

Drivers governing the seasonality of new particle formation in the Arctic

Dominic Heslin-Rees^{1,2}, Peter Tunved^{1,2}, Diego Aliaga¹, Janne Lampilahti³, Ilona Riipinen^{1,2}, Annica M. L. Ekman^{4,2}, Ki-Tae Park⁵, Martina Mazzini⁶, Stefania Gilardoni⁷, Roseline Thakur³, Kihong Park⁸, Young Jun Yoon⁹, Kitack Lee¹⁰, Mikko Sipilä³, Mauro Mazzola¹¹, Radovan Krejci^{1,2}

¹Department of Environmental Science (ACES), Stockholm, Sweden.

²Bolin Centre for Climate Research, Stockholm University, Stockholm, Sweden

³Institute for Atmospheric and Earth System Research (INAR), Helsinki, Finland.

⁴Department of Meteorology, Stockholm, Sweden.

⁵Hallym University, Chuncheon, Gangwon, Republic of Korea

⁶Institute of Atmospheric Science and Climate (CNR-ISAC), Bologna, Italy.

⁷Institute of Polar Sciences (CNR-ISP), Milan, Italy.

⁸School of Earth Sciences and Environmental Engineering, Gwangju Institute of Science and Technology, Buk-gu, Gwangju, Republic of Korea

⁹Korea Polar Research Institute, Yeosu-gu, Incheon, Republic of Korea

¹⁰Division of Environmental Science and Engineering, Pohang University of Science and Technology, Pohang, Republic of Korea

¹¹Institute of Polar Sciences, National Research Council (CNR-ISP), Bologna, Italy

Correspondence to: Dominic Heslin-Rees (dominic.heslin-rees@aces.su.se) and Radovan Krejci (radovan.krejci@aces.su.se)

45 **Abstract**

New particle formation (NPF) is the phenomenon wherein gaseous precursors form critical clusters of barely a few nanometres in diameter, after which, under favourable conditions these particles can grow to climate-relevant sizes. Here we present measurements from 2022 to 2024 of particle and ion number size distributions from the Zeppelin Observatory (ZEP), an Arctic research station situated on the western edge of Svalbard atop a mountain. NPF events begin in April and continue occurring into November. The events at the start of the NPF season (i.e. April/May) are considerably stronger (i.e. a larger production of nucleation mode particles) than compared with other months of the year. The peaks in NPF strength coincide with peaks in the accumulated solar insolation experienced by arriving air masses. During the summer period NPF events occur on 20-40% of days each month, however, there is a consistent decline starting in June. We show that the combined influence of solar insolation and the surface area of pre-existing aerosols (i.e. condensation sink, CS) are strong predictors for the likelihood of NPF. We develop a simplified predictive model which matches the frequency of NPF events identified via the classification schemes used in this study: we show that the ratio of solar insolation over CS corresponds well to the frequency of NPF events (R^2 of 0.78). We show that NPF events occur during the polar night (i.e. when the Sun does not pass above horizon), and that these events are linked to high altitude air masses. Furthermore, we detail the likely geographic origins of nucleation mode particles, as measured at ZEP. We show that NPF events are considerably more likely to originate from marine regions towards the west of Svalbard, particularly the Greenland Sea, which is the marine region with the greatest likelihood of originating air masses linked to an NPF day. We also show that NPF events lead to an increase in the number of Aitken mode particles, indicating that a significant proportion of the Aitken mode particles originate from NPF. Of the measured NPF events, 37% of them exhibited nucleation mode particles that grew beyond 25nm, a diameter representing the minimum activation diameter for particles to act as cloud condensation nuclei. Overall, we present a concise picture of the lifecycle of nucleation mode particles in the Arctic, including the effect wet scavenging has in reducing the condensation sink, which in turn encourages NPF events to occur.

Deleted: radiation

Deleted: .

Deleted: W

Formatted: Superscript

Deleted: speculate

Deleted: within the Arctic,

Deleted: the

Deleted: the

Deleted: presented th

Deleted: e

Deleted: that arriving air masses

Deleted: from this marine region would be

Deleted: We also remark on the proportion of the Aitken mode particles within the Arctic that could originate from NPF; w

Deleted: .

Deleted: We measure over 50

Deleted: where the

1. Introduction

New particle formation (NPF) describes the production of secondary aerosol particles from precursor gases to clusters of molecules a few nanometres in diameter. Newly formed particles belong to the nucleation mode (i.e. typically <10nm) and may grow through the continued coagulation and condensation of vapours into Aitken mode particles (i.e. tens of nanometres in diameter). The phenomenon of NPF has been observed globally (Kerminen et al., 2018), however, in each region the chemical characteristics and controlling source and sink processes leading to NPF may vary.

Deleted: (Kerminen et al., 2018)

Arctic aerosol seasonality is governed by a combination of atmospheric general circulation, meteorology, and several source and sink processes including the direct emission of local primary particles, NPF, the long-range transport of anthropogenic emissions from lower latitudes, and the removal of particles through wet scavenging (Garrett et al., 2011; Schmale et al., 2022; Tunved et al., 2013; Willis et al., 2018). One distinctive feature of the annual cycle is the Arctic haze in late winter and early spring, which is characterised by a sustained increase in the number of accumulation mode aerosol (i.e. particles typically around 100nm) (e.g. Shaw (1995)). Reduced wet scavenging and increased transport efficiency from anthropogenic source regions at lower latitudes during this part of the year give rise to this phenomenon (Garrett et al., 2011). The Arctic summertime, by contrast, is less influenced by long-range transported pollutants, and the increased precipitation effectively reduces the once pronounced accumulation mode. Moreover, increased photochemistry and biological activity (e.g. phytoplankton blooms) provide a source of precursor gases encouraging NPF and the growth of newly formed particles (Tunved, Ström and Krejci, 2013; Croft et al., 2016; Price et al., 2023); as a result, the smaller Aitken and ultrafine aerosol particles increase in number.

Deleted: (Garrett et al., 2011; Schmale et al., 2022; Tunved et al., 2013; Willis et al., 2018)

Deleted: Shaw (1995)

Deleted: (Garrett et al., 2011)

Deleted: Tunved, Ström and Krejci, 2013; Croft et al., 2016; Price et al., 2023)

Changes in aerosol size, abundance and chemical composition can perturb the radiative balance either indirectly by acting as cloud condensation nuclei (CCN), altering cloud formation, brightness, and lifetime or directly through their interaction with both short- and long-wave radiation (Li et al., 2022). The Arctic region can at times be CCN-limited, which means that small changes in CCN concentrations could potentially impact cloud properties via cloud-mediated indirect aerosol effects, which in this and similar environments could result in surface warming due to enhanced long-wave cloud forcing (Garrett et al., 2004; Garrett and Zhao, 2006; Mauritsen et al., 2011). Thus, local-regional scale production of aerosols during the summertime and changes to this production mechanism may have significant consequences for the Arctic (Kecorius et al., 2019). Still however, the impact of newly formed particles on Arctic cloud properties remains poorly quantified.

Formatted: Font: 10 pt, Font color: Black

Deleted: (Carslaw, 2022).

Deleted: (Garrett et al., 2004; Garrett and Zhao, 2006; Mauritsen et al., 2011)

Deleted: (Kecorius et al., 2019)

Depending on certain environmental factors, for example the rate of particle nucleation, the production rate and volatility of condensable vapours, and the condensation and coagulation sinks, to mention a few, these newly formed particles can reach climate-relevant sizes; In the Arctic, studies have suggested that particles as small as 25nm can participate in cloud formation by acting as CCN (Leaitch et al., 2016; Karlsson et al., 2020, 2021; Pöhlker et al., 2021; Gramlich et al., 2023; Motos et al., 2023), thus affecting cloud properties and climate (i.e., by influencing the radiative balance). Motos et al. (2023), predicted the smallest particle activation diameters in the summertime to be 20nm, with a modal average of 60 nm, but suggested that the dominant summertime Aitken mode originating from NPF may stay in the interstitial (inactivated) phase. Other studies have indicated that a substantial fraction of CCN in the Arctic are produced via NPF (Gordon et al., 2017; Merikanto et al., 2009).

Air ions are an additional focus area of aerosol effects, as they can play a role in nucleation (Hirsikko et al., 2011; Kirkby et al., 2011), and can influence aerosol particles through their formation and growth mechanism. Both positive and negative ions contribute to the production of newly formed particles in the Arctic, via ion-induced nucleation (see Beck et al., 2021 for negative ions, and Jokinen et al. (2018) for positive ions). The overall contribution of ions to NPF is disputed as quantifying their impact on nucleation rates in ambient conditions remains difficult. However, studies suggest that the dynamics of sub-2nm clusters is dominated by neutral clusters (Hirsikko et al., 2011; Kulmala et al., 2013). Ions can help stabilise clusters which form initially (Yu and Turco, 2001), with most particles formed from ion-induced nucleation reaching neutrality after growing beyond 2.5nm (Wagner et al., 2017).

Observations of ultrafine particles (~3-20nm) in the Arctic (onboard the Swedish icebreaker Oden, 70°N-85°N) were reported by Covert et al. (1996), and Wiedensohler et al. (1996). At Zeppelin Observatory (ZEP), Ström et al. (2003), and Tunved et al. (2013), presented measurements of small (>10nm) aerosol particles showing an increased concentration and frequency of formation events during the Arctic summer. Tunved et al. (2013), suggested that the summertime, coinciding with increased photochemical production of precursor gases, and the removal of larger pre-existing aerosol, generates favourable conditions in which NPF can occur. Other Arctic sites have been the focus of research into NPF, including Villum Research Station (hereafter Villum) (see Nguyen et al. (2016)) and Summit (see Ziemba et al. (2010)), both on Greenland, and Utqiagvik, Alaska (Kolesar et al., 2017a). Other studies, e.g. Brean et al. (2023), present NPF data from several locations, including Alert, Villum, Tiksi, ZEP, Gruebadet (Ny-Ålesund), and Utqiagvik; a commonality shared by all Arctic sites was the occurrence of NPF events during the summer, and also that events coincided with air masses generally influenced by marine regions, including at Utqiagvik, and Alert whereby a correlation was observed between biogenic methane

Deleted: (Leaitch et al., 2016; Karlsson et al., 2020, 2021; Pöhlker et al., 2021; Gramlich et al., 2023; Motos et al., 2023)

Deleted: Motos et al. (2023)

Deleted: (Gordon et al., 2017; Merikanto et al., 2009)

Deleted: (Hirsikko et al., 2011; Kirkby et al., 2011)

Deleted: both

Deleted: (see Beck et al., 2021 for negative ions)

Deleted: .

Deleted: (Hirsikko et al., 2011; Kulmala et al., 2013)

Deleted: (Yu and Turco, 2001)

Deleted: (Wagner et al., 2017)

Deleted: Covert et al. (1996)

Deleted: Wiedensohler et al. (1996)

Deleted: Ström et al. (2003)

Deleted: Tunved et al. (2013)

Deleted: Tunved et al. (2013)

Deleted: Observatory

Deleted: Nguyen et al. (2016)

Deleted: Ziemba et al. (2010)

Deleted: Barrow

Deleted: (Kolesar et al., 2017a)

Deleted: Brean et al. (2023)

Deleted: Barrow

Deleted: Barrow

suphonate (MSA-) and summertime particle number concentrations (Leaitch et al., 2013; Quinn et al., 2002). Beck et al. (2021), through a comparison of measurements from both Ny-Ålesund and Villum, showed that different nucleation mechanisms can also occur at different times of the year at these two locations; at Villum iodic acid (IA, HIO_3) was found to be the primary driver of NPF events during springtime, and sulphuric acid (SA, H_2SO_4) -ammonium (AM, NH_4^+) driven events were shown to occur in the summer. At Ny-Ålesund, NPF was driven by SA-AM during the springtime and highly oxygenated molecules (HOMs) during the summer. There is no evidence that anthropogenic precursors contribute to NPF at ZEP (Schmale and Baccarini, 2021). Instead, Dimethyl Sulphide (DMS), produced by phytoplankton, can serve as a precursor gas contributing to NPF. Marine regions with strong chlorophyll a concentrations and DMS production capacity have been linked to increased concentrations of nanoparticles, which suggests that marine biogenic sources are an important source of nucleating and condensing material (Lee et al., 2020). Observed DMS concentrations vary with airmass origin, season and residence time over different marine source regions. For example, it has been shown that air masses exposed more to the Greenland Sea, compared with the Barents Sea bring about higher DMS concentrations (Park et al., 2018). In the Arctic, there are low concentrations of nucleating agents as opposed to continental rural and urban locations (Karl et al., 2012; Pirjola et al., 2000). However, in spring and summer increased solar radiation, a lower total surface area of pre-existing particles (i.e. reduced condensation sink, CS), and coinciding phytoplankton blooms are shown to give rise to precursor gas concentrations sufficient to sustain ion-induced nucleation, and growth of NPF-induced particles beyond 20nm (Beck et al., 2021).

In this study, we present two and a half years' worth of measurements of particle and ion number size distributions, including three summer periods (in total data covers the period April 2022 – October 2024). Continuous measurements of sub-3nm up to ~850nm particles were performed at the ZEP, Svalbard, providing unique information of the initial stages of nucleation and the subsequent growth. NPF events were identified, formation rates were calculated, and subsequent growth was analysed. The seasonality and coinciding environmental parameters encouraging NPF events and subsequent growth have been explored in detail.

2. Materials and methods

2.1 Measurement site

Zeppelin observatory (ZEP) (78.90°N, 11.88°E, 474 m a.s.l.) is situated on the ridge of Mount Zeppelin, 2 km south from Ny-Ålesund research village on the western edge of the Norwegian Svalbard archipelago. ZEP represents regional background conditions in the Atlantic sector of the Arctic, as it is largely unaffected by local

Deleted: (Leaitch et al., 2013; Quinn et al., 2002)

Deleted: Beck et al. (2021)

Deleted:

Formatted: Subscript

Formatted: Subscript

Formatted: Subscript

Formatted: Superscript

Deleted: , whilst a

Formatted: Subscript

Deleted: sulphuric acid (SA) and ammonia

Deleted: (Schmale and Baccarini, 2021)

Deleted:

Deleted:

Deleted: Strong correlations between measured concentrations of DMS and the exposure of air masses to chlorophyll a have been presented

Deleted: (Lee et al., 2020)

Deleted: , which suggest marine DMS emissions are an important source of nucleating and condensing material

Deleted: (Park et al., 2018)

Deleted: (Karl et al., 2012; Pirjola et al., 2000)

Deleted: (Beck et al., 2021)

Deleted: ,

Deleted: whereby c

Deleted: have been

Deleted: from

Deleted: up to the higher end of the accumulation mode

235 air pollution due to its location well above Ny-Ålesund and the prevailing wind patterns driven by Kongsfjorden's
geography. ZEP is one of the most developed atmospheric observational sites in the Arctic where a broad set of
parameters has been measured for decades. It is part of numerous regional and global monitoring networks,
including the Global Atmosphere Watch (WMO/GAW), Integrated Carbon Observation System (ICOS), The
Aerosol, Clouds, and Trace Gases Research Infrastructure (ACTRIS) and the co-operative programme for
monitoring and evaluation of the long-range transmission of air pollutants in Europe (EMEP). For a detailed
240 description of the observatory, local meteorology and climatology, history and observational program see [Platt et
al. \(2022\)](#).

Deleted: Platt et al. (2022)

2.2 Instrumentation

For this study, an array of instruments has been used to measure aerosol and ion size distributions, between 0.8
and 850 nm including a Neutral Cluster and Air Ion Spectrometer (NAIS5, Airl Ltd. Tartu, Estonia), a
245 Nanoparticle Scanning Mobility Particle Sizer (Nano SMPS, TSI), and a Differential Mobility Particle Sizer
(DMPS) system. Particle and ion number size distributions and concentrations form the core data used in this
study. Several other supporting data sets were used for data analysis and interpretation and are described below.
This study covers the period from 2022-04-17 to 2024-10-04.

2.2.1 Differential Mobility Particle Sizer (DMPS) system

250 Particle number size distributions, from 5 to 850nm, were measured using a custom-made twin-DMPS system
with a closed loop sheath circulation and composed of two Hauke-type Differential Mobility Analysers (DMA).
DMPS-1, with a short (5.3 cm) DMA, measured aerosol size distribution from 5 to 57 nm using sample-to-sheath
air ratio close to 1:10. DMPS-2, with medium length DMA (28 cm), measured size distribution from 20 to 850 nm
using sample-to-sheath air ratio close to 1:5. The aerosol particles were counted using condensational particle
255 counters (CPCs) at a frequency of 1 Hz with the two CPCs. Particles > 3 nm were measured using the Ultrafine
CPC (UCPC) TSI model 3776 (cut-off of 2.5nm) [coupled to the DMPS-1](#). Particles > 10 nm [were measured](#) with
the CPC TSI model 3010 (cut-off of 10nm) [coupled to the DMPS-2](#). The two DMAs were run in a stepwise mode,
providing two particle number size distribution cycles approximately every 27 minutes. The overlap between the
size ranges measured by both DMPSs was used to check consistency and to merge both size distributions into one.
260 Additional information can be found in [Karlsson et al. \(2021\)](#). The DMPS measurements were corrected for
diffusional losses using the Particle Loss Calculator (PLC) according to [Von Der Weiden et al. \(2009\)](#) (see Fig.
S1 in the supplement for details).

Deleted: ; a TSI model 3010 was coupled to the DMPS-1
and a TSI model 3772 was coupled to the DMPS-2. The total
aerosol concentration was measured

Deleted: and p

Deleted:

Deleted: Karlsson et al. (2021)

Deleted: Von Der Weiden et al. (2009)

2.2.2 Neutral Cluster and Air Ion Spectrometer (NAIS)

The NAIS was utilised to measure concentrations of ions (charged particles and cluster ions) of both polarities between 0.8 and 40 nm, and particles between 2.5 and 40 nm. The NAIS is an aerosol mobility spectrometer; when the sample is left unmodified it detects naturally-charged ions and particles, and when the corona charger is used to charge the particle population it measures all particles including the uncharged (Manninen et al., 2011; Mirme et al., 2007; Mirme and Mirme, 2013). It consists of two mobility analyser columns, with a total of 25 electrometers per column. Its lower size thresholds and high temporal resolution make it highly suitable for observing the early stages of NPF. The NAIS provides very accurate size information, regardless of the choice of the inversion algorithm used (in this case, the 25-channel V14.1 inverter algorithm) (Wagner et al., 2016). The lowest detection limit for the NAIS in particle mode is set based on the diameter of the corona charger ions, and the inability of the electrical filter to remove all naturally charged particles (Manninen et al., 2011), however, there is a broad consensus that measurements of concentrations of particles below 2.5nm are inaccurate.

The NAIS at ZEP was installed with a very short metallic inlet, less than 1m long, to minimise diffusional losses. The inlet pointed slightly downward, and a metallic rain shield was attached at the end, to ensure that rain did not enter into the instrument. The NAIS inlet was heated to prevent ice blocking the inlet, and also to limit water condensing inside the tubing. The NAIS was serviced and cleaned once a year. Additional aspects of the NAIS observations, including the apparent overestimation of particle number concentrations when compared to DMPS measurements (see also Kangasluoma et al., 2020), and the loss of negative cluster ions can be found in section S2.6 of the supplementary material.

2.2.3 Complementary observations

A Nanoparticle Scanning Mobility Particle Sizer (Nano SMPS, TSI) comprising of a nano-differential mobility analyser (nano-DMA, TSI 3085, USA) and an ultrafine CPC (TSI 3776, USA), provided size distributions of nanoparticles (3–60nm). See Lee et al. (2020), for further details. Atmospheric Dimethyl Sulphide (DMS) was measured using an analytical system in which DMS is trapped and eluted, before being quantified using gas chromatography equipped with a pulsed-flame photometric detector. The detection limit is approximately 1.5 pptv given air samples of around 6 litres. For more details, see Jang et al. (2016), and Park et al. (2018). Meteorological parameters including visibility, relative humidity, wind speed and direction, ambient temperature and pressure were also further utilised throughout the study.

Deleted: (Manninen et al., 2011; Mirme et al., 2007; Mirme and Mirme, 2013)

Deleted: (Wagner et al., 2016)

Deleted: (Manninen et al., 2011)

Deleted: (see also Kangasluoma et al., 2020)

Deleted: Lee et al. (2020)

Deleted: Jang et al. (2016)

Deleted: Park et al. (2018).

