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A long-term satellite study of aerosol effects on convective clouds in Nordic background air

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A long-term satellite study of aerosol effects on convective clouds

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Abstract

Aerosol-cloud interactions constitute a major uncertainty in future climate predictions. This study combines 10 yr of ground-based aerosol particle measurements from 2 Nordic background stations (Vavihill and Hyytiälä) with MODIS (Moderate Resolution Imaging Spectroradiometer) satellite data of convective clouds. The merged data are used to examine the indirect aerosol effects on convective clouds over the Nordic countries. From the satellite scenes, vertical profiles of cloud droplet effective radius (r_e) are created by plotting r_e against cloud top temperature. The profiles have been divided according to aerosol loading but also modeled meteorological parameters from the ECMWF (European Centre for Medium-Range Forecasts). Furthermore, weather radar data from the BALTEX (Baltic Sea Experiment) and ground based precipitation measurements from several ground-based meteorological measurement stations have been investigated to determine whether aerosols affect precipitation intensity and amount.

Higher aerosol number concentrations result in smaller r_e throughout the entire profiles at both stations. Profiles associated with no or little precipitation have smaller droplets than those associated with more precipitation. Furthermore, an increase in aerosol loadings results in a suppression of precipitation rates, when the vertical extent of the clouds has been taken into account. Clouds with greater vertical extent have the highest precipitation rates and are most sensitive to aerosol perturbations. Nevertheless, meteorological parameters such as the vertical extent of the clouds, the atmospheric instability and the relative humidity in the lower atmosphere affect the amount of precipitation that reaches the ground more than the aerosols do. The combination of these ground-based and remote sensing datasets provides a unique long-term study of the effects of aerosols on convective clouds over the Nordic countries.

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1 Introduction

That aerosol particles have the ability to affect clouds has been known for several decades (Twomey, 1974). However, the magnitudes of the multiple effects and the exact mechanisms of how aerosols affect clouds are still associated with large uncertainties,

5 as well as how much these processes affect global climate. Aerosols in the atmosphere can also affect the climate through scattering and absorption of incoming solar radiation and to a minor degree by influencing outgoing thermal radiation. This is the called the direct aerosol effect, while anthropogenic aerosol effects on climate through interactions with clouds has been named the indirect effect (Lohmann and Feichter, 2005).

10 The latter includes increased cloud albedo due to smaller but more numerous droplets (Twomey, 1974), suppression of drizzle (Albrecht, 1989), an increase in cloud height (Pincus and Baker, 1994) and increased cloud lifetime (Ackerman et al., 2000). Furthermore, previous studies have revealed various aerosol effects on clouds depending on cloud type and meteorological conditions e.g. (Khain et al., 2008; Fan et al., 2009;

15 Lee et al., 2010).

Convective clouds are important for the global climate and hydrological cycle since they transport large amounts of energy vertically and produce intense precipitation. The influence of aerosols on convective clouds has been studied in several different manners. For example, aircraft investigations over the Amazonian rainforest have

20 shown that high aerosol loadings suppress precipitation and increase the level to which a cloud must grow before it can produce precipitation (Andreae et al., 2004; Freud et al., 2008a).

Moreover, numerous modeling studies on aerosol interactions with convective clouds have been performed. Ekman et al. (2007) used a model with an explicit aerosol mod-

25 ule and found that an increase in cloud condensation nuclei (CCN) concentrations decreases the precipitation rates and that ice nuclei (IN) has a substantial effect on the development of convective clouds. Furthermore, Fan et al. (2009) found that vertical wind shear determine whether aerosols enhance convective strength or not. In another

study, Khain et al. (2005) found that increased levels of CCN delay the onset of precipitation from the convective clouds and that this in turn affects the dynamics of the clouds, in particular the vertical wind velocity. Whether aerosols suppress or enhance precipitation has also been found to depend on the relative humidity (RH) of the air near cloud base (Khain et al., 2008). Lee et al. (2010) on the other hand found that clouds react differently to changes in the number of CCN depending on the atmospheric instability.

Convective clouds have also been studied using remote sensing techniques, mainly satellite-based sensors. A study over the Amazon found that higher aerosol optical depth (AOD) was associated with increased precipitation, cloud cover and cloud depth (Lin et al., 2006). Moreover, Rosenfeld and Lensky (1998) used satellite data to obtain vertical profiles of the cloud effective radius (r_e) and found significant differences between profiles in clean and polluted conditions. Applying the same technique, Freud et al. (2008b) studied convective clouds during one summer over southern Sweden. They found that the r_e profiles were affected by aerosol loading which was estimated using $PM_{0.5}$ (mass concentrations of airborne particles below 0.5 µm in diameter) measured at two ground-based stations in Sweden.

The modeling studies suggest that several parameters are important for the development of convective clouds and how aerosols affect them. Different investigations have found aerosols to both enhance and suppress precipitation, while other studies suggest that whether aerosols enhance or suppress precipitation depend on meteorological conditions. The dissimilar results clearly points to the fact that clouds are very complicated systems and that more research using different methods is needed to improve our knowledge about aerosol-cloud interactions.

