

**Ion-mediated  
nucleation as source  
of tropospheric  
aerosols**

F. Yu et al.

# Ion-mediated nucleation as an important global source of tropospheric aerosols

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Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

## Abstract

Aerosol nucleation events have been observed at a variety of locations worldwide, and may have significant climatic and health implications. While ions have long been suggested as favorable nucleation embryos, their significance as a global source of particles has remained uncertain. Here, an ion-mediated nucleation (IMN) mechanism, which incorporates new thermodynamic data and physical algorithms, has been integrated into a global chemical transport model (GEOS-Chem) to study ion mediated particle formation in the global troposphere. The simulated annual mean results have been compared to a comprehensive set of data relevant to new particle formation around the globe. We show that predicted annual spatial patterns of particle nucleation rates agree reasonably well with land-, ship-, and aircraft-based observations. Our simulations show that, globally, IMN in the boundary layer is largely confined to two broad latitude belts: one in the northern hemisphere ( $\sim 20^\circ \text{N}$ – $70^\circ \text{N}$ ), and one in the southern hemisphere ( $\sim 30^\circ \text{S}$ – $90^\circ \text{S}$ ). In the middle latitude boundary layer over continentals, the annual mean IMN rates are generally above  $10^4 \text{cm}^{-3} \text{day}^{-1}$ , with some hot spots reaching  $10^5 \text{cm}^{-3} \text{day}^{-1}$ . Zonally-averaged vertical distribution of IMN rates indicates that IMN is significant in the tropical upper troposphere, whole middle latitude troposphere, and over Antarctica. The ratio of particle number annual source strength due to IMN to those associated with primary particle emission suggests that IMN contribution is important. Further research is needed to reduce modeling uncertainties and understand the contribution of nucleated particles to the abundance of cloud condensation nuclei.

## 1 Introduction

Atmospheric particles perturb the Earth's energy budget directly by scattering and absorbing radiation and indirectly by acting as cloud condensation nuclei (CCN) and thus changing cloud properties and influencing precipitation. The aerosol indirect radiative

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### Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

forcing is largely determined by the number abundance of particles that can act as cloud condensation nuclei (CCN) (e.g., Twomey, 1977; Albrecht 1989; Charlson et al., 1992). The magnitude of the aerosol indirect radiative forcing is poorly constrained in climate models, and this represents the dominate uncertainty in assessing climate change (NRC, 2005; IPCC, 2007). To reduce the uncertainty in the calculation of aerosol radiative forcing and to improve our prognostic capability of Earth's climate, the key processes controlling the number size distributions of atmospheric aerosols have to be understood and properly incorporated in the large scale models. New particle formation frequently observed throughout the troposphere is an important source of atmospheric CCN and is one of key processes that need to be accurately represented in future generations of climate models (Ghan and Schwartz, 2007).

In the past several years, there are a growing number of studies looking into the new particle formation in the global atmosphere. With an empirical formula (nucleation rate  $J=2\times 10^{-6}\text{ s}^{-1} [\text{H}_2\text{SO}_4]$ , where  $[\text{H}_2\text{SO}_4]$  is sulfuric acid gas concentration in  $\text{cm}^{-3}$ ), Spracklen et al. (2006) studied the contribution of boundary layer nucleation events to total particle concentrations on regional and global scales. Using monthly means  $\text{SO}_2$  concentrations, parameterized OH diurnal cycle, daily mean temperature and relative humidity, Kazil et al. (2006) investigated the formation of sulfate aerosol in the marine troposphere (over oceans only) from neutral and charged nucleation of  $\text{H}_2\text{SO}_4$  and  $\text{H}_2\text{O}$ , by running the box model of Lovejoy et al. (2004) on grids embedded into 4 isobaric surfaces of the troposphere (925, 700, 550, and 300 hPa). Lucas and Akimoto (2006) evaluated in a 3-D global chemical transport model binary (Vehkamäki et al., 2002), ternary (Napari et al., 2002), and ion-induced nucleation (Lovejoy et al., 2004; Modgil et al., 2005) schemes.

Lucas and Akimoto (2006) found that the binary nucleation model of Vehkamäki et al. (2002) and ion-induced nucleation model of Lovejoy et al. (2004) predict new particle formation only in the colder upper troposphere, and their simulations also show that binary nucleation rates based on Vehkamäki et al. (2002) are generally several orders of magnitude higher than the ion-induced nucleation based on the parameter-

---

**Ion-mediated nucleation as source of tropospheric aerosols**F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

ization of Lovejoy et al. (2004)'s model. Recent studies indicate that the BHN model of Vehkamäki et al. (2002) may have overestimated the BHN rates by around three orders of magnitude (Hanson and Lovejoy, 2006; Yu, 2007). It is important to determine accurately the contribution of different nucleation mechanisms to the new particle production in the troposphere.

Based on an up-to-date kinetically consistent ion-mediated nucleation model (IMN) incorporating recently available thermodynamic data and schemes, Yu (2006a) showed that ions can lead to significant particle formation not only in the upper troposphere but also in the lower troposphere (including boundary layer). The involvement of ions in many boundary layer nucleation events has been recently confirmed by evolving charged cluster distributions and overcharging of freshly nucleated nanometer particles observed during nucleation events, although the relative importance of ion-mediated nucleation versus neutral nucleation under different atmospheric conditions needs to be further investigated (Iida et al., 2006; Hirsikko et al., 2007; Laakso et al., 2007; Yu, 2006b). New nanometer-sized particles are overcharged in more than 90% of the clear nucleation event-days sampled during spring 2005 in Hyytiälä, Finland, during the BACCI/QUEST IV field campaign (Laakso et al., 2007). Laakso et al. (2007) claim, based on an analytical analysis, that their measurements indicate a relatively small contribution of ion nucleation. By contrast, Yu (2006b), applying a different analytical approach, concludes that the same observations may indicate the dominance of IMN. Based on detailed kinetic simulations, Yu and Turco (2007) demonstrate that IMN can consistently explain the observed overcharging reported in Laakso et al. (2007).

The objective of this paper is to study the significance of IMN mechanism as a global source of new particles and the spatial distribution of nucleation zone. To achieve the objective, we integrated the IMN mechanism into a global chemical transport model (GEOS-Chem) and the simulated results are compared with land-, ship-, and aircraft-based measurements related to particle formation. The model and data used in this study are briefly described in Sects. 2 and 3, respectively. Section 4 presents modeling results and comparisons with measurements.

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## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Summary and discussion are given in Sect. 5.

## 2 GEOS-Chem model

To study particle nucleation in the global atmosphere, we include our IMN mechanism in the GEOS–Chem model which is a global 3-D model of atmospheric composition driven by assimilated meteorological observations from the Goddard Earth Observing System (GEOS) of the NASA Global Modeling Assimilation Office (GMAO). Meteorological fields include surface properties, humidity, temperature, winds, cloud properties, heat flux and precipitation. The GEOS-3 data, including cloud fields, have 6-h temporal resolution (3-h resolution for surface fields and mixing depths),  $1^\circ \times 1^\circ$  horizontal resolution, and 48 vertical sigma levels extending from the surface to approximately 0.01 hPa. The horizontal resolution can be degraded and vertical layers merged for computational efficiency. For the results presented in this paper, the GEOS-3 grid with  $2^\circ \times 2.5^\circ$  horizontal resolution and 30 vertical levels was used. The first 15 levels in the model are centered at approximately 10, 50, 100, 200, 330, 530, 760, 1100, 1600, 2100, 2800, 3600, 4500, 5500, and 6500 m above surface.

