Numerical simulations of the three-dimensional distribution of meteoric dust in the mesosphere and upper stratosphere

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1. Introduction

[1] Micrometeorites that ablate in the lower thermosphere and upper mesosphere are thought to recondense into nanometer-sized smoke particles and then coagulate into larger dust particles. Previous studies with one-dimensional models have determined that the meteoric dust size distribution is sensitive to the background vertical velocity and have speculated on the importance of the mesospheric meridional circulation to the dust spatial distribution. We conduct the first three-dimensional simulations of meteoric dust using a general circulation model with sectional microphysics to study the distribution and characteristics of meteoric dust in the mesosphere and upper stratosphere. We find that the mesospheric meridional circulation causes a strong seasonal pattern in meteoric dust concentration in which the summer pole is depleted and the winter pole is enhanced. This summer pole depletion of dust particles results in fewer dust condensation nuclei (CN) than has traditionally been assumed in numerical simulations of polar mesospheric clouds (PMCs). However, the total number of dust particles present is still sufficient to account for PMCs if smaller particles can nucleate to form ice than is conventionally assumed. During winter, dust is quickly transported down to the stratosphere in the polar vortex where it may participate in the nucleation of sulfate aerosols, the formation of the polar CN layer, and the formation of polar stratospheric clouds (PSCs). These predictions of the seasonal variation and resulting large gradients in dust concentration should assist the planning of future campaigns to measure meteoric dust.


[2] Estimates of the mass flux of meteoric material into the Earth’s atmosphere vary considerably [Gabrielli et al., 2004], but commonly used values are 16 ± 5 kt yr\(^{-1}\) [Hughes, 1978] and 40 ± 20 kt yr\(^{-1}\) [Love and Brownlee, 1993]. Most of these micrometeorites are thought to ablate in the lower thermosphere and upper mesosphere [Hunten et al., 1980; Kalashnikova et al., 2000], resulting in atomic metal layers observable by lidar between 80 and 110 km [Kane and Gardner, 1993; Plane, 2003]. The ablation products are likely to recondense into nanometer-sized smoke particles [Rosinski and Snow, 1961], which are extremely small and difficult to detect [Gelinas et al., 1998; Havnes et al., 2001; Lynch et al., 2005; Rapp et al., 2005; Strelnikova et al., 2007]. These small particles can coagulate into larger particles [Hunten et al., 1980] which may act as heterogeneous ice nuclei for polar mesospheric clouds (PMC) [Rapp and Thomas, 2006; Turco et al., 1982] and polar mesospheric summer echoes (PMSE) [Rapp and Lübken, 2001].

[3] Murphy et al. [1998] and Cziczo et al. [2001] have observed meteoric dust in ~50% of their samples of stratospheric aerosols. Meteoric dust may affect the size distribution and composition of the sulfate aerosol layer and subsequently affect the nucleation of polar stratospheric cloud (PSC) particles [Prather and Rodriguez, 1988]. Curtius et al. [2005] observed an increased fraction of stratospheric aerosols with a nonvolatile content inside the Arctic polar vortex (67%) compared to outside the vortex (24%). They concluded that meteoric dust is transported downward from the mesosphere inside the vortex and the increased fraction is therefore an indicator of mesospheric air inside the polar vortex. Meteoric dust descending in the polar vortex may also contribute to enhancements in the polar condensation nuclei layer observed in the winter around 30 km [Hofmann et al., 1988].

[4] Previous numerical simulations of meteoric dust formation and transport have been one-dimensional [Gabrielli et al., 2004; Hunten et al., 1980; Megner et al., 2006]. Megner et al. [2006] found that the size distribution of dust particles is sensitive to vertical velocity; however, the overall impact of transport is difficult to evaluate from a
one-dimensional simulation. Three-dimensional simulations of meteoric metals [Prather and Rodriguez, 1988] and chemical tracers [Fischer and O’Neill, 1993; Plumb et al., 2002] have shown that the meridional mesospheric circulation will transport material from the summer pole to the winter pole where it will be transported downward inside the polar vortex. For this study, we created a three-dimensional general circulation model with sectional microphysics (WACCM/CARMA) to investigate the combined impact of transport and coagulation on the distribution of meteoric dust in the mesosphere and upper stratosphere.