2.3 Data Analysis of aerosol measurements

For processing of the NAIS data the *nais-processor* was used (<https://github.com/jlpl/nais-processor>). The *nais-processor* Python package corrected for diffusional losses in the inlet (Gormley and Kennedy, 1948), applied an ion mode calibration (Wagner et al., 2016), converted ion mobility to particle diameter, remapped the distribution to a new size grid, and adjusted the data to standard conditions (273.15 K, 101325 Pa).

Further information on the removal of artefacts can be found in Sect. S2 of the supplementary material. In addition, a comparison between the particle concentrations measured by NAIS and the DMPS-system can be found in Sect. S2.5; we report a significant overestimation of particle number concentrations as measured by the NAIS in comparison to the DMPS-system, particularly at the smallest diameters (see Fig. S9-S11). It should be noted that the particle number size distributions presented in this study represent that of dry aerosol particle sizes. The measured particle size distributions for both the NAIS and DMPS were not corrected for hygroscopic growth, and thus do not reflect ambient conditions.

When assessing the qualitative and quantitative nature of NPF events, it is important to distinguish between on the one hand the production of newly formed particles (~2-4nm) and on the other hand subsequent particle growth. To qualify as an event, not only does formation of ultrafine particles need to be present, but also the subsequent growth of the particles. Growth can last for several hours and in some cases multiple days, with the latter timescale suggesting that the NPF events occur on a regional scale (Kecorius et al., 2019; Ström et al., 2009). In this study, we distinguish between (1) *on-site formation*, defined as occasions when the concentration of 2-4nm particles clearly rises above the background level; and (2) *formation and subsequent growth*, defined as a combination of *on-site formation* followed by subsequent growth. The latter is identified by visual inspection of the behaviour of the growing nucleation mode during the time it is present until when it disappears.

2.3.1 NPF Classification

New particle formation events were classified using schemes described by Dal Maso et al. (2005) and Aliaga et al. (2023). The classification schemes are used complementary. The classification by Dal Maso et al. (2005) focuses more on formation and growth, and is perhaps more intuitive to understand; however, it struggles with a large number of undefined cases. On the other hand, the nano-ranking analysis developed by Aliaga et al. (2023) can be used to explore the NPF intensity and focuses on the formation of nucleation mode particles as opposed to the subsequent growth. The Aliaga-method classifies NPF events into 3 groups base on the overall ranking of NPF intensity, avoiding the problem with unidentified events days (see Fig. S16 for the groups).

Deleted: (Gormley and Kennedy, 1948)

Deleted: (Wagner et al., 2016)

Deleted: 9

Deleted: 11

Deleted: (Kecorius et al., 2019; Ström et al., 2009)

Deleted: Dal Maso et al. (2005)

Deleted: Aliaga et al. (2023)

Deleted: Dal Maso et al. (2005)

Deleted: Aliaga et al. (2023)

Deleted: 4

Dal Maso classification system

In the [Dal Maso et al. \(2005\)](#), scheme an NPF event is defined if there exists a nucleation mode (3-25nm) which prevails for several hours and shows signs of growth; these events are further subdivided into three separate classes, namely, Class Ia and Ib, and Class II. Class I and Class II events are separated on the basis of whether the growth and formation rates, of the well-behaved size distribution, can be determined with good confidence. Class I is further divided into Class Ia and Class Ib depending on whether there are pre-existing particles obscuring the newly formed mode ([Dal Maso et al., 2005](#)). Undefined events are days where there are sporadic occurrences of nucleation-mode particles, and this classification is used to separate clear events from clear non-events. For examples of the different classifications and their respective averages see Fig. S12 and S13 respectively in the supplement.

Deleted: Dal Maso et al. (2005)

Deleted: (Dal Maso et al., 2005)

In this study, undefined days include days where the occurrence of nucleation mode particles (3-25nm) is linked to windblown snow and situations when ZEP is enveloped by clouds. For more details regarding these sampling conditions see Sect. S2.1 and S2.2 in the supplement for in-cloud-sampling and windblown snow events respectively. As a result, days classified as undefined are ones which experience bursts of nucleation mode particles, either from windblown snow (see Fig. S4), in-cloud conditions (see Fig. S2), from non-growing NPF or a combination of all three. Non-events are days that display no nucleation mode (3-25nm). It is important to note that this scheme assumes that particle formation occurs over a geographically wide area more or less simultaneously. Identification and classification were performed using daily surface plots of particle number size distribution based solely on NAIS data, that is, 2.5-40nm.

It should be noted that there is a degree of subjectivity to this classification approach, as to what is considered a certain class. The addition of the NAIS, in combination with the DMPS-system, made it easier to detect events using surface plots, especially when the growing mode struggled to surpass ~6nm. However, we were not able to identify any NPF events with the NAIS that we could not identify with the DMPS-system, as close inspection of the number concentration of the lowest size bins of the DMPS-system was sufficient to recognise NPF events.

Nanoparticle Ranking Analysis

The Nanoparticle Ranking Analysis method developed by [Aliaga et al. \(2023\)](#), provides a more continuous method to characterise atmospheric NPF, allowing users to gauge the strength of NPF. The nanoparticle ranking analysis was utilised using the daily maximum concentration of 2.8-5nm particles ($\Delta N_{\max, 2.8-5}$), from the negative channel of the NAIS, as the intensity parameter (note that no background value was used, unlike in [Aliaga et al. \(2023\)](#)).

Deleted: Aliaga et al. (2023)

Deleted: 2.82

Deleted: 2.82

Deleted: Aliaga et al. (2023)

due to the low concentrations measured at ZEP). In this study, periods which experienced windblown snow or in-cloud events were removed before applying the Nanoparticle Ranking method, as opposed to the [Dal Maso et al. \(2005\)](#) approach which classified them as undefined days. Special care was taken to clean the data before utilising the nanoparticle ranking analysis method, however it may still be the case that not all in-cloud and windblown snow events were removed successfully (see Sect. S2.1 and S2.2 in the supplement). Windblown and in-cloud events can influence the daily maximum without necessarily reflecting NPF activity. The strength of NPF activity was split into 3 groups depending on the intensity (i.e., $g1$, $g2$, $g3$), where $g3$ is the most intense event, followed by $g2$ and then $g1$; see [Aliaga et al. \(2023\)](#) for more details (see Fig. S16).

Deleted: Dal Maso et al. (2005)

2.3.2 Growth rates

The growth rate (GR) describes the rate of change of the modal diameter during an NPF event. For the GR estimation, NAIS and DMPS size distributions were merged at the maximum diameter of the NAIS size range (i.e. 40nm). The NAIS and DMPS data were both interpolated to 15-minute arithmetic means using linear interpolation. The start and end of each NPF event including their sequential growth (i.e. (2) *formation and subsequent growth*,) were estimated by visual inspection. The growing nucleation mode was isolated, and the GR was estimated for the full duration of the NPF event (see Fig. S17). For each size bin, we applied a Gaussian filter ($\sigma = 1$) and found the time in which the concentration of a particular size bin experienced the highest maximum rate of change. Essentially, the evolution of the nucleation mode was tracked using the timings for each identified peak in the rate of change of the growing nucleation mode diameter. The GR was calculated using a simple linear ordinary least squares (OLS) method.

Deleted: Aliaga et al. (2023)

Deleted: 4

Deleted: 5

It should be noted that the aforementioned issues concerning the overestimation of particle number concentrations for the NAIS measurements do not impact the calculation of GR as it only examines the changes to the diameter of the nucleation mode. Very few events grew beyond the limit of the NAIS (i.e. 40nm), hence the impact of a diameter jump from NAIS to DMPS was not considered. We tried to measure the GR of all NPF events (e.g. Class 1a, 1b and II), even though by definition Class II events lack the strength and consistency, making it difficult to calculate GRs ([Dal Maso et al., 2005](#)). The GRs for some Class II events were not estimated due to uncertain tracking of the nucleation growth.

Deleted: (Dal Maso et al., 2005)

2.3.3 Particle Formation Rates

The formation rates namely J_{3-7} , J_{7-25} , and J_{3-25} ($\text{cm}^{-3}\text{s}^{-1}$) were estimated using the following equation:

$$J_{i-j} = \frac{dN_{i-j}}{dt} + \text{CoagS} \times N_{i-j} + \frac{\text{GR}}{\Delta d_p} \times N_{i-j}, \quad (1)$$

where J_{i-j} is the formation rate of particles between diameters $i-j$ nm, CoagS is the coagulation sink of the pre-existing particle population, GR is the growth rate between $i-j$ nm, Δd_p is the difference in diameter between i and j nm and N_{i-j} is the total concentration of particles between i and j nm (Fig. S18 represents an example for calculating formation rates for particles between 3-7nm).

The Python GUI, *npf-event-analyzer*, was used to provide additional complementary estimations of the GRs and formation rates for the NPF events (i.e. in addition to the OLS fitting method). When using the *npf-event-analyzer* the max concentration method was used for the estimations of GR and J.

2.3.4. Condensation sink

The condensation sink (CS) is defined as the rate at which non-volatile vapours condense onto pre-existing particles (Kulmala et al., 2012). The twin DMPS system (5-850nm) was used to calculate CS, for SA vapour. The method for the determination of CS is described by Dal Maso et al. (2002), and uses the following equation:

$$\text{CS} = 2\pi D \int_i^{d_{p,i,\max}} d_{p,i} \beta(d_{p,i}) N(d_{p,i}) dd_{p,i}, \quad (2)$$

where $d_{p,i}$ is the diameter of a particle in size class i , N_i is the particle number concentration in the respective size class, and D is the diffusion coefficient of the condensing vapour (in this case condensing vapours were assumed to have H_2SO_4 diffusion properties). We used the transition regime correction factor β from Fuchs and Sutugin (1970).

2.3.5 Concentration of condensable vapours and source strength

The concentration of vapours (C_v) required to sustain the calculated GRs and their respective source rates (Q) were estimated using the following equations, and assuming the condensation of H_2SO_4 .

$$C_v = A \times dD_p/dt. \quad (3)$$

where A is the constant $1.37 \cdot 10^7 \text{ hcm}^{-3} \text{ nm}^{-1}$ representative for the molecular properties of SA and dD_p/dt is the rate of change of the nucleation mode i.e. GR. See Dal Maso et al. (2005) for more detail.

Deleted: and

Formatted: English (United Kingdom)

Deleted: (<https://github.com/jlpl/npf-event-analyzer>). T

Deleted: (Kulmala et al., 2012)

Deleted: Dal Maso et al. (2002)

Deleted: Fuchs and Sutugin (1970)

Deleted: ,

Deleted: , and CS is the condensation sink

Deleted: Dal Maso et al. (2005)

2.4 Airmass analysis

2.4.1 Transport model

Airmass back trajectory analysis was performed using the Hybrid Single-Particle Lagrangian Integrated Trajectory model (HYSPLIT V5.2.1) (Draxler et al., 1998; Stein et al., 2015). Ensemble back trajectories were initialised every hour starting at the latitude and longitude of ZEP, and at a height of 250m above ground level (AGL). The ensemble was generated by offsetting the meteorological grid point by one in the horizontal (i.e. $dx = 1.0$, $dy = 1.0$) and by the default offset of 0.01 sigma units in the vertical ($\sim 250\text{m}$), thus generating 27 back trajectories for all possible offsets in X, Y and Z. The starting height of 250m ensured that the starting location of all the ensemble members was at the surface or above. The back trajectories were calculated for 5 days back in time. Global Data Assimilation System (GDAS) $1^\circ \times 1^\circ$ archive data (<http://ready.arl.noaa.gov/archives.php>, last access: 08.08.2024) was used as the meteorological fields. No averaging of the back trajectories was performed, reducing them to single back trajectories.

The accumulated solar flux in the most recent 6 hours prior to arrival was used for the solar insolation term used throughout this study. The solar flux was taken from the original GDAS-derived HYSPLIT output. Furthermore, only endpoints within the mixed-layer were used as this is when air masses are influenced by the surface. The last 6 hours were used, as opposed to a longer duration, to place more significance on solar radiation closer to the receptor; an accumulated solar flux consisting for a longer duration could lead to misleading results.

2.4.2 Chlorophyll a data

Daily mapped chlorophyll a data derived from satellite observations (Aqua MODIS) (last accessed: 2024-10-04, downloaded from <https://oceancolor.gsfc.nasa.gov/>) were utilised along with the HYSPLIT output to estimate the chlorophyll a exposure. Exposure of the air masses to chlorophyll a was calculated based on the equation developed by Park et al. (2018), which linked chlorophyll a with DMS measurements on site at ZEP. Chlorophyll a exposure was essentially used as a proxy for DMS emissions. DMS in turn can form H_2SO_4 via SO_2 , although this was not explicitly addressed by Park et al. (2018). It should be noted that DMS is not only a precursor to H_2SO_4 , but also MSA. Park et al. (2018) utilised the relationship between DMS and air mass exposure to chlorophyll a to account for the differences in the species-specific phytoplankton population, between the Greenland Sea and the Barents Sea. For more details on the calculation of the chlorophyll exposure see Sect. 12.

Deleted: (Draxler et al., 1998; Stein et al., 2015)

Deleted:

Deleted:

Deleted: Park et al. (2018)

Deleted: Park et al. (2018)

Deleted: Park et al. (2018)

475 2.4.3 Nucleation site estimation

The method utilised here is similar to the NanoMap analysis developed by [Kristensson et al. \(2014\)](#), in which the geographic position of where new particle formation took place was estimated based on the particle number size distribution measurements during NPF events and HYSPLIT back trajectories. In this study, the duration of the nucleation mode was calculated. For every hour of the NPF event, back trajectories were initialised. The length at which the back trajectories were initialised for depended on how long the event had lasted. In short, i hours after the event begun, the back trajectory will have been initialised to run i hours backwards. So, for every hour except the start time the back trajectories were initialised for the following length of time: *back trajectory length* = *duration* - (*end time* - t), where $t \in$ (end to start times). Hence, for an NPF event that lasts 7 hours, 6 back trajectory ensembles were calculated for varying lengths from 7 hours to 1 hour long.

Deleted: Kristensson et al. (2014)

485 2.4.4 Air mass marine regions

The endpoints for every 5-day back trajectory ensemble were assigned to a marine region e.g. Arctic Ocean, Greenland Sea, Barents Sea, Norwegian Sea and North Atlantic Ocean (see Fig. S21 in supplement for defined regions). Days were assigned to a specific marine region based on the cumulative residence time in the specific sector. A threshold of 0.5 (50%) was used as qualifier. Days without clear source preference, i.e. the airmasses that did not reach the 50% threshold for either of the pre-defined source regions, were classified as mixed.

Deleted: 8

3. Results & Discussion

3.1 Seasonality and interannual variability

3.1.1 Annual cycle of NPF events

The 848 days of valid NAIS measurements (2022-04-17 to 2024-10-04) are classified according to the [Dal Maso et al. \(2005\)](#) classification system. 157 days are classified as NPF events, either Class Ia events (28), Class Ib (19) or Class II (110). A total of 438 days are classified as Undefined, and the remaining 271 days are considered to be either Non-event days (253) or consisting of too little data (i.e. defined as <12 hours of data per day) (18) (see Fig. 1). Overall, this meant that 19% of days with valid NAIS measurements experience an NPF event (i.e. Class Ia, Ib or II). The majority of days (52%) are classified as undefined due to the appearance of a non-growing nucleation mode that persisted for some hours. The undefined days are further sub-divided into those containing windblown snow events (19% of all data), in-cloud events (8% of all data) (see Fig. S6 in the supplement). Of the 438 Undefined days, 72 are a result of non-growing bursts (i.e. 17% of the Undefined days).

Deleted: Dal Maso et al. (2005)

The monthly frequency of NPF events (i.e. class Ia, Ib or II) shows events first appearing in April (see Fig. 1), reaching a peak, in terms of the occurrences per month, in May to July with approximately 30 - 40% of days experiencing NPF. The daily occurrence decreases during the month of June compared with the preceding and succeeding months, a consistent finding throughout the three years of data. After July, the frequency of events declines towards the end of summer and into autumn, where NPF events cease to be present beyond November. NPF events were observed as late in the year as November, which is interesting given that this is during the polar night when the Sun stays below the horizon (as witnessed by ZEP) and thus ZEP experiences negligible solar radiation. Multiple NPF events were observed during the polar night period (see Sect. 3.1.3 for more details). April 2022 appeared anomalous perhaps given that the measurements for that month were not an entire month as the measurements started on 17th April 2022.

Prior studies, at ZEP, which used slightly different event classification schemes, reported the occurrence of NPF events to be within the range of 18%-23% (Dallósto et al., 2017; Heintzenberg and Leck, 2012; Lee et al., 2020), Nieminen et al. (2018), reported NPF day frequencies of 14% (Mar-May), 34% (Jun-Aug) and 6.6% (Sep-Nov) and 0% (Dec-Feb). Lee et al. (2020), observed the highest frequency in terms of the formation and growth of N₃ particles in June (>40%), with the NPF frequency in May and July being ~40% and <40%. Dome-C, in Antarctica and Alert, on Greenland are two other polar sites with annual NPF frequencies between 10-20% (Nieminen et al., 2018), comparable to ZEP.

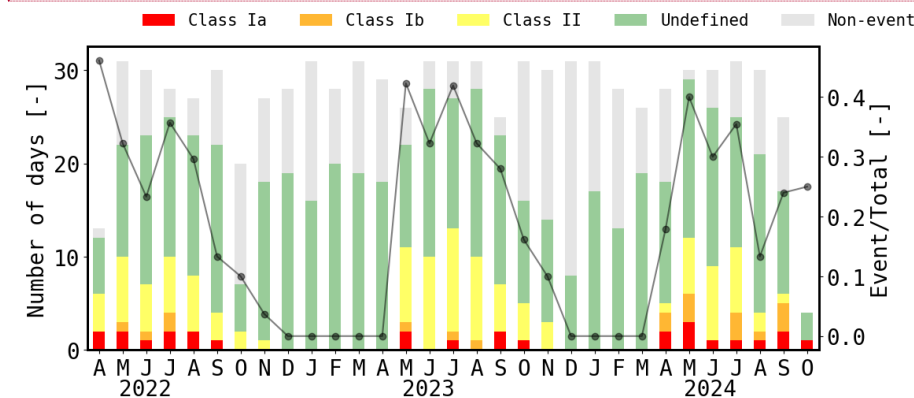


Fig 1: Number of days per month classified as either Class Ia (red), Ib (orange), II (yellow), Undefined (green) and Non-event (grey). Left-hand axis is the number of days classified based on the total number of valid measurement days. Right-hand side is the fraction of events (i.e. Class Ia, Ib, and II) over the total number of valid measurement

days (i.e. Class Ia, Ib, II, Undefined and Non-event). The fraction is given by Event/Total. Almost three years of Neutral cluster and Air Ion Spectrometer (NAIS) particle number size distribution data is classified using the Dal Maso et al. (2005) classification system.

Deleted: Classification of almost

Deleted: Dal Maso et al. (2005)

Deleted: Days in which there were no measurements were not included. Event/Total was the ratio of event days (i.e. Class Ia, Ib, and II) over the total number of days (i.e. Class Ia, Ib, II, Undefined and Non-event).

Formatted: Font: Not Bold

3.1.2 Interannual variability of main parameters

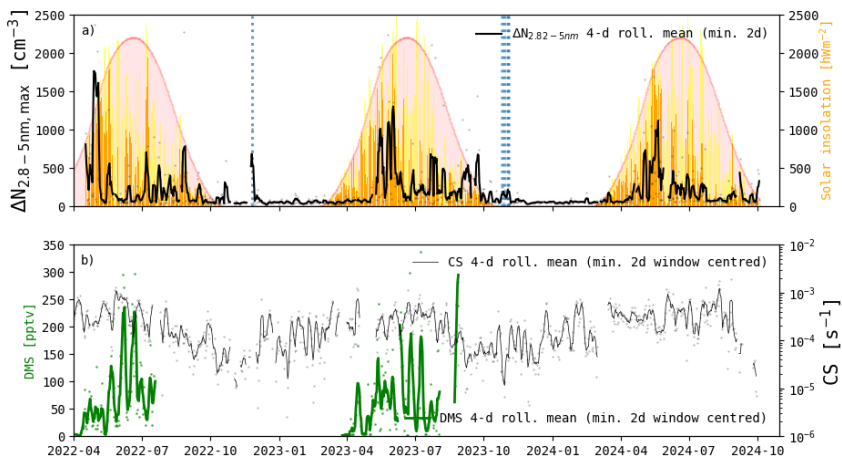
The interannual variability of the following main parameters, namely NPF intensity, solar insolation, DMS concentration and the CS, are explored in figures 2a-b. A proxy for the intensity of events, as opposed to the frequency, was determined by utilising elements of the nanoparticle ranking analysis developed by Aliaga et al. (2023). The maximum daily concentration in 2.8 – 5nm particles ($\Delta N_{max, 2.8-5}$), was used to represent the strength of NPF activity (shown to correlate well with calculated J). See Sect. 2.4.1 in the Methods for more details concerning the Nano ranking method and about the definition of solar insolation.