The GEOS-Chem model includes a detailed simulation of tropospheric ozone-NO<sub>x</sub>-hydrocarbon chemistry as well as of aerosols and their precursors (Park et al., 2004). In addition to sulfate and nitrate aerosols, the model also considers organic and elemental carbon aerosols (Park et al., 2003), dust (Fairlie et al., 2004), and sea salt aerosol (Alexander et al., 2005). Aero-sol and gas-phase simulations are coupled through sulfate and nitrate formation, heterogeneous chemistry (Evans and Jacob, 2005), aerosol effects on photolysis rates (Martin et al., 2003), and secondary organic aerosol formation. The ISORROPIA thermodynamic equilibrium model (Nenes et al., 1998) is used to calculate partitioning of total ammonia and nitric acid between the gas and aerosol phases. A detailed description of the model (including the treatment of various emission sources, chemistry and aerosol schemes) can be found in the model webpage (<http://www.as.harvard.edu/chemistry/trop/geos/index.html>).

### Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

---

**Ion-mediated nucleation as source of tropospheric aerosols**F. Yu et al.

---

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

The sulfur emission in GEOS-CHEM (Park et al., 2004) includes: 1) the fossil fuel and industrial emission (Benkovitz et al., 1996; Bey et al., 2001), 2) the gridded monthly aircraft and shipping emissions (Chin et al., 2000), 3) the biofuel emission based on the global biofuel CO emission from Yevich and Logan (2003), 4) the biomass burning emissions from Duncan et al. (2003), 5) the oceanic DMS emission calculated with an empirical formula from Liss and Merlivat (1986), and 6) the volcano emission from the database of Andres and Kasgnoc (1998). In the original version of GEOS-CHEM (v7-03-06), the fossil fuel and industrial emission is obtained by scaling the gridded, seasonally resolved inventory from the Global Emissions Inventory Activity (GEIA) for 1985 (Benkovitz et al., 1996) with updated national emission inventories and fuel use data (Bey et al., 2001). In this study, the SO<sub>2</sub> database from EDGAR 3.2, which fully considered the productions of energy, fossil fuel, biofuel, industrial processes, agriculture and waste handling (Olivier, 2001), is used to update the GEOS-CHEM anthropogenic sulfur emission data to year 2002. The global gridded scaling factor is derived according to the historical trend from 1990 to 2000. Due to the uncertainty of the emissions from sporadically erupting volcanoes, we only consider the continuously active volcanoes emission in this study. The emission from eruptive volcanoes is not considered in the study of Lucas and Akimoto (2006) as well.

### 3 Data relevant to atmospheric particle formation

#### 3.1 Land based measurements

New particle formation has been observed extensively at many locations around the globe. Kulmala et al. (2004) provides a comprehensive review of measurements relevant to the formation of particles in ambient atmosphere. All of the cases with defined particle formation rates as listed in Kulmala et al. (2004) are used in this study for comparison. Table 1 gives additional sets of particle formation data published since 2004, which are also considered for comparison in this study.

## 3.2 Ship based measurements

In a number of field campaigns, total concentrations of condensation nuclei (CN, diameter  $> \sim 12$  nm) and ultrafine condensation nuclei (UCN, diameter  $> \sim 3$  nm) in the surface layer of the ocean were continuously measured with CN counters during ship cruises. The average particle formation rates at a given day can be estimated based on the change (increase) rates in the UCN and CN concentration difference (i.e.,  $C_{\text{UCN}} - C_{\text{CN}}$ ) typically during the morning hours.

Table A1 in Appendix A gives the ship-based particle formation rates we have derived from the measurements obtained during the following field campaigns: RITS94, INDOEX99, ACE-Asia, ACE-2, ACE-1, NAURU99, NEAQS02, and NEAQS04. The original data were obtained from NOAA PMEL Atmospheric Chemistry Data Server (<http://saga.pmel.noaa.gov/data/>) where more information about the field campaigns can be found. In Table A1 we also include two additional sets of particle formation rates estimated from two published papers (Davison et al., 1996; Koponen et al., 2002).

## 3.3 Aircraft based measurements

Clarke and Kapustin (2002) published a survey of extensive aerosol data collected around the Pacific Basin during a number of field campaigns: Global Backscatter Experiment (GLOBE), First Aerosol Characterization Experiment (ACE-1), and Pacific Exploratory Mission (PEM)-Tropics A and B. The ultrafine condensation nuclei (UCN) counter was used to detect all particles larger than  $\sim 3$ – $4$  nm. The aircraft-based measurements considered in this study include the total UCN concentrations measured during GLOBE, ACE-1, and PEM-Tropics A and B (data obtained from Dr. Kapustin—about 146 600 10-second-average data points), as well as data from two more recent field missions: TRANsport and Chemical Evolution over the Pacific (TRACE-P), and the Intercontinental Chemical Transport Experiment-Phase A (INTEX-A). TRACE-P and INTEX-A (about 100 000 10-second-average data points) significantly increased the number of observations for the Northern Hemisphere. TRACE-P and INTEX-A

### Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

measurements were obtained from NASA's Global Tropospheric Experiment (GTE) database website.

#### 4 Simulations and comparisons with observations

The nucleation module used in this study is composed of look-up tables of pseudo-steady state nucleation rates under various conditions that are derived by running the detailed IMN model (Yu, 2006a). The ion-mediated nucleation rates ( $J_{\text{IMN}}$ ) depend on sulfuric acid vapor concentration  $[\text{H}_2\text{SO}_4]$ , relative humidity RH, and temperature T, ionization rate Q, and surface area of pre-existing particles  $S_0$  (i.e.,  $J_{\text{IMN}} = f([\text{H}_2\text{SO}_4], \text{RH}, T, Q, S_0)$ ). At given values of  $[\text{H}_2\text{SO}_4]$ , T, RH, Q, and  $S_0$ ,  $J_{\text{IMN}}$  can be accurately decided using the look-up tables with an efficient multiple-variable interpolation scheme.

In the current version of GEOS-Chem (v7-03-06),  $\text{H}_2\text{SO}_4$  vapor concentration ( $[\text{H}_2\text{SO}_4]$ ) is not explicitly resolved (all  $\text{H}_2\text{SO}_4$  gas produced is moved to particulate phase instantaneously). We have modified the code and now  $[\text{H}_2\text{SO}_4]$  is a prognostic variable. The change of  $[\text{H}_2\text{SO}_4]$  is determined by  $d[\text{H}_2\text{SO}_4]/dt = P - \text{CS} \times [\text{H}_2\text{SO}_4]$ , where P is the production rate of  $[\text{H}_2\text{SO}_4]$  from gas phase chemistry (mainly OH +  $\text{SO}_2$ ) and CS is the condensation sink for  $\text{H}_2\text{SO}_4$  gas associated with the condensation of  $\text{H}_2\text{SO}_4$  vapor on pre-existing particles. CS and  $S_0$  are calculated from the particle mass predicted in the GEOS-Chem and assumed particle sizes. The transport and deposition of  $\text{H}_2\text{SO}_4$  vapor are also taken into account in the model. The global ionization rates due to cosmic rays are calculated based on the schemes given in Usoskin and Kovaltsov (2006) and the contribution of radioactive materials from soil to ionization rates is parameterized based on the profiles given in Reiter (1992). We run the GEOS-Chem coupled with nucleation module for one year from 1 July 2001 to 30 June 2002. The time step for transport is 15 min and for chemistry (and nucleation) is 30 min. Figure 1 presents the simulated horizontal (averaged in seven lowest model layers representing the boundary layer) and vertical (zonal-averaged) spatial distributions of annual mean  $\text{SO}_2$  concentration, condensation sink, and  $[\text{H}_2\text{SO}_4]$ . The calculations of CS,  $[\text{H}_2\text{SO}_4]$ ,

### Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

and hence nucleation rates are only limited to grid boxes within GEOS-Chem's annual mean tropopause.