2. Model Description

[5] WACCM/CARMA is a combination of the Whole Atmosphere Community Climate Model (WACCM) general circulation model [Garcia et al., 2007] and a sectional microphysics model based upon the Community Aerosol and Radiation Model for Atmospheres (CARMA) [Toon et al., 1988; Turco et al., 1979]. A single column version of CARMA is implemented as a WACCM physics package. This means that the model’s state is passed to CARMA one column at a time, that CARMA calculates changes to the constituents within these columns from microphysical processes, and that these changes are then passed back to WACCM where they are used to adjust the model’s state. Each CARMA size bin is added as a unique WACCM constituent. WACCM is responsible for advection, vertical diffusion and wet deposition of all constituents, while CARMA is responsible for the production, sedimentation and coagulation of the meteoric dust constituents. Currently, CARMA only affects the dust size bins and the dust particles are not radiatively active. However, in future studies of ice particle formation, this model framework will allow CARMA to interact with other WACCM state variables like water vapor, generate heating rates and allow integration with the WACCM radiation code.

2.1. WACCM

[6] We used WACCM3 version 3.1.9 tag 9 at 4° × 5° resolution with 66 vertical levels and a model top near 140 km, which provides a vertical spacing of ~3.5 km in the mesosphere. This version has been used to model secular trends and the influence of the solar cycle in the middle and upper atmosphere [Garcia et al., 2007; Marsh et al., 2007]. For two simulations, we used a higher-resolution vertical grid with 125 levels and ~0.5 km spacing near the mesopause. WACCM uses the finite volume dynamical core, which is based upon the flux-form semi-Lagrangian transport scheme of Lin and Rood [Lin and Rood, 1996, 1997], and is mass conserving. WACCM’s vertical diffusion algorithm handles eddy and molecular diffusion. The molecular diffusion code was designed for gases and is unstable for the larger mass of the dust particles. Even for small (~1 nm) particles, the Brownian diffusion is significantly less important than eddy diffusion and sedimentation, so we disabled molecular diffusion for the dust particles. For 2 simulations, a smaller initial particle size of 0.2 nm was used. At this size, the Brownian diffusion approaches the importance of eddy diffusion and sedimentation, so small errors, typically less than 5% in mass density, number density and effective radius for altitudes below 100 km may have been introduced by making this assumption. WACCM incorporates a gravity wave parameterization to include the effect of vertically propagating gravity waves generated in the troposphere. The default tuning of the gravity wave algorithm [Garcia et al., 2007; Sassi et al., 2002] is a compromise between generating a good climatology in the stratosphere and in the mesosphere. We conducted sensitivity tests with a different choice for the source strength of the gravity waves to better simulate the dynamical structure of the mesopause.

2.2. CARMA

[7] We used CARMA 2.3 configured for one-dimensional columns using the same vertical grid as WACCM. Most runs used 21 size bins for dust particles with a mean radius from 1 to 100 nm and a doubling of particle volume between adjacent bins. Some runs used 28 size bins with radii from 0.2 to 100 nm also with the volume ratio doubling between bins. The initial particle size is a function of the coagulation [Hunten et al., 1980; Rosinski and Snow, 1961] and polymerization [Plane, 2003] that may occur in the region of the meteor trail. We base our cases on the approximate size of an iron oxide or silicon oxide molecule of 0.2 nm and a slightly larger size of 1 nm that assumes more initial recombination. We assume spherical particles and a dust particle bulk density of 2.0 g cm⁻³ [Hunten et al., 1980].

[8] The dust source function uses a production rate from Kalashnikova et al. [2000], which has production from 75 to 110 km with a peak around 83 km. All particles are placed into the smallest size bin and the production rate is scaled over the area of the Earth and the production altitude range to preserve the global mean mass influx at a rate of ~16 kt yr⁻¹. This global mass influx is based on the Kalashnikova et al. [2000] production rate for particles with a radius of 1.3 nm and is similar to the rates measured by Hughes [1978] and Gabrielli et al. [2004]. The dust source is assumed to be temporally and horizontally uniform. Meteoric particles that do not ablate because of their size are not included in the model. These particles represent about 20% of the total mass [Kalashnikova et al., 2000], are larger (5–50 μm radius) [Kalashnikova et al., 2000] and are expected to sediment rapidly [Hunten et al., 1980; Turco et al., 1981]. For simulation 6, we used double the default global mean mass influx for a rate of 32 kt yr⁻¹. This is the fraction of the meteoric influx that would be expected to ablate when using a mass influx rate of 40 kt yr⁻¹, the value measured by Love and Brownlee [1993].