This study of convective clouds combines 10 yr (2000–2009) of cloud data from the satellite based MODIS (Moderate Resolution Imaging Spectroradiometer) instruments with ground-based aerosol measurements from the stations Vavihill (Southern Sweden) and SMEARII (Hyytiälä, central Finland). Since both ground-based stations have extensive time series of aerosol measurements and the MODIS instruments have been in orbit for approximately a decade, a long-term study including over 600 satellite

scenes could be performed. No aerosol products from MODIS are used in the study since satellite scenes containing clouds are studied and no aerosol products are produced for cloud covered pixels. Several meteorological parameters have been investigated to determine which of these affect clouds the most and also whether the influence of aerosols on clouds depends on meteorological conditions. Furthermore, weather radar and ground-based precipitation data were analyzed to determine whether the precipitation intensity and amount of precipitation reaching the ground is affected by the aerosol particles. The study is performed at the mid-latitudes where the convective strength is weaker than in the tropics, where most studies of convective clouds have taken place. In addition, the aerosol loading in the current study is significantly lower than in regions affected by for example biomass burning, which has also often been the focus of studies aiming to determine how convective clouds are affected by aerosols. Hence, this is a unique long term study of how aerosols affect convective clouds in areas with relatively low aerosol loading and dynamical forcing.

2   Method and data

2.1   Measurement stations

Aerosol sub-micrometer size distribution data from the EUSAAR (EUropean Supersites for Atmospheric Aerosol Research) and ACTRIS (Aerosols, Clouds, and Trace gases Research InfraStructure Network) station Vavihill (Fig. 1) located in southern Sweden (56°01′ N, 13°09′ E, 172 m a.s.l.) has been used in the study. There are no local pollution sources close to the station, but the cities of Helsingborg, Malmö and Copenhagen are within 25–45 km from the station. A DMPS (Differential Mobility Particle Analyzer) that measure aerosol number concentrations of particles with a diameter between 3 and 900 nm has been in operation at the station since 2001. The DMPS instrument and the measurement station is described in more detail in Kristensson

et al. (2008). Data from a CCN counter (CCNC) starting May 2008 (Fors et al., 2011) have also been utilized in the study.

The second station used in this study is the Finish SMEAR II station (Hari and Kulmala, 2005) located in Hyytiälä (61°51′ N, 24°17′ E, 181 m a.s.l.). This is also a EU-SAAR and ACTRIS background station with no local sources of pollution nearby. Aerosol particle size distribution measurements (DMPS) have been ongoing since 1996 (Aalto et al., 2001), but only data from February 2000 and onward were used here since this is when the first MODIS instrument became operational. The DMPS at Hyytiälä measured particle number concentrations in the size range 3 to 500 nm until December 2004 when the upper limit was extended to 1000 nm. Furthermore, the Hyytiälä CCN-counter data (Sihto et al., 2011) from June 2008 and onwards and has also been utilized in this study.

Aerosol number concentrations are generally lower and have greater seasonal variability at these two Nordic stations compared to field monitoring stations in Central Europe (Asmi et al., 2011).

2.2 Satellite data

The MODIS instruments have been in orbit onboard the Terra and Aqua satellites since 1999 and 2002, respectively. Both satellites are polar orbiting and their daytime passages over the investigated region occur sometime between morning and early afternoon local time. The MODIS instruments measure radiance at 36 spectral channels. Numerous atmospheric, cloud and aerosol parameters are calculated from the calibrated raw data and are available as Level 2 products (King et al., 2003). In this study, both MODIS Level 1B and Level 2 Cloud Products (Platnick et al., 2003) data are utilized. From the Cloud Products, the following parameters have been used: the r_e at 3.7 μm, the cloud optical thickness (COT), the Cloud Phase Optical Properties, the Cirrus Reflectance and the Multilayer Cloud Flag. The 3.7 μm r_e and not the 2.1 μm r_e has been used here since it better captures the size of the droplets at the top of the clouds, which is required in this type of study of convective clouds (Rosenfeld et al., 2004).

The 3.7 μm r_e is also less sensitive to errors due to sub-pixel variability in reflectance (common for convective clouds) than the r_e at 2.1 μm (Zhang et al., 2012). The MODIS Level 2 cloud top temperature (CTT) is provided at a 5 km horizontal resolution, which is too coarse to investigate convective clouds. Therefore, the Level 1B data from band 31 (10.780–11.280 μm) and 32 (11.770–12.270 μm) have been used to calculate the CTT at a 1 km horizontal resolution.

Satellite scenes were investigated for days with aerosol number size distribution data available at Vavihill or Hyytiälä. An area approximately 90 km by 62 km large (Fig. 1) was examined for each station. To minimize influence of surface reflectance on the cloud retrievals, the chosen areas mainly contain dense forests which provide a dark homogeneous background in the satellite scenes. The satellite scenes were investigated manually to find convective clouds over the specified areas and one satellite scene per day was chosen for days with convective clouds. Satellite scenes containing convective clouds close to frontal passages or low pressure systems have however been excluded in order to ensure that only clouds fed by the boundary layer air were investigated. Furthermore, scenes recorded in late morning or early afternoon were favored, since convection typically needs several hours of daylight to develop. Also, scenes where the investigation areas were situated close to the middle of the scenes were preferred since these pixels are less affected by distortions that occur at the edge of the images.

2.3 Meteorological data and trajectory analysis

Several meteorological parameters from the European Centre for Medium-Range Forecasts (ECMWF) were examined to investigate how meteorological conditions, affect the clouds. The convective available potential energy (CAPE) has been used as a measure of the instability of the atmosphere. However, CAPE data is only available from October 2003 and hence not available for all cases included in the study. Furthermore, the vertical wind velocity at 500 hPa as well as the RH and specific humidity (SH) at 1000 hPa were also included. The ECMWF temperature and dew point temperature at 2 m were

acquired to calculate the lifting condensation level (LCL). The ECMWF meteorological parameters have all been obtained at 12:00 UTC.

Back trajectories calculated using Hysplit4 (Draxler and Hess, 1997) were used to determine origin of the air masses. The meteorological data employed in these calcula-
5 tions are from the National Centre of Environmental Predictions (NCEP) GDAS (Global Data Assimilation System). A center of gravity has been calculated for 72 h back trajectories and the azimuth and distance from the station to the center of gravity were derived.