Deleted: Aliaga et al. (2023)

Deleted: increase

Deleted: 2.82

Deleted: 2.82



Formatted Table

Formatted: Strikethrough

Fig 2: Timeseries a) presenting in black dots the intensity parameter, i.e. the daily maximum total number concentration of particles between 2.8 and 5 nm ($\Delta N_{max, 2.8-5}$), which corresponds to the strength of the atmospheric NPF event. The black line signifies the running 4-day (centred) mean for $\Delta N_{max, 2.8-5}$ and was applied with a minimum window of 4 days. The orange bar chart displays the daily-averaged solar insolation, which was defined as the accumulation of the most recent 6 hours of solar radiation experienced by air masses arriving at ZEP, within the mixed-layer (see Sect. 2.4.1 in the Methods for details). The faint yellow bars are for 5 days, regardless of if airmasses were within the mixed-layer. The shaded region in red is the average hourly clear skies global horizontal irradiance (GHI) (calculated based on the altitude, latitude and longitude of ZEP, using Python Package *pvlb* v0.9.0 <https://pvlb-python.readthedocs.io/en/v0.9.0/api.html#clear-sky>) multiplied by six. The timeseries b) displays all

Deleted: 2.82

Deleted: 1

Deleted: 4

valid dimethyl sulphide (DMS) measurements recorded at ZEP in green (typically, the instrument begins operation around late March or early April). In black, the condensation sink (CS) at ZEP is displayed, including a running mean (line) and daily means (dots). For the CS and DMS, a 4-day running (centred) mean is used with a minimum window of 4-days. The dashed blue lines represent the occurrence of the polar night events.

Deleted: 14

$\Delta N_{max 2.8-5}$ shows various peaks throughout the annual cycle, however, the one striking feature is that $\Delta N_{max 2.8-5}$ the first and highest peaks in $\Delta N_{max 2.8-5}$ occur in late spring and the beginning of summer (i.e. end of April in 2022, beginning of June in 2023, and towards the end of May in 2024) (see Fig. 2a)). For the annual cycle in terms of frequency (see Fig. 1), the first peak in occurrence matches the increased $\Delta N_{max 2.8-5}$ (Fig. 2a)). However, the succeeding peak in frequency (i.e. in July) coincides with reduced intensity ($\Delta N_{max 2.8-5}$). The seasonality of the $\Delta N_{max 2.8-5}$ value follows that of the solar insolation, and for the years 2023 and 2024, the maximum NPF strength (i.e. $\Delta N_{max 2.8-5}$) coincides with the greatest mean solar insolation; a much better relationship is observed when selecting for air masses within the mixed layer as this is when arriving air masses are influenced by the surface. The annual cycle for the calculated CS exhibits a sustained maximum in late winter and early spring due to the Arctic Haze, and the CS stays high into the summer before reaching a minimum in autumn, typically the cleanest part of the year (Tunved et al., 2013) (see Fig. 2b)). During the Arctic Haze period, long-range transported air masses from Eurasia bring aged aerosol contributing to higher CS. At the same time this period is characterised by higher sulphate and SO₂ concentrations, mainly of anthropogenic origin (Platt et al., 2022). One additional aspect is that it is clear that increases in $\Delta N_{max 2.8-5}$ occur during periods where there was a slight decrease in CS, below that of the seasonal CS average. For example, the peak in $\Delta N_{max 2.8-5}$ in April in 2022 (see Fig. 2b)), which occurs slightly prior to the maximum solar insolation, can potentially be explained by the reduction in the CS during that period.

Deleted: 2.82

Deleted: 2.82

Deleted: 2.82

Deleted: 2.82

Deleted: 2.82

Deleted: 2.82

Deleted: 2.82

Deleted: (Tunved et al., 2013)

Deleted: (Platt et al., 2022)

Deleted: 2.82

Deleted: 2.82

The solar intensity is a controlling factor in the production of SA (i.e. H₂SO₄) via the oxidation of SO₂. In the Arctic, DMS acts a significant source of sulphate, however other sulphur compounds can be oxidised e.g., carbon disulphide and hydrogen sulphide (Pei et al., 2021). The coagulation sink (CoagS) acts as a sink for the newly-formed particles, and the CS acts a sink for nucleating vapours, thus high CoagS and CS can inhibit NPF. NPF days with greater intensity (i.e., $\Delta N_{max 2.8-5}$) is skewed towards the end of spring and beginning of summer. Similarly, solar insolation is also shifted towards the start of summer, even though the theoretical solar maximum has a symmetrical distribution centred around the middle of summer (see Fig. 2a)). The skewed pattern in both the solar insolation and NPF intensity ($\Delta N_{max 2.8-5}$) can perhaps be explained by annual changes in cloud cover (see

Deleted:

Deleted: (among other potential precursor candidates)

Deleted: , whilst the

Deleted: CS

Deleted: coagulation sink (CoagS, CS can act as a proxy for CoagS)

Deleted: and

Deleted:

Deleted: 2.82

Deleted: , with a greater intensity

Deleted: 2.82

600 Maturilli and Ebell (2018), for the frequency occurrence of cloudy sky, increasing to ~80% during the summer from an April minimum of ~40-60%). Late autumn experiences numerous, but relatively weaker, NPF events, with peaks in $\Delta N_{max, 2.8-5}$ (~100-500 cm⁻³) every year at the end of the summer, despite the relatively small solar insolation. The transition period between summer and haze (i.e. Oct. – Jan.) is typically the cleanest part of the year, and thus exhibits the lowest CS (Tunved et al., 2013). During this part of the year the small amount of solar insolation and vapours present are enough to trigger NPF, possibly as a result of the low CS.

605 Finally, DMS is the main of source of sulphur for clean marine aerosol particles, and constitutes an important compound in regards to NPF via its oxidation products. Numerous peaks in DMS are observed during the summertime, with a maximum observed in July (see Fig. 2b)). Elevated concentrations of DMS at ZEP coincide with air masses which traverse over marine regions containing chlorophyll a (see Fig. S30 – S37 in supplement, but also noted by Park et al. (2018b)).

610 3.1.3 Events with minimal solar insolation (i.e. Polar night events)

Contrary to expectations, during the polar night (i.e. when the sun does not cross the horizon at Ny-Ålesund) a number of NPF events are observed. As shown earlier, solar insolation is clearly one of the driving parameters for NPF occurrence and intensity, and Ny-Ålesund in November does not receive any direct solar radiation (see Fig. S19 for examples of the polar night events). For the November polar night events, air mass back trajectories indicate that air masses have to travel at least half a day backwards before reaching Ny-Ålesund to experience direct solar radiation. In total, five types of these events are observed, and are detailed in Table 1.

Table 1: Polar night events observation from 2022-04-17 to 2024-10-04

Polar night day	Datetime	Duration [hrs]	GR [nm hr ⁻¹]
1.	2022-11-25 - 2022-11-26	15	0.38
2.	2023-10-25 - 2023-10-26	23	0.41
3	a. 2023-10-28 01:00 – 2023-10-29	36	0.25
	b. 2023-10-28 12:00 – 2023-10-29	27	
4.	2023-11-01 - 2023-11-02	28	0.32
5.	2023-11-02 -2023-11-03	8	0.78

Deleted: Maturilli and Ebell (2018)

Formatted: Font: Times New Roman

Deleted: 2.82

Deleted: ,

Deleted: (Tunved et al., 2013)

Deleted: considering the

Deleted: not shown here

Deleted: Park et al. (2018b)

Deleted: 6

Deleted: i.e. 1. 2022-11-25 - 2022-11-26, 2. 2023-10-25 - 2023-10-26, 3ab. 2023-10-28 (x2, 01:00 12:00) – 2023-10-29, 4. 2023-11-01 - 2023-11-02, 5. 2023-11-02 -2023-11-03.

Formatted Table

Deleted: events

Formatted: Font: 8 pt

Formatted: Superscript

Formatted: Font: 10 pt

Formatted: Font: 10 pt

Formatted: List Paragraph, Left

Formatted: List Paragraph, Numbered + Level: 1 + Numbering Style: a, b, c, ... + Start at: 1 + Alignment: Left + Aligned at: 0,63 cm + Indent at: 1,27 cm

Formatted: Font: 10 pt

635 The polar night NPF event starting on the 2023-10-25 appears to be a transported event (TE, see [Dada et al. \(2018\)](#),
 for details on the definition of a transported event) as we do not observe the entire growth of freshly nucleated
 particles; instead the growing nucleation mode begins at 4 nm. The extremely low accumulated solar radiation
 experienced by the arriving air masses during these events means that there should be little available nucleating
 vapours (less oxidation before reaching ZEP). Two other Arctic sites namely [Utqiagvik](#), Alaska and Tiksi, Russia,
 also observe NPF events without the sites experiencing any solar radiation ([Asmi et al., 2016](#); [Kolesar et al.,
 2017b](#)). However, in these two cases the suggestion is that anthropogenic emissions of semi-volatile gases from
 nearby oil fields or the build-up of anthropogenic emissions during the haze period can explain these events; an
 interpretation which is not applicable for ZEP.

640 The vertical component of the arriving back trajectories associated with these polar night events shows that air
 masses originated from higher altitudes (Figure S17, in the supplement), and that these air masses descend before
 arriving at ZEP. One possible explanation for these types of events could be that precursor gases, originally
 emitted from the surface, are transported to high altitudes where the CS is sufficiently low enough to allow them
 to survive long enough to be nucleated on descent to ZEP. The air masses travel further south and higher in altitude,
 645 where the small amount of solar radiation potentially allows for photochemical production of nucleating vapours
 to build up.

3.1.4 Start times and duration throughout the year

The median duration of NPF events, including their subsequent growth, is approximately ~10 hours and seems to
 be fairly consistent across the annual cycle, with a decrease towards the autumn (see Fig. 3a)). Of all events, 75%
 650 > 7 hours, and 90% > 4 hours. During the sunlit period, the starting time of events is typically around 09:00 (UTC)
 for the period April to August. From September onwards, the onset of the observed NPF shifts to later in the day
 (Fig. 3b)). For the period August to October, onset of NPF is approximately 8 hours after sunrise at Ny Ålesund
 (see the gradient of the two slopes in Fig. 3b) remains constant but the start time is shifted by 8 hours). This [could](#)
 reflect that during the NPF season there is a delay of approximately 8 hours from the minimum solar intensity to
 655 the onset of NPF, and [this could](#) be related to the time it takes for precursor gases to [accumulate](#).
[Lee et al. \(2020\)](#) suggested that the duration of NPF at ZEP was approximately 6-7 hours, with the longest duration
 in summer. In addition, NPF start times were 13:00 – 14:00 (local time, which is 11:00 – 12:00 UTC in the summer
 time). The start times observed at ZEP, here in this study, were typically 2 hours earlier, perhaps as a result of the
 slightly lower minimum particle diameter detection and not the same definition of the onset of nucleation.

Deleted: ¶

Deleted: Dada et al. (2018)

Deleted: We find that these events coincide with air masses mainly arriving from lower latitudes (i.e. further south).

Deleted: Utqiagvik

Deleted: (Asmi et al., 2016; Kolesar et al., 2017b)

Deleted: s perhaps

Deleted: minimum

Deleted: might

Deleted: the availability of

Deleted: build up in the mixed layer.

Deleted: Lee et al. (2020)

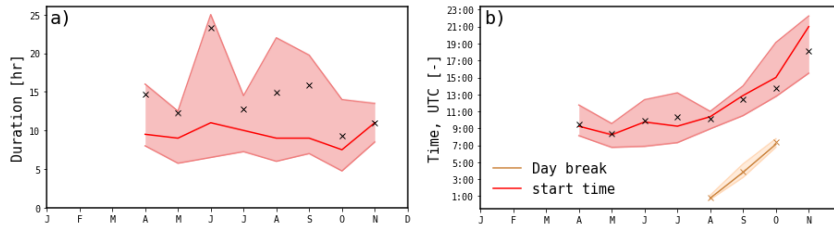
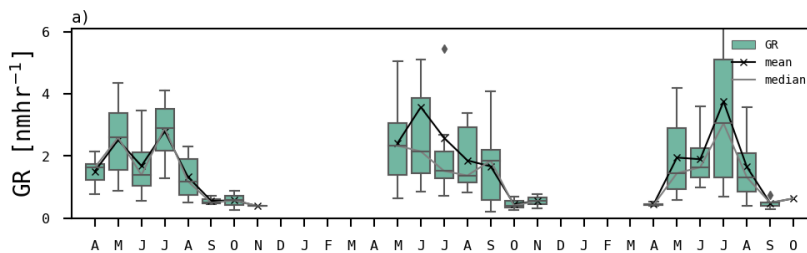


Fig 3: (a) the duration of the entire growth of newly formed particles from the first measurable size bin i.e. 2.5nm to the largest size bin (b) the average start time of the observed new particle formation (NPF) (monthly mean displayed by the cross and median by the red line), along with a shaded region representing the 25th and 75th percentiles. Along with the estimated time the global horizontal irradiance (GHI) goes beyond zero (i.e. daybreak). The average time of day break for the months April until July is not applicable, since the sun is constantly above the horizon (as experienced by ZEP). Time is given in terms of coordinated universal time (UTC).

3.1.5 Seasonality of Growth Rates and formation rates

The average, mean (median), GRs for all the events is 2.2 (1.5) nmhr⁻¹. For class Ia, Ib, and II events the average GRs are 1.4 (1.0), 1.7 (1.2), 2.6 (1.2) nmhr⁻¹ respectively. The largest GRs occur during July 4.0 (2.1) nm hr⁻¹. The polar night events exhibit a significant reduced GR of 0.43 (0.35) nmhr⁻¹ (see Fig. 4 a). Using the *npf-event-analyzer* for the Class Ia events the GRs are 0.9 (1.1), 2.1 (2.6), 4.1 (4.4) nmhr⁻¹ for 3-7, 3-25 and 7-25 nm diameter ranges.

The GRs presented here agree well with earlier estimates of GRs by [Nieminen et al. \(2018\)](#), reporting a GR_{10-25nm} at ZEP within the range of 1.2 -1.6 nmhr⁻¹. [Kerminen et al. \(2018\)](#), report a median GR for the Arctic of 2.3 (0.23-4.1, 5th and 95th percentiles) nm hr⁻¹, the site type (i.e. globally) with the smallest GRs.



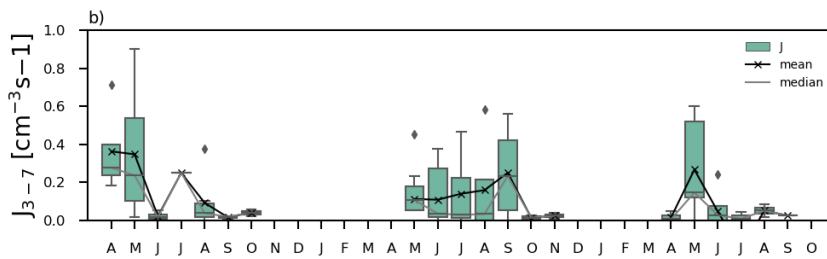


Fig 4: **a)** the calculated growth rates (GRs) **b)** calculated median formation rates for particles between 3 and 7nm ($J_{med, 3-7}$) throughout the measurement period are represented by box-whiskers, which display the monthly growth rate averages and include the median and 25th - 75th percentiles. The green box-whiskers refer to the average GRs/ J_{3-7} s from the start of the NPF until it stops growing. The black line cross represents the mean, and the grey lines join the median values. Diamonds represent outliers.

Deleted: T

Deleted: throughout

685 Formation rates for $J_{median, 3-7nm}$ are $0.14 \pm 0.23 \text{ cm}^{-3}\text{s}^{-1}$ (mean $\pm$ std) for all events in which calculations of J and GR could be ascertained (see Fig. 4 b) for the seasonality). Formation rates are generally greater in May with 2023 not exhibiting this maximum in May. The formation rates for J_{3-7nm} , J_{3-25nm} , J_{7-25nm} are estimated to be 0.28 ± 0.33 , 0.40 ± 0.35 and $0.31 \pm 0.28 \text{ cm}^{-3}\text{s}^{-1}$ (mean $\pm$ std) for class Ia events (using the max concentration method and *npf-event-analyzer* see Sect. 2.3.3 in the Methods).

690 Previous studies also estimate J for various diameters at Ny-Ålesund and ZEP. Beck et al. (2021), reports a $J_{1.5}$ equal to $0.27 \text{ cm}^{-3}\text{s}^{-1}$ (Beck et al., 2021). Nieminen et al. (2018), presents the seasonal medians for J_{nuc} of nucleation-mode particles (10-25nm) varying from $0.08 \text{ cm}^{-3}\text{s}^{-1}$ in spring, to $0.032 \text{ cm}^{-3}\text{s}^{-1}$ in summer and to $0.0066 \text{ cm}^{-3}\text{s}^{-1}$ in autumn. Lee et al. (2020), gives J_{3-7} , J_{7-25} , J_{3-25} values ranging between 0.001-0.54, 0.003-0.5, 0.007-0.61 $\text{cm}^{-3}\text{s}^{-1}$ respectively, and the averages for J_{3-7} , J_{7-25} , J_{3-25} are 0.04, 0.09 and $0.12 \text{ cm}^{-3}\text{s}^{-1}$. Kerminen et al. (2018), report a median Arctic J (i.e. varying size ranges) of $0.51 \text{ cm}^{-3}\text{s}^{-1}$.

Deleted: Beck et al. (2021)

Deleted: (Beck et al., 2021)

Deleted: Nieminen et al. (2018)

Deleted: Lee et al. (2020)

Deleted: Kerminen et al. (2018)

695 In comparison to GRs and Js at other sites (e.g. Kerminen et al. (2018); Nieminen et al., (2018)), we see that processes within the Arctic, as observed at ZEP, are significantly slower. In urban environments, compared with more remote ones, estimations for J are significantly larger due to the larger coagulation sink that needs to be overcome (Cai and Jiang, 2017); here, the coagulation sink is much lower than in urban environments.

Deleted: Kerminen et al. (2018); Nieminen et al., (2018)

Deleted: (Cai and Jiang, 2017)

3.1.6 Seasonality of Condensable Vapours

The concentration of condensable vapours (Cv) required to sustain the GRs is estimated using equation (3) and the calculated growth rates for the various classes of events (i.e., Class Ia, Ib and II). We show that Cv increases until July before decreasing during the autumn and into the winter months. It is striking to note that the seasonal variation is similar to that of the sum of the measured concentrations of potential aerosol precursor gases measured in Ny-Ålesund during a 2017 campaign (see Fig. 5). The estimated Cv is also of a similar magnitude to the combined sum of measured HOMs, sulphuric acid (H₂SO₄), methanesulphonic acid (MSA) and iodic acid (HIO₃). For Ny-Ålesund, Beck et al. (2021) report increasing IA, SA and MSA concentrations from the start of the sunlight period, and high concentrations prevailing before decreasing around mid-June. HOMs concentrations averaged around $\sim 1.8 \cdot 10^6 \text{ cm}^{-3}$ during the spring before increasing to $2.3 \cdot 10^6 \text{ cm}^{-3}$ in June. As a ratio of the total available precursor vapours HOMs account for approximately 0.78 of the entire concentration of precursor vapours after June and prior to June this fraction approximately 0.24 (see Fig. S24). It can therefore be argued that the composition of condensable vapours undergoes a dramatic transformation around the start of June, and the composition of newly formed particles during the initial spring peak is different to that of the subsequent peaks after June (see Fig. 1 & 2). Boyer et al. (2024) measured the concentrations of condensable vapours in Arctic, however at higher latitudes ($>80^\circ\text{N}$). The estimated Cv was on average 5.9 times greater than the combined concentrations of the measured precursor gases given in Boyer et al. (2024) (see Fig. S23 in supplement). At higher latitudes ($>80^\circ\text{N}$), there is potentially less photochemistry and biological activity, moreover, the potential sources of precursor gases are more limited (Brean et al., 2023), and may be more influenced by the sea ice cover, further reducing an available and giving rise to the difference between the estimated Cv at ZEP and total concentrations during MOSiAC.