[9] CARMA includes routines for sedimentation and coagulation of particles. The fall velocities are calculated by assuming a Stokes flow with size corrections from Fuchs [1964]. Sedimentation is solved by using the piecewise parabolic method of Coletta and Woodward [1984]. Since WACCM handles vertical advection and eddy diffusion, no additional diffusion of aerosol particles is added by CARMA. Coagulation coefficients are calculated to include Brownian, convective and gravitational effects. A sticking
coefficient of 1 is used, which means that particles are assumed to stick together once they collide. The sticking coefficient and coagulation coefficients could be affected by particle charging [Hunten et al., 1980; Jensen and Thomas, 1991; Reid, 1997], intermolecular forces like van der Waals forces [Hunten et al., 1980] and magnetic dipole interactions [Saunders and Plane, 2006], which could either reduce or enhance coagulation. Megner et al. [2006] identified coagulation rates as a significant source of uncertainty in their meteoric dust model and it remains so in these simulations. CARMA calculates the effect of coagulation using the numerical approach described in Jacobson et al. [1994].

3. Results

We will show results from 6 simulations including a control run with a 10-year duration and 5 sensitivity tests with 4-year durations. The definition of each simulation is shown in Table 1. All simulations except simulations 5 and 6, the 125 level cases, start with the same initial condition and no of them start with any meteoric dust. The 125 level cases have an initial condition interpolated from a slightly different reference simulation; however, these differences should not have an influence on the results. Figure 1 shows the total mass of dust in the atmosphere as a function of height and time for the control run. A semiannual pattern in dust mass exists in the mesosphere because of seasonal variations in the mesospheric meridional circulation. An annual pattern exists in the stratosphere and troposphere, because of the enhanced mass transported down in the southern hemisphere winter relative to the northern hemisphere winter and because of mixing during the relatively slow descent through the stratosphere. The model nearly reaches steady state in the entire atmosphere after 10 years and reaches steady state above 30 km within 3 years (Figure 2). Since we are not modeling the interaction of dust with sulfate aerosols and water vapor, we do not expect the model to properly represent the state of the dust particles in the troposphere or lower stratosphere. Therefore the analysis that follows is restricted to heights above 30 km. In the control run, we have averaged the last 7 years of data.

![Dust Mass, 10 Years](image)

**Figure 1.** Evolution of the vertical distribution of the mass of meteoric dust in the atmosphere during the 10 years of control simulation (left) and larger view of the last 2 years (right).

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Duration, years</th>
<th>Particle Radius, nm</th>
<th>Levels</th>
<th>$\tau^*_b$, Pa</th>
<th>Notes</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 (Control)</td>
<td>10</td>
<td>1–100</td>
<td>66</td>
<td>0.006</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>4</td>
<td>1–100</td>
<td>66</td>
<td>0.006</td>
<td>no sedimentation</td>
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<td>66</td>
<td>0.006</td>
<td>no coagulation</td>
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</tr>
<tr>
<td>4</td>
<td>4</td>
<td>0.2–100</td>
<td>125</td>
<td>0.002</td>
<td></td>
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</tr>
<tr>
<td>5</td>
<td>4</td>
<td>0.2–100</td>
<td>125</td>
<td>0.002</td>
<td>2× meteoric influx</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>4</td>
<td>0.2–100</td>
<td>125</td>
<td>0.002</td>
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</table>

*aAll of the simulations include the Kalashnikova et al. [2000] source function applied to the smallest size bin and wet deposition based on Barth et al. [2000]. Most simulations include the microphysical processes of coagulation, sedimentation, and wet deposition. Simulation 5 has a meteoric influx that is twice the mass of the other cases. The background gravity wave source strength is controlled by the $\tau^*_b$ parameter in the WACCM gravity wave parameterization [Garcia et al., 2007; Sassi et al., 2002]. The default for $\tau^*_b$ is 0.006 Pa, while a value of 0.002 Pa creates a more realistic mesopause.

