside (>20 °C) generally provides sufficient decrease of RH. The SMPS sample flow RH only in exceptional cases exceeded 35%. An algorithm according to Pfeifer et al. (2014) was used to invert the SMPS measurements. The resulting particle number size distribution series were quality controlled to ensure correct functioning of the instrument and absence of

186 influence from local pollution from near-by vehicles or by the military camp. Data was
187 excluded from further analysis when these conditions were not met.

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190 **2.3 Concentrations of Gaseous Pollutants**

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192 O₃ and NO_x were measured using gas analyzers (API photometric O₃ analyser (M400),
193 API chemiluminescence NO_x analyzer (M200AU)). NO_x data was available for most of
194 2011-2012, whereas O₃ data was available throughout most of the study period.

195

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197 **2.4 Particulate Matter Properties**

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199 In the period from May 2011 to August 2013, observations of the aerosol light absorption
200 coefficient were conducted using the multiangle absorption photometer (MAAP, Model
201 5012 Thermo Scientific), described in detail in Petzold and Schönlinner (2004). More
202 information can be found in Massling et al. (2015). Soot particle aerosol mass
203 spectrometer (Aerodyne, SP-AMS) was deployed for four months over the period
204 February-June 2015. Further details can be found in Nielsen et al. (2017) (manuscript in
205 preparation).

206

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208 **2.5 Meteorological Data**

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210 Wind speed and wind direction from a sonic anemometer were available from April 2011 to
211 April 2013 (Sonic anemometer (METEK, USA-1)). Data coverage is poorer during the
212 winter months due to frost on the anemometer.

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219 **2.6 Cloud Condensation Nuclei**

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221 Aerosol CCN activity was measured by a Cloud Condensation Nuclei Counter (CCN-100,
222 DMT) during two field studies in 2016. The first campaign was from April to May and the
223 second from August to September 2016. The CCN counter measurement cycle included
224 10 settings of supersaturation (SS) in the range 0.1-1.0% SS and one at maximum
225 reachable SS (~2% SS). The temperature gradient in the CCN column was allowed to
226 stabilize for 5 min before recording measurements. However, 15 min. stabilizing time was
227 used when resuming to 0.1% SS from the highest reachable SS. The measurement
228 recording time was 5 min in all cases. The total CCN measurement cycle duration thus
229 was 120 min. At the highest reachable SS all particles above 25 nm are assumed to
230 activate as CCN. The CCN concentration at this SS was utilized to calibrate the total
231 detectable particle concentration by the CCN counter, relative to that inferred from the
232 SMPS. The instrument SS was calibrated at the beginning and end of each of the two field
233 studies at 0.1-0.47% SS, resulting in four total SS calibrations during the whole CCN
234 measurement period. Following the conclusion of the campaigns, additional SS
235 calibrations were undertaken to verify that the calibrations made during the field studies
236 were linear and valid ranging up to 1% SS. During these SS calibrations, CCN activation
237 D_p ($D_{p,crit}$) of monodisperse ammonium sulfate aerosol was determined. The E-AIM model
238 (Clegg et al., 1992; Wexler and Clegg, 2002) and the Köhler equation (Köhler, 1936) was
239 used to calculate the corresponding real SS (SS_{calc}), which then was compared to the set
240 SS on the CCN counter (SS_{set}). Combining the data from all performed SS calibrations
241 yielded a linear relationship between SS_{calc} and SS_{set} . This was determined by linear least
242 squares fitting. CCN data was quality controlled on the basis of achievement and stability
243 of column temperature gradients, tolerances to temperature differences inside the
244 instrument and stability of sample- and sheath air flows. By assuming chemical
245 homogeneity of the measured aerosol population, critical activation diameters were
246 calculated by sequential downwards integration of SMPS number size distributions until
247 the following condition was satisfied (equation 1) (Kristensen et al., 2016; Jurányi et al.,
248 2010):

249

$$\int_{Dp_{crit}}^{Dp_{max}} n_N(Dp)dDp = N_{CCN} \quad (\text{equation 1})$$

250 Where Dp_{max} is the maximum Dp measured by the SMPS, n_N is the particle number size
251 distribution and N_{CCN} is the corresponding measured CCN number concentration. Internal
252 particle losses of the CCN counter were accounted for when determining Dp_{crit} . This was
253 done by applying a particle size dependent transmission curve for the CCN counter,
254 published by Rose et al. (2010) to the SMPS number size distributions.

255

256

257 **2.7 Aerosol Size Distribution Statistical Analysis**

258

259 SMPS data from a total of 1,717 days distributed over the 7 years were examined by k-
260 means analysis according to the methodology in Beddows et al. (2009). Eight clusters
261 were selected based on the best compromise between Silhouette Width (0.43) and Dunn
262 Index ($1.6 \cdot 10^{-3}$). The time series of these clusters were inspected to see if any could be
263 merged and they proved to be separable. To each day of the data period the respective
264 dominant cluster was assigned. Of the identified eight clusters, three are related to a
265 dominant and distinct accumulation mode aerosol. The climate-relevant characteristics of
266 these accumulation mode clusters are discussed in detail in this study. In total, the
267 accumulation mode clusters are dominant on 56% of the days in the data period, while the
268 remaining five clusters were dominant 44% of the time. The five remaining clusters are
269 related to ultrafine particle modes (<100 nm) and are described elsewhere (Dall'Osto et
270 al., 2017b). Size distribution parameters were obtained by fitting of log-normal functions to
271 the average distributions. For each accumulation mode cluster, the average and median
272 values of all measured parameters were calculated from the data obtained during the days
273 the respective cluster was observed. To increase robustness towards outliers, the upper
274 and lower 1-percentile from all utilized datasets were removed.

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283 Results and Discussion

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285 3.1 K-means Clustering Results