Figures 1a and 1b shows that the high SO<sub>2</sub> concentration zones are generally confined to source regions. The annual mean SO<sub>2</sub> surface layer concentrations in large areas of eastern United States, Europe, eastern China, Indian, Mexico, and Chile are above 1 ppb with some hot spots above 3 ppb. Vertically, high SO<sub>2</sub> concentration (zonal average >0.1 ppb, note the difference in the scale of Fig. 1a and Fig. 1b) zone can reach up to around 700 hpa (sigma = ~0.7) with the highest concentration limited to below 800 hpa in the northern hemisphere. The relatively low SO<sub>2</sub> concentration in middle and upper tropical (~20° S–30° N) troposphere is probably due to relatively weak SO<sub>2</sub> sources and high scavenging rate associated with convection and precipitation. The relatively high SO<sub>2</sub> concentration over Antarctica is mainly associated with DMS emission near the Antarctic coast. The extension of high SO<sub>2</sub> zone from surface up to 300 hpa around latitude of 30° S is associated with mountain uplifting of anthropogenic SO<sub>2</sub> emission in Chile and direct injection of SO<sub>2</sub> into middle troposphere from the continuously active volcano Lascar in Chile (2400 Mg-SO<sub>2</sub>/day, 23.32° S, 67.44° W, elevation 5.6 km).

It is clear from Figs. 1c and 1d that large areas of eastern and southern Asia, western Europe, eastern United States, southern America, and Africa have high CS associated with anthropogenic emission, biomass burning, and dust emission. Vertically, the high CS zone centered around 30° N extends to ~600–700 hpa. In contrast to Arctic region which is influenced by regional transport of particle pollutants, the CS around Antarctica is very low. In addition to sea salt emission, the CSs over oceans adjacent to continentals are significantly affected by transported particles. The concentration of H<sub>2</sub>SO<sub>4</sub> vapor (Figs. 1e, 1f) is determined by its production rate (mainly controlled by SO<sub>2</sub> and OH concentration) and loss rate (condensation sink). The highest [H<sub>2</sub>SO<sub>4</sub>] regions are confined to areas of high SO<sub>2</sub> concentration, high annual irradiance flux, and low CS. In regions of higher SO<sub>2</sub> as well as higher CS, it appears that the increased production dominate and thus [H<sub>2</sub>SO<sub>4</sub>] are generally higher. Vertically, [H<sub>2</sub>SO<sub>4</sub>] gen-

---

## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

erally decreases with altitude due to more rapidly decrease of  $\text{SO}_2$  with altitude. The relatively high  $[\text{H}_2\text{SO}_4]$  in tropical upper troposphere is due to the very lower CS calculated in the model.

Figure 2 shows the predicted annual mean nucleation rates averaged within the seven lowest model layers ( $\sim 0$ – $930$  m) representing the atmospheric boundary layer. Also given for comparison are average particle formation rates derived from various surface-based measurements (refer to Sect. 3 for details). Observed nucleation events typically last for  $\sim 3$  h a day, and thus an observed average nucleation rate of 1 particle  $\text{cm}^{-3}\text{s}^{-1}$  is equivalent to roughly  $10^4$  particles  $\text{cm}^{-3}\text{day}^{-1}$ . We have used this equivalence to cross-calibrate the color bars in the figure.

Our simulations show that, globally, nucleation in the boundary layer is largely confined to two broad latitude belts: one in the northern hemisphere ( $\sim 20^\circ\text{N}$ – $70^\circ\text{N}$ ), and one in the southern hemisphere ( $\sim 30^\circ\text{S}$ – $90^\circ\text{S}$ ). In the boundary layer, nucleation rates over continentals are generally much higher than those over oceans. In the middle latitude boundary layer over continentals, the annual mean new particle production rates are generally above  $10^4$   $\text{cm}^{-3}\text{day}^{-1}$ , with some hot spots reaching  $10^5$   $\text{cm}^{-3}\text{day}^{-1}$ . Over middle latitude oceanic boundary layer, the annual mean new particle production rates are generally below  $2500$   $\text{cm}^{-3}\text{day}^{-1}$ . Most boundary layer nucleation events in the northern hemisphere (except over remote ocean areas, and Greenland) are associated with anthropogenic  $\text{SO}_2$  emissions; in the southern hemisphere, nucleation is triggered both by oceanic DMS and anthropogenic  $\text{SO}_2$ . Owing to higher temperatures, nucleation rates in the boundary layer at tropical latitudes ( $30^\circ\text{S}$ – $30^\circ\text{N}$ , except some regions with high  $\text{SO}_2$  source) are negligible even though  $\text{H}_2\text{SO}_4$  gas concentrations are at medium level (see Fig. 1c). Particle formation over the Antarctica occurs mainly during the austral summer season. The simulations also indicate that nucleation induced by anthropogenic  $\text{SO}_2$  emission contributes to particle abundances in the southern hemisphere. The high nucleation zone along the Chile coast in South American appears to be a significant source of new particle in the southern hemisphere. The relatively higher nucleation rate over Antarctica is due to lower CS, colder

---

## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

temperatures, and higher ionization rates. By contrast, nucleation rates in the Arctic region ( $\sim 70^\circ\text{N}$ – $90^\circ\text{N}$ ) are much lower due to relatively high concentrations of pre-existing particles associated with regional pollution (Arctic haze).

Most of the land-based measurements were taken in Western Europe and North America. Ship data, which span the major ocean basins, show negligible particle formation over tropical seas ( $\sim 30^\circ\text{S}$ – $30^\circ\text{N}$ ). It should be noted that the model results represent annual mean nucleation rates (averaged over periods that may or may not include nucleation events) for each  $2^\circ \times 2.5^\circ$  grid cell, while the observations represent average “apparent” particle formation rates based on measured particle concentrations (mostly of sizes  $\sim 3\text{ nm}$  or larger) detected during nucleation events at specific locations. The fraction of freshly nucleated particles ( $\sim 1.5\text{ nm}$ ) that can grow to measurable size depends on the local growth rate and coagulation lifetime. While the comparison between simulations and observations shown in Fig. 2 is qualitative and limited, it is the first of the kind and Fig. 2 shows that, overall, the predicted spatial pattern of aerosol formation agrees quite well with measurements. The comparison also reveals regions with high predicted nucleation rates in middle-western United States, Canada, Middle East, Eastern Europe, Greenland, Asia, Chile, and Antarctica where nucleation measurements are sparse. Measurements in these regions would therefore be useful for improving our understanding of particle nucleation in the global atmosphere.