**Table 1.** Definitions of the Control Run and Sensitivity Tests

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but in the sensitivity tests, we only use the last year of the 4-year run. Comparisons between the control run and the sensitivity tests are made using the fourth year of each simulation; however, the confidence levels associated with the differences between simulations are based on the variance in the last 7 years of the control run.

3.1. Control Run

Our control simulation uses the same WACCM configuration as Marsh et al. [2007] using solar maximum and 1995 greenhouse gas conditions and an initial dust particle radius of 1 nm. The simulation was run for 10 years and the average of the last 7 years for January and July is shown in Figure 3. There is a large seasonal variation in dust concentrations caused by the mesospheric circulation that goes from the summer pole to the winter pole. These two solstice seasons are the extremes cases, with a more uniform distribution of dust occurring in-between. It can also be seen that the mass density is highest over the winter pole and increases downward into the stratosphere. There is a peak in effective radius at low latitudes and altitudes because the circulation ascends from the summer pole to the low latitudes in the winter hemisphere and thus only large particles with higher sedimentation velocities will be able to descend in the tropics. Other than near the dust source altitude, where no vortex was identified, the dust is largely confined to the vortex. The peak of the dust mass takes about a month, from April to May to descend through the mesosphere; however, it takes an additional four months, until September, for the peak mass to descend to 30 km.

To study the vortex in more detail, the vortex identification technique of Harvey et al. [2002] is used to locate the vortex in the last year of the data. This technique defines the vortex edge kinematically by separating the flow field into strain and rotational components and then marking the vortex as a region bounded by a stream function contour that is dominated by rotational cyclonic flow. In the stratosphere, this technique agrees well with the widely used Nash et al. [1996] vortex definition, which is based upon maximum potential vorticity gradients. The structure of the polar vortices identified by using Harvey et al. [2002] transitions smoothly across the stratopause and identifies the polar vortices well into the upper mesosphere. Figure 4.
shows the area-weighted average mass mixing ratio over the entire hemisphere of all the meteoric dust inside and outside of the vortex for both hemispheres. The seasonal transport of dust between hemispheres because of the meridional circulation can be seen clearly both inside and outside of the vortex between potential temperatures of 4000 and 5000 K, which corresponds to heights of about 70 to 80 km. At lower potential temperatures, dust can be seen descending into the stratosphere mostly inside the vortex from May through October in the Southern Hemisphere and from December through April in the Northern Hemisphere.

Figure 5 shows the ratio of the dust mass mixing ratio inside the vortex to that outside the vortex. This figure highlights that the dust is well confined while it descends inside the vortex, particularly in the southern hemisphere winter from 700 to 2000 K. While there are large seasonal variations in the dust mass mixing ratio above 4000K, no significant enhancement is seen inside the vortex relative to outside of the vortex. In the southern hemisphere, a weak "vortex" with enhanced dust concentrations persists throughout the year between 500 and 1000 K. Curtius et al. [2005] measured the fraction of stratospheric aerosols with a nonvolatile component and found a ratio of 3:1 between the inside and the outside of the vortex. Their measurements were made between 10 January 2003 and 19 March 2003 at potential temperatures of 300 to 500 K near Kiruna, Sweden at a latitude of ~68°N. While not measuring exactly the same thing, in Figure 5 we get a ratio ~3:1 at a similar time and location. Figure 5 indicates that much larger dust gradients should be measurable at other times and locations.

Hunten et al. [1980] did some of the fundamental modeling work on meteoric dust using one-dimensional simulations with an earlier version of CARMA using conditions from midlatitudes. In Figure 6, we compare our results for the 1 nm case to theirs presented in their Figures 4 and 5. Hunten et al. [1980] is most relevant to midlatitudes, but our annual averages at high latitude are most similar to their results. They significantly overestimate the concentration and surface area density below 75 km.