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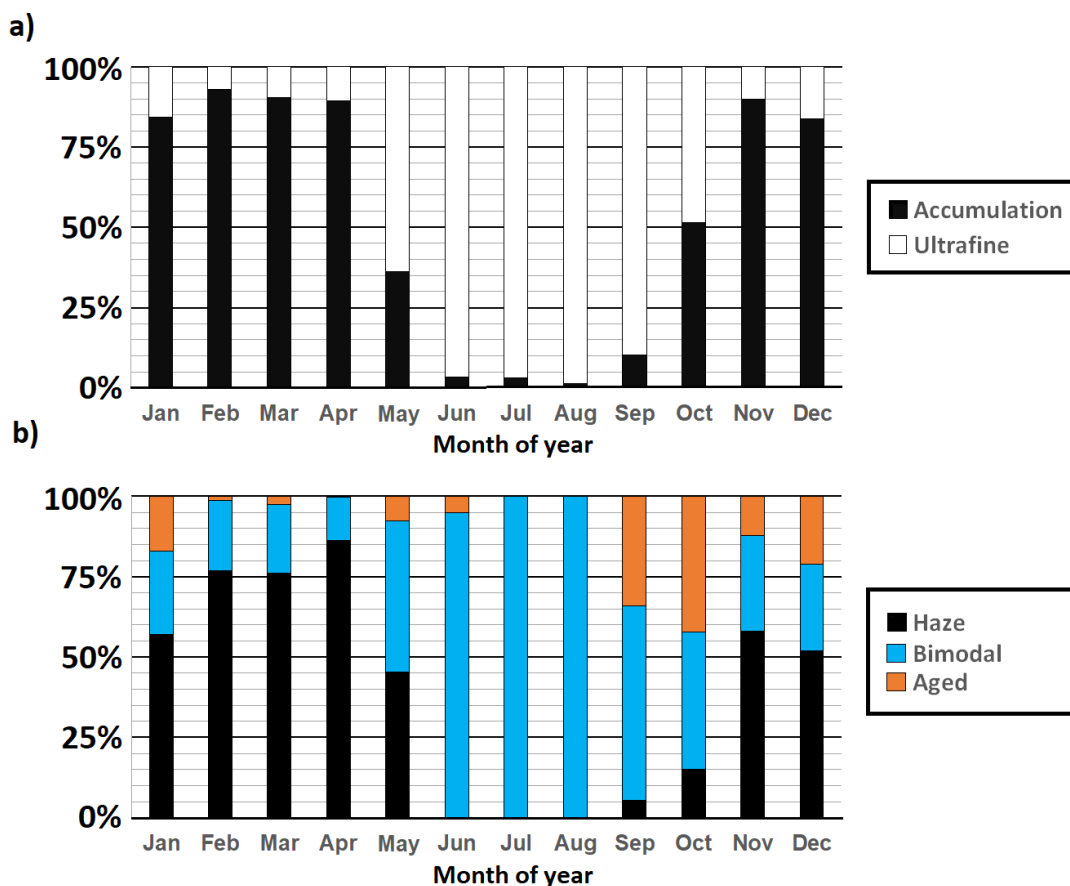
287 **Fig. 1a** shows the annual variation of the sum of the three accumulation categories with
288 respect to their abundance relative to the sum of the five ultrafine ones (Dall'Osto et al.,
289 2017b). The main difference between the ultrafine and the accumulation categories is in
290 having the majority of $N_{9-915nm}$ in particle sizes lower and higher than 100nm, respectively.
291 It can clearly be seen that the accumulation categories dominate the winter times
292 (January-March, 89-91%), whereas they present a minimum in summer months (June-
293 August, 1-5%). **Fig. 1b** shows the seasonality of each accumulation category, represented
294 by the occurrence of each accumulation category during each month of the year. The
295 three particle size distribution accumulation mode categories show very different
296 seasonality for multiple reasons, including different meteorology and different biological
297 ocean activity throughout the year as well as different anthropogenic influences over time.
298 The three accumulation mode categories are termed *Haze*, *Bimodal* and *Aged*, based
299 mainly on the temporal trends and aerosol size distributions. Additionally, it is important to
300 note that chemical and physical parameters associated with each individual category are
301 presented in the following sections - supporting the assigned terminology.

302

303 **Fig. 1b** shows that during July and August - where ultrafine categories dominate the
304 particle number size distribution - the remaining accumulation aerosols mode categories
305 consist solely of the *Bimodal* category. By contrast, during the months September and
306 October its occurrence decreases while the *Aged* category reaches its maximum relative
307 occurrence. The *Aged* category is largely absent from February to August. In the months
308 November to April the *Haze* category is dominant, reaching its maximum occurrence in
309 April.

310

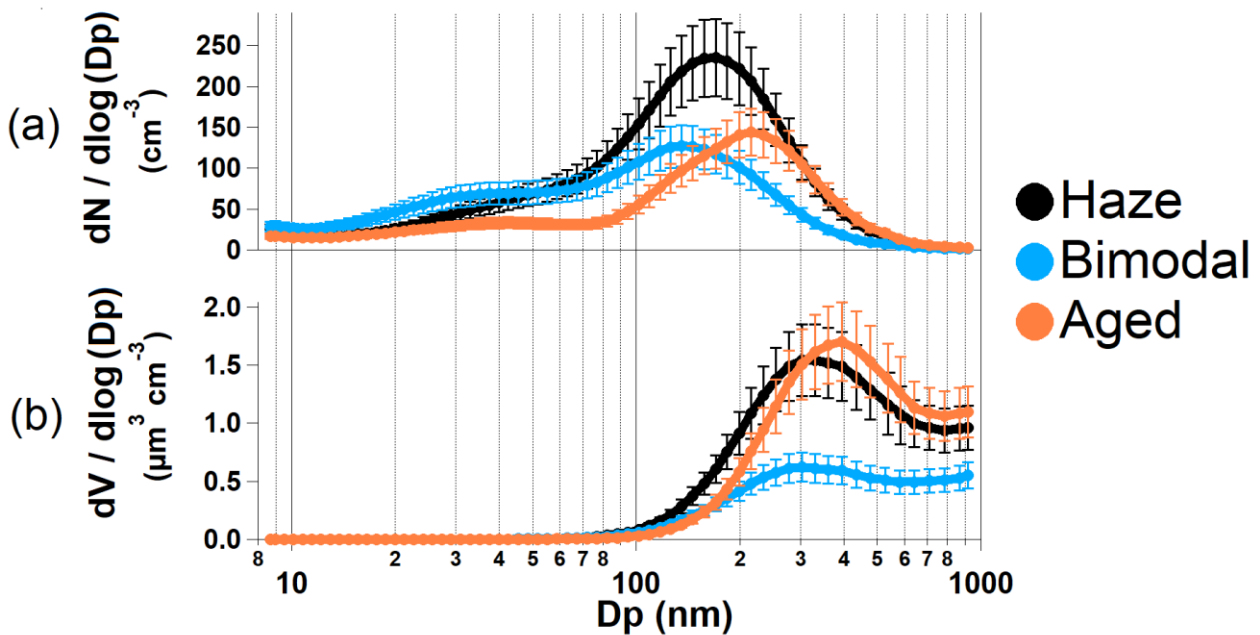
311 The aerosol number and volume size distributions of the three accumulation aerosol
 312 categories are shown in **Fig. 2a** and **2b**, respectively. The *Haze* category appears with the
 313 highest total number concentration of the three categories. Its number size distribution
 314 peaks at $D_p = 173$ nm, and is unimodal in appearance. The *Bimodal* category depicts a
 315 larger mode that peaks at $D_p = 150$ nm, which accounts for 53% of its average total
 316 particle number concentration. The smaller mode around $D_p = 38$ nm accounts for the
 317 remaining 47% of average total particle number concentration. The *Aged* category is
 318 unimodal, with the maximum number concentration at $D_p = 213$ nm. The size range of our
 319 SMPS measurements was limited to the maximum size bin of 915 nm, leaving the particle
 320 volume size distributions incomplete (**Fig. 2b**). It appears that the *Haze* and *Aged*
 321 categories have roughly equal total volume concentrations in the measured range,
 322 whereas the *Bimodal* category had a significantly lower volume concentration in this range.
 323