Deleted: T

Deleted: found in the high Arctic during the MOSiAC campaign (Boyer et al., 2024), namely sulphuric acid (H₂SO₄), methanesulphonic acid (MSA) and iodic acid (HIO₃). The sum of these three species increases during the spring, peaks in May and then decreases towards the end of the year (see Fig. 5). Overall, the concentration of the calculated Cv is on average 5.9 times greater than the combined concentrations of the measured precursor gases (H₂SO₄, MSA and HIO₃). For Ny-Ålesund, Beck et al. (2021) report increasing IA, SA and MSA concentrations from the start of the sunlight period, and high concentrations prevailing before decreasing around mid-June. HOM concentrations are very low during spring and increase in May, however it still shows their importance in contributing to Cv.

Formatted: Superscript

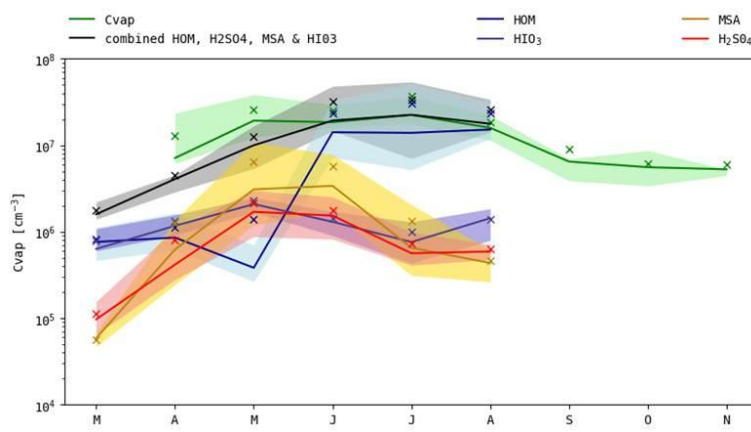
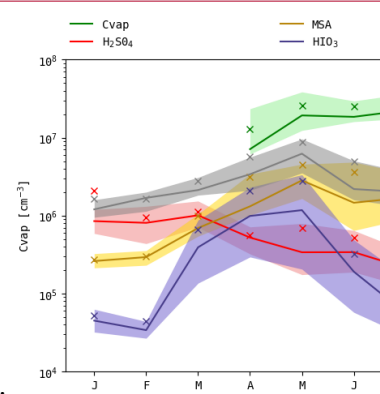


Fig 5: Estimated concentration of condensable vapours (Cv, green) compared with measured concentrations: the monthly means (crosses) and medians (thick line), along with the 25th and 75th percentiles (shaded region) for both Cv and Ny-Ålesund measurements. Arctic precursor vapours namely sulphuric acid (H₂SO₄, red), methanesulphonic acid (MSA, gold) and iodic acid (HIO₃, purple) and combined (grey) from Beck et al. (2021) are displayed. The estimated Cv measurements cover the years 2022 to 2024, whilst the Svalbard measurements from Beck et al. (2021) cover the period 2017-03-22 to 2017-08-23.

3.2 Positive and negative ions during NPF

The polarity of air ions, which aid in nucleation, provides information about the composition of the newly formed aerosol particles. Cluster air ions are typically defined between 0.8-1.7nm, whilst intermediate ions range from 1.6-7.4nm. Intermediate ions are created in the initial stage of nucleation and are typically formed when neutral nanoparticles encounter cluster ions and acquire their charge (Tammet et al., 2014). Negative ions can be linked to IA and SA to name a couple of compounds. Unfortunately, due to an instrumental problem related to a consistent loss of the negative cluster ions experienced during the measurement campaign (see Sect. S2.6 in the supplement) it was not possible to measure the concentration of negative ions less than 2nm.

The total number of positive (N_{2-4nm}^+) and negative (N_{2-4nm}^-) small intermediate ions (2-4nm) within the growing mode are calculated for each NPF event. In May, at the start of the NPF season, we observe the smallest fraction of positive small intermediate ions compared to the total number of small positive intermediate ions (i.e. the mean and median ratio, $N_{2-4nm}^+ / (N_{2-4nm}^- + N_{2-4nm}^+) \approx 0.2$). The low $N_{2-4nm}^+ / \text{total}$ ratio in May perhaps relates to negatively charged molecules dominating the initial stages of ion-induced nucleation (i.e., compared with positive ions). Moreover, the low ratio in May could be related to the greater NPF intensity (i.e. larger production of



Deleted:

Formatted Table

Deleted: The monthly means (crosses) and medians (thick line), along with the 25th and 75th percentiles (shaded region) for both Cv and Ny-Ålesund measurements. The estimated concentration of equivalent condensational vapours (Cv, green), and the concentrations of measured

Formatted: Justified

Formatted: Font: 9 pt, Font color: Black

Deleted: Beck et al. (2021)

Formatted: Font: Not Bold

Deleted: , ...overing...the years 2022 to 2024, whilst the Svalbard measurements from Beck et al. (2021) cover the periods...2017-03-22 to 2017-08-23..

Deleted: Considering the concentrations of precursor vapours from Boyer et al. (2024) are measured at higher latitudes (>80°N), where there is less photochemistry and biological activity, the calculated Cv and combined measured concentrations of precursors are not too different in magnitude. The sources of available precursors are considered to be more limited where the MOSiAC campaign was sampling (Brean et al., 2023), and may be more influenced by the sea ice cover. Furthermore, the calculated Cv presented here is only available for days with NPF, while the results from Boyer et al. (2024) represent average concentrations regardless of if NPF is observed or not. Hence, it is natural to expect higher concentrations than when NPF takes place, which in turn may explain as to why the calculated Cv is higher

Formatted

Deleted: ¶

Deleted: (Tammet et al., 2014)

Deleted: highest...ratio ...f positivenegative...small intermediate ions compared to the total number of small positive intermediate ions (i.e. the ...ean and median ratio, $N_{2-4nm}^+ / (N_{2-4nm}^- + N_{2-4nm}^+) \approx 0.2$).

Deleted: =...0.24 and 1.4 respectively.... Higher ...he low $N_{2-4nm}^+ / \text{total}$ ratio in May perhaps relates to the...presence of... negatively charged molecules dominating participating in ...he initial stages of ion-induced

negative nucleation mode particles, in May). However, towards the end of the NPF season (i.e. Sep. - Nov.) there is a shift towards a growing nucleation mode (e.g., 2-4nm) more dominated by positive small intermediate ions, as opposed to negative small intermediate ions (see Fig. 6). The tendency for small intermediate ions (2-4nm) to exhibit a positive charge appears from September onwards, which is also when there begins to be limited solar insolation and the occurrence of polar night events (i.e. late Oct. and Nov.), it is also, and perhaps more importantly, when the composition of condensable vapours changes to being dominated by HOMs (see Sect. 3.1.6.). However, it needs to be noted that there are few events during September and beyond (29 in total). Despite the lack of information about the cluster ions, the shift in the overall charge of the small intermediate ions, from more negative to more positive, potentially hints at a seasonal change in the vapours contributing to the initial formation of critical clusters.

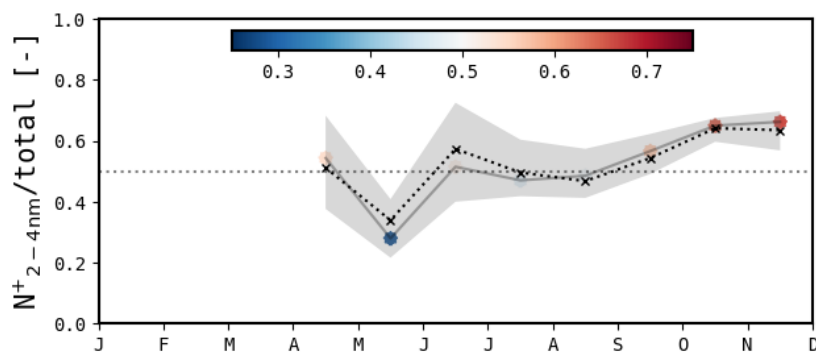


Fig 6: For all new particle formation (NPF) events the monthly ratio of the total number concentration of positively charged small intermediate ions (2-4nm) over the total concentration of both positively and negatively charged small intermediate ions (2-4nm) ($N_{2-4nm}^+/N_{2-4nm}^+ + N_{2-4nm}^-$). N_{2-4nm}^+ and N_{2-4nm}^- includes only ion concentrations within the identified growing mode between 2 and 4nm. The mean monthly $N_{2-4nm}^+/total$ is presented as a cross, and the median is displayed by coloured circles, blue for an overall negative charge (ratio < 0.5) and red for an overall positive charge (ratio > 0.5). The grey shading displays the 25th – 75th percentiles. The dashed black line represents an overall neutral charge of small intermediate ions. It should be noted that there may be some instrumental biases between the two polarities, however, the background concentrations (during no-event periods) display $N_{2-4nm}^+/N_{2-4nm}^+ + N_{2-4nm}^-$ ratios close to unity, ~1.02.

As mentioned above, Beck et al. (2021) reports higher concentrations of precursor gases with low proton affinity (e.g., MSA, SA, HSO_4^-) during the month of May, while highly oxygenated organic molecules (HOMs) display

Deleted:)

Deleted: T

Deleted: The consistent loss of the cluster ions in the negative polarity (see Sect. S2.6 in the supplement) means that the ratio between the concentrations of the two polarities for ions within that size range (i.e. <2nm) is not explored.

Deleted: this

Deleted: negatively

Deleted: /

Deleted: /

Deleted: N_{2-4nm}^-

Deleted: i

Deleted: >1

Deleted: <1

Deleted: The number of events which make up the monthly values are written above each respective cross.

Formatted: Font: 9 pt, Bold

Formatted: Font: Bold

Formatted: Font: 9 pt, Bold

Formatted: Font: 9 pt

Deleted: ¶

Deleted: Beck et al. (2021)

higher concentrations than SA and MSA during April and from June towards September. Beck et al. (2021) suggest that the events in May were likely initiated by negative ion-induced NPF and relate to sulphuric acid and ammonia nucleation.

Deleted: Beck et al. (2021)

Ammonia (NH_4^+) is a possible positively charged cluster ion which could play a role in the shift in polarity (Kirkby et al., 2016), however, it needs to be stressed that further work is needed to ascertain what is the reason for this shift in the polarity. We were unable to compare the overall polarity charges for small intermediate ions with concentrations of NH_4^+ . However, the shift from more negative to more positive small intermediate ions seems to coincide with the increase in concentrations of HOMs relative to the total concentration of other condensable vapours as also indicated above (see Fig. S24).

Deleted: The ratio seems to reflect the varying seasonal NPF strength (i.e. production of nucleation mode particles). It should be noted that there may be some instrumental biases between the two polarities, however, the background concentrations (during no-event periods) display $N_{2-4\text{nm}}/N_{>2-4\text{nm}}$ ratios close to unity, ~ 1.02 .

Formatted: Font: Times New Roman

Deleted: (Kirkby et al., 2016)

Deleted: it there is

Deleted: a

Deleted: of the air ions

Formatted: Not Superscript/ Subscript

Deleted: .

3.3 Source regions and geo-spatial extent

NPF events are assigned to various marine regions using the method described in Sect. 2.4.5. Air masses that traverse over the Arctic Ocean coincide with the highest number of NPF events (57 in total or 37% of NPF events; see Table 1). 37% of air masses are classified as originating from the Arctic Ocean, the largest marine region in terms of contribution. The air masses that mainly originate over the Greenland Sea result in 29 NPF events and show the highest probability of an NPF event occurring, with 27% of the Greenland Sea air masses being linked to an NPF day. Air masses originating over the Arctic Ocean and Barents Sea have a 19% and 13% likelihood, respectively.

Table 2: The frequency of occurrence of air masses from different sectors compared with the number of events. Both numbers are given in days, followed by a ratio of the two (i.e. Likelihood). Although not shown in table, back trajectories traversed over the Kara Sea on two occasions. The average CS, solar insolation (both mixed-layer and all), maximum daily increase in $2.8 - 5\text{nm}$ particles ($\Delta N_{\text{max},2.8-5}$) and proportion of back trajectory within mixed layer are given for each marine sector. The averages for these variables area also given during the NPF days and presented in brackets.

Field Code Changed

Deleted: 1

Deleted: 2.82

Formatted: Font color: Auto

Deleted: 2.82

Formatted: Font color: Auto

Formatted: Superscript

Deleted: 2.82

Deleted: h

Air mass	Region's frequency (days)	Number of NPF events (days)	Likelihood	CS [s^{-1}]	Solar insolation [h W m^{-2}]	$\Delta N_{\text{max},2.8-5}$ [cm^{-3}]	Proportion in ML
Arctic Ocean	313	57	0.19	0.20 (0.19)	185 (378) 368 (753)	239 (776)	0.63 (0.60)
Greenland Sea	109	29	0.27	0.25 (0.25)	327 (442) 666 (991)	254 (665)	0.60 (0.54)

Barents Sea	69	9	0.13	0.36 (0.29)	118 (211) 597 (1100)	93 (204)	0.32 (0.28)
Norwegian Sea	44	7	0.16	0.25 (0.43)	232 (418) 618 (1330)	136 (382)	0.48 (0.34)
Mixed	303	51	0.17	0.23	184	147	0.54 (0.50)
Total	840	153	0.18	0.23	200	190	0.57 (0.53)

The analysis is expanded, using the method outlined in Sect. 2.4.4, by estimating the location of the formation of newly formed particles. The most common location for nucleation to occur is off the western coast of Svalbard, i.e., in the Greenland Sea and Arctic Ocean (see Fig. 7). Furthermore, the Greenland Sea, south-west of ZEP, as opposed to the Barents Sea, is home to more nucleation sites. It is also clear that the vast majority of nucleation sites are estimated to occur over the ocean. The estimated nucleation sites are almost all within a $1 \cdot 10^6$ km radius of ZEP, and the vast majority are fairly close to ZEP (see Fig. 7), a finding which can also be inferred from the observation that the majority of events begin growing from 2.5 nm. Furthermore, from the air mass back trajectory analysis, we gather that air masses typically reside between sea level and the altitude of ZEP (i.e. 0-474 m, see Fig. S20). There is little difference between the altitudes of the arriving air masses for the NPF events compared to all days.

Deleted: 17

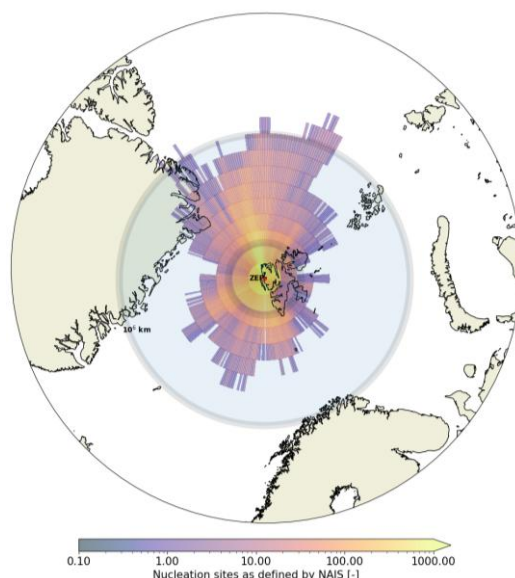


Fig 7: Total number of estimated nucleation sites per grid cell. Unit is based on the number of back trajectories initialised and does not reflect a physical value. Sites of nucleation for class Ia, Ib and II events are estimated by using HYSPLIT to trace the likely origin of particles produced during NPF events, and initialising the back trajectories based on the duration of the NPF event (see Sect. 2.4.4).

DMS emissions are closely linked to phytoplankton dynamics and abundance, and typically occur following phytoplankton biomass maxima (Galí and Simão, 2015). Numerous studies have tried to link chlorophyll a, which is used as a proxy for phytoplankton biomass, with DMS emissions to the atmosphere (Jang et al., 2019). Park et al. (2018) note a strong correlation between DMS concentrations measured at ZEP and air masses traversing over the surrounding ocean with high chlorophyll a. They find that the relationship between DMS concentrations and air masses likely exposed to high DMS emissions are stronger for air masses traversing over the Greenland Sea, compared with the Barents Sea. Therefore, suggesting that the Greenland Sea has a DMS production capacity three times greater than the Barents Sea. By replicating the method detailed in Park et al. (2018), we also find good correlations between chlorophyll a exposure and DMS, however, not for all years, and not for the years related to this study (i.e. 2022 and 2023). In the study by Park et al. (2018), the years 2010, 2014 and 2015 were included, however, 2022 and 2023 did not yield such strong correlations. We were unable to replicate the findings presented in Park et al. (2018) for the years 2022 and 2023 (see Fig. S30-37), however, we could replicate the

Deleted: (Galí and Simão, 2015)

Deleted: (Jang et al., 2019)

Deleted: At ZEP,

Deleted: Park et al. (2018)

Deleted: Park et al. (2018)

Deleted: Park et al. (2018)

findings for 2010 (see Fig. S30). We can only speculate about the reasons for the interannual variation between the relationship between chlorophyll a and DMS. We argue that because of the large spatial and temporal interannual variability, possible geo-spatial changes in the DMSP-rich phytoplankton species in the neighbouring seas, a wide range of DMS oxidation timescales depending on photochemistry and available solar insolation (Ghahreman et al., 2019), can all potentially influence this relationship. Furthermore, the influence of various other meteorological and environment factors that affect DMS oxidation and the lifetime of precursor gases, and sea ice and cloud cover could lead to inaccuracies in HYSPLIT outputs and chlorophyll a satellite retrievals. In addition, direct correlations between atmospheric DMS and chlorophyll a exposure are not detected around Iceland, however, a high atmospheric DMS to chlorophyll a exposure coincides with areas containing DMSP-rich phytoplankton species (Lee et al., 2023).

The Greenland Sea is connected to a higher likelihood of NPF events; compared with the Barents Sea there is a distinct difference between both seas in terms of the proportion of air masses coinciding with NPF events, and the number of nucleation sites estimated to arise from each sea (see Fig. 7). The combined effect of a larger number of air masses coming from the west and an increased NPF-likelihood linked to western air masses means that we observe more NPF events off the western edge of Svalbard. This difference could be linked to the finding that the DMS production capacity in the Greenland Sea is estimated to be three times greater than that of the Barents Sea, due to more DMSP-containing phytoplankton species (e.g. Phaeocystis) (Park et al., 2018). However, another distinct difference between these two source regions is that air masses, which traversed over the Greenland Sea, experienced more solar insolation and a relatively lower CS average, compared to the air masses coming over the Barents Sea (see Table 1). The combination of both high solar insolation and relatively low CS, could also explain why NPF likelihood is higher for the Greenland Sea as opposed to Barents Sea. One further difference is that the Barents Sea has a much larger sea ice coverage compared with the Greenland Sea, and thus there is potentially less open ocean in contact with the lower atmosphere during the NPF season. The reduced availability of DMS given the greater sea ice coverage could additionally help explain the differences in NPF likelihood of the Greenland and Barents Seas. The Arctic Ocean is another NPF source region (see Fig. 7 and Table 1). Considering that the Arctic Ocean sector is considerably more covered by sea ice then the NPF precursors are potentially being emitted from sea ice, e.g. iodine and sympagic algae. The Arctic Ocean contributes significantly, in terms of the portion of arriving air masses, between March to May and September to October. HIO_2 and HOMs are both significant contributors prior to May and HOMs increases in concentration after June.

Deleted: (Ghahreman et al., 2019)

Deleted: and t

Deleted: ,

Deleted: we are unable to replicate the findings presented in Park et al. (2018) for the years 2022 and 2023 (not shown here).

Deleted: (Lee et al., 2023)

Deleted: By averaging over large enough areas and time periods, distinctive DMS-to-chlorophyll exposure ratios may present themselves for different marine source regions, as the time scales involved in the oxidation of gaseous precursors influence the results less.

Deleted: (Park et al., 2018)

Formatted: Subscript

Deleted: typically

1015 4.1 Predicting NPF likelihood with condensation sink and solar radiation

1020 In Sect. 3.1.2, we show that NPF events and increases in NPF intensity, typically occur during periods of high solar insolation. High solar insolation seems to be a prerequisite for NPF, as it leads to the photochemical production of nucleating compounds. NPF events also seem to require the total surface area of pre-existing aerosol particles (i.e., CS) to decrease below seasonal averages (see Fig. 2). The distribution of solar insolation is statistically significantly different for periods experiencing events as opposed to non-events. For CS it depends on the classification scheme used (as demonstrated using the Mann-Whitney U test, see Fig. S29 a) – b) in supplement). NPF events require the concentration of nucleating vapours to exceed a critical concentration, and as the CS acts as a sink of those highly condensable vapours, low CS values promote the potential for nucleation. In this section, we argue that solar insolation and CS can be used as the two central parameters to determine, with high certainty the likelihood that a given day will experience an NPF event. For the definition of the term solar insolation refer to Sect. 2.4.1 in the methods.