2.4 Precipitation data

10 Precipitation data from ground-based meteorological stations, henceforth GBP, for the area surrounding Vavihill have been obtained from the Swedish Meteorological and Hydrological Institute (SMHI). The locations of the stations have varied somewhat over the years, but simultaneous data from at least 11 stations are available for the entire time period investigated. The time resolution is however only on a daily basis. For the
15 area around Hyytiälä, there are unfortunately not as many precipitation measurements available. Precipitation data from the SMEAR station itself have been used together with data from 3 FMI (Finish Meteorological Institute) stations. The data from the 2 of the 3 FMI stations do not cover the entire investigation period and hence this dataset is considerably smaller than that from Vavihill. The GBP data for Vavihill is measured
20 from 6:00 a.m. local standard time to 6:00 a.m. the next morning, while at Hyytiälä the measurements are from 8:00 a.m. to 8:00 a.m. the next day.

Weather radar data have been obtained from the BALTEX (Baltic Sea Experiment) Radar Data Centre for both areas investigated in this study. Composite images from approximately 30 C-band weather radars in the Nordic countries, Poland and Germany
25 are generated every 15 min (Michelson, 2006; Michelson and Sunhede, 2004). The images have a 2 km horizontal resolution and contain the reflectivity factor in dBZ (referred to as dbzc here). The same areas used in the analysis of the satellite data (Fig. 1) was analyzed in the radar composite images. Ideally, the radar images could have been

put on top of the satellite images to determine which clouds that were precipitating. However, since the coordinates of the images are obtained with different positioning systems the images could be shifted with respect to each other. In addition, since the radar images are generated every 15 min, there is a time lag between the radar and satellite observations, even though the lag is maximum 5 min. Hence the entire areas marked in Fig. 1 were used to compare the radar and satellite data. For the dbzc images, only the pixels containing precipitation were used to calculate an average rain rate for each scene. In addition to the dbzc data, 3 h accumulated precipitation (rr3h), generated 8 times per day from the radar data (Michelson et al., 2000; Koistinen and Michelson, 2002) has also been analyzed here. The rr3h scene covering the time period to which the satellite scene belongs was used for each case. For the rr3h scenes, all pixels are used to calculate the average rain amount since it is interesting to investigate the mean amount of precipitation that has fallen in the entire region during 3 h.

2.5 Cloud profile retrievals

The method to obtain vertical profiles of clouds from a satellite scene developed by Rosenfeld and Lensky (1998), where the r_e is plotted against the CTT, was applied in this study. This method is assumes that convective clouds within a satellite image have cloud tops at different heights which represent clouds during different stages of their development. The validity of this assumption has been verified by both aircraft measurements (Freud et al., 2008a) and satellite measurements (Lensky and Rosenfeld, 2006). The 30th percentile of the variation in r_e with height has been studied here since this represents clouds early in their development, which are less influenced by ice formation (Freud et al., 2008b). Clouds with less ice formation are preferred, since the measurements at both stations focuses on CCN. Similar results were however obtained when the median profile was used instead. The same procedure has been applied to COT data to obtain vertical profiles of this parameter.

Certain pixels have been removed from the dataset to ensure that the study focuses on convective clouds. If pixels have a COT less than 7, the r_e retrieval becomes uncertain due to possible surface contaminations and therefore these pixels have been removed. Pixels containing cirrus clouds overlaying convective clouds have uncertain cloud retrievals and hence pixels flagged with a multilayer cloud flag were not included in the analysis either. Furthermore, single layers of cirrus clouds are also problematic since the COT seems to decrease with height when they are included in the profiles. Pixels with low COT and high cirrus reflectance have therefore also been removed before creating the profiles.

Once the profiles were created some further criteria were applied. For profiles containing vertical gaps greater than 7 °C, pixels at higher altitude in the clouds were removed from the profile. This was done to ensure that high level clouds in the satellite images, not connected to the cumulus cloud area were not included in the study. Moreover, only profiles with a vertical extent greater or equal to 7 °C has been included in the study such that CTT varies significantly in the satellite scene, hence that the investigated clouds are indeed convective. The temperature at the LCL (Inman, 1969) was calculated from the ECMWF temperature and dew point temperature at 2 m and compared to the cloud base temperature (T_B) derived from MODIS data (assuming that the highest CTT within a satellite scene equals T_B) to ensure that the clouds did not form at altitudes far greater than convective clouds could be anticipated at. Additionally, the r_e and COT are expected to increase with height in convective clouds and hence the profiles that do not fulfill this criteria are not included in the analysis.

Rosenfeld and Gutman (1994) used satellite and radar data to show that an r_e of 14 μm can be seen as the threshold limit for where the cloud droplets become large enough to be influenced by coalescence processes and produce precipitation in warm convective clouds. The difference between the cloud base temperature (T_B) and the temperature where the r_e cloud profiles reaches 14 μm (ΔT_{14}) is therefore a useful parameter to characterise the profiles with merely one number (Freud et al., 2008b;

Lensky and Drori, 2007). Since not all clouds reach an r_e of 14 µm the number of cases used in the analysis of this parameters is somewhat less than for the other parameters.

3 Results

In total, 388 profiles from Vavihill and 295 from Hyytiälä fulfill all the criteria described above and are included in the study. Table 1 shows the amount of days included in each step of the analysis. MODIS r_e data were not available during all days with DMPS data since the retrieval requires daylight which is not always present at the satellite overpass over these regions in winter. This is especially noticeable in the amount of days excluded for Hyytiälä station which is located further north. At both stations, only about half of the scenes containing convection were included as profiles in the study. Many scenes were excluded because the convection was too shallow to produce profiles with high enough vertical extent while other scenes contained contamination from non-convective cloud types and some may have been misclassifications of convective clouds.