The IMN rate is limited by the local ionization rate, roughly  $\sim 10\text{ ion-pairs cm}^{-3}\text{s}^{-1}$  in the continental surface layer, and  $\sim 2\text{ ion-pairs cm}^{-3}\text{s}^{-1}$  over oceans (and snow or ice). Most observed particle production rates (Kulmala et al., 2004, also see Table 1) fall below the background ionization-rate limit ( $\sim 20\text{ ions/cm}^3\text{s}$  over continental sites). Comparing the predictions and observations in Fig. 2, it appears that IMN can account for much of the observed particle formation near Earth’s surface. An obvious exception is the extremely high rate of particle formation (well above  $1000/\text{cm}^3\text{s}$  and up to  $\sim 10^5/\text{cm}^3\text{s}$ ) observed in the clean marine coastal environment at Mace Head (O’Dowd et al., 1998). It seems that these anomalously high nucleation rates are linked to the occurrence of low tides and may be associated with homogeneous nucleation of io-

---

## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

dine species (O'Dowd et al., 2002). Some measurements also yield particle formation rates that exceed the assumed IMN ionization limit. In this case, possible explanations include: (1) homogeneous nucleation mechanisms that involve other species, which remain to be identified; (2) nucleation in exhaust streams that are not fully diluted prior to sampling, where binary homogeneous nucleation can lead to very high levels of nanoparticles (Du and Yu, 2006); (3) enhanced particle formation at locations where the ambient ionization rate exceeds  $\sim 10$  ion-pairs/cm<sup>3</sup>s—indeed, some measurements indicate that ionization rates near the surface can exceed 100 ion-pairs/cm<sup>3</sup>s due to the accumulation of radon gas in the nocturnal boundary layer (Dhanorkar and Kamra, 1994); (4) the inferred high rates of particle formation based on ultrafine particle concentrations may be a result of rapid mixing of particles formed elsewhere (Stanier et al., 2004).

Figure 3a shows GOES-IMN simulated annual-mean zonally-averaged nucleation rates as a function of sigma (=pressure/surface pressure) and latitude. It is clear that while nucleation rates are generally small in the lower tropical atmosphere, very high nucleation rates are predicted in middle and upper tropical air layers associated with very low temperature, high ionization rate, and lower condensation sink. High nucleation rates are also obvious in the whole mid-latitude troposphere (25° N–75° N) in the northern hemisphere although the nucleation rates decrease with altitude. Nucleation appears to be negligible in most northern part of the troposphere ( $\sim 75^\circ$  N and north). In the southern hemisphere, nucleation in the middle troposphere over Antarctica is significant and nucleation zone extends to  $\sim 60^\circ$  S. Ito (1993) reported that bimodal size distribution with a trough at around 20 nm in diameter was observed at Syowa station (69° S, 39° 35'E) in almost all the days from August to December in 1978. Deshpande and Kamra (2004) observed very high concentrations (as high as 10<sup>4</sup>/cm<sup>3</sup>) of nucleation mode particles around 10 nm in diameter associated with subsidence of mid-tropospheric air at the Indian Antarctica station, Maitri (70° 45'S, 11° 44'E). It appears that these measurements support our simulations which indicate the existence of a nucleation zone over the Antarctica. The nucleation zone in the lower troposphere around

---

**Ion-mediated nucleation as source of tropospheric aerosols**F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

30° S is primarily a result of anthropogenic SO<sub>2</sub> emissions (also see Fig. 1). The strong nucleation zone in the middle to upper troposphere around 30° S appears to be mainly associated with the continuously active volcano Lascar in Chile which injects 2400 Mg of SO<sub>2</sub> per day at an altitude of around 5.6 km.

The total concentrations of particles larger than ~3 nm have been measured at various altitudes, latitudes, and longitudes with aircraft-based ultrafine condensation nuclei (UCN) counters. While it is difficult to derive in situ particle formation rates directly from these data owing to rapid changes in air mass, UCN concentrations nevertheless can be used as indicators of nucleation, since high UCN concentrations are generally associated with large nucleation rates. Figure 3b summarizes the zonally-averaged latitudinal and vertical distributions of total UCN concentrations measured during a number of field campaigns covering a wide range of areas and seasons (see Sect. 3.3). The high UCN regions in the upper troposphere and northern mid-latitude troposphere, and lower UCN in tropical lower troposphere are consistent with corresponding high or low nucleation rates in Fig. 3a (keeping in mind that the nucleation rates in Fig. 3a represent zonal and temporal averages, while the UCN concentrations in Fig. 3b represent measurements at selected locations and times). While the comparison between Fig. 3a and Fig. 3b should be considered qualitative, it is the first attempt to compare global nucleation zones with aircraft-based UCN measurements. Figure 3 indicates that the IMN mechanism appears to capture the vertical spatial patterns in the UCN distribution for the regions where sufficient data are available to discern larger-scale patterns. High concentrations of ultrafine particles were also observed during the upper systematic tropospheric transequatorial Africa flights (Heintzenberg et al., 2003). These data are not included in Fig. 3b but are generally consistent with Fig. 3. Aircraft-based measurements at higher latitudes in both hemispheres are currently lacking to verify our model predictions.

The general agreement between simulations and observations demonstrated above strongly supports the important role of IMN in generating new particles in global troposphere. Figure 4 compares the annual mean IMN rates integrated over the low-

---

## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

est 3 km of atmosphere (e.g., the source strength due to IMN,  $SS_{\text{IMN0-3}}$ ,  $\#/cm^2/day$ ) with the annual mean source of primary particles due to emissions (e.g., the source strength of primary emissions,  $SS_{\text{primary}}$ ,  $\#/cm^2/day$ ) in terms of the ratio of  $SS_{\text{IMN0-3}}$  to  $SS_{\text{primary}}$ . The results in Fig. 4 clearly indicate that IMN is a significant source of particles throughout the lower troposphere. At high latitudes ( $\sim 30^\circ\text{N}$ – $90^\circ\text{N}$ ,  $30^\circ\text{S}$ – $90^\circ\text{S}$ ), the ratio exceeds 10 over oceans, and lies between  $\sim 10$  and  $\sim 300$  over land. In the tropics ( $30^\circ\text{S}$ – $30^\circ\text{N}$ ),  $SS_{\text{IMN0-3}}/SS_{\text{primary}}$  is generally between 0.1 and 10, although some spots have very high ratio and some others spots have very low ratio.

In discussing the relative contribution of secondary particle formation versus primary particle emission to climate active particles, we should keep in mind that the diameters of freshly nucleated particles are just a few nanometers, while those of primary particles are generally greater than 50 nm. The fraction of nucleated particles that grow to CCN sizes depends on the local growth rates (and, hence, the precursor vapor concentrations), and on the concentration of pre-existing particles. Pierce and Adams (2007) found that the probability of a nucleated particle generating a CCN varies from  $<0.1\%$  to  $>90\%$  in different regions of the atmosphere, and falls between 5% and 40% for a large fraction of nucleated particles in the boundary layer. Clearly, with these statistics in mind, IMN is very likely to be a significant source of particles that impact climate. It should be noted that the ratios shown in Fig. 4 do not include the contributions of new particle formation in the middle and upper troposphere. Some of the particles nucleated in the middle and upper troposphere will contribute to the climate effective particles due to their relatively long lifetime (against scavenging by pre-existing particles), although the particle growth rates in these regions are typically small. The evolution of nucleated particles into CCN should be analyzed using a size-resolved aerosol microphysical model coupled to global code like GEOS-Chem.