The majority of measurements feature the combination of both relatively low solar insolation and low CS ($<1000 \text{ hW m}^{-2}$ and $<0.5 \cdot 10^{-3} \text{ s}^{-1}$, see bottom-left of Fig. S25 b) in supplement), and on very few occasions there are measurements with a combination of high solar insolation ($>1000 \text{ hW m}^{-2}$) and low CS ($<0.5 \cdot 10^{-3}$) (see top-left of Fig. S25 b)). Nonetheless, we show that when the combination of high solar insolation ($>1000 \text{ hW m}^{-2}$) and relatively low CS ($<0.5 \cdot 10^{-3} \text{ s}^{-1}$, i.e. top-left of Fig. S25 b)), arises, an NPF event that day is likely to occur (as defined by Aliaga et al. (2023), see Fig. 8). We find that periods exhibiting high solar insolation ($>1000 \text{ hW m}^{-2}$) and relatively low CS ($<0.5 \cdot 10^{-3} \text{ s}^{-1}$) are linked to more NPF days. Overall, the combinations of CS and solar insolation which promote NPF are relatively rare, compared to the entire set of measurements. The probability of occurrence is derived from the ratio of the number of measurements for each combination of CS and solar insolation during an NPF day to the number of valid data points for each of these combinations (where both sets of data are composed of hourly measurements) (see Sect. S10.1 in the supplement for more details on the normalisation approach)).

Deleted: Above,

Formatted: Font: 10 pt

Deleted: The

Deleted: showed that solar insolation statistically significantly alters the distributions of NPF intensity and frequency. The

Deleted: reach higher than

Deleted: s

Deleted: ing

Deleted: 4

Deleted: 4

Formatted: Superscript

Deleted:

Deleted: 8a

Deleted: 2

Deleted: 4

Deleted: 8a

Deleted:)

Deleted: 4

Deleted: 4

Deleted:)

Deleted: 8b

Deleted:)

Deleted: it occurs during an

Deleted: Aliaga et al. (2023)

Deleted: 4

Deleted: 4

Deleted: (

Deleted: for each of these combinations

Deleted: Fig.

Deleted: 8b

Deleted: ¶

We define a region of interest (ROI) (represented using the dashed line, separating the top left-hand corner in figure 9b). The ROI represents a region where the likelihood that a set of measurements is linked to an NPF day is on average, mean (median), 66.3 (68.4) % (ROI consists of 16% of data). In the ROI, the combination of low CS and plentiful solar insolation provide the most favourable conditions for NPF days to occur, and on the contrary, outside of the ROI, NPF days are much less likely to occur (i.e. bottom-right of Fig. 8b)). Moreover, the region outside of the ROI with low solar insolation and low CS might be associated with increased cloudiness as the incident radiation is reduced by the presence of clouds, and large accumulation mode particles are reduced due to the effect of wet scavenging. The region with low solar insolation and high CS could be related to Arctic Haze conditions.

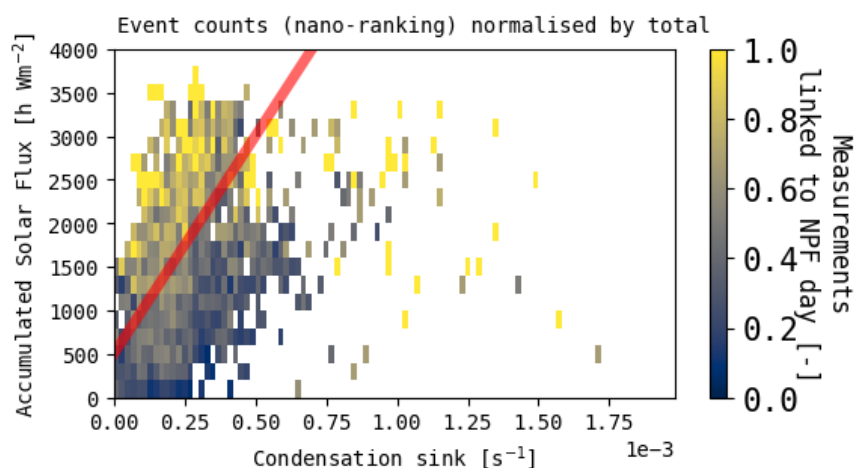


Fig. 8: Grid mesh where the data is binned by accumulated solar insolation and condensation sink (CS) (CS between $0 - 10^{-3} \text{ s}^{-1}$ every $2 \cdot 10^{-5} \text{ s}^{-1}$ and solar insolation between $0 - 4 \cdot 10^4 \text{ h W m}^{-2}$ every 200 h W m^{-2}). Each grid box displays the likelihood, calculated by normalising the data set for the event days (defined using Aliaga et al. (2023)) by the total count for each given bin (i.e. see Fig. S25 a) and b) in supplement)). Likelihood is the fraction of measurement combinations lined to an NPF day. The red line defines the region of interest (ROI) whereby data on the left of it is considered to have the ideal conditions for NPF i.e. high accumulated solar flux and relatively low CS. The equation for the red line is: $\text{Solar insolation} = 5 \cdot 10^6 \cdot \text{CS} + 500$. The grids required a count greater than 1.

We define a region of interest (ROI) (represented using the red line, separating the top left-hand corner in figure 8). The ROI is defined using a simple straight line to separate ideal and non-ideal NPF conditions from each other. It should be stressed that it should not be used as an empirical relationship, defining the occurrence of NPF, instead it should be used as a guideline to highlight the general conditions most suitable for the promotion of NPF. In the ROI, the combination of low CS and plentiful solar insolation provide the most favourable conditions for NPF days to occur, and on the contrary, outside of the ROI, NPF days are much less likely to occur (i.e. bottom-right of Fig. 8).

Moreover, the region outside of the ROI with low solar insolation and low CS might be associated with increased cloudiness as the incident radiation is reduced by the presence of clouds, and large accumulation mode particles are reduced due to the effect of wet scavenging. The region with low solar insolation and high CS could be related to Arctic Haze conditions.

Formatted Table

Formatted: Font: 9 pt, Font color: Black

Commented [DHR2]: You can show this – by colouring by cloudiness/brightness

Deleted: ¶

The ROI is defined indicatively using a simple straight line to separate ideal and non-ideal NPF conditions from each other. It should not be used as an empirical relationship, defining the occurrence of NPF. Instead it should be used as a guideline to highlight the general conditions most suitable for the promotion of NPF. ¶

The ROI defined in Figure 8 is statistically significantly different from the non-ROI region on the probability that an NPF day occurs given such solar insolation and CS measurements occurring. The probabilities which arise when normalisation is done via the Nano-ranking classification scheme, as opposed to the Dal Maso et al. (2005) scheme, are more distinct between the ROI and non-ROI regions (see Fig. 9). Here, the statistical significance between the two regions is conducted using the non-parametric Mann-Whitney U test (where H_0 representing two identical distributions is rejected). In addition, the ROI and non-ROI represented 23% and 77% of the data respectively; for the ROI 62.5% of the hourly data were linked to NPF days, as opposed to 23.8% in the non-ROI. The total number of Nano_{events} is 229 (g2=178, g3=51, see Fig.S16 in supplement). Approximately, hourly data points from 98% of the days with NPF events can be found in the non-ROI part, as opposed to 55% of the days with NPF events in the ROI.

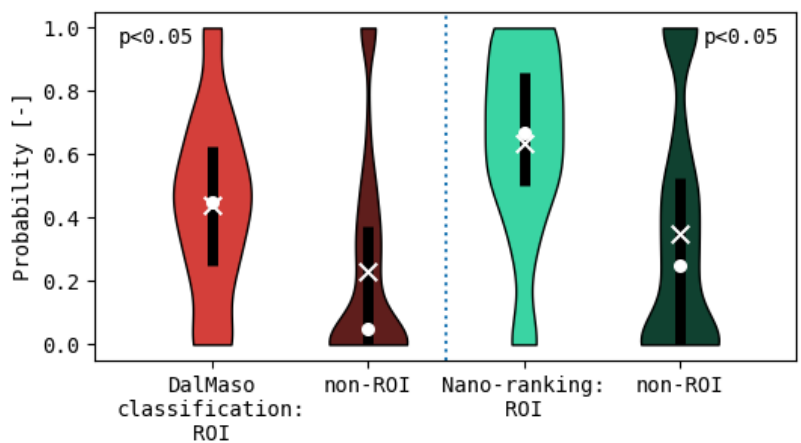


Fig 9: Violin plots representing the distribution of the probabilities as defined by the normalisation plot (see Fig. 8). The red violins on the left-hand-side represent the probability distributions related to the Dal Maso et al. (2005) classification and the green violin on the right-hand side represent the probability distributions according to the Nano-ranking approach.

The data within the ROI and when they occur can be used to try and estimate the likelihood of a day experiencing an NPF event. A time series representing the daily likelihood of NPF occurrence (i.e. $P_{NPF,D}$) can be generated from the ROI data (see Fig. 10) by labelling hourly data points as either 1 (i.e. within the ROI) or 0 (i.e. outside of the ROI) and then taking daily averages before applying a rolling average, to create the daily likelihood, $P_{NPF,D}$.

Formatted Table

Formatted: Font: 9 pt, Bold

Formatted: Font: 9 pt, Bold

Formatted: Font: 9 pt, Bold

Formatted: Font: 9 pt, Font color: Black

Formatted: Font: 9 pt, Bold

Deleted: 9

Deleted: counting the number of weekly data points within the ROI and dividing it by the total number of weekly data points either within or outside of the ROI (i.e. all). The weekly probabilities that measurements fall within the ROI are interpolated

$P_{NPF,D}$ generally, matches the frequency of NPF events as defined by Aliaga et al. (2023) or Dal Maso et al. (2005) (see Fig. 10), however, there are some major discrepancies. Generally, the predicted frequency of occurrence (i.e., defined by the ROI) is significantly lower than the observations. The spring and autumn peaks in $P_{NPF,D}$ and the event frequencies line up well in terms of the number of peaks, however, 2023 displays essentially only peak in $P_{NPF,D}$. Using $P_{NPF,D}$ we can replicate the distinctive decline in the NPF frequency around July in 2022 and 2024. However, in July 2023, the decline in NPF events in the observations is followed by an increase (i.e. two more sequential peaks in the observations) which is not captured by $P_{NPF,D}$. Still, it is striking that the two parameters alone (CS and solar insolation) are able to serve as predictors for NPF probability, by representing the balance between the production and removal of NPF-precursors. The solar insolation reflects the photochemical production potential (likely via the production of hydroxyl radicals). However, the parameter combination does not account for the reactants required to produce the nucleating vapours (which may be SO₂ (likely via DMS) and/or iodine compounds). Hence, deviations like those observed during July 2023 are expected as the conceptual model is partly incomplete due to a lack of adequate information regarding the original reactants. This said, however, the NPF frequency and likelihood largely seem limited by the balance between photochemical production potential and CS, and this in turn is suggestive of, to some degree, a self-regulating mechanism regarding the number of particles present in the nucleation and Aitken modes. There exist more sophisticated models to predict NPF (Kuang et al., 2010).

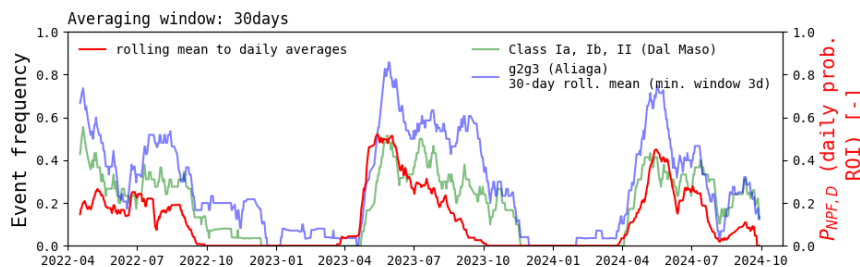


Fig 10: The red line is $P_{NPF,D}$, the estimated daily likelihood of measurements occurring in the region of interest (ROI). The ROI is based on the relation between CS and 6-hour solar accumulation and is defined in Fig. 8). The green and blue lines represent the rolling averages of the NPF daily event frequencies using the Dal Maso et al. (2005) and Aliaga et al. (2023) classification respectively. The green and blue curves are present here to be able to compare to $P_{NPF,D}$ (daily prob. ROI).

Deleted: Aliaga et al. (2023)

Deleted: Dal Maso et al. (2005)

Deleted: 9

Deleted: .

Deleted: through

Deleted: , though in the summer (around July), the event frequency is high although daily $P_{NPF,D}$ is low

Deleted: rapid

Deleted: (Kuang et al., 2010).

Formatted Table

Formatted: Font: 9 pt, Font color: Black

Formatted: Font: 9 pt, Font color: Black

4.1.1 Ratio of source and sink: ratio of solar irradiance

In section 4.1 we defined a region of interest, whereby measurements within this ROI provide an indication of the conditions likely to lead to an NPF day. By observing when these conditions occur offers one approach, however, the ability to predict NPF day occurrence can be improved by simply using the proxies for a source and a sink, and taking their respective ratio (i.e., Solar insolation/CS) (see Fig. 11). For a rolling average window of 30 days the R^2 , comparing solar insolation/CS and the nano-ranking frequency, is 0.72 (see Fig. S26 in supplement).

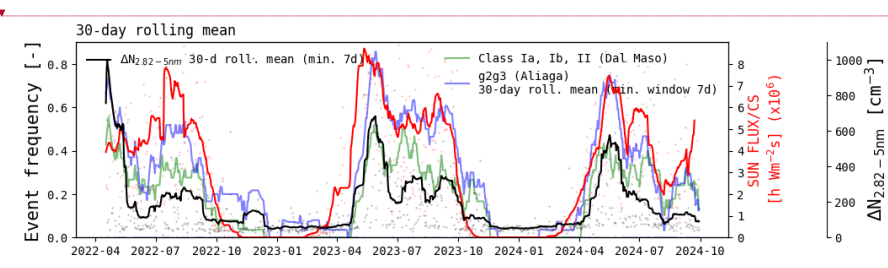


Fig 11: The red line is the solar insolation divided by the condensation sink (Solar insolation/CS) with a 30-day rolling mean. The intensity parameter ($\Delta N_{2.82-5nm}$). The green and blue lines represent the rolling averages of the NPF daily event frequencies using the Dal Maso et al. (2005) and Aliaga et al. (2023) classification respectively. The green and blue curves are present to be able to compare to the red line representing a proxy for the balance between sources and sinks.

Figure 11 demonstrates that by using a simplified parameter i.e., solar insolation/CS, albeit with a large rolling averaging window, we achieve a good agreement with both event frequencies (i.e., classified using the Dal Maso et al. (2005) or Aliaga et al. (2023) approach). Overall, the relationship between Solar insolation/CS varies depending on the averaging window utilised (see Fig. 12). The coefficient of determination (R^2) comparing solar insolation/CS and the daily event frequency (i.e., rolling average applied) increases and begins to plateau after a window of 2 weeks is applied (14 days, $R^2=0.65$) (see Fig. 12). Considering that a degree of averaging needs to be performed as the daily event frequencies are given as 1s (event) and 0s (non-event), then it seems justified that the solar insolation/CS parameter offers a robust predictive capability. There are some slight discrepancies between the solar insolation/CS parameter and the event frequencies, particularly at the start and end of the measurement period; the discrepancies could be due to the NPF seasons having started already or not having ended.

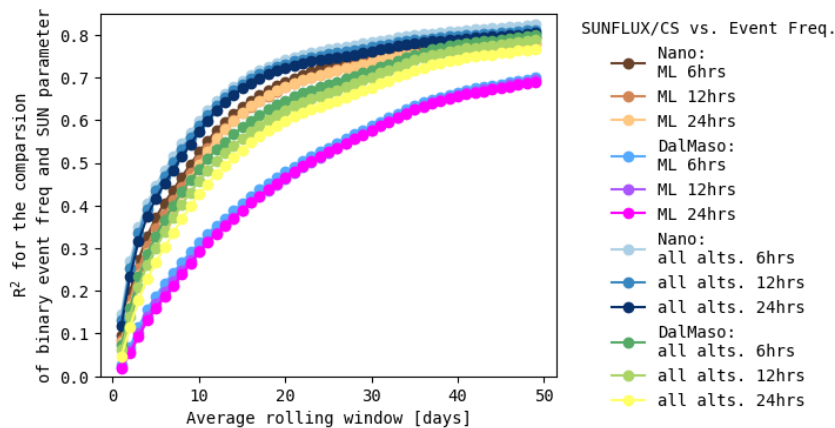


Fig 12: The coefficient of determination is displayed as a function of the size of the rolling window average (in number of days). The length of the back trajectory varies from the first 6, 12 and 24 hours. Each length is given a slightly different shade. A rolling average is applied to the accumulated solar insolation, and the event frequency. Parts of the back trajectory which only traversed through the mixed-layer are labelled as ML, whilst the entire back trajectory is given by all alts. The brown curves represent the comparison between the solar insolation within the mixed-layer part of the back trajectory and the event frequency as defined by the Nano-ranking approach. Each curve is compared to the size of the averaging window; blue and pink curves are the comparison between the event frequency defined using the Dal Maso classification, blue curves the

4.2 Dynamics driving NPF and growth

During NPF events there is an excess availability of condensable precursors, e.g. SA, which contribute to the growth of newly formed and pre-existing particles, thus increasing CS. The increased CS enhances the removal of condensable and nucleating vapours. As a result, a typical day with sustained NPF can lead to a relatively high CS ($>0.5 \cdot 10^{-3} \text{ s}^{-1}$), although the events tend to begin at low CS values ($\sim 0.1 \cdot 10^{-3} \text{ s}^{-1}$) (see Fig. 13). It is worth noting, that during an event, not only do the nucleation mode particles grow in size, but also all other existing particles grow as well. Practically, this means that, the sink is doubled during the NPF-event. Consequently, and assuming steady state, this would mean that the source rate of the condensing species must double to sustain the same concentration of condensing material.

Formatted: Centered

Formatted Table

Formatted: Font: 9 pt, Bold

Formatted: Justified

Formatted: Font: 9 pt, Bold

Formatted: Font: Bold

Formatted: Font: Bold

Formatted: Font: 9 pt, Bold

Formatted: Font: 9 pt, Bold

Formatted: Font: Bold

Formatted: Font: 9 pt, Bold

Deleted: ¶

Deleted: 4

Deleted: 0

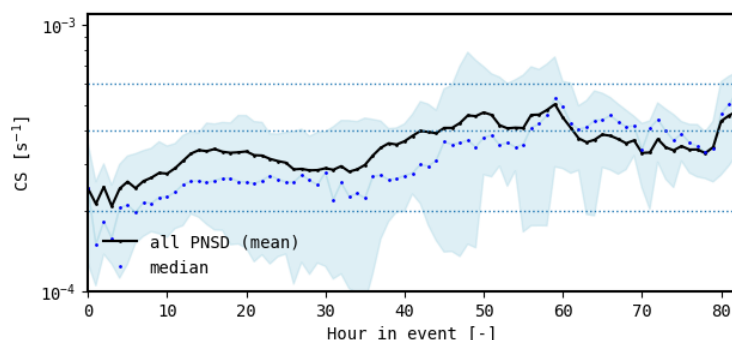


Fig. 13: The calculated condensation sink (CS) during NPF events. It should be noted that the average NPF typically lasts for 10 hours; however, in some cases events can persist for days until the newly formed mode is removed via wet scavenging or advected away from ZEP. The black line is the mean and the blue is the median. The blue shaded region represents 25th-75th percentiles. The dashed blue lines display CS values of 0.1, 0.2 and 0.4·10⁻³ s⁻¹ and are there to guide the reader.

Deleted: 0

The effect of this additional increase in CS during NPF events can be illustrated by using the ROI. The increase in CS shifts the ambient conditions away from the defined ROI ($>0.5 \cdot 10^{-3} \text{ s}^{-1}$) (see Fig. 8), thus potentially suppressing any subsequent NPF events. This mechanism will prevent further NPF, if the increase in CS during the NPF event is large enough to limit the availability of precursors. This mechanism also helps to explain why the combination of low CS and high solar insolation measurements are seemingly rare occurrences, as these conditions are soon followed by NPF, and increasing CS.

Deleted: T

Deleted: his

Deleted: ,

Deleted: 4

Deleted: b)

During the Arctic Haze period in late winter and early spring, high concentrations of accumulation mode aerosol dominate the contribution to the CS. Reduction in accumulation mode aerosol at the end of the Arctic Haze season (i.e., end of May) should result in decreased CS, but ZEP continues to experience relatively high CS. We argue that this relatively high sustained CS is the result of particle formation and growth within the Arctic; the increasing frequency of NPF events during the summer months increases CS, and maintains the comparatively high CS throughout the summer and the most intense part of the NPF season (see Fig. 2b)), additionally Tunved et al. (2013). Towards the late summer and autumn, the solar insolation has decreased substantially, and both marine surface water productivity and photochemical production potential are declining. Together with low transport efficiency from anthropogenic sources at lower latitudes and efficient wet removal, the aerosol reaches a minimum with respect to CS and number during this part of the year.