The profiles are compared to each other by normalizing them with respect to T_B. To determine which parameters affect the r_e, the profiles have been sorted according to several different variables, not all shown here. Plotting all the profiles in one graph is not possible due to the large amount of profiles and they were hence divided into 6 bins according to the chosen parameter, with approximately the same number of profiles in each bin. One profile for each bin was then obtained by calculating a median value of all profiles in the bin for each temperature level. Moreover, at least 7 profile values were required at each level for a median value to be calculated.

The aerosol and trajectory parameters have been averaged over 5 h, centered at the time of the satellite overpass. However, using daily or daytime averaging instead did not affect the results to a large extent. Since CCN counter data from Vavihill and Hyytiälä was only available during less than two of the investigated years, the amount of particles above 80 nm, henceforth N_{80}, has been used as a proxy for CCN. This was

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the size range that best correlated with the cloud parameters. When N_{80} is compared to the CCN data (Table 2), it best correlates to a supersaturation of 0.4 % which is a reasonable value for convective clouds (Andreae, 2009).

3.1 Air mass origin and seasonal variability

The back trajectory analysis suggests that air masses arriving from the south to Vavihill and Hyytiälä generally contain higher particle number concentrations than air masses arriving from other directions (Fig. 2a and b). These events with high number concentrations do however not occur very frequently and the high aerosol loadings are expected since air masses that arrive from the south pass over areas with substantial anthropogenic emissions. Westerly air masses are very dominant at Vavihill while at Hyytiälä there is a larger directional variation and most air masses arrive from the southwest or north. The meteorological parameters used in this study do not seem to vary significantly with air mass origin (Fig. 2c to h).

The meteorological parameters do however vary significantly with season, see Fig. 3c to h. The T_B, CAPE and the SH show clear annual variation, with higher values in summer. There is also an annual variation in particle number concentrations in Hyytiälä (Fig. 2b), while for at Vavihill the concentrations are more consistent during the year (Fig. 2a). That the particle number concentrations in Hyytiälä are generally higher in summer indicates that N_{80} is influenced by biogenic secondary organic aerosols (BSOA), which reach high concentrations during this season (Kulmala et al., 2011; Leaitch et al., 2011). This is further supported by the relatively strong correlation between N_{80} and the ECMWF reanalysis 2 m temperature for Hyytiälä ($r = 0.64$). Additionally, the T_B and N_{80} are better correlated in Hyytiälä than in Vavihill (Table 3).

3.2 Effective radius profiles

Figure 4 shows the r_e profiles from Vavihill, Hyytiälä and both datasets combined, divided according to aerosol loading, T_B, CAPE and SH. It is clear that increased aerosol

number concentrations decrease the cloud droplet sizes at Vavihill, Hyytiälä and both stations combined (Fig. 4a to c). The profiles line up nicely with aerosol loading at almost all levels of the clouds, especially for the data from Vavihill. A previous study of convective clouds over southern Sweden, although only for one summer, also showed that clouds are sensitive to aerosol loading measured at ground-based stations (Freud et al., 2008b). Lin et al. (2006) created similar cloud profiles over South America but they sorted their profiles according to satellite retrieved AOD. The size of the r_e in their profiles did however not decrease with increasing AOD. In another study of cloud profiles, Koren et al. (2005) created vertical cloud profiles with the cloud top pressure on the y-axis instead of the CTT. This latter study covered clouds over the North Atlantic and showed that the r_e decreased with increasing AOD throughout the entire profiles, similar results to this study.

The r_e profiles at lower levels are not affected by the T_B which can be seen in Fig. 4d–f. However, at 10 to 15 °C above T_B when the r_e starts increasing faster with height, the warmer clouds have a smaller r_e than the colder clouds. This is most clearly visible when both datasets are used, Fig. 4f, and is thought to be related to at which altitude the clouds become mixed phase clouds. When the mixed phase is reached, coalescence and mixed-phase precipitation formation cause the droplets to start growing rapidly with height, hence increasing the r_e substantially (Rosenfeld and Lensky, 1998). In Fig. 4d–f it is visible that the onset of this zone is dependent on the T_B since the profiles are separated at the cloud levels where the r_e starts growing from approximately 12 µm. The clouds with the highest T_B enter the mixed phase zone at the highest altitude above T_B.

The profiles were also divided according to meteorological parameters from the ECMWF. The instability of the atmosphere does not seem to affect the droplet sizes in this dataset as can be seen in Fig. 4g and 4i. However, the cloud profiles do seem to be affected by the humidity at 1000 hPa. A lower SH results in smaller droplet sizes at lower levels of the profiles (Fig. 4j–l) and dividing the profiles according to RH produces similar results (not shown). The profiles were divided according to a few more parame-

ters such as the vertical velocity at 500 hPa (w), the vertical extent of the profiles (dT), the time the satellite image was retrieved, but since the profiles were not sensitive to these parameters the results are not shown here.

ΔT_{14} is the parameter that describes at what temperature below T_B that the r_e reaches 14 µm. The ΔT_{14} could be calculated for 40 % of the r_e profiles at Vavihill, and for Hyytiälä this number was 44 %. The remaining profiles do not reach 14 µm. To investigate if the aerosols are suppressing the cloud droplet sizes and inhibiting the profiles from reaching an r_e of 14 µm, the histogram of N_{80} for profiles that reach 14 µm has been compared to the N_{80} histogram for profiles which do not reach 14 µm, see Fig. 5. At Vavihill, Fig. 5a, the group of profiles that do not reach 14 µm have a higher average (with a 99.9 % confidence interval according to a Welch t test) N_{80} value than the group of profiles that do reach 14 µm. Hence, high aerosol loading may prevent some profiles from reaching an r_e of 14 µm. The Hyytiälä distributions also indicate that N_{80} is lower for the profiles with ΔT_{14} but the distributions are not significantly different from each other.