324 **Fig. 1.** (a) Ultrafine (<100nm) and Accumulation (>100nm) aerosol category occurrence
 325 presented as monthly averages (period 2010-2017). (b) Annual variation of the three
 326 accumulation mode aerosol categories only.
 327
 328

329 The aerosol size distributions presented in **Fig. 2** are obtained from the k-means clustering
 330 carried out with daily resolution. In general, aerosol number size distributions appear very
 331 stable over the days, allowing a classification of these distributions based on 24 hour
 332 averages. This can also be supported by the fact that same aerosol categories often
 333 appear in consecutive days. A strong variation in daily aerosol number size distribution is
 334 only observed during nucleation events, because this is the nature of the underlying
 335 process. The calculated average condensation sink (CS) (Dall'Osto et al., 2013) presented
 336 by the three accumulation categories was $1.12 \cdot 10^{-3}$, $0.71 \cdot 10^{-3}$ and $0.89 \cdot 10^{-3} \text{ s}^{-1}$ for the
 337 *Haze*, *Bimodal* and *Aged*, respectively. These daily average CS values are well above
 338 those calculated for days characterized by ultrafine categories ($1-8 \cdot 10^{-4} \text{ s}^{-1}$) (Dall'Osto et
 339 al., 2017b).

340



341

342 **Fig. 2.** Aerosol (a) number and (b) volume size distributions for the three accumulation
 343 categories.

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346

3.2 Gas Concentrations and Meteorological Parameters

347

348 Daily median gaseous concentrations were calculated and compared with the appearance
 349 of accumulation aerosol categories (**Table 1**). NO_x was always found near the detection
 350 limit of 0.2 ppb. O_3 concentration medians were 32.9, 34.6 and 28.7 ppb for the categories

351 *Haze*, *Bimodal* and *Aged*, respectively. Median O₃ and NO_x concentrations did not show
 352 any diurnal profile, indicating that there is no varying influx of polluted air from the station
 353 premises to the measurement site. The above mentioned factors suggest that the
 354 examined accumulation categories are probably attributable to regional and long range
 355 transport, and not to local pollution.

356 **Table 1** also shows the average meteorological parameters observed for each category.
 357 The *Haze* aerosol category was observed during episodes of high atmospheric pressure,
 358 low temperatures, relatively dry conditions and a high amount of incoming shortwave
 359 radiation. The *Bimodal* category was observed during episodes with higher temperature
 360 and lower atmospheric pressure, while the *Aged* category was observed at the highest
 361 temperatures, highest relative humidity, highest wind speed and lowest pressure. We find
 362 statistically significant differences (Wilcoxon signed-rank, $z < 0.05$) between all average
 363 parameters except for the irradiance and pressure of the *Bimodal* and *Aged* categories,
 364 and between the average wind speed of the *Haze* and *Aged* categories. These findings
 365 agree well with the expectation that Arctic haze is most prominent during late winter and
 366 early spring, when conditions are usually sunny and cold. The *Aged* category on the other
 367 hand is usually observed during cloudy and humid weather conditions during autumn or
 368 during winter, when sunlight is absent. We calculated about 10,300 air mass back
 369 trajectories aiming to shed light on possible source regions. A “modal” air mass back
 370 trajectory for each number size distribution cluster was calculated by averaging all the
 371 back trajectories calculated with arrival dates corresponding to days with assignment of a
 372 k-Mean cluster. Using HYSPLIT4 (with revision made in February 2016), five day
 373 backward trajectories were calculated from 2010 to 2016 using arrival hours of 00:00,
 374 06:00, 12:00 and 18:00 and an arrival height of 10 m. Unfortunately, no robust differences
 375 were found among the accumulation categories.

376

Accumulation aerosol category	O ₃ (ppb)	T (° C)	RH (%)	Radiation (W/m ²)	Pressure (hPa)	Wind speed (m/s)
<i>Haze</i>	32.9 (±6.4)	-18.9 (±6.2)	74.1 (±6.6)	60.4 (±110.6)	1017.4 (±8.7)	3.1 (±2.2)
<i>Bimodal</i>	34.6 (±8.2)	-16.0 (±10.0)	74.4 (±8.1)	48.0 (±110.9)	1013.7 (±10.2)	3.0 (±2.3)
<i>Aged</i>	28.7 (±7.6)	-12.2 (±9.7)	80.0 (±4.1)	18.2 (±72.3)	1012.2 (±10.0)	3.3 (±2.7)

377 **Table 1**

378 Median Ozone concentration and average meteorological parameters (Temperature,
379 Relative Humidity, Radiation, Pressure and Wind speed) for the three accumulation
380 aerosol categories.

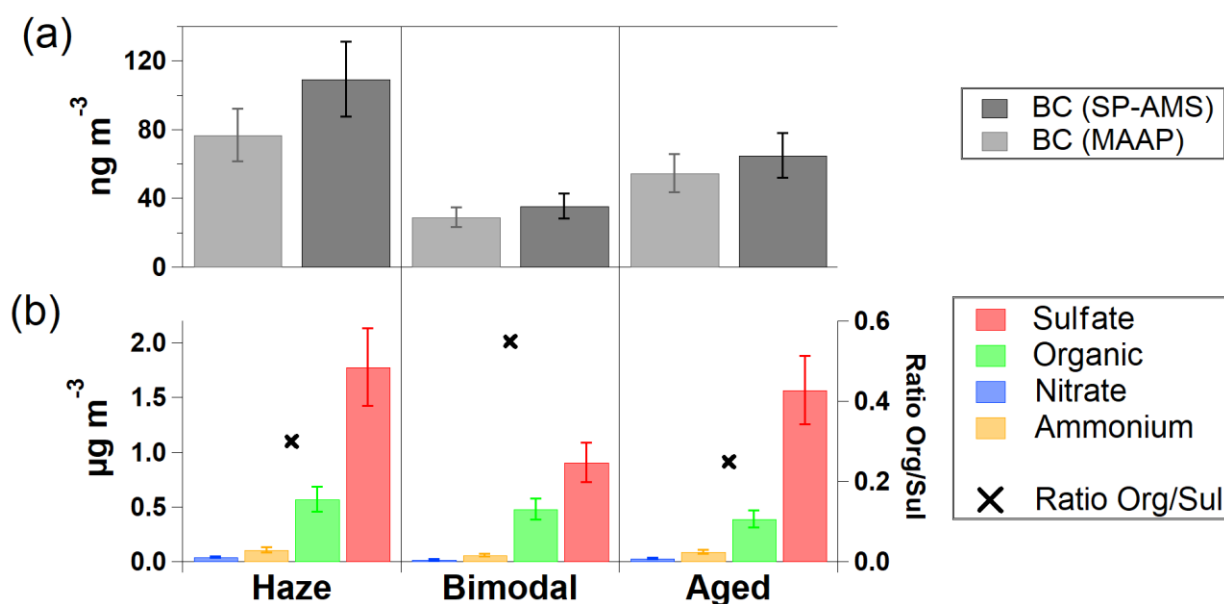
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384 **3.3 Chemical Composition**