Deleted: Tunved et al. (2013)

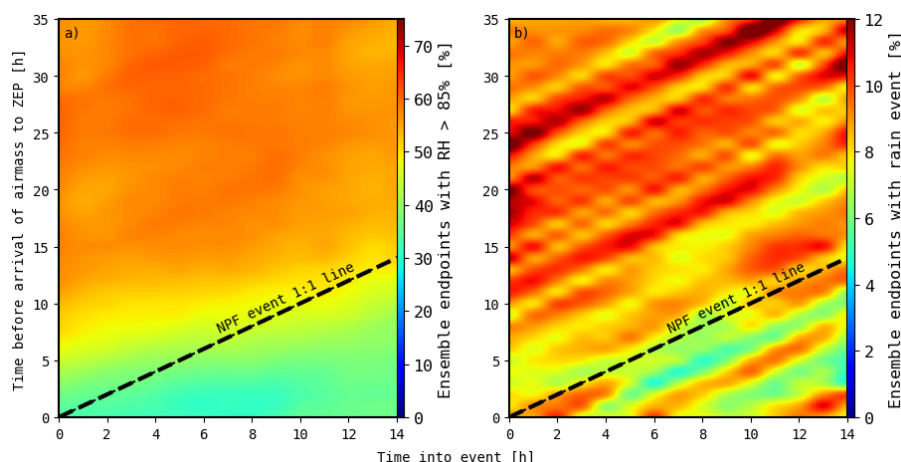


Fig 14: Air mass history for a) percentage of back trajectory endpoints within an ensemble with a relative humidity above 85, and b) percentage of endpoints within ensemble which experienced a rainfall event (i.e. rainfall > 0mm). Y-axis is the number of hours backwards before the air masses arrive at ZEP, while the x-axis is the number of hours into the event e.g. if one event started at 6am and another at 9am, their start times would both be 0 hours. The black line is a 1:1 line which represents the start of the NPF event such that the grid points towards the top-left represent endpoints prior to the NPF event, whilst grid points towards the bottom-right represent endpoints during the NPF event. The x-axis has been sliced based on the typical duration of events (most events are <14 hours). This figure consists of the air mass history of Class Ia and Ib NPF events.

Deleted: 1

In the Arctic environment, wet removal is a key factor responsible for the rapid reduction in CS (Garrett et al., 2011). For class Ia and Ib NPF events, we can observe that prior to the start an NPF event there is an increased likelihood of cloudy conditions (RH >85%) and rainfall events (see Figs 14 a-b). Figures 14 a-b demonstrate how wet scavenging helps to effectively reduce the CS and provide more suitable conditions for NPF events to occur.

Deleted: (Garrett et al., 2011)

Deleted: 1

Deleted: 1

The changes in the likelihood of cloudy conditions and rainfall prior to and post the start of an NPF event is slight, but significantly different, especially considering the duration prior to the event is longer and the likelihood of any rainfall should be compounded.

Any reduction in CS will shift the system towards the ROI, and as a result the system will rapidly respond to this reduction by producing new particles and grow the already existing ones. If suitable gaseous precursors and solar insolation are present, NPF will take place and the cycle will be reinitialised. This conceptual model, whilst highly

1225 simplified, offers substantial predictive value regarding the timing and frequency of NPF events (see Fig. 10). It
 1230 does not, however, account for the original source of condensable and nucleating species nor does it consider cloud
 activation and the in-cloud chemistry of Aitken mode particles in non-precipitating clouds.

4.3 NPF contribution to Aitken and potential CCN

Here, we argue that new particle formation is the main contributor to the Aitken mode in the Arctic; through
 1230 frequent NPF events and the subsequent growth of nucleation mode particles, NPF activity is able to provide a
 significant proportion of the overall concentration of Aitken mode particles. We show in Sect 3.1.2 that at ZEP
 the production of nucleation mode particles is largest at the end of spring, however, the production of newly formed
 particles persists throughout the summer months and into the autumn (see Fig. 2a)). During NPF events, including
 formation and subsequent growth the total number concentration of Aitken mode particles (5-70nm) is 3.6 times
 1235 the amount compared with pre-event conditions (see Fig. 16). These events are relatively frequent (30 - 40% of
 summer days experience NPF, (see Fig. 1). Fundamentally, though, it is difficult to use in-situ measurements at
 a single location to provide an estimation of the fraction of Aitken mode particles which are produced via NPF;
 for the Aitken mode particles at ZEP which are not directly linked to NPF activity, we cannot rule out the
 possibility that they are derived from NPF events outside of the vicinity of ZEP and then transported downwind.

1240 At ZEP, NPF and the subsequent growth that follows, results in particles large enough to potentially be activated
 and become CCN; on numerous occasions (57 times) we observed newly formed particles able to grow beyond
 25nm while in the same regional event (see Fig. 15b), before being scavenged. Here, we use 25nm as the lowest
 bound for particles to act as CCN, however, it should be stated that this limit requires strong updrafts and extremely
 hygroscopic particles for particles to be activated. We observe that after ~30 hours of growth, we reach potential
 1245 CCN-sizes at ZEP (see Fig. 15a)). In addition, 11, 8, and 3 NPF-produced particles above 60, 70 and 80nm
 respectively were observed (see Fig. 15b)). It is important to note that once produced these particles most likely
 continue to grow downwind of ZEP by condensation and in-cloud processing, so a large fraction of the Aitken
 mode should inevitably reach CCN size in due time, as long as they are not otherwise scavenged.

Deleted: 9

Deleted: 3

Commented [DH3]: Change to the exact numbers

Deleted: >

Deleted: X

Deleted: 0

Deleted: 2

Deleted: 2

Deleted: XX

Deleted: XX

Deleted: XX

Deleted:

Deleted: 2

Deleted: as well

Deleted: downstream

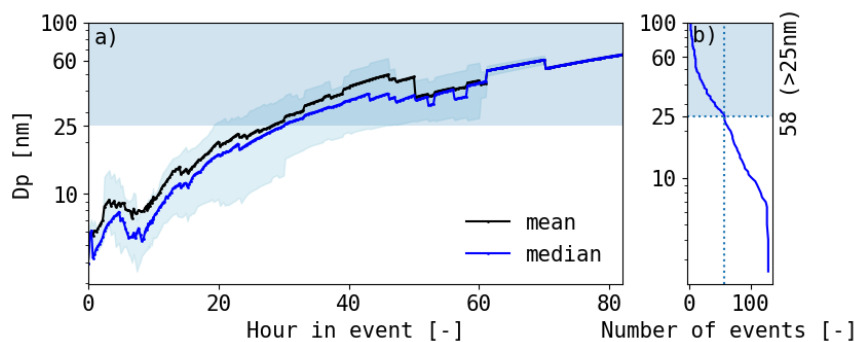


Fig 15: Change in diameter during all class Ia, Ib and II NPF events. The blue shaded region covers 25nm – 60nm. (a) shows the average change in the diameter of the nucleation mode over time. The black and blue lines signify the mean and median diameters respectively. The blue shaded region displays the distance between the 25th and 75th percentiles respectively (b) displays the number of NPF events against the maximum diameter reached during growth. The horizontal dashed line marks 25nm whilst the vertical dashed line marks the number of NPF events which reached 25nm.

It is difficult to ascertain the proportion of NPF-derived Aitken mode particles that actually become CCN before being scavenged or deposited, and estimating an exact fraction is beyond the scope of this study. However, previous modelling and measurement studies can help to shed some light on the likelihood that NPF events as measured from ZEP can contribute to the CCN budget. Jung et al. (2018) demonstrate that at supersaturations (ss) greater than 0.4% there is a considerable number of CCN measured during the summer months at ZEP. Moreover, the changes in sources of CCN-sized aerosol particles, from mainly anthropogenic to NPF-derived, during the Haze to summer transition can be seen in the reduction of the geometric mean diameter. For summer, Jung et al. (2018) suggest a significant number of particles >10nm are activated into cloud droplets, as the activation ratio during summer is around 0.5 for fairly high ss, i.e. >0.6%. In addition, Karlsson et al. (2021) show that Aitken mode-sized cloud residuals are fairly common in the summer time (i.e. July onwards); the transition from accumulation mode-dominated to Aitken mode-dominated aerosol did not lead to a significant reduction in the concentration of cloud residuals, indicating that particles as small as 20nm can be activated. Motos et al. (2023) suggest that particles >20nm are activated in the summer, with a modal activation diameter of 60nm showing that Aitken mode particles act as CCN.

Formatted: Justified

Deleted: 2

Formatted: Superscript

Deleted: (

Deleted: ion

Deleted: Jung et al. (2018)

Deleted: Jung et al. (2018)

Deleted: Karlsson et al. (2021)

Deleted: Motos et al. (2023)

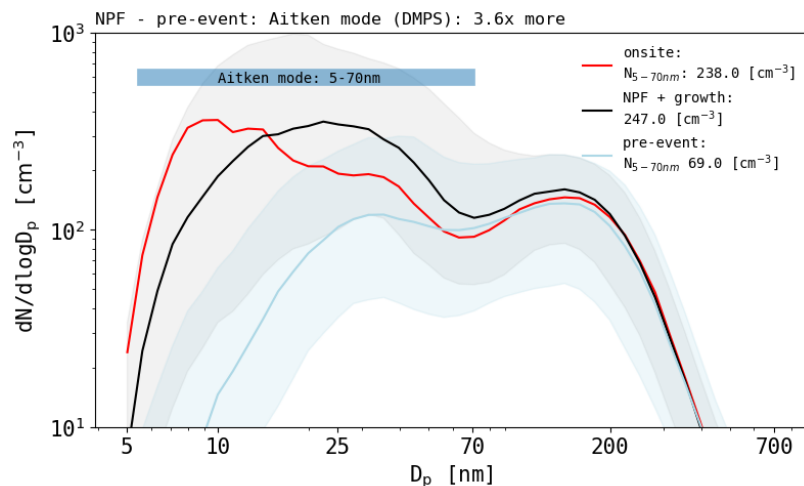


Fig 16: The particle number size distributions (DMPS) for periods before the NPF events (1-2 hours before event started) (pre-event, light blue), during NPF and growth (black) and on-site nucleation (production of 2-4nm particles, red, on-site formation).

Deleted: 3

5. Conclusion

In this study, we explore the annual cycle of atmospheric new particle formation (NPF) events from almost 3-years' worth of measurements (2022-2024). In keeping with previous studies, e.g., [Tunved et al. \(2013\)](#) and [Ström et al. \(2009\)](#), the Arctic summer and more specifically the end of spring and start of summer represent a peak in NPF intensity as observed at Zeppelin Observatory (ZEP). We find that this peak in the strength of NPF events coincides roughly with the maximum solar insolation (defined as the last 6 hours of accumulated solar radiation along HYSPLIT back trajectories); solar insolation allows for the increased availability of precursors by encouraging oxidation. The NPF frequency during the summer remains fairly consistent, with an estimated 20-30% of days experiencing an NPF event. Towards the end of summer, the decline in the amount of solar insolation reduces the NPF intensity. In addition, it is also apparent that the condensation sink (CS i.e. linked to the total surface area of pre-existing aerosol) plays an important role, as the NPF intensity is more pronounced when there are reductions in CS. For NPF events, the minimum CS was observed at the start of the events and signifies that a reduced condensation sink required for events to begin. This study is the first to utilise the Nano-ranking method in the Arctic. It should be stated that special care needs to be taken in treating wind-blown events separately from NPF events. Furthermore, this study also highlights potential problems when sampling lines are submerged in

Deleted: Tunved et al. (2013)

Deleted: Ström et al. (2009)

Deleted: within the mixed-layer arriving at ZEP

1305 dense clouds. It is recommended that relative humidity, visibility and/or web cam footage is used to help screen for these in-cloud events when sampling at high altitudes.

Deleted: typically begin when there is less of a sink, in regard to precursors.

Typically, as the Arctic transitions from the Haze season to summer, cloud coverage and precipitation increase, which in turn, via wet scavenging, reduces the once predominant accumulation mode. However, the increased cloud cover, not only reduces the CS, but also reduces the solar insolation. Hence, the transition to summertime both promotes and discourages NPF. The latter stages of the summer and the autumn provide less solar insolation; the peak in solar insolation is skewed more towards the start of summer. These competing effects can explain why we observe such a reduction in NPF intensity as the summer progresses. The dip in frequency and intensity that occurs around July coincides with a decrease in brightness (solar insolation over potential solar radiation). It is the interplay between both solar insolation and condensation sink that can potentially give rise to NPF events, and thus cloudiness will also play a role in NPF activity given interplay with CS and solar radiation. We can only speculate that a warming climate may appear more similar to the late Arctic summer, being both warmer and wetter. Some studies indicate that cloud cover in the Arctic could potentially increase (Barton and Veron, 2012), or has (Eastman and Warren, 2010; Francis et al., 2009; Kay and Gettelman, 2009; Palm et al., 2010; Vihma et al., 2008), with periods of rapid sea ice loss.

Deleted: the

Deleted: time

Deleted: and this

Deleted: via wet scavenging

Deleted: the

Deleted: towards the end

Deleted: (Barton and Veron, 2012)

Deleted: (Eastman and Warren, 2010; Francis et al., 2009; Kay and Gettelman, 2009; Palm et al., 2010; Vihma et al., 2008)

Deleted: e

Deleted: a

Deleted: 9

1320 One main finding from this study is that we can, with rather good accuracy, predict the occurrence of an NPF day (Fig. S26 in the supplement) by using CS and solar insolation as the two main predictive parameters. Periods exhibiting low CS and high solar insolation provide the most ideal conditions for NPF events and the occurrence of these periods can reproduce the observed frequency of NPF events. It should be noted, though, that other factors play a role. The sources to the precursors of the condensable vapours, as well as additional meteorological and environmental parameters such as sea-ice cover, chlorophyll a concentration and the varying contributions from different source regions all can influence the occurrence and strength of NPF. However, from this study we argue that both solar insolation and CS can be used to create a rather simple predictive model, simply by taking the ratio of these two terms. The Arctic perhaps offers a unique environment, where the relationship between solar insolation and NPF is potentially stronger, due to the typically lower CS and lower overall solar radiation intensity (Bousiotis et al., 2021).

Deleted: .

Deleted: (Bousiotis et al., 2021)

1330 Quantifying the contribution of CCN from NPF on low-level clouds is beyond the scope of this study. However, we are able to demonstrate that periods of formation and subsequent growth lead to a significant increase in the proportion of the total number of Aitken mode particles. In combination with this, we show that on many occasions

1350 during NPF events, the nucleation mode particles grow to potential CCN sizes once they are formed (~~57~~ NPF
events lead to diameters >25nm). With the correct conditions, i.e. a relatively strong updraft, the formation of
particles >25nm could lead to droplet formation and in turn alter cloud properties.

Given the high frequency of onsite NPF and the large potential area for formation, it is not unlikely that a majority
of Aitken mode particles ~~originate~~ from NPF within or close to the Arctic region. Provided sufficient time for
1355 growth up to CCN-sizes, it seems clear that NPF events constitute a significant source of CCN in the Arctic. For
a more detailed picture of the aerosol budget in the Arctic and a more quantitative account of the NPF-contribution
to the overall Arctic CCN-population, large scale modelling evaluated against observations and combined with
process studies are required.

1360 **Code availability**

Code will be made available on GitHub when published.

Data availability

Data will be on the Bolin Centre database when published (<https://bolin.su.se/data/>).

Author Contributions

1365 Dominic Heslin-Rees (DHR), Radovan Krejci (RK), and Peter Tunved (PT) worked together to develop the
research questions and design the study. DHR performed data analysis and wrote the paper together with assistance
from RK and PT. There was good input and advice along the way from Ilona Riipinen (IR), Annica Ekman
(AK), Diego Aliaga (DA) and Martina Mazzini (MMi). Janne Lampilahti (JL) developed many of the NAIS
specific Python packages and showed DHR how to clean and service the NAIS. Stefania Gilardoni (SG) provided
1370 comments. SG, Mikko Sipilä (MS), Mauro Mazzola (MMa) and Roseline Thakur (RT) helped with the
Gruvabadet work. Ki-Tae Park (KTP) and Kitack Lee (KT) maintained and operated the Zeppelin DMS analyzer
are part of a collaborative effort between KOPRI and POSTECH research team. Kihong Park (KP) and Young
Jun Yoon (YJY) provided the Nano-SMPS data.

Acknowledgements

1375 This study would not be possible without the work done by the research engineers Birgitta Noone, Tabea Henning,
and Ondrej Tesar from ACES and the staff from the Norwegian Polar Institute (NPI). NPI provide invaluable on-

Commented [DHR4]: Exact number

Deleted: >

Deleted: 0

Deleted: in fact

1380 site support including attending to the NAIS. NPI is also acknowledged for substantial long-term support in
maintaining the measurements at Zeppelin Observatory (ZEP). We would also like to thank Kai Rosman for
developing the software we used for the instrumentation at ZEP. We would like to thank Ove Hermansen from
the Norwegian Institute for Air Research (NILU) for providing the ambient meteorological data.

Financial Support

1385 Aerosol size distribution measurements are supported by the Swedish EPA. Swedish participation was supported
by Swedish Environmental Protection Agency (Naturvårdsverket), Knut and Alice Wallenberg foundation
(KWA) and FORMAS funding agency.

Knut-and-Alice-Wallenberg Foundation within the Arctic Climate Across Scales (Project No. \,2016.0024), the
Swedish EPA's (Naturvårdsverket) Environmental monitoring program (Miljöövervakning), the Swedish
1390 Research Council FORMAS (Project "Interplay between water, clouds and Aerosols in the Arctic," \# 2016-
01427)

The aerosol observations at the Zeppelin station have been supported by the KAW Stiftelse (grant
no. 2016.0024), the Swedish Environmental Protection agency (Naturvårdsverket), and the ACTRIS Sweden
project supported by the Swedish Research Council

1395 The KAW project CLIVE and VR financed infrastructure project ACTRIS-Sweden.

■

Deleted: ¶

¶

Formatted: Font: 11 pt, Bold

References

- Aliaga, D., Tuovinen, S., Zhang, T., Lampilahti, J., Li, X., Ahonen, L., Kokkonen, T., Nieminen, T., Hakala, S., Paasonen, P., Bianchi, F., Worsnop, D., Kerminen, V.-M., and Kulmala, M.: Nanoparticle ranking analysis: determining new particle formation (NPF) event occurrence and intensity based on the concentration spectrum of formed (sub-5 nm) particles, *Aerosol Research*, 1, 81–92, <https://doi.org/10.5194/ar-1-81-2023>, 2023.
- Asmi, E., Kondratyev, V., Brus, D., Laurila, T., Lihavainen, H., Backman, J., Vakkari, V., Aurela, M., Hatakka, J., Viisanen, Y., Uttal, T., Ivakhov, V., and Makshtas, A.: Aerosol size distribution seasonal characteristics measured in Tiksi, Russian Arctic, *Atmos Chem Phys*, 16, 1271–1287, <https://doi.org/10.5194/acp-16-1271-2016>, 2016.
- Barton, N. and Veron, D.: Response of clouds and surface energy fluxes to changes in sea-ice cover over the Laptev Sea (Arctic Ocean), *Clim Res*, 54, 69–84, <https://doi.org/10.3354/cr01101>, 2012.
- Beck, L. J., Sarnela, N., Junninen, H., Hoppe, C. J. M., Garmash, O., Bianchi, F., Riva, M., Rose, C., Peräkylä, O., Wimmer, D., Kausiala, O., Jokinen, T., Ahonen, L., Mikkilä, J., Hakala, J., He, X. C., Kontkanen, J., Wolf, K. K. E., Cappelletti, D., Mazzola, M., Traversi, R., Petroselli, C., Viola, A. P., Vitale, V., Lange, R., Massling, A., Nøjgaard, J. K., Krejci, R., Karlsson, L., Zieger, P., Jang, S., Lee, K., Vakkari, V., Lampilahti, J., Thakur, R. C., Leino, K., Kangasluoma, J., Duplissy, E. M., Siivola, E., Marbouti, M., Tham, Y. J., Saiz-Lopez, A., Petäjä, T., Ehn, M., Worsnop, D. R., Skov, H., Kulmala, M., Kerminen, V. M., and Sipilä, M.: Differing Mechanisms of New Particle Formation at Two Arctic Sites, *Geophys Res Lett*, 48, <https://doi.org/10.1029/2020GL091334>, 2021a.
- Beck, L. J., Sarnela, N., Junninen, H., Hoppe, C. J. M., Garmash, O., Bianchi, F., Riva, M., Rose, C., Peräkylä, O., Wimmer, D., Kausiala, O., Jokinen, T., Ahonen, L., Mikkilä, J., Hakala, J., He, X. C., Kontkanen, J., Wolf, K. K. E., Cappelletti, D., Mazzola, M., Traversi, R., Petroselli, C., Viola, A. P., Vitale, V., Lange, R., Massling, A., Nøjgaard, J. K., Krejci, R., Karlsson, L., Zieger, P., Jang, S., Lee, K., Vakkari, V., Lampilahti, J., Thakur, R. C., Leino, K., Kangasluoma, J., Duplissy, E. M., Siivola, E., Marbouti, M., Tham, Y. J., Saiz-Lopez, A., Petäjä, T., Ehn, M., Worsnop, D. R., Skov, H.,

Kulmala, M., Kerminen, V. M., and Sipilä, M.: Differing Mechanisms of New Particle Formation at Two Arctic Sites, *Geophys Res Lett*, 48, <https://doi.org/10.1029/2020GL091334>, 2021b.