Table 3 can be used to investigate how ΔT_{14} is affected by cloud and meteorological parameters. ΔT_{14} is strongest correlated with N_{80} and dT. Larger ΔT_{14} means that droplets need to reach higher altitudes to obtain an r_e of 14 µm, hence that the droplets are generally small. The positive correlations between ΔT_{14} and N_{80} show that higher aerosol number concentrations result in smaller droplet sizes, just like the profile examination showed. The strong correlation between ΔT_{14} and dT is expected since the vertical extent of the profile also limits the ΔT_{14} and this correlation is hence an artifact of the method. The w is negatively correlated to ΔT_{14} which is also linked to the vertical extent of the profiles. Large negative values of w ($\mathrm{Pa\,s^{-1}}$) mean that there is a stronger upward motion of the atmosphere which favors greater vertical extent of the profiles. T_B is positively correlated with ΔT_{14} which is not surprising since a higher T_B elevate the level of the mixed-phase zone where droplets starts to grow rapidly.

The COT profiles were also divided according the same parameters but neither aerosols nor meteorological parameters were found to separate the profiles from each other to any great degree.

3.3 Precipitation

The r_e profiles have also been divided according to the mean amount of precipitation measured at ground level (Fig. 6a–c), mean radar reflectivity factor (Fig. 6d–f) and mean 3 h accumulated precipitation calculated from radar data (Fig. 6g–6i). The separation of the profiles according to the different precipitation datasets display quite similar results. Profiles with no or almost no precipitation have smaller droplet sizes and smaller vertical extent compared to cases with precipitation. Furthermore, for profiles that produce no or very little precipitation, the mixed-phase zone is not visible since there is no strong increase in r_e with height.

In Table 3, the correlation coefficients for the precipitation parameters are shown. The dT, CAPE, w, RH and SH have the strongest correlations with the precipitation parameters at both stations and all have positive correlations except w. Only low, hardly significant correlations with different signs are found between N_{80} and the precipitation parameters. The meteorological conditions hence control the amount of precipitation, rather than the aerosol loading. The strongest correlations at both stations are between the precipitation parameters and dT indicating that the most important parameter for precipitation is the vertical extent of the clouds. This is expected since convective clouds which reach a greater altitude generally produce greater amount of precipitation than clouds with a lower vertical extent. dT is calculated by taking the temperature difference between the warmest and coldest pixels included in the profile retrieval.

The precipitation parameters are barely significantly correlated with ΔT_{14} but all correlation coefficients are positive. However, when the precipitation amounts from the profiles that do not reach 14 μm are compared to those that do, it is clear that profiles with a ΔT_{14} produce more precipitation that those without (Fig. 7). This supports the

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results by Rosenfeld and Gutman (1994) that 14 μm is a useful threshold to estimate if warm clouds are precipitating.

Previous studies have found that the response of the precipitation to aerosol loading depends on the LWP of the clouds (Sorooshian et al., 2009; Li et al., 2011) but when this dataset was divided according to a geometrical average LWP for each scene, no significant difference in the correlations between the aerosol and precipitation parameters was found. There are however strong positive correlations between the LWP and precipitation. Modeling studies have found that how the clouds respond to aerosol loading depend on factors such as vertical wind velocities (Khain et al., 2008), RH near the surface (Khain et al., 2008) and CAPE (Lee et al., 2010). The data was divided into subsets according to these parameters but no significant changes in the way the aerosols affected the precipitation was found.

However, when the data is divided into 4 intervals of dT, the negative correlation between dbzc and N_{80} increase in magnitude which can be seen in Fig. 8a–c and Table 4. If the days when there is no precipitation detected by the radars are omitted the negative correlations become even stronger and most intervals have significant correlations (solid lines) which is shown in Fig. 8d–f. There is a lot of scatter in the data but the results indicate that increased aerosol loading reduce the precipitation intensity. The clouds with the highest dT are associated with the most intense precipitation and these clouds also seem to be most sensitive to aerosol effects since they show the greatest decrease in radar reflectivity with increasing aerosol loading.

4 Discussion and summary

The aim of this study was to investigate how convective clouds over Scandinavia are affected by aerosol loading. The profile and correlation analysis show that r_e is sensitive to the aerosol number concentration measured at ground level. The aerosol particles affect the profiles from the base throughout the vertical extent of the profiles since the profiles line up nicely according to aerosol loading at all levels. (Fig. 4a–c). The differ-

ence between the profiles divided according to N_{80} for Vavihill is larger than those for Hyytiälä (Fig. 4a and b). This is in line with the magnitude and spread of N_{80}, which is greater at Vavihill than at Hyytiälä. The 4 middle N_{80} intervals at Hyytiälä only differs by approximately 600 particles cm^{-3} while at Vavihill this number is almost 1000. There is also a stronger positive correlation between N_{80} and SH at Hyytiälä (Table 3) which could mask some of the effect the aerosols have on the droplets. A low SH can reduce the size of the droplets as can be seen in Fig. 4j–l. However, the different sensitivities to N_{80} could also be caused by dissimilarities in the aerosol particle properties between the two stations. Vavihill is more closely situated to continental Europe with more anthropogenic aerosol emissions while Hyytiälä is more affected by BSOA (Kerminen et al., 2012). That these BSOA are not as efficient CCN (Kulmala et al., 2011) as the more aged anthropogenic particles at Vavihill could also explain the variation in how much the aerosols affect the r_e between the two stations.