385

386 Long-term (two year period between 2011 and 2013) MAAP (Multi Angle Absorption
387 Photometer) measurements of black carbon (BC) mass concentration were compared with
388 our accumulation aerosol classification (Fig 3a). The highest average BC concentration of
389 77 ng m^{-3} was found concurrent with the *Haze* category. By contrast, the lowest (27 ng m^{-3})
390 was found with the *Bimodal* category. Intermediate values were found for the *Aged*
391 cluster (55 ng m^{-3}). This trend is in agreement with five months of SP-AMS measurements
392 conducted in 2016 (Fig 3a). Arctic marine air masses are expected to be associated with
393 pristine clean conditions, with BC concentrations smaller than 15 ng m^{-3} . We conclude
394 from the BC data that all three accumulation categories occurred under perturbed natural
395 conditions, which to some extent are influenced by anthropogenic emissions. Previously
396 measured atmospheric black carbon concentrations in Northeast Greenland averaged
397 $67 \pm 71 \text{ ng m}^{-3}$ in winter and $11 \pm 9 \text{ ng m}^{-3}$ in summer (Massling et al., 2015). Our study
398 shows that aerosol number size distributions characterized by dominant accumulation
399 modes are associated with average BC concentrations in the range of 27-77 ng m^{-3} .

400



401 **Fig. 3.** (a) Aerosol black carbon mass derived from the MAAP and SP-AMS. (b) AMS-
 402 derived sulfate, organic, nitrate and ammonium mass as a function of the three aerosol
 403 accumulation categories.
 404
 405

406 Additional information can be obtained by comparing our accumulation aerosol categories
 407 with aerosol chemical measurements obtained by SP-AMS. **Fig. 3b** shows average mass
 408 concentrations obtained for nitrate, sulfate, organics and ammonium. The *Haze* category
 409 presented high concentrations of sulfate ($1.79 \mu\text{g m}^{-3}$) and organics, ($0.57 \mu\text{g m}^{-3}$). This
 410 category is also associated with the highest average concentrations of nitrate ($0.04 \mu\text{g m}^{-3}$)
 411 and ammonium ($0.11 \mu\text{g m}^{-3}$). By contrast, the *Bimodal* category concurs with lower
 412 concentrations of sulfate ($0.91 \mu\text{g m}^{-3}$), organics ($0.48 \mu\text{g m}^{-3}$), nitrate ($0.02 \mu\text{g m}^{-3}$) and
 413 ammonium ($0.06 \mu\text{g m}^{-3}$). The *Aged* category features $1.57 \mu\text{g m}^{-3}$ and $0.39 \mu\text{g m}^{-3}$ of
 414 sulfate and organics, respectively, and $0.03 \mu\text{g m}^{-3}$ and $0.09 \mu\text{g m}^{-3}$ of nitrate and
 415 ammonium, respectively. The very low nitrate concentrations for the *Bimodal* and *Aged*
 416 categories are below the limit of quantification. A statistically significant difference
 417 (Wilcoxon signed-rank, $z < 0.05$) exists between the basic chemical composition of the
 418 three categories, except for nitrate, ammonium and organics between the *Haze* and the
 419 *Aged* aerosols.

420

421 Previous studies at Villum Research Station observed a positive correlation ($R^2 = 0.72$)
 422 between BC and sulfate concentrations over the years 2011 to 2013 (Massling et al.,
 423 2015), suggesting that the transport of combustion-derived BC-rich particles to the Arctic

424 was accompanied by aging of the aerosols through condensational processes. It has also
425 previously been shown that the sulfate concentrations observed at the VRS station are
426 dominantly affected by anthropogenic emissions and to a lesser extent by sea spray
427 (Heidam et al., 2004; Nguyen et al., 2013). However, our new analysis with a much higher
428 resolution (hourly aerosol number size distributions, versus previously 7 days off-line filter
429 measurements) allows to separate the three accumulation mode categories. For all the
430 analyzed chemical components, the largest aerosol loadings are found in the *Haze*
431 category, and the lowest in the *Bimodal* category. Additional important information can be
432 drawn from the ratios among different chemical components. For example, the ratios of
433 organics to BC in the *Haze* and *Aged* categories were very similar (7.1 and 7.4). By
434 profound contrast, the ratio was 18 for the *Bimodal* aerosol category. A similar pattern was
435 found for the ratio of organics to sulfate average mass concentrations (**Fig. 3b**): the ratio of
436 the *Bimodal* category (0.55) is higher than the ratios of the other two categories (*Haze* and
437 *Aged*) (0.25-0.3). Therefore, the *Bimodal* aerosols are enriched with organic matter and
438 anti-correlation with BC content. Due to the usually low concentrations of aerosols over the
439 inner Arctic pack ice area in summer, biogenic natural particle sources have been
440 emphasized to be more important than transport from continental sources. Biogenic
441 primary ultrafine aerosols include micro-colloids shown to behave as polymeric gels (Chin
442 et al., 1998; Leck and Bigg, 2005; Orellana et al., 2011). These are produced by
443 phytoplankton and sea ice algae as biological secretions. A number of studies have also
444 reported in situ formation of secondary new aerosols in the Arctic, which mostly involve
445 new particle formation from natural emissions of volatile species and their subsequent
446 oxidation to low volatility compounds (Dall'Osto et al., 2017a; Nguyen et al., 2016; Tunved
447 et al., 2013). It is likely that the organic enrichment detected in the *Bimodal* cluster results
448 from the combination of primary and secondary aerosols of biological origin.