Figure 4. The annual variation in the vertical profile of the meteoric dust mass mixing ratio inside (left) and outside (right) the polar vortex for the Northern (top) and Southern (bottom) Hemispheres.
during the summer at mid to high latitudes relative to our results. Our results also show an enhancement during the winter at high latitudes and a significant depletion at low latitudes during all seasons. In Figure 7, we compare our size distributions at 30, 60 and 90 km to those of Hunten et al. [1980]. At 90 km, we have fewer large particles than did Hunten et al. [1980] in the summer because of the mesospheric circulation, but overall the distributions are similar. At 60 km, the summer size distributions show a large depletion in small particles, but a slight enhancement in large particles relative to Hunten et al. [1980] because of ascending air in the polar region during the summer. At 30 km, the distribution is shifted toward larger particles than Hunten et al. [1980] found because of the enhanced dust concentrations from the mesospheric circulation and the confinement within the polar vortex.

[14] It is very difficult to detect meteoric dust particles, because of their small size and their remote location. Several groups [Gelinas et al., 1998; Havnes et al., 2001; Lynch et al., 2005; Rapp et al., 2005] have attempted to detect charged dust particles with rocket soundings and Strelnikova et al. [2007] have attempted to detect meteoric dust using incoherent scatter radar (ISR) data. Figure 8 shows a comparison of our data from the control run and simulation 5 and 6 with the dust observations from Lynch et al. [2005] made on 7 March 2002 and 15 March 2002 from Poker Flats, Alaska. Simulations 5 and 6 have a higher vertical resolution than the control run and also have a smaller initial particle radius. Simulation 6 has twice the mass of meteoric influx of the control run and simulation 5. The interpretation of charge particle data from rocket soundings is controversial with significant uncertainties in the aerodynamic properties of the inlet, the sensitivity of the detector, and what fraction of the dust particle size distribution is charged [Rapp et al., 2007]. To compare their charged particle measurements to our particle concentrations, we have scaled up their results assuming that only 5% of the total particles were charged and that the particles they detected were ~1 to 1.5 nm in radius [Gelinas et al., 2005]. This charging assumption probably underestimates the number of charged nanometer size dust particles, but preliminary aerodynamic studies indicate that the instrument is sensitive to 1 to 1.5 nm dust particles at altitudes above 83 km and that data below this level is suspect (Gelinas, personal communication 2008). Therefore while this comparison should only be considered qualitative, we see that the observations are bracketed by the control run and simulation 5, suggesting the best fit would be obtained with an initial radius larger than 0.2 nm or with an increased meteoric influx. There is good agreement in the altitude of the upper edge of the dust distributions, and the overall shape and slope of the profiles from 83 to 95 km; however, the exact positioning of the simulated profiles depends on the radius range used for the comparison. The curves shift lower in altitude and concentration if larger particles are assumed to have been detected.

[15] Strelnikova et al. [2007] analyzed data acquired on 10 September 2006 from the ISR at the Arecibo Observatory and have detected a signal that they attribute to charged meteoric dust particles. Like the rocket measurements, this retrieval is subject to particle charging uncertainties and they also have to make assumptions about the temperature, neutral air density and the mass of the positive ions that also contribute to the ISR signal. However, this technique is free of the aerodynamic issues that affect the rocket measurements. Figure 9 shows a comparison of the dust effective radius and number density from Strelnikova et al. [2007] with the 10 September daily average data for the grid point containing Arecibo from the control run and simulations 5 and 6. The retrieved radius is between the effective radius from the simulations with a 0.2 nm and 1 nm initial radius, suggesting that the initial particle radius would need to be ~0.75 nm for the best agreement of the simulation and the observations. Below ~90 km, the retrieved dust number density appears to be limited by the available charge; however, above this level the electron density increases.
and the retrieved dust number density decreases. This suggests that the number of dust particles and not the available charge is limiting the retrieved signal. The retrieved dust number density is 2 to 3 orders of magnitude lower than that from the 0.2 nm simulations. The 1 nm simulation has dust number densities above 90 km that are similar to the values from Strelnikova et al. [2007], again suggesting that the initial particle radius may be closer to 1 nm than 0.2 nm. An increased meteoric influx also increases the effective radius from the simulations; however, the mass would have to be increased substantially for the 0.2 nm simulations to match the data and this would increase the number density, which is already too large. Increased coagulation rates beyond those assumed in the simulations would both increase the effective radius and decrease the number density allowing for an initial radius smaller than 0.75 nm. We include all dust particles in the effective radius calculation. If the larger particles are charged preferentially, this will shift the retrieved effective radius to higher values.