None of the meteorological parameters that we have investigated here seem to affect the r_e profiles as much as the aerosol particles do even though the cloud base temperature affect at what height the clouds reach the mixed phase. As mentioned previously, the SH do affect the droplet sizes, especially at lower SH where the droplet sizes are reduced. The entire effect of the SH may not be visible in the data because of the positive correlation with N_{80}.

Previous studies (Freud et al., 2008a; Koren et al., 2005; Li et al., 2011) have found that higher aerosol loading leads to higher cloud tops. No correlations between N_{80} and dT was found in our dataset (Table 3), not even after the dataset have been divided into subsets of variables found to control dT here. dT is instead best correlated to CAPE and the vertical winds at 500 hPa.

The precipitation results from the profile analysis showed that clouds with smaller droplets produce less precipitation, while the correlation analysis showed small or non-significant correlations between the precipitation parameters and N_{80}. Instead, dynamical and meteorological variables like dT, w, RH, SH and also CAPE at Vavihill are strongest correlated to the precipitation parameters (Table 3). However, when taking

dT into account, the increased aerosol loading is found to decrease precipitation inten- sity (Fig. 8) and clouds with a high dT are generally more sensitive to aerosol loading (Table 4). The large scatter in the data in Fig. 8 is expected since precipitation rates from clouds formed in very different conditions is being compared.

The decrease in precipitation with increasing aerosol loading is not present for the other two precipitation datasets. This could be caused by the larger amount of small values in these two datasets which makes correlations harder to detect. It could also be due to the fact that these two datasets represent the accumulated precipitation over a longer time period and not the intensity. The aerosols may not reduce the total amount of precipitation produced by the clouds over longer time periods but still affect the precipitation rates. Most of these satellite and radar images are from early in the afternoon and it could be that the aerosols decrease the intensity at this time of the day but cause the clouds to grow taller and produce more precipitation later in the afternoon as has been observed in other studies (Andreae et al., 2004; Li et al., 2011; Koren et al., 2012). There is however no correlation between dT and N_{80} in this dataset which point against such an invigoration of the clouds by the aerosols.

It is remarkable that the profiles associated with little or no precipitation show no rapid growth in r_e with altitude and never reaches 14 μm (Fig. 5) which is the limit for warm precipitation found by Rosenfeld and Gutman (1994). Moreover, that the mixed-phase zone is missing for the r_e profiles with no or little precipitation confirms that it is indeed mixed-phase and ice processes that are responsible for the convective precipitation at these latitudes. The results in Fig. 7 also support previous work (Rosenfeld and Gutman, 1994) even though quite a few profiles without a ΔT_{14} are associated with radar echoes Fig. 7c and d. This could be explained by large variations in r_e and/or radar echoes within the scene, the choice of 30th percentile profiles rather than median profiles (Sect. 2.5) and that the dataset is not restricted to warm clouds.

This study combines a variety of datasets from ground-based measurements to satellite cloud retrievals and ECMWF model data which have surely contributed to the low correlation between the different parameters in Table 3. Another reason for the low

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correlations is that clouds, and the environments that they form in, are highly variable systems affected by a range of parameters. The found correlations, often significant, are however encouraging and prove that these multiplatform approaches are very useful in investigations of clouds. The combination of datasets the clearly show that r_e profiles can be separated according to aerosol number concentrations measured at the ground and modeled meteorological parameters. Warm air can hold more moisture and in Table 3 it can be seen that the T_B, which is a satellite derived parameter, correlate strongly with the SH at 1000 hPa, which is a model parameter. Furthermore, the precipitation datasets, from different platforms, all have significant correlations with parameters known to affect precipitation such as dT (from satellite data) and RH (modeled parameter). These different datasets hence complement each other well and produce interesting results when combined.

In this investigation of convective clouds over Sweden and Finland, in conditions with low aerosol and dynamical forcing, aerosols are found to decrease the size of the droplets throughout the vertical extent of the clouds. The results also show that the meteorological parameters investigated here affect the cloud droplet sizes to a lesser degree than the aerosols. However, the meteorological parameters rather than aerosols control the amount of precipitation that reaches the ground. The aerosols are nevertheless found to decrease precipitation intensity when cloud with similar vertical extent is compared. Performing a similar study with data from other stations with long term ground-based size distributions measurements in other meteorological and aerosol conditions would be very interesting. Moreover, it would also be very interesting to compare these results to cloud model simulations of convective clouds over this region.

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Table 1. Number of days in different steps of the data analysis. The columns from left to right show the number of days with: DMPS data, DMPS data and MODIS scenes simultaneously, convection present in the scene according to the manual investigation of the scenes and approved profiles included in the study.

	DMPS	DMPS + MODIS	Convection	Profiles
Vavihill	2180	2129	800	388
Hyytiälä	3588	2863	500	295

Table 2. Correlation coefficients between CCNC data at different supersaturation and N_{80}.

	0.1 %	0.2 %	0.4 %	0.7 %	1 %
Vavihill	0.80	0.92	0.94	0.88	0.84
Hyytiälä	0.79	0.92	0.98	0.97	0.88

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Table 3. Correlation coefficients for Vavihill (top right) and Hyytiälä (bottom left). N_{80} (cm^{-3}) is the amount of particles with a diameter above 80 nm, T_B (°C) is the cloud base temperature, dT (°C) is the temperature difference between cloud base and cloud top, CAPE (Jkg^{-1}) is the convective available potential energy, w (Pas^{-1}), RH (%) is the relative humidity, SH (gkg^{-1}) is the specific humidity, ΔT_{14} (°C) is the temperature difference between the r_e profiles' cloud base and the 14 μm level, GBP (mmday^{-1}) is short for ground-based precipitation, dbzc is mean radar reflectivity factor (dBZ) and rr3h is the accumulated precipitation between over a 3 h interval retrieved from weather radars (mm). The stars indicate at what confidence interval the correlation coefficients are significant.