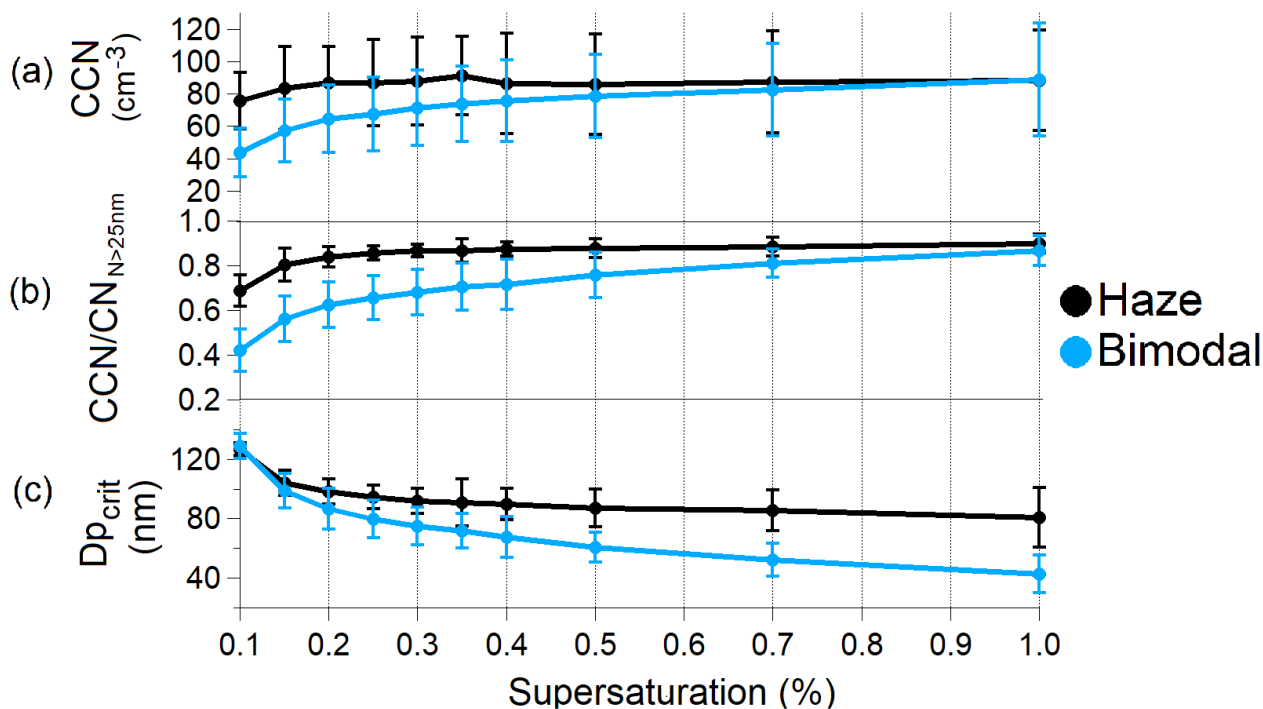
449
450

451 **3.3 CCN Properties**

452

453 The net climate impact of atmospheric aerosols depends on their number size distribution
454 and chemical composition (Rosenfeld, 2006). In particular, particle size (Anttila et al.,
455 2012; Dusek et al., 2006) have been found to be the greatest controlling factor in cloud

456 condensation nuclei (CCN) efficiency. As the Arctic often is a CCN-limited regime, the
457 variability of even low concentrations of CCN is important (Mauritsen et al., 2011).
458 We calculated average daily CCN concentrations corresponding to each accumulation
459 aerosol cluster. **Fig. 4a** shows the CCN concentrations at different supersaturations (%)
460 for two of the three clusters. *Aged* aerosols did not have enough associated CCN data and
461 are not discussed hereafter. The *Haze* cluster presented total CCN concentrations of
462 around $80 \pm 20 \text{ cm}^{-3}$ at all supersaturations. By contrast, a clear gradient was seen for the
463 *Bimodal* cluster, with CCN concentrations increasing from 43 cm^{-3} to 88 cm^{-3} as SS
464 increased from 0.1 to 1.0%. No statistically significant difference was found between CCN
465 concentrations for the two different accumulation clusters at SS above 0.3%. This is very
466 interesting, given the higher $N_{9-915\text{nm}}$ in *Haze* than in *Bimodal* aerosols (**Fig. 2a**). Most
467 CCN-active particles are typically sized between 50 and 150 nm diameter; traditionally a
468 typical cloud condensation nucleus (CCN) is considered to have a minimum diameter of
469 about 100 nm. Considering this threshold size, we find an average $N_{>100\text{nm}}$ concentration of
470 109 cm^{-3} and 54 cm^{-3} in the *Haze* and *Bimodal* clusters, respectively; whereas
471 concentrations of $N_{<100\text{nm}}$ are similar (54 cm^{-3} and 59 cm^{-3} , respectively; **Fig. 2a**). **Fig. 4b**
472 shows the CCN activated fraction (i.e. the ratio of CCN activated at a given
473 supersaturation over the total particle concentration at sizes $D_p > 25 \text{ nm}$) for the two
474 aerosol clusters. For the *Haze* aerosols, the CCN activated fraction ($\text{CCN}/\text{CN}_{25\text{nm}}$)
475 increases from 0.68 to 0.81 when SS increases from 0.1 to 0.4%; thereafter it increases
476 only slowly to 0.85 at a maximum SS of 1.0%. By contrast, the CCN activated fraction of
477 the *Bimodal* aerosols increases from 0.42 to 0.71 with SS increasing from 0.1 to 0.4%, and
478 from 0.71 to 0.88 when SS increases from 0.5% to 1.0%. As the smaller mode accounts
479 for 47% of the *Bimodal* particle number, this implies that these smaller particles must
480 contribute to the CCN concentrations even at relatively low supersaturations. Indeed, **Fig.**
481 **4c** shows the critical activation diameter as calculated by equation 1. The $D_{p\text{crit}}$ of *Haze*
482 accumulation mode aerosols slightly decreases across increasing SS, from 129 nm at SS
483 0.1% to 91 nm at SS 0.35% to 82 nm at SS 1.0%. By contrast, the $D_{p\text{crit}}$ of *Bimodal*
484 aerosols decreases more sharply from 128nm at SS 0.1% to 71 nm at SS 0.35% to as
485 small as 42 nm at SS 1%.
486



487 **Fig. 4.** CCN properties for the *Haze* and *Bimodal* category as function of supersaturation.
 488 (a) Total CCN concentration. (b) Fraction of CCN activated particles larger than 25 nm. (c)
 489 Critical CCN activation diameter Dp_{crit} .
 490

491
 492 Given the different average BC concentrations (22 and 75ng m⁻³, respectively), *Haze* can
 493 be associated mainly with Arctic Haze anthropogenic events, while *Bimodal* being
 494 representative of summer months and mostly of natural imprinting.

495
 496 Aerosol particles smaller than 100 nm in diameter are often considered too small to
 497 activate to cloud droplets. This result comes from the assumption that the cooling
 498 mechanisms are not efficient enough to generate supersaturations required to activate the
 499 smaller particles in liquid clouds, thus the kelvin effect acts as the limiting factor from a
 500 microphysical perspective (Browse et al., 2014; Garrett et al., 2004; Leaitch et al., 2013;
 501 Zhao and Garrett, 2015). However, in the clean environment often found in the Arctic
 502 during summer, the absence of larger particles may lower water uptake rates during
 503 droplet formation, which increases SS, thus enabling smaller particles to become cloud
 504 droplets. In addition, we found relatively low activation diameters for arctic aerosols with
 505 strong abundance in summer months. Our study strongly supports recent findings by
 506 Leaitch et al. (2016) that 20-100 nm particles from natural sources can have a broad
 507 impact on CCN numbers in Arctic environments.

508

509

510 3.3 Summary, Implications and Conclusions

511

512 Mass concentrations of atmospheric aerosols in the Arctic are higher during winter
513 compared to summer due to differences in transport of anthropogenic particles and wet
514 scavenging (Stohl, 2006). By contrast, total aerosol number concentrations in the Arctic
515 are often found similar throughout the period of March-September (Tunved et al., 2013).
516 However, the high concentrations in spring (March–April) are almost exclusively governed
517 by accumulation mode aerosols, while the high summer concentrations are associated
518 with elevated numbers of Aitken mode particles and frequent new particle formation
519 events. So far, differences within the accumulation mode aerosol types over seasons have
520 not been studied, and this was the main objective of our study. Based on k-means cluster
521 analysis of seven years of aerosol number size distributions, we identified eight distribution
522 aerosol size categories. Five were associated with aerosol modes dominated by the
523 ultrafine (<100nm) sizes and are described elsewhere (Dall’Osto et al., 2017b). The
524 remaining three aerosol categories were dominated by the accumulation mode particles
525 (>100 nm) and were named *Haze* (dominant 32% of the time), *Aged* (14%) and *Bimodal*
526 (6%). We found the accumulation mode categories to comprise the ambient aerosol
527 number size distributions at the high Arctic site of Villum research Station more than half of
528 the time.