3.2. Importance of Sedimentation

Figure 10 shows the results of simulation 2, which does not include the sedimentation of dust particles. Sedimentation plays an important role in defining the sharp drop off in particle concentration above 90 km, since particle fall velocities are larger than the vertical wind in this region (Figure 11). Sedimentation also plays a role in the descent of particles in the mesosphere and stratosphere. Without sedimentation, the particles take longer to descend in the polar vortex, which leads to increased mass densities in the mesosphere and decreased mass densities in the upper

Figure 6. Comparison of the average vertical profile of concentration and surface area density of meteoric dust between Hunten et al. [1980] and the control simulation. Averages are calculated in different latitude ranges for the year (left), DJF (center), and JJA (right).
stratosphere. The differences between the simulations in the stratosphere are less than 30% of the total mass density, so sedimentation is not as important as vertical advection for the transport of meteoric dust to the stratosphere. The model assumes that the dust particles are spherical. If the larger particles have a fluffier shape [Saunders and Plane, 2006], then the fall velocities will be reduced and sedimentation would play a smaller role in the lower mesosphere and stratosphere than indicated by these results.

3.3. Importance of Coagulation

The results of simulation 3, which does not include coagulation, are similar to the results from simulation 2 (Figure 10) in the mesosphere and stratosphere. This similarity indicates that it is the sedimentation of large particles formed by coagulation that is important in the stratosphere and lower mesosphere, while it is the sedimentation of small particles that is important for defining the sharp gradient in number concentration near 90 km (Figure 3). Coagulation is enhanced in the polar regions because of the increased concentration of dust from the meridional circulation and the subsequent confinement of dust in the polar vortex. We assume a sticking coefficient of 1.0 in the model; however, particle charging in the lower thermosphere and upper mesosphere [Jensen and Thomas, 1991; Reid, 1997] and dipole–dipole interactions [Saunders and Plane, 2006] could impact the choice of the sticking coefficient and the coagulation kernels. Further study is needed to reduce the uncertainty in the effect of the coagulation process upon the distribution of meteoric dust particles.

3.4. Impact of a Smaller Initial Dust Particle Size

Lacking observations, there is uncertainty as to what the particle size is after the ablation and initial recombination in the meteor trail. Rosinski and Snow [1961] indicate that most of the ablated material in a meteor trail diffuses rapidly into the background atmosphere without condensing. This gives rise to global layers of metal atoms that, over a period of days, form stable oxides, carbonates and silicates making a molecular scale (~0.2 nm) the most realistic initial particle radius. Hunten et al. [1980] suggest that uncertainties in the size of the meteor trail, the rate the trail spreads and the coagulation rate could allow for a larger initial particle size. Hunten et al. [1980] use initial radii ranging from 0.2 nm to 10 nm in their simulations. Here we only consider initial radii of 1 nm and smaller.

Figure 12 shows the results of simulation 4, which assumes an initial particle radius of 0.2 nm and a production rate that scales with radius to conserve the total mass being produced at ~16 kt yr\(^{-1}\). The increase in the small particle production rate causes an increase in particle concentration near the source region. Because the transport of dust away
Figure 8. Comparison of particle concentrations of simulations 1, 5, and 6 with and rocket soundings by Lynch et al. [2005] from Poker Flats, Alaska. Their observations of charged particles have been converted into dust particles by assuming that 5% of the dust particles are charged and that the particle radius is \( \sim 1 \) to 1.5 nm [Gelinas et al., 2005; Lynch et al., 2005]. The solid line with squares is the average of the last 7 years of the control run taken from the 4° × 5° model grid point containing Poker Flats using a daily average for the date indicated. The solid lines with the open and solid circles are sampled in a similar way from the fourth year of simulations 5 and 6, respectively. The vertical resolution of the control run near the mesopause is \( \sim 3.5 \) km, while simulations 5 and 6 have an increased resolution of \( \sim 0.5 \) km. The points on the solid lines indicate the midpoint altitudes of the model levels. The dashed lines are the observations.