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**Ion-mediated nucleation as source of tropospheric aerosols**F. Yu et al.

---

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

## 5 Summary and discussion

The magnitude of the aerosol indirect radiative forcing is poorly constrained in climate models, and this is the dominant uncertainty in assessing climate change. The aerosol indirect radiative forcing is largely determined by the number abundance of particles that can act as cloud condensation nuclei (CCN). A clear understanding of the contribution of new particle formation and growth to CCN abundance, which is essential to properly assess the influences of aerosols on climate, depends on our ability to predict accurately the rates of new particle formation in large-scale models. Significant theoretical and experimental progresses have been made in last couple of years with regard to the role of ions in the formation of tropospheric particles.

In this study, we integrate a recently updated ion-mediated nucleation (IMN) mechanism into a global chemical transport model (GEOS-Chem) to investigate the significance of IMN mechanism as a global source of new particles and the spatial distribution of nucleation zone. We run the GEOS-Chem coupled with nucleation mechanism for one year from 1 July 2001 to 30 June 2002, using GEOS-3 grid with  $2^\circ \times 2.5^\circ$  horizontal resolution and 30 vertical levels. The time step for chemistry (and nucleation) is 30 min. Our simulations indicate that IMN can lead to significant new particle production.

Horizontally, a comparison of simulated annual mean particle formation rates in boundary layer with a comprehensive dataset of land- and ship- based nucleation measurements suggests that IMN mechanism may be able to account for many of the observed nucleation events. Vertically, the simulated high and low regions of annual-mean zonally-averaged nucleation rates appears to be consistent with high and low zones of UCN concentrations measured during a number of aircraft-based field campaigns. While the comparison between simulations and observations shown in this study is qualitative and limited, it is the first of the kind and, overall, the predicted spatial pattern of aerosol formation agrees quite well with measurements. The comparison also reveals regions with high predicted nucleation rates where nucleation measurements are sparse and thus identifies the regions where possible future nucleation

### Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

measurements should be carried out to improve our understanding of particle nucleation in the global atmosphere.

Particle formation rates are sensitive to  $[\text{H}_2\text{SO}_4]$ . One of major uncertainties in our simulated results is associated with the accuracy of the calculated  $[\text{H}_2\text{SO}_4]$ . In addition to the uncertainty in the simulated  $\text{SO}_2$  concentrations which depend on emission, transport, and loss processes, the uncertainty in the condensation sink (CS) estimated from the simulated mass and assumed size of particles of different types also influence the accuracy of  $[\text{H}_2\text{SO}_4]$ . In addition, the contribution of nucleation mode particles to CS is not considered in current model. To resolve the issue and to study the contribution of nucleation to CCN in different global environments, we will include size-resolved aerosol microphysics processes in GEOS-Chem in our future study.

Similar to other nucleation schemes that have been used in the global models to predict new particle formation, the IMN mechanism is subject to uncertainty as well. First, the thermodynamic data and physical algorithms used in the IMN model have limitation and uncertainties. Second, species other than  $\text{H}_2\text{SO}_4$  and  $\text{H}_2\text{O}$  (such as  $\text{NH}_3$ ,  $\text{HNO}_3$ , and organics) may affect the properties of small clusters and the nucleation rates in the real atmosphere. These uncertainties may imply that the IMN contribution to new particle formation in the troposphere could be either higher or lower than what we presented in this study. It should be noted that other nucleation mechanisms may also contribute to tropospheric new particle formation. In addition to improve the nucleation mechanisms through theoretical development, laboratory and field studies, and quantum calculations for small clusters, further research is also needed on the contributions of different nucleation mechanisms to global source of new particles. More detailed and comprehensive comparisons of model predictions with relevant data obtained in various field campaigns will be helpful to assess the successfulness of various nucleation mechanisms in explaining the observed nucleation events and to identify the areas for further improvement in the existing theories.

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## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

## Appendix A

Table A1 gives particle formation rates over oceanic surface derived from ship-based condensation nuclei (CN, diameter  $> \sim 12$  nm) and ultrafine condensation nuclei (UCN, diameter  $> \sim 3$  nm) measurements.

5 *Acknowledgements.* This study is supported by the NOAA/DOC under grant NA05OAR4310103 and NSF under grant 0618124. Z. Wang acknowledges the support of the CAS International Partnership Program for Creative Research Teams, the National 973 Project (2005CB422205) and NSFC (40533017). The GEOS-Chem model is managed by the Atmospheric Chemistry Modeling Group at Harvard University with support from the  
10 NASA Atmospheric Chemistry Modeling and Analysis Program. We thank V. N. Kapustin and A. Clark for providing the aircraft-based UCN data. TRACE-P and INTEX-A data are from NASA Global Tropospheric Experiment (GTE) database website. The original ship-based data were obtained from NOAA PMEL Atmospheric Chemistry Data Server.

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F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

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**Ion-mediated nucleation as source of tropospheric aerosols**

F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

---

**Ion-mediated  
nucleation as source  
of tropospheric  
aerosols**F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

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ACPD

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---

## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

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---

**Ion-mediated nucleation as source of tropospheric aerosols**

F. Yu et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

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**Ion-mediated nucleation as source of tropospheric aerosols**

F. Yu et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

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15

ACPD

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---

**Ion-mediated nucleation as source of tropospheric aerosols**

F. Yu et al.

---

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

**Table 1.** Measurements of particle formation events in addition to those listed in the review paper by Kulmala et al. (2004). J3 and J10 refer to the “apparent” formation rates of 3 nm and 10 nm particles during the nucleation periods, respectively (in  $\text{cm}^{-3}\text{s}^{-1}$ ) (refer to Turco et al., 1998, for a discussion of apparent nucleation rates).