	N_{80}	T_B	dT	CAPE	w	RH	SH	ΔT_{14}	GBP	dbzc	Vavihill rr3h
N_{80}	N_{80}	0.26***	0.12*	0.13*	−0.16**	0.00	0.20***	0.50***	−0.03	−0.07	0.02
T_B	0.53***	T_B	0.13**	0.16**	−0.08	0.02	0.77***	0.30***	0.10	0.00	0.10
dT	0.17**	0.17**	dT	0.42***	−0.36***	0.18***	0.27***	0.45***	0.38***	0.37***	0.53***
CAPE	0.07	0.27***	0.38***	CAPE	−0.38***	0.17**	0.42***	0.26**	0.28***	0.18**	0.37***
w	−0.10	−0.05	−0.47***	−0.26***	w	−0.26***	−0.28***	−0.30***	−0.27***	−0.13*	−0.30***
RH	−0.05	0.39***	0.42***	0.39***	−0.36***	RH	0.25***	0.01	0.31***	0.29***	0.27***
SH	0.48***	0.80***	0.34***	0.41***	−0.24***	0.65***	SH	0.11	0.30***	0.17**	0.25***
ΔT_{14}	0.50***	0.40***	0.49***	0.03	−0.20*	0.05	0.27**	ΔT_{14}	0.09	0.14	0.20*
GBP	0.16**	0.15*	0.52***	0.25***	−0.38***	0.38***	0.33***	0.22*	GBP	0.30***	0.51***
dbzc	−0.10	−0.14*	0.49***	0.21**	−0.22***	0.36***	0.12	0.16	0.27***	dbzc	0.47***
rr3h	0.04	0.08	0.69***	0.21**	−0.36***	0.50***	0.30***	0.27**	0.61***	0.58***	rr3h
Hyytiälä											

* = 95 %
** = 99 %
*** = 99.9 %

Table 4. Slopes of trend lines in Fig. 8 ± the 95 % confidence intervals of the slopes, correlation coefficients and p values for the dT intervals (°C) in Fig. 8. The upper part of table is values for the dataset including zero-values of dbcz and lower part of table are for the dataset without zero-values of dbcz.

Vavihill dT interval	Slope × 10^4	r	p	Hyytiälä dT interval	Slope × 10^4	r	p	Both dT interval	Slope × 10^4	r	p
8–12	−10.3 ± 43	−0.05	0.64	8–12	−51.0 ± 60	−0.18	0.10	8–12	−25.1 ± 34	−0.11	0.15
12–16	−16.0 ± 39	−0.09	0.42	12–15	−97.8 ± 81	−0.30	0.02	12–15	−28.2 ± 34	−0.14	0.11
16–22	−48.5 ± 31	−0.31	0.00	15–19	−44.1 ± 46	−0.24	0.06	15–21	−64.3 ± 23	−0.39	0.00
22–64	−40.0 ± 35	−0.25	0.02	19–65	−47.1 ± 50	−0.22	0.07	21–65	−47.5 ± 28	−0.27	0.00
8–12	−18.5 ± 25	−0.15	0.15	8–12	−42.0 ± 35	−0.27	0.02	8–12	−27.3 ± 20	−0.21	0.01
12–16	−8.4 ± 25	−0.08	0.51	12–15	−76.4 ± 53	−0.37	0.01	12–15	−20.3 ± 23	−0.16	0.08
16–22	−34.4 ± 25	−0.28	0.01	15–19	−44.1 ± 46	−0.24	0.06	15–21	−49.8 ± 19	−0.37	0.00
22–64	−45.6 ± 24	−0.40	0.00	19–65	−75.9 ± 36	−0.46	0.00	21–65	−58.4 ± 20	−0.44	0.00