Figure 9. Comparison of the vertical profiles of the simulated daily average meteoric dust effective radius (left) and number density (right) for 10 September at Arecibo (18.3°N 67.8°W) with ISR retrievals from Strelnikova et al. [2007]. The x’s are the data from Strelnikova et al. [2007]. The solid lines with the open and solid circles are from the fourth year of simulations 5 and 6, respectively, and have an initial radius of 0.02 nm with 1× and 2× meteoric influx. The dashed line with the open square is the average of the last 7 years of the control run and has an initial radius of 1 nm. The dotted line is a corresponding electron density profile at Arecibo (Rapp, personal communication, 2008).
from the summer pole by the meridional circulation is relatively rapid, coagulation cannot quickly create 1 nm particles so there is a reduction in the number of particles 1 nm or larger in the summer mesopause region relative to the 1 nm initial size case. This reduction in the effective radius causes a slower descent of mass into the stratosphere leaving increased mass densities in the mesosphere; although, this change is smaller than the variation in the last seven years of the control run. Vertical profiles of dust concentration, surface area density and size distributions have spatial and temporal differences compared to those of Hunten et al. [1980] for their 0.2 nm results in a pattern similar to those previously shown for the control run. The dust size distribution and number concentrations are sensitive to the initial size of the particles as they emerge from the meteor trail. An initial particle size of 0.2 nm is consistent with a recent laboratory experiment and associated modeling by Saunders and Plane [2006] to mimic smoke production from metal and silicon atoms in the presence of ozone and water; however, it is inconsistent with comparisons to the ISR data from Strelnikova et al. [2007] that suggest a larger initial radius. Additional observations are needed to constrain the model.

### 3.5. Implications for PMC Nucleation

Meteoric dust is one of the leading candidates to be the source of condensation nuclei (CN) needed for the...
formation of PMCs \cite{RappandThomas, 2006; Turco et al., 1982}. \textcolor{red}{Rapp and Thomas} [2006] have reviewed current theories about the nucleation of PMCs. \textcolor{red}{Turco et al.} [1982] and \textcolor{red}{Rapp and Thomas} [2006] indicate that concentrations of several hundred particles per cm\(^3\) with a radius of at least 1 to 1.5 nm are necessary to form PMCs under typical summer mesopause conditions. However, this requirement is not based upon observations and rests on numerous assumptions about the nucleation process including the particle charge, morphology and surface properties. Large temperature variations by gravity waves could permit ice nucleation on the more numerous smaller dust particles. A sufficient concentration of large particles are present in the control run to nucleate under the traditional assumptions, but reductions in the number concentration of large particles are seen when starting with a smaller initial radius and when the gravity wave parameterization is tuned to make a more realistic mesopause. In simulation 5, we investigate the availability of meteoric dust CN by using a higher vertical resolution that provides \(\sim \)0.5 km spacing near the mesopause, an initial dust particle radius of 0.2 nm, and the WACCM gravity wave parameterization tuned for a

\textbf{Figure 12.} Zonal average plots for July of the concentration of particles with 1 nm or larger radius, mass density, and effective radius for simulation 4, the 0.2-nm initial radius case. The percentage difference between these results and the control simulation are shown (right). The stippled area on the difference plots indicates where the difference is not significant to the 90\% confidence level relative to the control run.
more realistic mesopause height and temperature by decreasing the gravity wave source strength to 0.002 Pa. Our results from simulation 5 show that the dust concentrations of particles 1 nm and larger are 15 to 30 cm$^{-3}$ in the summer mesopause region with maximum values ~3 km above the temperature minimum and that there are very few particles below the mesopause (Figure 13). These seasonal patterns are fairly symmetric. There are widespread

![Figure 13](image)

**Figure 13.** Zonal average plots for January and July of the concentration of particles with 1 nm or larger radius for simulation 6, which has enhanced vertical resolution and a more realistic mesopause thermal structure than the control run. The solid lines are temperature contours spaced every 10 K from 130 to 160 K.