Location	Time period	# of events	J3	J10	Growth rates (nm/hr)	references
Rochester, USA (43°10'N, 77°36'W)	Dec 2001–Dec 2003	307		3	5–12	Jeong et al., 2004, 2006; Hopke and Utell, 2005;
Värriö, Finland (67°46'N, 29°35'E)	1998–2002	147		0.1	0.5–10	Vehkamäki et al., 2004
Sumas Mountain (49°03'N, 122°15'W)	13 Aug 2001–1 Sep 2001	5	5		5–10	Mozurkewich et al., 2004
Antarctica (70°45'S, 11°44'E)	10 Jan 1997–24 Feb 1997	14		0.1–0.8		Deshpande and Kamra, 2004
Pittsburgh, USA (40°26'N, 79°59'W)	July 2001–June 2002	107	1–10			Stanier et al., 2004
Po Valley, Italy (44°39'N, 11°37'E)	24 Mar 2002–24 Aug 2004	304	~ 7		0.3–22.2	Laaksonen et al., 2005; Hamed et al., 2007
New Delhi, Indian (28°35'N, 77°12'E)	26 Oct 2002–09 Nov 2002	8	~ 7.3		11.6–18.1	Mönkkönen et al., 2005
Santa Ana, Mexico (19°11'N, 98°59'W)	10 Apr 2003–20 Apr 2003	2	~ 10		4.7	Dunn et al., 2004
CENICA, Mexico (19°21'N, 99°04'W)	2 May 2003–11 May 2003	3	~ 4			Dunn et al., 2004
Tumbarumba, Australia (35°40'S, 148°15'E)	2005	~ 100	0.15		1–6.5	Suni et al., 2006
Mukteswar, Himalaya Mt. (29°31'N, 79°39'E)	23 March 2006–7 June 2006	23	0.4			Lihavainen et al., 2006
Anmyeon, S. Korea (36°22'N, 126°19'E)	Jan 2005–Dec 2005	24	1.5			Lee et al., 2006
Beijing, China (39°55'N, 116°25'E)	Mar 2004–Feb 2005	~ 170	~ 1.5		0.1–13.5	Wehner et al., 2006; Wu et al., 2007.
Gosan, S. Korea (33°17'N, 126°10'E)	11 Mar 2005–08 Apr 2005	~ 6	~ 1.5			Yum et al., 2006
Houston, USA (29°54'N, 95°20'W)	22 Aug 2004–29 Aug 2004	~ 8	~ 2			Fan et al., 2006
Marseille, France (43°19'N, 5°42'E)	1 Jul 2002–19 Jul 2002	4	3–5.3		2–8	Petäjä et al., 2007
Athens, Greece (38°9'N, 23°45'E)	11 Jun 2003–26 Jun 2003	7	1.3–6.5		1.2–9.9	Petäjä et al., 2007
St. Louis, USA (38°36'N, 90°09'W)	1 Apr 2001–31 May 2003	155	8–14		4.7	Qian et al., 2007
Pear River Delta, China (22°36'N, 113°36'E)	3 Oct 2004–5 Nov 2004	4	4–6		6.8–13.8	Gong et al., 2007

## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

**Table A1.** Ship-based particle formation rates ( $J_{\text{obs}}$ ) derived from measurements obtained during 10 ship cruise campaigns. The data in this table are plotted in Fig. 2 of the main text.

Latitude (degree)	Longitude (degree)	$J_{\text{obs}}$ (#/cm <sup>3</sup> s)						
<b>(1) RITS94</b>			<b>(2) INDOEX99</b>			<b>(3) ACE-ASIA</b>		
49.98	-128.20	1.5E-02	27.37	-60.62	7.6E-04	34.03	-174.90	2.7E-03
51.49	-131.20	4.6E-02	24.48	-56.26	9.5E-04	31.76	178.00	3.1E-03
54.90	-139.80	1.6E-02	21.99	-52.59	1.9E-03	31.33	173.50	2.0E-04
50.46	-140.00	1.5E-04	18.99	-48.23	3.7E-03	32.39	168.10	4.5E-03
44.84	-140.00	0.0E+00	15.94	-43.90	2.5E-03	32.95	165.30	4.3E-03
39.99	-140.00	0.0E+00	13.72	-40.78	2.6E-03	34.20	162.80	0.0E+00
36.31	-140.00	2.8E-03	10.79	-36.73	1.6E-03	33.00	158.10	0.0E+00
32.24	-140.00	5.6E-03	7.66	-32.46	8.1E-03	32.74	155.20	1.6E-03
28.27	-140.00	3.4E-03	1.93	-25.01	0.0E+00	33.00	143.90	1.0E-01
22.95	-140.00	6.6E-05	-0.64	-22.07	0.0E+00	33.01	141.50	0.0E+00
17.62	-140.00	0.0E+00	-3.86	-18.39	0.0E+00	33.20	139.90	8.3E-02
12.14	-140.00	7.6E-05	-7.55	-14.13	1.6E-03	32.82	136.70	2.2E-02
5.42	-140.00	4.9E-05	-10.36	-10.89	0.0E+00	31.96	133.50	2.9E-02
-0.07	-140.00	4.5E-04	-13.90	-6.74	5.5E-03	30.72	131.50	1.4E-01
-3.97	-140.00	0.0E+00	-17.53	-2.44	8.1E-05	31.70	127.70	5.5E-02
-4.95	-140.20	2.6E-04	-23.82	5.25	6.1E-03	33.45	128.60	6.2E-02
-4.85	-140.40	9.2E-04	-25.98	7.98	6.9E-04	35.47	131.80	0.0E+00
-4.75	-140.50	7.8E-04	-28.97	11.80	6.0E-03	38.06	133.60	3.8E-02
-4.62	-140.90	1.1E-03	-31.47	15.13	9.1E-03	38.97	134.50	1.7E-02
-12.93	-141.70	1.7E-03	-34.22	18.07	0.0E+00	37.92	131.00	5.8E-03
-14.99	-145.60	0.0E+00	-34.53	22.64	1.0E-02	37.53	130.00	0.0E+00
-19.14	-149.50	2.5E-03	-33.58	27.55	1.2E-02	35.01	130.00	1.9E+00
-23.49	-149.10	2.0E-03	-32.30	30.71	0.0E+00	35.74	132.50	2.8E+00
-28.24	-148.40	1.9E-03	-30.32	35.53	0.0E+00	33.84	129.50	0.0E+00
-31.45	-145.40	1.5E-03	-28.23	40.52	4.1E-03	32.52	128.40	3.0E-01
-35.39	-145.10	9.6E-03	-26.15	45.39	5.8E-04	31.36	126.40	2.6E-02
-38.88	-144.80	0.0E+00	-24.42	49.38	4.8E-04	33.01	128.00	8.1E-02
-46.02	-143.00	5.5E-02	-19.45	57.59	9.9E-04	31.22	131.40	1.9E-01
-49.61	-141.70	0.0E+00	-15.34	58.43	3.7E-04	33.13	135.40	6.3E-01
-55.01	-139.30	6.1E-03	-11.07	59.27	1.2E-03	<b>(4)ACE2</b>		
-61.27	-135.60	9.0E-03	-7.50	60.59	2.0E-04	36.94	-9.40	0.0E+00
-67.28	-130.30	1.7E-02	-3.12	65.12	2.7E-03	35.67	-10.78	5.9E-03
-67.02	-122.90	4.7E-03	15.71	69.81	2.3E-02	35.44	-8.97	7.1E-02
-69.00	-113.00	1.0E-02	17.44	68.43	1.2E-02	35.92	-9.00	9.9E-02
-68.56	104.90	0.0E+00	16.55	67.00	0.0E+00	38.22	-12.58	2.0E-03
-68.95	-94.26	5.0E-04	10.76	67.00	1.3E-03	40.40	-14.01	0.0E+00
-67.27	-79.36	8.1E-03	7.08	71.43	1.5E-03	37.27	-14.86	0.0E+00
-66.81	-72.76	0.0E+00	4.98	73.48	4.4E-03	37.00	-8.91	8.2E-02
-64.78	-64.08	4.7E-03	1.59	75.00	7.4E-03	37.16	-9.05	8.0E-02
-64.71	-63.10	2.9E-02	-2.76	75.00	0.0E+00	37.16	-9.05	0.0E+00
-58.01	-63.88	3.9E-02	4.10	73.76	1.9E-02	38.85	-10.67	0.0E+00
-53.34	-67.03	2.4E-01	3.08	77.74	0.0E+00	39.12	-11.65	0.0E+00