Fig. 1. Map showing the locations of the stations and the selected areas studied in the satellite images.
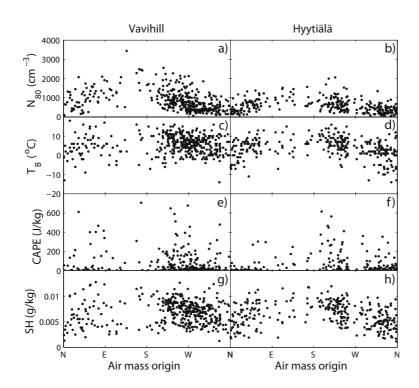
Fig. 2. The figure show how the following parameters vary with air mass origin: N_{80} (row 1), T_B (row 2), CAPE (row 3) and SH (row 4). The data from Vavihill is presented in column 1 and the data from Hyytiälä in column 2. N_{80} is the number concentration of particles with a diameter above 80 nm, T_B is the cloud base temperature, CAPE is the convective available potential energy and SH is the specific humidity at 1000 hPa. Only data from the 388 and 295 selected cloud scenes for Vavihill and Hyytiälä respectively, are shown.
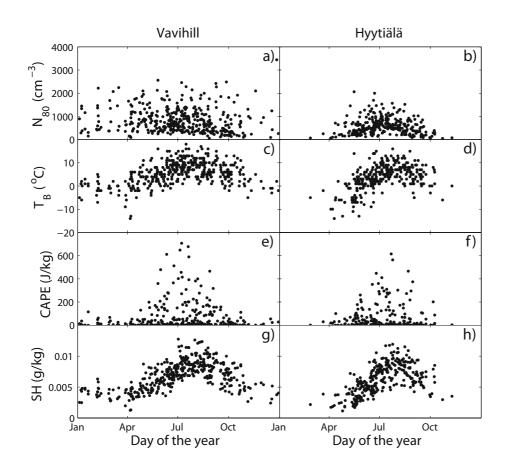
Fig. 3. The figure shows how the parameters in Fig. 2 vary over the year.
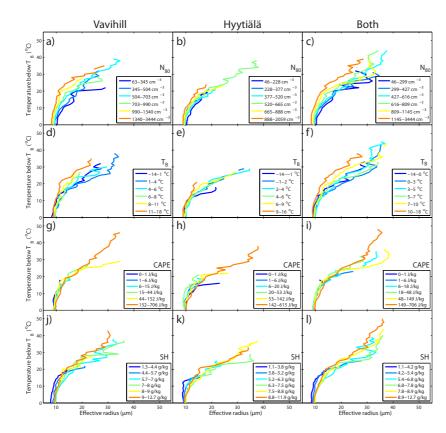
Fig. 4. Cloud effective radius is plotted as a function of temperature below that at cloud base to create vertical effective radius profiles. The figure shows average effective radius profiles for 6 percentiles of the data according to the N_{80} (row1), T_B (row 2), CAPE (row 3) and SH (row 4). The profiles from Vavihill are shown in column 1, from Hyytiälä in column 2 and both data sets combined are shown in column 3. N_{80} is the number concentration of particles with a diameter above 80 nm, T_B is the cloud base temperature, CAPE is the convective available potential energy and SH is the specific humidity at 1000 hPa.
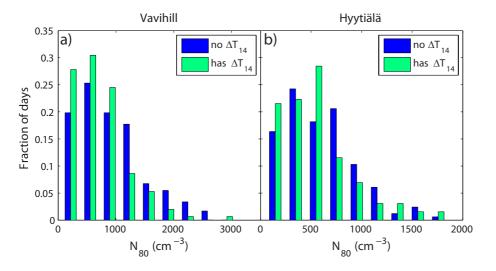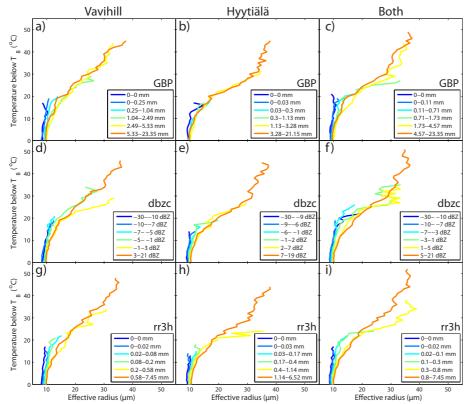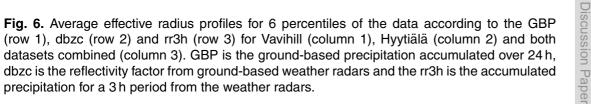
Fig. 5. Histogram of N_{80} levels for cases with profiles that reach $14\,\mu$m (has ΔT_{14}) and profiles that don't reach $14\,\mu$m (no ΔT_{14}) for Vavihill **(a)** and Hyytiälä **(b)**.

Fig. 6. Average effective radius profiles for 6 percentiles of the data according to the GBP (row 1), dbzc (row 2) and rr3h (row 3) for Vavihill (column 1), Hyytiälä (column 2) and both datasets combined (column 3). GBP is the ground-based precipitation accumulated over 24 h, dbzc is the reflectivity factor from ground-based weather radars and the rr3h is the accumulated precipitation for a 3 h period from the weather radars.

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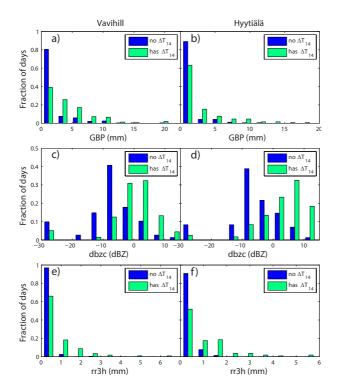
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Fig. 7. Histograms of the amount of ground-based precipitation accumulated over 24 h (**a** and **b**), reflectivity factor from ground-based weather radars (**c** and **d**) and the accumulated precipitation for a 3 h period from the weather radars (**e** and **f**) for cases with profiles that reach 14 µm (has ΔT_{14}) and profiles that don't reach 14 µm (no ΔT_{14}). The figures on the left (**a, c, e**) are for the Vavihill dataset and (**b, d, f**) are for the Hyytiälä dataset.
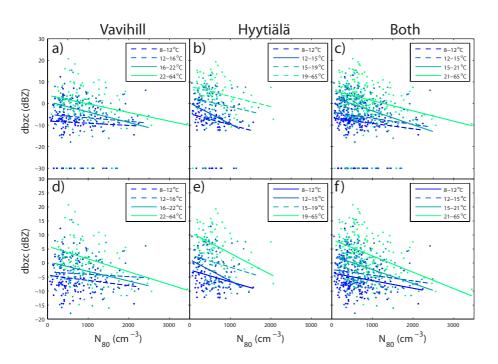
Fig. 8. Radar reflectivity vs aerosol loading for Vavihill (**a** and **d**), Hyytiälä (**b** and **e**) and both datasets combined (**c** and **f**). Each dataset have been divided into 4 subsets according to dT and a trend line calculated for each dataset. Shown in the legend are the limits of dT for each subset. Lines that are solid have a p value smaller than 0.05 while dashed lines have a p value greater than 0.05. p values below 0.05 mean that the correlation is significant with a 95 % confidence interval. Days when there is no precipitation detected by the radars are included in subfigures **(a–c)** and omitted in subfigures **(d–f)**.