![Figure 14](image)

**Figure 14.** Comparison of the average size distribution of meteoric dust during the summer at high latitude between Hunten et al. [1980] (thick line) at 90 km and simulation 6 (thin lines) at altitudes of 90, 88, and 95 km
areas with large depletions relative to Hunten et al. [1980] and to our other simulations, particularly below the mesopause. The dust particle size distribution at several levels near the summer polar mesopause from simulation 5 is compared to Hunten et al. [1980] in Figure 14. There is a significant depletion of particles at high latitude in the summer relative to Hunten et al. [1980]; however, the ascending winds do appear to loft a small amount of larger particles. The meteoric dust concentration, mass density and size distribution are compared to Megner et al. [2006] in Figures 15 and 16. The summer depletion and winter enhancement in concentration and mass density are not as large as those seen in Megner et al. [2006], particularly for particles that are 1 nm and larger in the summer. This simulation shows that size distributions like Hunten et al. [1980] often used in PMC models, significantly overstate the number of 1 nm and larger particles available and that particle concentrations are extremely depleted below the mesopause. However, summer conditions are not as depleted as indicated by Megner et al. [2006], who used extreme vertical velocities from CHEM2D [Summers et al., 1997] for their winter and summer wind profiles. If meteoric dust is a CN for PMC particles, nucleation is more likely to occur at or above the mesopause and nucleation may require higher supersaturations than is typically assumed. Sedimentation of PMC particles containing meteoric dust could further reduce the concentration of larger dust particles in the nucleation region. It is possible that meteoric dust may interact with other gases and aerosols to create larger particles that would facilitate PMC nucleation [Mills et al., 2005]. The concentration of large dust particles at the summer pole would also be larger than in simulation 5, if

Figure 15. Comparison of the average vertical profile of concentration and mass density of meteoric dust during the summer and winter between Megner et al. [2006] and simulation 6 at high latitude in the Northern Hemisphere.

Figure 16. Comparison of the average size distribution of meteoric dust between 80 and 90 km during the summer and winter between Megner et al. [2006] and simulation 6 at high latitude in the Northern Hemisphere.
there were a larger meteoric influx. Simulation 6 uses a meteoric influx of 32 kt yr$^{-1}$, which results in a distribution of 1 nm and larger dust particles in the polar summer mesopause similar to that shown in Figure 13, but with a maximum value in the region above the mesopause of 120–250 cm$^{-3}$. Because coagulation is a nonlinear process, increasing the meteoric influx by a factor of 2 results in a 8-fold increase in the concentration of large particles, which approaches the values traditional assumed to be necessary for PMC nucleation.

4. Summary and Future Work

[22] These are the first three-dimensional microphysical simulations of meteoric dust and they show a strong seasonal pattern with meteoric dust particles being transported horizontally from the summer pole to the winter pole where they are quickly transported down into the stratosphere within the polar vortex. The increased concentration of dust particles in the vortex creates larger particles in the stratosphere than previously suggested by one-dimensional simulations and may make meteoric dust important to the polar CN layer and the sulfate aerosol layer. Any impact on the characteristics of sulfate aerosols in the winter stratosphere could also affect PSC nucleation. While the few published observations of meteoric dust contain a number of uncertainties and assumptions, the comparison of the model with the data from Lynch et al. [2005] and Srelnikova et al. [2007] suggests that either the initial particle radius may be as large as 0.75 nm or that the coagulation of smaller particles is more efficient than we have assumed. The prediction of a seasonal variation that creates areas of high concentrations and high-concentration gradients may be useful for the planning of future campaigns to measure meteoric dust.

[23] The calculated depletion of dust particles in the summer mesopause region may reduce the concentration of large particles near the mesopause below the number traditionally thought to be needed for PMC nucleation; although, the amount and location of the depletion depends on the initial particle radius, the gravity wave parameterization and the mass of the meteoric influx. Uncertainties in the coagulation efficiency could also have an impact on the number of larger particles available. Additional dust observations and laboratory studies are needed to constrain the model. For studies of the summer mesopause, additional tuning and validation of the WACCM gravity wave parameterization is necessary. We have constructed a PMC model using WACCM/CARMA and the dust source described here, which we plan to use to evaluate the potential for meteoric dust particles to be a significant source of condensation nuclei for PMCs. We intend to compare the model results to observations from the recently launched Aeronomy of Ice in the Mesosphere satellite mission, as well as other PMC climatologies.

[24] Based upon radar observations, Janches et al. [2006] have suggested that there are temporal and spatial variations in the meteoric influx. Some of this variation may get averaged out as the dust is accumulated and transported; however, it may be important in areas where dust is depleted such as near the summer mesopause. We plan to use a dust source function derived from Janches et al. [2006] and Fentzke and Janches [2008] to investigate the sensitivity of the dust distribution to temporal and spatial variations in the source function.

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References


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