529

530 Accumulation mode categories presented very distinct chemical and physical properties
531 across seasons. *Haze* accumulation aerosols show a single mode size distribution (150
532 nm), peaking in February-April; *Aged* aerosols show a single mode at 213 nm, peaking in
533 September-October; and *Bimodal* aerosols show two modes at 38 nm and 150 nm,
534 peaking in June-August. A first conclusion that can be drawn from the current study is that
535 a typical accumulation mode does not exist, and profound differences are found especially
536 between the Arctic Haze period and the summer period. Surprisingly the largest Dp
537 accumulation mode is not straightly associated with the Arctic Haze, but it is found during
538 fall and associated with humid conditions. A clear unimodal distribution is never found in
539 accumulation aerosols during the months of July and August (and most of June); rather, it

540 is mostly *Bimodal*, implying that aerosols originating from long-range transport to the Arctic
541 coexist with a smaller mode formed locally to regionally.

542

543 Considerable attention has been given to the role of anthropogenic and biomass burning
544 (BB) particles as warming agents in the Arctic (UNEP, 2011). Black carbon contributes to
545 Arctic warming, yet sources of Arctic BC and their geographic origins remain uncertain (Xu
546 et al., 2017). BC particles and especially aged BC particles affect the radiation budget
547 directly by scattering and absorbing incoming solar radiation (Massling et al., 2015).
548 Sensitivity simulations (Xu et al., 2017) suggest that anthropogenic emissions in eastern
549 and southern Asia have the largest effect on the Arctic BC column burden both in spring
550 (56%) and annually (37%). By investigating the relationship between aerosol categories
551 and black carbon concentration, we have previously demonstrated that pristine clean
552 conditions ($BC < 18 \text{ ng m}^{-3}$) co-occur with ultrafine-sized dominating aerosols (Dall'Osto et
553 al., 2017b). The present analysis of accumulation aerosol categories shows that number
554 size distributions characterized by dominant accumulation modes are associated with
555 average BC concentrations in the range of 27-77 ng m^{-3} , i.e., associated to a varying
556 extent with contributions of anthropogenic pollutants mainly originated from northern
557 Eurasia. Only when lower-end BC concentrations occur over the summer (in the order of
558 27 ng m^{-3}) the *Bimodal* category, characterized by smaller aerosols of biogenic origin,
559 becomes dominant.

560

561 When including chemical components in the analysis, further conclusions can be drawn.
562 Sulfate is the dominant component in the Arctic haze (Massling et al., 2015; Udisti et al.,
563 2016); indeed we found the highest concentrations associated with the *Haze* category.
564 However, sulfate is also produced by the oxidation of dimethyl sulfide (DMS) (Simo, 2001).
565 DMS is of marine origin, and is produced in the upper ocean via interactions of multiple
566 biological processes (Gali and Simo, 2015). During the Arctic summer, the impact of the
567 anthropogenic source is lower (42%), with a contribution comparable to that coming from
568 biogenic emissions (35%), which reaches absolute and percentage values about two times
569 higher than those measured in spring (Udisti et al., 2016). Nevertheless, sulfate
570 concentrations observed at Villum Research Station in summer time are relatively small
571 compared to winter and spring when anthropogenic sources are the main contributor

572 (Heidam et al., 1999). The highest ratio of organics/sulfate and organics/BC among the
573 accumulation mode aerosol categories was found in the *Bimodal* one. It is probable that
574 such organic enhancement is associated with the smaller mode (38 nm). The origin of this
575 mode is likely to be a combination of secondary aerosol formation of marine biological
576 origin from the open waters between ice floes (Dall'Osto et al., 2017a, b), but also from
577 fragmentation and/or dispersion of primary marine polymer gels also originated in water
578 adjacent to the ice (Leck and Bigg, 2005; Orellana et al., 2011).

579

580 Cloud condensation nuclei are a functionally important fraction of the atmospheric aerosol,
581 because they influence cloud microphysical and radiative properties, and consequently the
582 aerosol indirect radiative forcing (IPCC, 2014). Low level clouds are one of the major
583 players controlling the radiative balance in the Arctic. At the most fundamental level,
584 understanding the processes that determine cloud properties from microscale to global
585 scale requires information of which particles actually form cloud droplets under various
586 conditions. The effect of the background aerosol on liquid clouds has been identified as
587 one of the most important factors for reducing uncertainty in the aerosol cloud albedo
588 effect (Carslaw et al., 2013). Moreover, the effectiveness of particles smaller than 100 nm
589 for cloud droplet nucleation is a large factor in that uncertainty. During summer the Arctic is
590 thought to be relatively free of anthropogenic influence, which means that dominantly only
591 particles from natural sources determine cloud droplet formation. This study shows that,
592 despite anthropogenic influence is maintained moderate through most of the summer,
593 natural sources have indeed a significant impact on particle number, and on facilitating
594 aerosol activation to cloud droplets and thus cloud formation. Further integrated studies
595 with joint multi-component observations are warranted.

596

597

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599

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626 **References**

- 627
628 AMAP (2011), AMAP: Snow, Water, Ice and Permafrost in the Arctic (SWIPA), Climate Change
629 and the Cryosphere, 538 pp, OSLO.
- 630 Anttila, T., D. Brus, A. Jaatinen, A. P. Hyvarinen, N. Kivekas, S. Romakkaniemi, M. Komppula,
631 and H. Lihavainen (2012), Relationships between particles, cloud condensation nuclei
632 and cloud droplet activation during the third Pallas Cloud Experiment, *Atmospheric
633 Chemistry and Physics*, 12(23), 11435-11450, doi:10.5194/acp-12-11435-2012.
- 634 Asmi, E., et al. (2016), Aerosol size distribution seasonal characteristics measured in Tiksi, Russian
635 Arctic, *Atmospheric Chemistry and Physics*, 16(3), 1271-1287, doi:10.5194/acp-16-
636 1271-2016.
- 637 Beddows, D. C. S., M. Dall'Osto, and R. M. Harrison (2009), Cluster Analysis of Rural, Urban, and
638 Curbside Atmospheric Particle Size Data, *Environmental Science & Technology*,
639 43(13), 4694-4700, doi:10.1021/es803121t.
- 640 Browse, J., K. S. Carslaw, S. R. Arnold, K. Pringle, and O. Boucher (2012), The scavenging
641 processes controlling the seasonal cycle in Arctic sulphate and black carbon aerosol,
642 *Atmospheric Chemistry and Physics*, 12(15), 6775-6798, doi:10.5194/acp-12-6775-
643 2012.