Bousiotis, D., Brean, J., Pope, F. D., Dall'Osto, M., Querol, X., Alastuey, A., Perez, N., Petäjä, T., Massling, A., Klenø Nøjgaard, J., Nordstrøm, C., Kouvarakis, G., Vratolis, S., Eleftheriadis, K., Niemi, J. V., Portin, H., Wiedensohler, A., Weinhold, K., Merkel, M., Tuch, T., and Harrison, R. M.: The effect of meteorological conditions and atmospheric composition in the occurrence and development of new particle formation (NPF) events in Europe, *Atmos Chem Phys*, 21, 3345–3370, <https://doi.org/10.5194/acp-21-3345-2021>, 2021.

Boyer, M., Aliaga, D., Quéléver, L. L. J., Bucci, S., Angot, H., Dada, L., Heutte, B., Beck, L., Duetsch, M., Stohl, A., Beck, I., Laurila, T., Sarnela, N., Thakur, R. C., Miljevic, B., Kulmala, M., Petäjä, T., Sipilä, M., Schmale, J., and Jokinen, T.: The annual cycle and sources of relevant aerosol precursor vapors in the central Arctic during the MOSAiC expedition, *Atmos Chem Phys*, 24, 12595–12621, <https://doi.org/10.5194/acp-24-12595-2024>, 2024.

Brean, J., Beddows, D. C. S., Harrison, R. M., Song, C., Tunved, P., Ström, J., Krejci, R., Freud, E., Massling, A., Skov, H., Asmi, E., Lupi, A., and Dall'osto, M.: Collective geographical ecoregions and precursor sources driving Arctic new particle formation, *Atmos Chem Phys*, 23, 2183–2198, <https://doi.org/10.5194/acp-23-2183-2023>, 2023.

Cai, R. and Jiang, J.: A new balance formula to estimate new particle formation rate: Reevaluating the effect of coagulation scavenging, *Atmos Chem Phys*, 17, 12659–12675, <https://doi.org/10.5194/acp-17-12659-2017>, 2017.

Covert, D. S., Wiedensohler, A., Aalto, P., Heintzenberg, J., McMurry, P. H., and Leck, C.: Aerosol number size distributions from 3 to 500 nm diameter in the arctic marine boundary layer during summer and autumn, *Tellus B Chem Phys Meteorol*, 48, 197–212, <https://doi.org/10.3402/tellusb.v48i2.15886>, 1996.

Croft, B., Martin, R. V., Richard Leaitch, W., Tunved, P., Breider, T. J., D'Andrea, S. D., and Pierce, J. R.: Processes controlling the annual cycle of Arctic aerosol number and size distributions, *Atmos Chem Phys*, 16, 3665–3682, <https://doi.org/10.5194/acp-16-3665-2016>, 2016.

Dada, L., Chellapermal, R., Buenrostro Mazon, S., Paasonen, P., Lampilahti, J., E Manninen, H., Junninen, H., Petäjä, T., Kerminen, V. M., and Kulmala, M.: Refined classification and characterization of atmospheric new-particle formation events using air ions, *Atmos Chem Phys*, 18, 17883–17893, <https://doi.org/10.5194/acp-18-17883-2018>, 2018.

Dall'osto, M., Beddows, D. C. S., Tunved, P., Krejci, R., Ström, J., Hansson, H. C., Yoon, Y. J., Park, K. T., Becagli, S., Udisti, R., Onasch, T., Ódowd, C. D., Simó, R., and Harrison, R. M.: Arctic sea ice melt leads to atmospheric new particle formation, *Sci Rep*, 7, <https://doi.org/10.1038/s41598-017-03328-1>, 2017.

Dal Maso, M., Kulmala, M., Lehtinen, K. E. J., Mäkelä, J. M., Aalto, P., and O'Dowd, C. D.: Condensation and coagulation sinks and formation of nucleation mode particles in coastal and boreal forest boundary layers, *Journal of Geophysical Research Atmospheres*, 107, <https://doi.org/10.1029/2001JD001053>, 2002.

Dal Maso, M., Kulmala, M., Riipinen, I., Wagner, R., Hussein, T., Aalto, P. P., Lehtinen, K. E. J., Maso, D., Kulmala, M., Riipinen, M., Wagner, I., Hussein, R., Aalto, T., and Lehtinen, P. P. &: Formation and growth of fresh atmospheric aerosols: eight years of aerosol size distribution data from SMEAR II, Hyytiälä, Finland, 2005.

Draxler, R. R., Spring, S., Maryland, U. S. A., and Hess, G. D.: An Overview of the HYSPLIT_4 Modelling System for Trajectories, Dispersion, and Deposition, *Australian Meteorological Magazine*, 295–308 pp., 1998.

Eastman, R. and Warren, S. G.: Interannual variations of arctic cloud types in relation to sea ice, *J Clim*, 23, 4216–4232, <https://doi.org/10.1175/2010JCLI3492.1>, 2010.

Francis, J. A., Chan, W., Leathers, D. J., Miller, J. R., and Veron, D. E.: Winter Northern Hemisphere weather patterns remember summer Arctic sea-ice extent, *Geophys Res Lett*, 36, <https://doi.org/10.1029/2009GL037274>, 2009.

Fuchs, N. A. and Sutugin, A. G.: Highly Dispersed Aerosols, Ann Arbor Science Publishers, Michigan, 1970.

Galí, M. and Simó, R.: A meta-analysis of oceanic DMS and DMSP cycling processes: Disentangling the summer paradox, *Global Biogeochem Cycles*, 29, 496–515, <https://doi.org/10.1002/2014GB004940>, 2015.

Garrett, T. J. and Zhao, C.: Increased Arctic cloud longwave emissivity associated with pollution from mid-latitudes, *Nature*, 440, 787–789, <https://doi.org/10.1038/nature04636>, 2006.

Garrett, T. J., Zhao, C., Dong, X., Mace, G. G., and Hobbs, P. V.: Effects of varying aerosol regimes on low-level Arctic stratus, *Geophys Res Lett*, 31, <https://doi.org/10.1029/2004GL019928>, 2004.

Garrett, T. J., Brattström, S., Sharma, S., Worthy, D. E. J., and Novelli, P.: The role of scavenging in the seasonal transport of black carbon and sulfate to the Arctic, *Geophys Res Lett*, 38, <https://doi.org/10.1029/2011GL048221>, 2011.

Ghahreman, R., Gong, W., Galí, M., Norman, A.-L., Beagley, S. R., Ayodeji, A., Zheng, Q., Lupu, A., Lizotte, M., Levasseur, M., and Leaitch, W. R.: Dimethyl sulfide and its role in aerosol formation and growth in the Arctic summer – a modelling study, *Atmos Chem Phys*, 19, 14455–14476, <https://doi.org/10.5194/acp-19-14455-2019>, 2019.

Gordon, H., Kirkby, J., Baltensperger, U., Bianchi, F., Breitenlechner, M., Curtius, J., Dias, A., Dommen, J., Donahue, N. M., Dunne, E. M., Duplissy, J., Ehrhart, S., Flagan, R. C., Frege, C., Fuchs, C., Hansel, A., Hoyle, C. R., Kulmala, M., Kürten, A., Lehtipalo, K., Makhmutov, V., Molteni, U., Rissanen, M. P., Stozkhov, Y., Tröstl, J., Tsagkogeorgas, G., Wagner, R., Williamson, C., Wimmer, D., Winkler, P. M., Yan, C., and Carslaw, K. S.: Causes and importance of new particle formation in

the present-day and preindustrial atmospheres, *Journal of Geophysical Research: Atmospheres*, 122, 8739–8760, <https://doi.org/10.1002/2017JD026844>, 2017.

Gormley, P. G. and Kennedy, M.: Diffusion from a Stream Flowing through a Cylindrical Tube, Source: *Proceedings of the Royal Irish Academy. Section A: Mathematical and Physical Sciences*, 163–169 pp., 1948.

Gramlich, Y., Siegel, K., Haslett, S. L., Freitas, G., Krejci, R., Zieger, P., and Mohr, C.: Revealing the chemical characteristics of Arctic low-level cloud residuals - in situ observations from a mountain site, *Atmos Chem Phys*, 23, 6813–6834, <https://doi.org/10.5194/acp-23-6813-2023>, 2023.

Heintzenberg, J. and Leck, C.: The summer aerosol in the central Arctic 1991–2008: Did it change or not?, *Atmos Chem Phys*, 12, 3969–3983, <https://doi.org/10.5194/acp-12-3969-2012>, 2012.

Hirsikko, A., Nieminen, T., Gagné, S., Lehtipalo, K., Manninen, H. E., Ehn, M., Hörrak, U., Kerminen, V. M., Laakso, L., McMurry, P. H., Mirme, A., Mirme, S., Petäjä, T., Tammet, H., Vakkari, V., Vana, M., and Kulmala, M.: Atmospheric ions and nucleation: A review of observations, <https://doi.org/10.5194/acp-11-767-2011>, 2011.

Jang, E., Park, K. T., Jun Yoon, Y., Kim, T. W., Hong, S. B., Becagli, S., Traversi, R., Kim, J., and Gim, Y.: New particle formation events observed at the King Sejong Station, Antarctic Peninsula - Part 2: Link with the oceanic biological activities, *Atmos Chem Phys*, 19, 7595–7608, <https://doi.org/10.5194/acp-19-7595-2019>, 2019.

Jang, S., Park, K. T., Lee, K., and Suh, Y. S.: An analytical system enabling consistent and long-term measurement of atmospheric dimethyl sulfide, *Atmos Environ*, 134, 217–223, <https://doi.org/10.1016/j.atmosenv.2016.03.041>, 2016.

Jokinen, T., Sipilä, M., Kontkanen, J., Vakkari, V., Tisler, P., Duplissy, E.-M., Junninen, H., Kangasluoma, J., Manninen, H. E., Petäjä, T., Kulmala, M., Worsnop, D. R., Kirkby, J., Virkkula, A., and Kerminen, V.-M.: Ion-induced sulfuric acid-ammonia nucleation drives particle formation in coastal Antarctica, *Sci. Adv*, 2018.

Jung, C. H., Yoon, Y. J., Kang, H. J., Gim, Y., Lee, B. Y., Ström, J., Krejci, R., and Tunved, P.: The seasonal characteristics of cloud condensation nuclei (CCN) in the arctic lower troposphere, *Tellus B Chem Phys Meteorol*, 70, 1–13, <https://doi.org/10.1080/16000889.2018.1513291>, 2018.

Kangasluoma, J., Cai, R., Jiang, J., Deng, C., Stolzenburg, D., Ahonen, L. R., Chan, T., Fu, Y., Kim, C., Laurila, T. M., Zhou, Y., Dada, L., Sulo, J., Flagan, R. C., Kulmala, M., Petäjä, T., and Lehtipalo, K.: Overview of measurements and current instrumentation for 1–10 nm aerosol particle number size distributions, <https://doi.org/10.1016/j.jaerosci.2020.105584>, 1 October 2020.

Karl, M., Leck, C., Gross, A., and Pirjola, L.: A study of new particle formation in the marine boundary layer over the central Arctic Ocean using a flexible multicomponent aerosol dynamic model, *Tellus B Chem Phys Meteorol*, 64, <https://doi.org/10.3402/tellusb.v64i0.17158>, 2012.

Karlsson, L., Krejci, R., Koike, M., Ebell, K., and Zieger, P.: The role of nanoparticles in Arctic cloud formation, <https://doi.org/10.5194/acp-2020-417>, 29 May 2020.

Karlsson, L., Krejci, R., Koike, M., Ebell, K., and Zieger, P.: A long-term study of cloud residuals from low-level Arctic clouds, *Atmos Chem Phys*, 21, 8933–8959, <https://doi.org/10.5194/acp-21-8933-2021>, 2021.

Kay, J. E. and Gettelman, A.: Cloud influence on and response to seasonal Arctic sea ice loss, *Journal of Geophysical Research Atmospheres*, 114, <https://doi.org/10.1029/2009JD011773>, 2009.

Kecorius, S., Vogl, T., Paasonen, P., Lampilahti, J., Rothenberg, D., Wex, H., Zeppenfeld, S., Van Pinxteren, M., Hartmann, M., Henning, S., Gong, X., Welti, A., Kulmala, M., Stratmann, F., Herrmann, H., and Wiedensohler, A.: New particle formation and its effect on cloud condensation nuclei abundance in the summer Arctic: A case study in the Fram Strait and Barents Sea, *Atmos Chem Phys*, 19, 14339–14364, <https://doi.org/10.5194/acp-19-14339-2019>, 2019.

Kerminen, V. M., Chen, X., Vakkari, V., Petäjä, T., Kulmala, M., and Bianchi, F.: Atmospheric new particle formation and growth: Review of field observations, <https://doi.org/10.1088/1748-9326/aadf3c>, 27 September 2018.

Kirkby, J., Curtius, J., Almeida, J., Dunne, E., Duplissy, J., Ehrhart, S., Franchin, A., Gagné, S., Ickes,
 1550 L., Kürten, A., Kupc, A., Metzger, A., Riccobono, F., Rondo, L., Schobesberger, S., Tsagkogeorgas,
 G., Wimmer, D., Amorim, A., Bianchi, F., Breitenlechner, M., David, A., Dommen, J., Downard, A.,
 Ehn, M., Flagan, R. C., Haider, S., Hansel, A., Hauser, D., Jud, W., Junninen, H., Kreissl, F., Kvashin,
 A., Laaksonen, A., Lehtipalo, K., Lima, J., Lovejoy, E. R., Makhmutov, V., Mathot, S., Mikkilä, J.,
 Minginette, P., Mogo, S., Nieminen, T., Onnela, A., Pereira, P., Petäjä, T., Schnitzhofer, R., Seinfeld,
 1555 J. H., Sipilä, M., Stozhkov, Y., Stratmann, F., Tomé, A., Vanhanen, J., Viisanen, Y., Vrtala, A.,
 Wagner, P. E., Walther, H., Weingartner, E., Wex, H., Winkler, P. M., Carslaw, K. S., Worsnop, D.
 R., Baltensperger, U., and Kulmala, M.: Role of sulphuric acid, ammonia and galactic cosmic rays in
 atmospheric aerosol nucleation, *Nature*, 476, 429–435, <https://doi.org/10.1038/nature10343>, 2011.
 Kirkby, J., Duplissy, J., Sengupta, K., Frege, C., Gordon, H., Williamson, C., Heinritzi, M., Simon,
 1560 M., Yan, C., Almeida, J., Trostl, J., Nieminen, T., Ortega, I. K., Wagner, R., Adamov, A., Amorim,
 A., Bernhammer, A. K., Bianchi, F., Breitenlechner, M., Brilke, S., Chen, X., Craven, J., Dias, A.,
 Ehrhart, S., Flagan, R. C., Franchin, A., Fuchs, C., Guida, R., Hakala, J., Hoyle, C. R., Jokinen, T.,
 Junninen, H., Kangasluoma, J., Kim, J., Krapf, M., Kurten, A., Laaksonen, A., Lehtipalo, K.,
 Makhmutov, V., Mathot, S., Molteni, U., Onnela, A., Perakyla, O., Piel, F., Petaja, T., Praplan, A. P.,
 1565 Pringle, K., Rap, A., Richards, N. A. D., Riipinen, I., Rissanen, M. P., Rondo, L., Sarnela, N.,
 Schobesberger, S., Scott, C. E., Seinfeld, J. H., Sipilä, M., Steiner, G., Stozhkov, Y., Stratmann, F.,
 Tomé, A., Virtanen, A., Vogel, A. L., Wagner, A. C., Wagner, P. E., Weingartner, E., Wimmer, D.,
 Winkler, P. M., Ye, P., Zhang, X., Hansel, A., Dommen, J., Donahue, N. M., Worsnop, D. R.,
 Baltensperger, U., Kulmala, M., Carslaw, K. S., and Curtius, J.: Ion-induced nucleation of pure
 1570 biogenic particles, *Nature*, 533, 521–526, <https://doi.org/10.1038/nature17953>, 2016.
 Kolesar, K. R., Cellini, J., Peterson, P. K., Jefferson, A., Tuch, T., Birmili, W., Wiedensohler, A., and
 Pratt, K. A.: Effect of Prudhoe Bay emissions on atmospheric aerosol growth events observed in
 Utqiagvik (Barrow), Alaska, *Atmos Environ*, 152, 146–155,
<https://doi.org/10.1016/j.atmosenv.2016.12.019>, 2017a.

Kolesar, K. R., Cellini, J., Peterson, P. K., Jefferson, A., Tuch, T., Birmili, W., Wiedensohler, A., and Pratt, K. A.: Effect of Prudhoe Bay emissions on atmospheric aerosol growth events observed in Utqiagvik (Barrow), Alaska, *Atmos Environ*, 152, 146–155, <https://doi.org/10.1016/j.atmosenv.2016.12.019>, 2017b.

Kristensson, adam, Johansson, martin, swietlicki, erik, Kivekäs, niku, hussein, tareq, nieminen, tuomo, Kulmala, markku, and Dal maso, miikka: nanomap: Geographical mapping of atmospheric new-particle formation through analysis of particle number size distribution and trajectory data, 2014.

Kuang, C., Riipinen, I., Sihto, S. L., Kulmala, M., McCormick, A. V., and McMurry, P. H.: An improved criterion for new particle formation in diverse atmospheric environments, *Atmos Chem Phys*, 10, 8469–8480, <https://doi.org/10.5194/acp-10-8469-2010>, 2010.

Kulmala, M., Petäjä, T., Nieminen, T., Sipilä, M., Manninen, H. E., Lehtipalo, K., Dal Maso, M., Aalto, P. P., Junninen, H., Paasonen, P., Riipinen, I., Lehtinen, K. E. J., Laaksonen, A., and Kerminen, V. M.: Measurement of the nucleation of atmospheric aerosol particles, *Nat Protoc*, 7, 1651–1667, <https://doi.org/10.1038/nprot.2012.091>, 2012.

Kulmala, M., Kontkanen, J., Junninen, H., Lehtipalo, K., Manninen, H. E., and Nieminen, T.: Direct Observations of Atmospheric Aerosol Nucleation, *Science* (1979), 339, 943–946, <https://doi.org/10.1126/science.1227385>, 2013.

Leaitch, W. R., Sharma, S., Huang, L., Toom-Sauntry, D., Chivulescu, A., Macdonald, A. M., Von Salzen, K., Pierce, J. R., Bertram, A. K., Schroder, J. C., Shantz, N. C., Chang, R. Y. W., and Norman, A. L.: Dimethyl sulfide control of the clean summertime Arctic aerosol and cloud, *Elementa*, 1, <https://doi.org/10.12952/journal.elementa.000017>, 2013.

Leaitch, W. R., Korolev, A., Aliabadi, A. A., Burkart, J., Willis, M. D., Abbatt, J. P. D., Bozem, H., Hoor, P., Köllner, F., Schneider, J., Herber, A., Konrad, C., and Brauner, R.: Effects of 20–100nm particles on liquid clouds in the clean summertime Arctic, *Atmos Chem Phys*, 16, 11107–11124, <https://doi.org/10.5194/acp-16-11107-2016>, 2016.