**Ion-mediated  
nucleation as source  
of tropospheric  
aerosols**

F. Yu et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Table A1. Continued.

Latitude (degree)	Longitude (degree)	$J_{\text{obs}}$ (#/cm <sup>2</sup> s)									
<b>(4) ACE2</b>			-40.81	144.20	2.5E-01	36.97	-76.44	9.5E-02	43.00	11.92	1.0E-03
36.30	-9.86	3.2E-01	-44.28	141.10	0.0E+00	<b>(8) NEAQSO4</b>			-39.53	12.76	1.0E-04
36.77	-9.84	3.5E-02	-45.14	141.20	0.0E+00	42.42	-70.65	2.0E+00	-35.32	13.88	1.0E-04
34.21	-10.50	0.0E+00	-47.47	147.00	8.7E-03	42.51	-70.56	2.3E-01	-32.84	16.44	1.0E-04
33.40	-14.40	0.0E+00	-44.96	144.90	2.1E-02	43.72	-66.20	2.9E-01	-30.61	18.15	1.0E-04
30.74	-14.15	0.0E+00	<b>(6) NAURU99</b>			43.00	-66.80	7.6E-02	-27.90	15.28	1.0E-04
30.54	-11.10	8.0E-03	-10.82	135.30	5.8E-03	42.46	-70.35	4.6E-01	-24.93	12.12	1.0E-04
33.15	-9.70	1.0E-01	-10.78	140.00	0.0E+00	42.36	-71.03	3.7E+00	-21.23	8.39	1.0E-04
35.57	-8.26	5.8E-02	-9.59	145.50	2.6E-03	42.78	-70.71	2.3E+00	-17.78	4.94	1.0E-04
37.17	-9.35	2.7E-02	-10.42	152.50	4.5E-03	42.41	-70.74	2.0E-01	-14.07	1.78	1.0E-04
39.22	-11.29	4.7E-02	-1.91	164.40	0.0E+00	42.38	-70.86	4.0E-01	-9.88	-2.52	1.0E-04
39.75	-10.87	1.0E-01	-1.92	164.40	0.0E+00	42.74	-70.70	7.1E-01	-5.19	-7.12	1.0E-04
37.91	-9.78	5.8E-02	-1.92	164.40	2.6E-03	42.74	-70.73	6.2E-01	-1.24	-10.85	1.0E-04
36.71	-8.94	3.1E-02	-1.92	164.40	3.7E-03	43.17	-69.26	1.5E-01	2.71	-14.30	1.0E-04
36.35	-9.36	0.0E+00	-0.52	166.70	7.1E-03	44.32	-67.10	2.5E-02	6.66	-17.46	1.0E-04
38.46	-11.78	0.0E+00	-0.52	166.90	3.6E-02	44.00	-66.59	1.2E-01	11.36	-18.91	1.0E-04
<b>(5) ACE1</b>			-0.55	166.90	5.2E-02	43.16	-70.47	1.3E-01	15.00	-18.93	4.1E-02
40.44	-135.70	0.0E+00	-0.50	166.90	4.5E-03	42.45	-70.82	6.6E-01	18.56	-18.90	6.0E-02
36.17	-141.00	5.2E-03	-0.56	167.00	0.0E+00	42.80	-70.51	1.4E-02	21.26	-18.39	1.0E-02
31.99	-145.80	0.0E+00	-0.57	167.00	0.0E+00	42.80	-70.63	5.9E-02	25.47	-16.98	1.0E-04
27.85	-150.40	0.0E+00	<b>(7) NEAQSO2</b>			42.97	-70.51	0.0E+00	29.92	-15.28	1.0E-04
23.80	-154.80	3.8E-04	34.45	-76.11	9.6E-03	42.67	-69.77	3.8E-01	36.12	-12.74	5.7E-01
19.19	-157.00	5.0E-02	38.84	-72.78	4.7E-03	42.82	-70.74	7.7E-02	41.32	-9.91	6.4E-01
19.12	-159.10	0.0E+00	40.49	-73.87	1.2E+00	43.32	-70.17	2.8E-01	45.53	-7.07	5.0E-01
12.34	-160.00	1.0E-03	40.44	-73.80	5.9E-01	43.09	-70.44	3.9E-01	48.01	-3.94	4.0E-01
7.39	-160.00	3.1E-03	40.82	-68.92	3.2E-02	43.52	-70.07	9.9E-01	<b>(10) Davidson et al., 1996</b>		
2.49	-160.00	1.2E-04	42.46	-70.79	8.1E-01	43.65	-69.90	2.3E-01	-68.00	0.00	3.2E+00
-3.68	-160.00	7.1E-04	43.02	-70.33	5.8E-02	44.39	-67.62	4.7E-01	-69.00	-4.00	3.2E+00
-8.34	-160.00	3.1E-04	43.01	-70.66	6.5E-01	43.64	-69.41	4.2E-01	-70.00	-8.00	3.2E+00
-14.14	-160.00	1.0E-03	42.94	-70.72	2.3E-01	43.62	-70.12	1.4E-01	-71.00	-12.00	3.2E+00
-20.51	-160.00	5.4E-02	42.83	-70.71	8.3E-01	42.55	-68.38	2.3E-02	-72.00	-16.00	3.2E+00
-25.33	-160.00	1.2E-0	42.75	-70.59	4.2E-01	42.43	-70.50	4.6E+00	-73.00	-20.00	3.2E+00
-30.76	-161.60	2.2E-02	43.46	-70.21	1.3E-01	42.37	-71.05	1.6E+00			
-32.11	-164.30	7.7E-04	43.40	-69.37	0.0E+00	42.63	-69.61	0.0E+00			
-34.58	-169.50	0.0E+00	43.02	-70.67	8.1E-01	42.87	-70.77	4.4E-01			
-37.61	-176.60	1.1E-02	42.32	-70.73	9.7E-01	<b>(9) Koponen et al., 2002</b>					
-35.53	178.80	0.0E+00	42.40	-70.77	6.4E-01	-69.50	4.06	1.0E-01			
-35.15	177.10	1.5E-01	42.97	-70.65	2.4E-01	-66.49	3.47	7.4E-02			
-34.40	172.10	0.0E+00	42.79	-70.59	9.3E-02	-63.30	3.78	4.0E-02			
-42.04	150.80	3.0E-02	43.02	-70.68	1.8E-01	-59.59	4.00	1.0E-02			
-50.29	155.90	9.8E-03	43.01	-70.67	8.2E-02	-55.14	4.83	5.5E-02			
-47.75	145.40	0.0E+00	42.79	-70.54	1.2E-01	-51.92	6.25	4.6E-02			
-49.99	138.30	0.0E+00	43.02	-70.64	1.7E-01	-49.69	8.52	3.4E-02			
-42.88	140.80	5.7E-02	39.06	-72.99	1.1E-01	-47.21	11.08	1.8E-02			
-41.08	143.30	2.6E-02	36.82	-75.84	5.1E-02	-44.98	11.64	8.0E-03			

## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

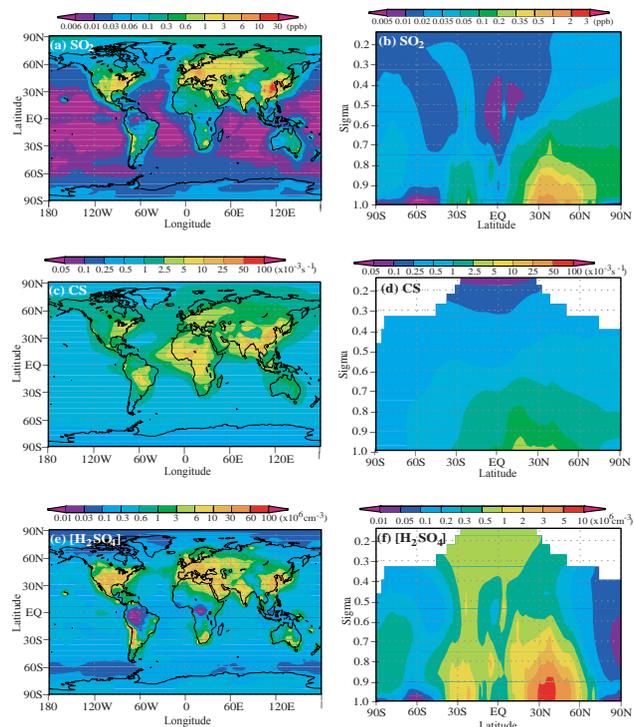
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Printer-friendly Version

Interactive Discussion

## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.