644 Browse, J., K. S. Carslaw, G. W. Mann, C. E. Birch, S. R. Arnold, and C. Leck (2014), The
645 complex response of Arctic aerosol to sea-ice retreat, *Atmospheric Chemistry and*
646 *Physics*, *14*(14), 7543-7557, doi:10.5194/acp-14-7543-2014.

647 Carslaw, K. S., et al. (2013), Large contribution of natural aerosols to uncertainty in indirect
648 forcing, *Nature*, *503*(7474), 67-71, doi:10.1038/nature12674.

649 Chin, W. C., M. V. Orellana, and P. Verdugo (1998), Spontaneous assembly of marine dissolved
650 organic matter into polymer gels, *Nature*, *391*(6667), 568-572.

651 Clegg, S. L., K. S. Pitzer, and P. Brimblecombe (1992), Thermodynamics of Multicomponent,
652 Miscible, Ionic-Solutions .2. Mixtures Including Unsymmetrical Electrolytes, *Journal*
653 *of Physical Chemistry*, *96*(23), 9470-9479, doi:10.1021/j100202a074.

654 Croft, B., R. V. Martin, W. R. Leitch, P. Tunved, T. J. Breider, S. D. D'Andrea, and J. R. Pierce
655 (2016), Processes controlling the annual cycle of Arctic aerosol number and size
656 distributions, *Atmospheric Chemistry and Physics*, *16*(6), 3665-3682.

657 Dall'Osto, M., X. Querol, A. Alastuey, C. O'Dowd, R. M. Harrison, J. Wenger, and F. J. Gomez-
658 Moreno (2013), On the spatial distribution and evolution of ultrafine particles in
659 Barcelona, *Atmospheric Chemistry and Physics*, *13*(2), 741-759, doi:10.5194/acp-13-
660 741-2013.

661 Dall'Osto, M., et al. (2017a), Arctic sea ice melt leads to atmospheric new particle formation,
662 *Scientific Reports*, *7*(1), 3318, doi:10.1038/s41598-017-03328-1.

663 Dall'Osto, M., R. Lange, C. Geels, D. C. S. Beddows, R. M. Harrison, R. Simo, D. Boertmann, H.
664 Skov, and A. Massling (2017b), Open pack ice drives new particle formation in North
665 East Greenland, *Scientific Reports*, *Under review*.

666 Dall'Osto, M., et al. (2017c), Antarctic sea ice region as a source of biogenic organic nitrogen in
667 aerosols, *Scientific Reports*, *7*, doi:10.1038/s41598-017-06188-x.

668 Dusek, U., et al. (2006), Size matters more than chemistry for cloud-nucleating ability of aerosol
669 particles, *Science*, *312*(5778), 1375-1378, doi:10.1126/science.1125261.

670 Engvall, A. C., R. Krejci, J. Strom, R. Treffeisen, R. Scheele, O. Hermansen, and J. Paatero (2008),
671 Changes in aerosol properties during spring-summer period in the Arctic troposphere,
672 *Atmospheric Chemistry and Physics*, *8*(3), 445-462.

673 Freud, E., R. Krejci, P. Tunved, R. Leitch, Q. T. Nguyen, A. Massling, H. Skov, and L. Barrie
674 (2017), Pan-Arctic aerosol number size distributions: seasonality and transport
675 patterns, *Atmospheric Chemistry and Physics*, *17*(13), 8101-8128, doi:10.5194/acp-
676 17-8101-2017.

677 Gali, M., and R. Simo (2015), A meta-analysis of oceanic DMS and DMSP cycling processes:
678 Disentangling the summer paradox, *Global Biogeochemical Cycles*, *29*(4), 496-515,
679 doi:10.1002/2014gb004940.

680 Garrett, T. J., C. Zhao, X. Dong, G. G. Mace, and P. V. Hobbs (2004), Effects of varying aerosol
681 regimes on low-level Arctic stratus, *Geophysical Research Letters*, *31*(17),
682 doi:10.1029/2004gl019928.

683 Heidam, N. Z., J. Christensen, P. Wahlin, and H. Skov (2004), Arctic atmospheric contaminants in
684 NE Greenland: levels, variations, origins, transport, transformations and trends 1990-
685 2001, *Science of the Total Environment*, *331*(1-3), 5-28,
686 doi:10.1016/j.scitotenv.2004.03.033.

687 Heidam, N. Z., P. Wahlin, and J. H. Christensen (1999), Tropospheric gases and aerosols in
688 northeast Greenland, *Journal of the Atmospheric Sciences*, *56*(2), 261-278,
689 doi:10.1175/1520-0469(1999)056<0261:Tgaain>2.0.Co;2.

690 IPCC (2014), Climate Change 2014: Impacts, Adaptation, and vulnerability.

691 Iziomon, M. G., U. Lohmann, and P. K. Quinn (2006), Summertime pollution events in the Arctic
692 and potential implications, *Journal of Geophysical Research-Atmospheres*, *111*(D12),
693 doi:10.1029/2005jd006223.

694 Jurányi, Z., M. Gysel, E. Weingartner, P. F. DeCarlo, L. Kammermann and U. Baltensperger
695 (2010), Measured and modelled cloud condensation nuclei number concentration at
696 the high alpine site Jungfraujoch, *Atmospheric Chemistry and Physics*, *10*(16), 7891-
697 7906, doi: 10.5194/acp-10-7891-2010

698 Kohler, H. (1936), The nucleus in and the growth of hygroscopic droplets., *Transactions of the*
699 *Faraday Society*, *32*(2), 1152-1161, doi:10.1039/tf9363201152.

700 Kristensen, T. B., T. Müller, K. Kandler, N. Benker, M. Hartmann, J. M. Prospero, A.
701 Wiedensohler and F. Stratmann (2015), Properties of cloud condensation nuclei
702 (CCN) in the trade wind marine boundary layer of the western North Atlantic,
703 *Atmospheric Chemistry and Physics*, *16*(4), 2675-2688, doi: 10.5194/acp-16-2675-
704 2016

705 Latham, T. L., A. J. Beyersdorf, K. L. Thornhill, E. L. Winstead, M. J. Cubison, A. Hecobian, J. L.
706 Jimenez, R. J. Weber, B. E. Anderson, and A. Nenes (2013), Analysis of CCN activity
707 of Arctic aerosol and Canadian biomass burning during summer 2008, *Atmospheric*
708 *Chemistry and Physics*, *13*(5), 2735-2756, doi:10.5194/acp-13-2735-2013.

709 Law, K. S., and A. Stohl (2007), Arctic air pollution: Origins and impacts, *Science*, *315*(5818),
710 1537-1540, doi:10.1126/science.1137695.

711 Leaitch, W. R., L. Huang, A. M. Macdonald, S. Sharma, D. Toom-Sauntry, K. von Salzen, and J. R.
712 Pierce (2013), A Comparison of Measurements and Global Model Simulations of the
713 Atmospheric Aerosol at Two Remote Sites, *Nucleation and Atmospheric Aerosols*,
714 *1527*, 511-514, doi:10.1063/1.4803318.

715 Leaitch, W. R., et al. (2016), Effects of 20-100nm particles on liquid clouds in the clean
716 summertime Arctic, *Atmospheric Chemistry and Physics*, *16*(17), 11107-11124,
717 doi:10.5194/acp-16-11107-2016.

718 Leck, C., and E. K. Bigg (2005), Biogenic particles on the surface microlayer and overlaying
719 atmosphere in the central Arctic Ocean during summer, *Tellus Series B-Chemical and*
720 *Physical Meteorology*, *57B*, 305-316.

721 Massling, A., I. E. Nielsen, D. Kristensen, J. H. Christensen, L. L. Sorensen, B. Jensen, Q. T.
722 Nguyen, J. K. Nojgaard, M. Glasius, and H. Skov (2015), Atmospheric black carbon
723 and sulfate concentrations in Northeast Greenland, *Atmospheric Chemistry and*
724 *Physics*, *15*(16), 9681-9692, doi:10.5194/acp-15-9681-2015.

725 Mauritsen, T., et al. (2011), An Arctic CCN-limited cloud-aerosol regime, *Atmospheric Chemistry*
726 *and Physics*, *11*(1), 165-173, doi:10.5194/acp-11-165-2011.

727 Nguyen, Q. T., M. Glasius, L. L. Sorensen, B. Jensen, H. Skov, W. Birmili, A. Wiedensohler, A.
728 Kristensson, J. K. Nojgaard, and A. Massling (2016), Seasonal variation of
729 atmospheric particle number concentrations, new particle formation and atmospheric
730 oxidation capacity at the high Arctic site Villum Research Station, Station Nord,
731 *Atmospheric Chemistry and Physics*, *16*(17), 11319-11336, doi:10.5194/acp-16-
732 11319-2016.

733 Nguyen, Q. T., H. Skov, L. L. Sorensen, B. J. Jensen, A. G. Grube, A. Massling, M. Glasius, and J.
734 K. Nojgaard (2013), Source apportionment of particles at Station Nord, North East
735 Greenland during 2008-2010 using COPREM and PMF analysis, *Atmospheric*
736 *Chemistry and Physics*, *13*(1), 35-49, doi:10.5194/acp-13-35-2013.

737 O'Neill, N. T., O. Pancrati, K. Baibakov, E. Eloranta, R. L. Batchelor, J. Freemantle, L. J. B.
738 McArthur, K. Strong, and R. Lindenmaier (2008), Occurrence of weak, sub-micron,

739 tropospheric aerosol events at high Arctic latitudes, *Geophysical Research Letters*,
740 35(14), doi:10.1029/2008gl033733.

741 Orellana, M. V., P. A. Matrai, C. Leck, C. D. Rauschenberg, A. M. Lee, and E. Coz (2011), Marine
742 microgels as a source of cloud condensation nuclei in the high Arctic, *Proceedings of
743 the National Academy of Sciences of the United States of America*, 108(33), 13612-
744 13617, doi:10.1073/pnas.1102457108.

745 Petzold, A., and M. Schonlinner (2004), Multi-angle absorption photometry - a new method for the
746 measurement of aerosol light absorption and atmospheric black carbon, *Journal of
747 Aerosol Science*, 35(4), 421-441, doi:10.1016/j.jaerosci.2003.09.005.

748 Pfeifer, S., W. Birmili, A. Schladitz, T. Muller, A. Nowak, and A. Wiedensohler (2014), A fast and
749 easy-to-implement inversion algorithm for mobility particle size spectrometers
750 considering particle number size distribution information outside of the detection
751 range, *Atmospheric Measurement Techniques*, 7(1), 95-105, doi:10.5194/amt-7-95-
752 2014.

753 Ramanathan, V., P. J. Crutzen, J. T. Kiehl, and D. Rosenfeld (2001), Atmosphere - Aerosols,
754 climate, and the hydrological cycle, *Science*, 294(5549), 2119-2124,
755 doi:10.1126/science.1064034.

756 Rose, D., A. Nowak, P. Achtert, A. Wiedensohler, M. Hu, M. Shao, Y. Zhang, M. O. Andreae, and
757 U. Poschl (2010), Cloud condensation nuclei in polluted air and biomass burning
758 smoke near the mega-city Guangzhou, China - Part 1: Size-resolved measurements
759 and implications for the modeling of aerosol particle hygroscopicity and CCN activity,
760 *Atmospheric Chemistry and Physics*, 10(7), 3365-3383.

761 Rosenfeld, D. (2006), Atmosphere - Aerosols, clouds, and climate, *Science*, 312(5778), 1323-1324,
762 doi:10.1126/science.1128972.

763 Simo, R. (2001), Production of atmospheric sulfur by oceanic plankton: biogeochemical, ecological
764 and evolutionary links, *Trends in Ecology & Evolution*, 16(6), 287-294,
765 doi:10.1016/S0169-5347(01)02152-8.

766 Stohl, A. (2006), Characteristics of atmospheric transport into the Arctic troposphere, *Journal of
767 Geophysical Research-Atmospheres*, 111(D11), doi:10.1029/2005jd006888.

768 Tunved, P., J. Strom, and R. Krejci (2013), Arctic aerosol life cycle: linking aerosol size
769 distributions observed between 2000 and 2010 with air mass transport and
770 precipitation at Zeppelin station, Ny-Alesund, Svalbard, *Atmospheric Chemistry and
771 Physics*, 13(7), 3643-3660, doi:10.5194/acp-13-3643-2013.

772 Udisti, R., et al. (2016), Sulfate source apportionment in the Ny-Alesund (Svalbard Islands) Arctic
773 aerosol, *Rendiconti Lincei-Scienze Fisiche E Naturali*, 27, 85-94, doi:10.1007/s12210-
774 016-0517-7.

775 UNEP (2011), United Nations Environment Programme: Near-term Climate Protection and Clean
776 Air Benefits: Actions for Controlling Short-Lived Climate Forcers, available at:
777 <http://www.unep.org/publications/ebooks/SLCF/> (last access: 29 August 2016),
778 Nairobi, Kenya.

779 Wexler, A. S., and S. L. Clegg (2002), Atmospheric aerosol models for systems including the ions
780 H⁺, NH₄⁺, Na⁺, SO₄²⁻, NO₃⁻, Cl⁻, Br⁻, and H₂O, *Journal of Geophysical Research-
781 Atmospheres*, 107(D14), doi:10.1029/2001jd000451.

782 Wiedensohler, A., et al. (2012), Mobility particle size spectrometers: harmonization of technical
783 standards and data structure to facilitate high quality long-term observations of
784 atmospheric particle number size distributions, *Atmospheric Measurement
785 Techniques*, 5(3), 657-685, doi:10.5194/amt-5-657-2012.

- 786 Xu, J. W., et al. (2017), Source attribution of Arctic black carbon constrained by aircraft and surface
787 measurements, *Atmospheric Chemistry and Physics*, 17(19), 11971-11989,
788 doi:10.5194/acp-17-11971-2017.
- 789 Zhao, C. F., and T. J. Garrett (2015), Effects of Arctic haze on surface cloud radiative forcing,
790 *Geophysical Research Letters*, 42(2), 557-564, doi:10.1002/2014gl062015.