1600 [Lee, H., Lee, K., Rene Lunder, C., Krejci, R., Aas, W., Park, J., Park, K. T., Yong Lee, B., Yoon, Y. J., and Park, K.: Atmospheric new particle formation characteristics in the Arctic as measured at Mount Zeppelin, Svalbard, from 2016 to 2018, Atmos Chem Phys, 20, 13425–13441, <https://doi.org/10.5194/acp-20-13425-2020>, 2020.](#)

1605 [Lee, K., Kim, J. S., Park, K. T., Park, M. J., Jang, E., Gudmundsson, K., Olafsdottir, S. R., Olafsson, J., Yoon, Y. J., Lee, B. Y., Kwon, S. Y., and Kam, J.: Observational evidence linking ocean sulfur compounds to atmospheric dimethyl sulfide during Icelandic Sea phytoplankton blooms, Science of the Total Environment, 879, <https://doi.org/10.1016/j.scitotenv.2023.163020>, 2023.](#)

[Li, J., Carlson, B. E., Yung, Y. L., Lv, D., Hansen, J., Penner, J. E., Liao, H., Ramaswamy, V., Kahn, R. A., Zhang, P., Dubovik, O., Ding, A., Lacis, A. A., Zhang, L., and Dong, Y.: Scattering and absorbing aerosols in the climate system, <https://doi.org/10.1038/s43017-022-00296-7>, 1 June 2022.](#)

1610 [Manninen, H. E., Franchin, A., Schobesberger, S., Hirsikko, A., Hakala, J., Skromulis, A., Kangasluoma, J., Ehn, M., Junninen, H., Mirme, A., Mirme, S., Sipilä, M., Petäjä, T., Worsnop, D. R., and Kulmala, M.: Characterisation of corona-generated ions used in a Neutral cluster and Air Ion Spectrometer \(NAIS\), Atmos Meas Tech, 4, 2767–2776, <https://doi.org/10.5194/amt-4-2767-2011>, 2011.](#)

1615 [Maturilli, M. and Ebell, K.: Twenty-five years of cloud base height measurements by ceilometer in Ny-Ålesund, Svalbard, Earth Syst Sci Data, 10, 1451–1456, <https://doi.org/10.5194/essd-10-1451-2018>, 2018.](#)

1620 [Mauritsen, T., Sedlar, J., Tjernström, M., Leck, C., Martin, M., Shupe, M., Sjogren, S., Sierau, B., Persson, P. O. G., Brooks, I. M., and Swietlicki, E.: An Arctic CCN-limited cloud-aerosol regime, Atmos Chem Phys, 11, 165–173, <https://doi.org/10.5194/acp-11-165-2011>, 2011.](#)

[Merikanto, J., Spracklen, D. V., Mann, G. W., Pickering, S. J., and Carslaw, K. S.: Atmospheric Chemistry and Physics Impact of nucleation on global CCN, Atmos. Chem. Phys, 8601–8616 pp., 2009.](#)

1625 Mirme, A., Tamm, E., Mordas, G., Vana, M., Uin, J., Mirme, S., Bernotas, T., Laakso, L., Kulmala, M., and Hirsikko, A.: A wide-range multi-channel air ion spectrometer, *Boreal Environment Research*, 12, 247–264, 2007.

Mirme, S. and Mirme, A.: The mathematical principles and design of the NAIS - A spectrometer for the measurement of cluster ion and nanometer aerosol size distributions, *Atmos Meas Tech*, 6, 1061–1071, <https://doi.org/10.5194/amt-6-1061-2013>, 2013.

1630 Motos, G., Freitas, G., Georgakaki, P., Wieder, J., Li, G., Aas, W., Lunder, C., Krejci, R., Pasquier, J. T., Henneberger, J., David, R. O., Ritter, C., Mohr, C., Zieger, P., and Nenes, A.: Aerosol and dynamical contributions to cloud droplet formation in Arctic low-level clouds, *Atmos Chem Phys*, 23, 13941–13956, <https://doi.org/10.5194/acp-23-13941-2023>, 2023a.

1635 Motos, G., Freitas, G., Georgakaki, P., Wieder, J., Li, G., Aas, W., Lunder, C., Krejci, R., Pasquier, J. T., Henneberger, J., David, R. O., Ritter, C., Mohr, C., Zieger, P., and Nenes, A.: Aerosol and dynamical contributions to cloud droplet formation in Arctic low-level clouds, *Atmos Chem Phys*, 23, 13941–13956, <https://doi.org/10.5194/acp-23-13941-2023>, 2023b.

Nguyen, Q. T., Glasius, M., Sørensen, L. L., Jensen, B., Skov, H., Birmili, W., Wiedensohler, A., Kristensson, A., Nøjgaard, J. K., and Massling, A.: Seasonal variation of atmospheric particle number concentrations, new particle formation and atmospheric oxidation capacity at the high Arctic site Villum Research Station, Station Nord, *Atmos Chem Phys*, 16, 11319–11336, <https://doi.org/10.5194/acp-16-11319-2016>, 2016.

1640 Nieminen, T., Kerminen, V. M., Petäjä, T., Aalto, P. P., Arshinov, M., Asmi, E., Baltensperger, U., Beddows, D. C. S., Paul Beukes, J., Collins, D., Ding, A., Harrison, R. M., Henzing, B., Hooda, R., Hu, M., Hörrak, U., Kivekäs, N., Komsaare, K., Krejci, R., Kristensson, A., Laakso, L., Laaksonen, A., Richard Leaitch, W., Lihavainen, H., Mihalopoulos, N., Németh, Z., Nie, W., O'Dowd, C., Salma, I., Sellegri, K., Svenningsson, B., Swietlicki, E., Tunved, P., Ulevicius, V., Vakkari, V., Vana, M., Wiedensohler, A., Wu, Z., Virtanen, A., and Kulmala, M.: Global analysis of continental boundary

layer new particle formation based on long-term measurements, *Atmos Chem Phys*, 18, 14737–14756, <https://doi.org/10.5194/acp-18-14737-2018>, 2018.

Palm, S. P., Strey, S. T., Spinhirne, J., and Markus, T.: Influence of Arctic sea ice extent on polar cloud fraction and vertical structure and implications for regional climate, *Journal of Geophysical Research Atmospheres*, 115, <https://doi.org/10.1029/2010JD013900>, 2010.

Park, K. T., Lee, K., Kim, T. W., Yoon, Y. J., Jang, E. H., Jang, S., Lee, B. Y., and Hermansen, O.: Atmospheric DMS in the Arctic Ocean and Its Relation to Phytoplankton Biomass, *Global Biogeochem Cycles*, 32, 351–359, <https://doi.org/10.1002/2017GB005805>, 2018a.

Park, K. T., Lee, K., Kim, T. W., Yoon, Y. J., Jang, E. H., Jang, S., Lee, B. Y., and Hermansen, O.: Atmospheric DMS in the Arctic Ocean and Its Relation to Phytoplankton Biomass, *Global Biogeochem Cycles*, 32, 351–359, <https://doi.org/10.1002/2017GB005805>, 2018b.

Pei, Q., Saikawa, E., Kaspari, S., Widory, D., Zhao, C., Wu, G., Loewen, M., Wan, X., Kang, S., Wang, X., Zhang, Y. L., and Cong, Z.: Sulfur aerosols in the Arctic, Antarctic, and Tibetan Plateau: Current knowledge and future perspectives, <https://doi.org/10.1016/j.earscirev.2021.103753>, 1 September 2021.

Pirjola, L., O’Dowd, C. D., Brooks, I. M., and Kulmala, M.: Can new particle formation occur in the clean marine boundary layer?, *Journal of Geophysical Research: Atmospheres*, 105, 26531–26546, <https://doi.org/10.1029/2000JD900310>, 2000.

Platt, S. M., Hov, Ø., Berg, T., Breivik, K., Eckhardt, S., Eleftheriadis, K., Evangeliou, N., Fiebig, M., Fisher, R., Hansen, G., Hansson, H. C., Heintzenberg, J., Hermansen, O., Heslin-Rees, D., Holmén, K., Hudson, S., Kallenborn, R., Krejci, R., Krognes, T., Larssen, S., Lowry, D., Myhre, C. L., Lunder, C., Nisbet, E., Nizzetto, P. B., Park, K. T., Pedersen, C. A., Pfaffhuber, K. A., Röckmann, T., Schmidbauer, N., Solberg, S., Stohl, A., Ström, J., Svendby, T., Tunved, P., Tørnkvist, K., Van Der Veen, C., Vratolis, S., Yoon, Y. J., Yttri, K. E., Zieger, P., Aas, W., and Tørseth, K.: Atmospheric composition in the European Arctic and 30 years of the Zeppelin Observatory, Ny-Ålesund, *Atmos Chem Phys*, 22, 3321–3369, <https://doi.org/10.5194/acp-22-3321-2022>, 2022.

Pöhlker, M. L., Zhang, M., Campos Braga, R., Krüger, O. O., Pöschl, U., and Ervens, B.: Aitken mode particles as CCN in aerosol- And updraft-sensitive regimes of cloud droplet formation, *Atmos Chem Phys*, 21, 11723–11740, <https://doi.org/10.5194/acp-21-11723-2021>, 2021.

Price, R., Baccarini, A., Schmale, J., Zieger, P., Brooks, I. M., Field, P., and Carslaw, K. S.: Late summer transition from a free-Tropospheric to boundary layer source of Aitken mode aerosol in the high Arctic, *Atmos Chem Phys*, 23, 2927–2961, <https://doi.org/10.5194/acp-23-2927-2023>, 2023.

Quinn, P. K., Miller, T. L., Bates, T. S., Ogren, J. A., Andrews, E., and Shaw, G. E.: A 3-year record of simultaneously measured aerosol chemical and optical properties at Barrow, Alaska, *Journal of Geophysical Research: Atmospheres*, 107, <https://doi.org/10.1029/2001jd001248>, 2002.

Schmale, J. and Baccarini, A.: Progress in Unraveling Atmospheric New Particle Formation and Growth Across the Arctic, *Geophys Res Lett*, 48, <https://doi.org/10.1029/2021GL094198>, 2021.

Schmale, J., Sharma, S., Decesari, S., Pernov, J., Massling, A., Hansson, H. C., Von Salzen, K., Skov, H., Andrews, E., Quinn, P. K., Upchurch, L. M., Eleftheriadis, K., Traversi, R., Gilardoni, S., Mazzola, M., Laing, J., and Hopke, P.: Pan-Arctic seasonal cycles and long-term trends of aerosol properties from 10 observatories, *Atmos Chem Phys*, 22, 3067–3096, <https://doi.org/10.5194/acp-22-3067-2022>, 2022.

Shaw, G. E.: The Arctic Haze Phenomenon, 2403–2414 pp., [https://doi.org/https://doi.org/10.1175/1520-0477\(1995\)076<2403:TAHP>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0477(1995)076<2403:TAHP>2.0.CO;2), 1995.

Stein, A. F., Draxler, R. R., Rolph, G. D., Stunder, B. J. B., Cohen, M. D., and Ngan, F.: NOAA's hysplit atmospheric transport and dispersion modeling system, <https://doi.org/10.1175/BAMS-D-14-00110.1>, 1 December 2015.

Ström, J., Umegård, J., Tørseth, K., Tunved, P., Hansson, H. C., Holmén, K., Wismann, V., Herber, A., and König-Langlo, G.: One year of particle size distribution and aerosol chemical composition measurements at the Zeppelin Station, Svalbard, March 2000–March 2001, *Physics and Chemistry of the Earth*, 28, 1181–1190, <https://doi.org/10.1016/j.pce.2003.08.058>, 2003.

Ström, J., Engvall, A. C., Delbart, F., Krejci, R., and Treffeisen, R.: On small particles in the Arctic summer boundary layer: Observations at two different heights near Ny-Ålesund, Svalbard, *Tellus B Chem Phys Meteorol*, 61 B, 473–482, <https://doi.org/10.1111/j.1600-0889.2008.00412.x>, 2009.

Tammet, H., Komsaare, K., and Hörrak, U.: Intermediate ions in the atmosphere, *Atmos Res*, 135–136, 263–273, <https://doi.org/10.1016/j.atmosres.2012.09.009>, 2014.

Tunved, P., Ström, J., and Krejci, R.: Arctic aerosol life cycle: Linking aerosol size distributions observed between 2000 and 2010 with air mass transport and precipitation at Zeppelin station, Ny-Ålesund, Svalbard, *Atmos Chem Phys*, 13, 3643–3660, <https://doi.org/10.5194/acp-13-3643-2013>, 2013.

Vihma, T., Jaagus, J., Jakobson, E., and Palo, T.: Meteorological conditions in the Arctic Ocean in spring and summer 2007 as recorded on the drifting ice station Tara, *Geophys Res Lett*, 35, <https://doi.org/10.1029/2008GL034681>, 2008.

Wagner, R., Manninen, H. E., Franchin, A., Lehtipalo, K., Mirme, S., Steiner, G., Petäjä, T., and Kulmala, M.: On the accuracy of ion measurements using a Neutral cluster and Air Ion Spectrometer, 2016.

Wagner, R., Yan, C., Lehtipalo, K., Duplissy, J., Nieminen, T., Kangasluoma, J., Ahonen, L. R., Dada, L., Kontkanen, J., Manninen, H. E., Dias, A., Amorim, A., Bauer, P. S., Bergen, A., Bernhammer, A. K., Bianchi, F., Brilke, S., Buenrostro Mazon, S., Chen, X., Draper, D. C., Fischer, L., Frege, C., Fuchs, C., Garmash, O., Gordon, H., Hakala, J., Heikkinen, L., Heinritzi, M., Hofbauer, V., Hoyle, C. R., Kirkby, J., Kürten, A., Kvashnin, A. N., Laurila, T., Lawler, M. J., Mai, H., Makhmutov, V., Mauldin, R., Molteni, U., Nichman, L., Nie, W., Ojdanic, A., Onnela, A., Piel, F., Quéléver, L. L. J., Rissanen, M. P., Sarnela, N., Schallhart, S., Sengupta, K., Simon, M., Stolzenburg, D., Stozhkov, Y., Tröstl, J., Viisanen, Y., Vogel, A. L., Wagner, A. C., Xiao, M., Ye, P., Baltensperger, U., Curtius, J., Donahue, N. M., Flagan, R. C., Gallagher, M., Hansel, A., Smith, J. N., Tomé, A., Winkler, P. M., Worsnop, D., Ehn, M., Sipilä, M., Kerminen, V. M., Petäjä, T., and Kulmala, M.: The role of ions in

new particle formation in the CLOUD chamber, *Atmos Chem Phys*, 17, 15181–15197,

<https://doi.org/10.5194/acp-17-15181-2017>, 2017.

Von Der Weiden, S.-L., Drewnick, F., and Borrmann, S.: *Atmospheric Measurement Techniques*

Particle Loss Calculator-a new software tool for the assessment of the performance of aerosol inlet systems, *Atmos. Meas. Tech*, 479–494 pp., 2009.

Wiedensohler, A., Covert, D. S., Swietlicki, E., Aalto, P., Heintzenberg, J., and Leck, C.: Occurrence

of an ultrafine particle mode less than 20 nm in diameter in the marine boundary layer during Arctic

summer and autumn, *Tellus B Chem Phys Meteorol*, 48, 213–222,

<https://doi.org/10.3402/tellusb.v48i2.15887>, 1996.

Willis, M. D., Leaitch, W. R., and Abbatt, J. P. D.: Processes Controlling the Composition and

Abundance of Arctic Aerosol, <https://doi.org/10.1029/2018RG000602>, 1 December 2018.

Yu, F. and Turco, R. P.: From molecular clusters to nanoparticles: Role of ambient ionization in

tropospheric aerosol formation, *Journal of Geophysical Research Atmospheres*, 106, 4797–4814,

<https://doi.org/10.1029/2000JD900539>, 2001.

Ziemba, L. D., Dibb, J. E., Griffin, R. J., Huey, L. G., and Beckman, P.: Observations of particle

growth at a remote, Arctic site, *Atmos Environ*, 44, 1649–1657,

<https://doi.org/10.1016/j.atmosenv.2010.01.032>, 2010.

Deleted: Aliaga, D., Tuovinen, S., Zhang, T., Lampilahti, J., Li, X., Ahonen, L., Kokkonen, T., Nieminen, T., Hakala, S., Paasonen, P., Bianchi, F., Worsnop, D., Kerminen, V.-M., and Kulmala, M.: Nanoparticle ranking analysis: determining new particle formation (NPF) event occurrence and intensity based on the concentration spectrum of formed (sub-5 nm) particles, *Aerosol Research*, 1, 81–92, <https://doi.org/10.5194/ar-1-81-2023>, 2023.¶

Asmi, E., Kondratyev, V., Brus, D., Laurila, T., Lihavainen, H., Backman, J., Vakkari, V., Aurela, M., Hatakka, J., Viisanen, Y., Uttal, T., Ivakhov, V., and Makshtas, A.: Aerosol size distribution seasonal characteristics measured in Tiksi, Russian Arctic, *Atmos Chem Phys*, 16, 1271–1287, <https://doi.org/10.5194/acp-16-1271-2016>, 2016.¶

Barton, N. and Veron, D.: Response of clouds and surface energy fluxes to changes in sea-ice cover over the Laptev Sea (Arctic Ocean), *Clim Res*, 54, 69–84, <https://doi.org/10.3354/cr01101>, 2012.¶

Beck, L. J., Sarnela, N., Junninen, H., Hoppe, C. J. M., Garmash, O., Bianchi, F., Riva, M., Rose, C., Peräkylä, O., Wimmer, D., Kausiala, O., Jokinen, T., Ahonen, L., Mikkilä, J., Hakala, J., He, X. C., Kontkanen, J., Wolf, K. K. E., Cappelletti, D., Mazzola, M., Traversi, R., Petroselli, C., Viola, A. P., Vitale, V., Lange, R., Massling, A., Nøjgaard, J. K., Krejci, R., Karlsson, L., Zieger, P., Jang, S., Lee, K., Vakkari, V., Lampilahti, J., Thakur, R. C., Leino, K., Kangasluoma, J., Duplissy, E. M., Siivola, E., Marbouti, M., Tham, Y. J., Saiz-Lopez, A., Petäjä, T., Ehn, M., Worsnop, D. R., Skov, H., Kulmala, M., Kerminen, V. M., and Sipilä, M.: Differing Mechanisms of New Particle Formation at Two Arctic Sites, *Geophys Res Lett*, 48, <https://doi.org/10.1029/2020GL091334>, 2021.¶

Bousiotis, D., Brean, J., Pope, F. D., Dall’Osto, M., Querol, X., Alastuey, A., Perez, N., Petäjä, T., Massling, A., Klenø Nøjgaard, J., Nordström, C., Kouvarakis, G., Vratolis, S., Eleftheriadis, K., Niemi, J. V., Portin, H., Wiedensohler, A., Weinhold, K., Merkel, M., Tuch, T., and Harrison, R. M.: The effect of meteorological conditions and atmospheric composition in the occurrence and development of new particle formation (NPF) events in Europe, *Atmos Chem Phys*, 21, 3345–3370, <https://doi.org/10.5194/acp-21-3345-2021>, 2021.¶

Boyer, M., Aliaga, D., Quéléver, L. L. J., Bucci, S., Angot, H., Dada, L., Heutte, B., Beck, L., Duetsch, M., Stohl, A., Beck, I., Laurila, T., Sarnela, N., Thakur, R. C., Miljevic, B., Kulmala, M., Petäjä, T., Sipilä, M., Schmale, J., and Jokinen, T.: The annual cycle and sources of relevant aerosol precursor vapors in the central Arctic during the MOSAiC expedition, *Atmos Chem Phys*, 24, 12595–12621, <https://doi.org/10.5194/acp-24-12595-2024>, 2024.¶

Brean, J., Beddows, D. C. S., Harrison, R. M., Song, C., Tunved, P., Ström, J., Krejci, R., Freud, E., Massling, A., Skov, H., Asmi, E., Lupi, A., and Dall’osto, M.: Collective geographical ecoregions and precursor sources driving Arctic new particle formation, *Atmos Chem Phys*, 23, 2183–2198, <https://doi.org/10.5194/acp-23-2183-2023>, 2023.¶

Cai, R. and Jiang, J.: A new balance formula to estimate new particle formation rate: Reevaluating the effect of coagulation scavenging, *Atmos Chem Phys*, 17, 12659–12675, <https://doi.org/10.5194/acp-17-12659-2017>, 2017.¶

Chen, X., Virkkula, A., Kerminen, V. M., Manninen, H. E., Busetto, M., Lanconelli, C., Lupi, A., Vitale, V., Del Guasta, M., Grigioni, P., Vaanänen, R., Duplissy, E. M., Petäjä, T., and Kulmala, M.: Features in air ions measured by an air ion spectrometer (AIS) at Dome C, *Atmos Chem Phys*, 17, 13783–13800, <https://doi.org/10.5194/acp-17-13783-2017>, 2017.¶