**Fig. 1.** Horizontal (a, c, e: average over first seven model layers above Earth's surface) and vertical (b, d, f: sigma=pressure/surface pressure) distributions of annual mean values of SO<sub>2</sub> mixing ratio, condensation sink (CS), and [H<sub>2</sub>SO<sub>4</sub>]. We run the GEOS-Chem coupled with nucleation model for one year from 1 July 2001 to 30 June 2002. The time step for transport is 15 min and for chemistry (and nucleation) is 30 min. The calculations of CS, [H<sub>2</sub>SO<sub>4</sub>], and hence nucleation rates are only limited to grid boxes below GEOS-Chem's annual mean tropopause.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

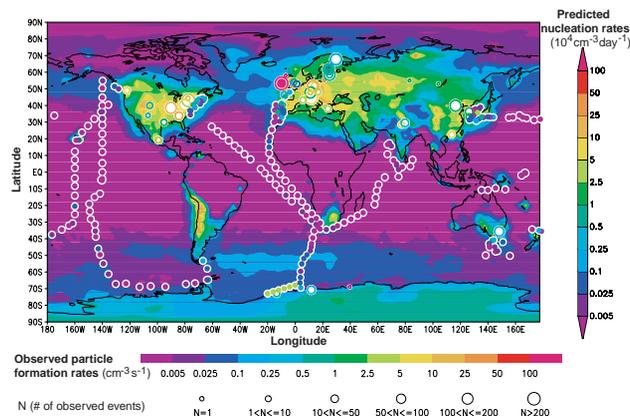
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Printer-friendly Version

Interactive Discussion

## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.



**Fig. 2.** Simulated global distribution of annual mean nucleation rates averaged within the boundary layer (0–930 m) (refer to the color bar on the right). Corresponding measured particle formation rates from ship observations are shown as color-filled circles (refer to the lower color bar, which gives the average local nucleation rate, typically over a window of several hours). Measurements over land are indicated by unfilled circles, where circle size defines the number of nucleation events reported (refer to the scale at the bottom of the figure), while color gives the average nucleation rate over the event ensemble. Land-based nucleation data prior to 2004 are taken from Kulmala et al. (2004); after 2004, publications listed in Table 1 are used. Ship-based nucleation rates are derived from the recorded time-series of ultrafine particle concentrations at sizes between  $\sim 3 \text{ nm}$  and  $\sim 12 \text{ nm}$  (Appendix Table A1). Assuming that typical nucleation events last for 3 h, an observed nucleation rate of  $1 \text{ particle cm}^{-3} \text{ s}^{-1}$  is equivalent to roughly  $10^4 \text{ particles cm}^{-3} \text{ day}^{-1}$ ; this equivalence has been used to cross-calibrate the color bars in the figure.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

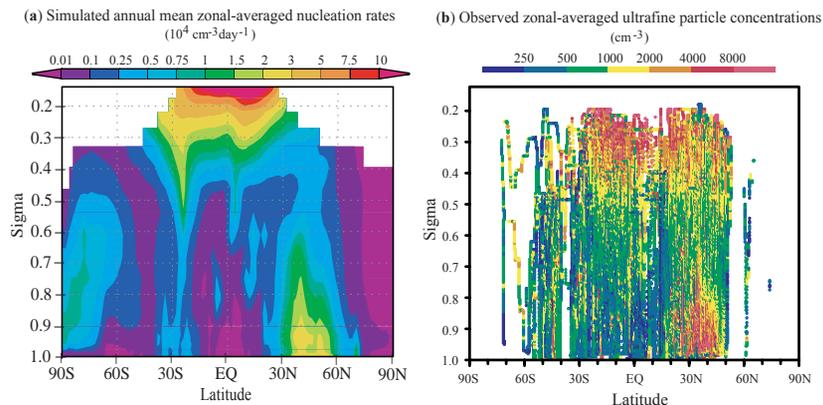
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Printer-friendly Version

Interactive Discussion

## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.

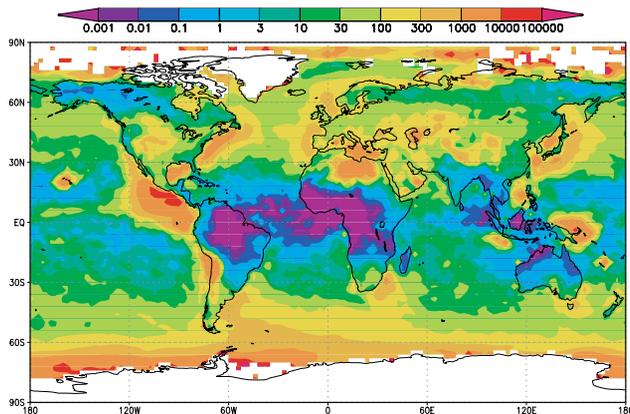


**Fig. 3.** (a) Meridional structure of predicted annual-mean zonal-averaged nucleation rates calculated using GEOS-Chem coupled to an ion-mediated nucleation sub-model. (b) Zonally-averaged latitudinal and vertical distributions of total ultrafine condensation nuclei (UCN) concentrations measured in situ (Sect. 3.3). In panel (3b), to avoid overlap owing to the large number of data points, the pressure-latitude cross section is divided into a  $5 \text{ mb} \times 0.5^\circ$  grid, wherein all data, at all longitudes, are averaged. The average value for each grid point is represented as a color-coded circle. The observed UCN concentrations have also been normalized to standard conditions (1 atm, 298 K).

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

## Ion-mediated nucleation as source of tropospheric aerosols

F. Yu et al.



**Fig. 4.** The ratio of annual mean IMN rates integrated within the lowest 3 km of atmosphere (e.g., the source strength due to IMN,  $SS_{\text{IMNO-3}}$ ,  $\#/cm^2/day$ ) to the annual mean rate of emission of primary particles (e.g., source strength due to primary aerosol emission,  $SS_{\text{primary}}$ ,  $\#/cm^2/day$ ). The primary aerosols considered in GEOS-Chem, and their corresponding assumed (fixed) sizes (radius), are: dust (4 sizes: 0.7, 1.5, 2.5, 4  $\mu\text{m}$ ), sea salt (3 sizes: 0.732, 5.67  $\mu\text{m}$ , and an ultrafine sea salt mode with a radius of 40 nm), black carbon aerosol (39 nm), and organic carbon particles (70 nm).

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion