Temporal variability of dust mobilization and concentration in source regions

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[1] The time-varying properties of dust mobilization and concentration variability in source areas are studied in model simulations to understand the processes that control entrainment of dust into the atmosphere. Two different meteorological reanalysis data sets (National Centers for Environmental Prediction/National Center for Atmospheric Research and NASA Data Assimilation Office) and two different source parameterizations are used for the analysis, which is done for six different main dust source regions. The results show that ∼35–70% variance of dust mobilization is associated with diurnal variability of dust mobilization in the major dust sources regions, and this is similar using different meteorological analyses and source parameterizations. Synoptic-scale variability is responsible for 6–50% of the dust mobilization, with more variation between different meteorology and parameterizations. The variability of dust concentrations in source regions is very sensitive to source meteorology and parameterization, with the synoptic variability responsible for 30–50% of the variability in some configurations or 6–40% in other configurations. Diurnal variability in concentrations varies from 20–50 to 25–80% depending on the model configuration. Analyses suggest that dust mobilization and dust concentration are more influenced by synoptic variability in Australia and east Asia than in North Africa and Arabian dust source regions, consistent with observations of cold fronts and cyclone movements systems driving dust mobilization and transport in Australia and east Asia source regions. Differences in variability between different meteorological data sets and source parameterizations suggest that possible dust feedbacks onto the dust cycle may be sensitive to source parameterization and model.

INDEX TERMS: 0305 Atmospheric Composition and Structure: Aerosols and particles (0345, 4801); 0312 Atmospheric Composition and Structure: Air/sea constituent fluxes (3339, 4504); 0322 Atmospheric Composition and Structure: Constituent sources and sinks; 0368 Atmospheric Composition and Structure: Troposphere—constituent transport and chemistry; KEYWORDS: mineral aerosol, temporal variability, radiative forcing


1. Introduction

[2] Mineral aerosols are suggested to play an important role in climate forcing by altering the radiation balance in the atmosphere [e.g., Miller and Tegen, 1999], which, in turn, can feed back onto the dust cycle [Perlwitz et al., 2001]. In addition, mineral aerosols can provide surface area for heterogeneous reactions in the atmosphere and can significantly affect the cycles of atmospheric species [e.g., Dickerson et al., 1997; Dentener et al., 1996]. Finally, mineral aerosol deposition impacts nutrient cycles in ocean and terrestrial ecosystems and thereby could play an important role in modulating the global carbon cycle [e.g., Martin, 1990; Archer et al., 2000; Watson et al., 2000; Chadwick et al., 1999]. There are major uncertainties in the modeling of dust aerosol due to the sparseness of the observational data and gaps in our understanding of the physical and chemical properties of mineral aerosol. Mineral aerosol entrainment into the atmosphere (called mobilization) is sensitive to a wide range of factors including the composition of the soils, the soil moisture content, surface condition, and wind speed and may be modulated by human activities and land degradation. Because detailed
soil and source information is not available, there are differences in the source parameterizations used in different models. In addition, while several models use reanalysis winds [e.g., Schulz et al., 1996; Ginoux et al., 2001; Tegen et al., 2002], in areas with few meteorological observations such as desert areas it is likely that important meteorological parameters such as surface wind speeds are not well observed but rather are sensitive to forecast center model.

Figure 1a shows the yearly mean dust mobilization (yearly mean data averaged from daily data for BASE case in 1995), and Figure 1b shows the dust mobilization variability (standard deviation of daily averaged mobilization divided by annual mean mobilization) in each grid box, showing that dust mobilization is highly variable temporally, with standard deviations more than a factor of 10 larger than the mean. Previous studies show North African dust undergoing large variation with different timescales (interannual, seasonal, and synoptic) [e.g., Westphal et al., 1987; Marticorena et al., 1997; Schulz et al., 1996; Jones et al., 2003; Luo et al., 2003; Mahowald et al., 2003]. Jones et al. [2003] show that 20% of the variability of dust mobilization is associated with African Easterly Wave (AEW) activities in North Africa during the summer while 8% is due to interannual variability [Mahowald et al., 2003]. What processes control the rest of the variability in dust mobilization? What contributions are associated with diurnal cycles of dust mobilization and dust concentration? What timescale controls variability of other dust sources areas, such as the Arabian Peninsula (43°–56°E, 20°–30°N), Australia (118°–138°E, 27°–31°S), and east Asia (75°–80°E, 35°–45°N; 100°–110°E, 37°–45°N)? In order to understand these questions we extend the previous studies [Jones et al., 2003; Mahowald et al., 2003] by using model simulations from 1995 [Luo et al., 2003] and spectral analysis techniques [Bingham et al., 1967; Jones et al., 2003; Madden and Julian, 1971; Mitchell, 1966; Madden and Julian, 1971; Wallace and Chang, 1969] to quantitatively evaluate the variability associated with different timescales. The results of Miller et al. [2004] suggest that much of the feedback of dust onto climate and dust sources is due to changes in the surface heating budget, which is associated with the diurnal variability in dust, highlighting the importance of understanding the relative importance of diurnal and synoptic variability in driving the dust cycles.

In section 2 we describe the model simulations conducted for this paper. In section 3 we show the results of

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**Table 1. Sensitivity Case Description**

<table>
<thead>
<tr>
<th>Source area topographic lows</th>
<th>GDD</th>
<th>BASE</th>
<th>GDN</th>
<th>BDD</th>
</tr>
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<td>GDD</td>
<td>DEAD</td>
<td>GOCART</td>
<td>DEAD</td>
<td>GOCART</td>
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<td>Data for mobilization calculation</td>
<td>NCEP</td>
<td>DAO</td>
<td>DAO</td>
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<td>Data for transport calculation</td>
<td>NCEP</td>
<td>DAO</td>
<td>DAO</td>
<td>NCEP</td>
</tr>
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</table>

*From Ginoux et al. [2001].
*From Zender et al. [2003].
spectra and statistics analyses. In section 4 we summarize the results of this study.

2. Model Description

For this study we use simulations described in more detail by Luo et al. [2003], using the Model of Atmospheric Transport and Chemistry (MATCH), [Rasch et al., 1997; Mahowald et al., 1997], coupled with the Desert Entrainment and Deposition (DEAD) model [Zender et al., 2003]. This mobilization scheme is based on the wind tunnel and in situ studies of Iversen and White [1982], Marticorena and Bergametti [1995], Gillette et al. [1997], and Fecan et al. [1999]. It is similar to those used by Tegen and Fung [1994], Mahowald et al. [1999], Guelle et al. [2000], Ginoux et al. [2001], and Tegen et al. [2002] in that it is based on a wind threshold velocity and has a wind speed cubed relationship for dust mobilization, but the details of the mobilization are slightly different in each case. The DEAD module mobilization scheme [Zender et al., 2003] is based on the friction velocity (wind stress at the surface), not the wind velocity on the surface level in the model, consistent with the original derivations [e.g., Gillette and Passi, 1988]. The wind stress is a function of the wind speed, roughness length, and the atmospheric stability at the surface. The surface roughness length in dust-producing regions is set to the globally uniform value of 100 μm, a value more typical of erodible soil beds [Gillette et al.,

![Figure 2](image1.png)

**Figure 2.** Time series of (a) dust mobilization and (b) dust surface concentration in May 1995 for a location in the Sahara in the model (24°N, 12°W); solid line for BASE and dotted line for GDD cases.

![Figure 3](image2.png)

**Figure 3.** Spectral variance of (a) dust mobilization (mbl), (b) dust surface concentration (dst), (c) surface wind (wnd), and (d) wind friction velocity (wndf) in summer in west Saharan source area for BASE case. Smoothed solid line represents the background red noise spectrum, and dashed line represents the 95% significance level.
Figure 3
1997] than the large-scale roughness lengths for bare ground (~5 cm) used in general circulation models [e.g., Bonan et al., 2002]. The mass flux of saltating particles depends on the excess of the wind friction speed over the threshold wind friction velocity for saltation. More details about the source parameterizations are available from Zender et al. [2003] and Luo et al. [2003]. Because accurate soil characteristics are not available globally at sufficient resolution, we assume that soils are replete with particles optimally sized to initiate saltation, and we use a factor to describe what fraction of each grid box consists of easily erodible soils. In this study, we follow Ginoux et al. [2001], Zender et al. [2003], and Mahowald et al. [2002] and assume that all topographic lows with little vegetation and low soil moisture are dust sources, using the time-independent source area from Ginoux et al. [2001], which includes only nonvegetated low-lying regions (see Figure 1a). There is some debate about the sources of desert dust [e.g., Prospero et al., 2002; Mahowald et al., 2002; Luo et al., 2003; Mahowald et al., 2003; Mahowald and Dufresne, 2004; Tegen et al., 2004]; however, some modeling studies have indicated low sensitivity to the exact location of sources [e.g., Luo et al., 2003]. In this study, the mobilization (or emission of dust into the atmosphere) is derived from multiplying the source area by the mobilization calculated using the DEAD dust module [Zender et al., 2003]. The fractions of the clay (dust particles smaller than 1 μm) and silt sizes are different for different soil types at each location. Owing to the uncertainties in the available soil texture data, following Tegen and Fung [1994] and Ginoux et al. [2001], we chose a globally constant particle size distribution; the fractions are 0.1 for the class 0.1–1 μm and 0.3 for the classes 1–2.5, 2.5–5.0, and 5.0–10 μm, where the sizes are diameters of the particles. Within each bin we assume lognormal distributions in aerosol sizes [Zender et al., 2003].

Both dry and wet deposition are included as loss processes for the mineral aerosols. Dry deposition processes are simulated following Seinfeld and Pandis [1996] and include turbulent deposition and gravitational settling, with the latter dominating for large particles. The efficiency of mineral aerosol removal by rain is usually described by the scavenging ratio Z, which is defined by $Z = C_{\text{rain}}/C_{\text{air}}$, where $C_{\text{rain}}$ is the dust concentration in rain in grams of dust per kilogram of rainwater, and $C_{\text{air}}$ is the aerosol concentration in air in grams of dust per kilogram of air. We chose a wet scavenging efficiency of 750 for all aerosol sizes and for convective and stratiform precipitation [Tegen and Fung, 1994], which lies with the range of measurements for clay-size particles and results in a wet deposition lifetime of the dust of ~12 days.

Transport is calculated using the MATCH off-line chemical transport model [Rasch et al., 1997; Mahowald et al., 1997], which has been shown to simulate wet and dry convective mixing, in addition to large-scale precipitation processes, while using the National Centers for Environmental Prediction (NCEP) reanalysis [Mahowald et al., 1997]. The horizontal resolution of the model is T62 (~1.9° × 1.9°) and 28 vertical levels from surface to 10 mbar (the same as the resolution of the NCEP reanalysis made available at the National Center for Atmospheric Research.

For sensitivity studies, simulations are also conducted using the Global Ozone Chemistry Aerosol Radiation Transport (GOCART) emissions scheme and Data Assimilation Office (DAO) data sets similar to Ginoux et al. [2001]. The dust mobilization of GOCART based on empirical formulation by Gillette and Passi [1988] differs from the DEAD scheme in details but also is based on the 10 m wind height instead of the wind friction velocity and thus is less sensitive to atmospheric stability. More details about the scheme are available from Ginoux et al. [2001], and more details about our implementation are available from Luo et al. [2003].

Our BASE case used NCEP/MATCH/DEAD, the same as Luo et al. [2003]. In the sensitivity case (GDD) we implemented GOCART’s mobilization [Ginoux et al., 2001; Luo et al., 2003] and used the NASA DAO winds for both mobilization and transport. Case BDD used BASE mobilization methodology (DEAD from Zender et al. [2003]) but with mobilization calculated by DAO winds and with DAO transport. Case GDN used GOCART’s mobilization scheme and mobilization calculated by DAO, but transport was driven by NCEP winds. Table 1 summarized four cases we used in the study. The transport, dry, and wet remove processes are the same in all cases.

In order to explore the dust diurnal variability we archive the dust mobilization, concentration, and some related parameters at every time step (2400 s), realizing that the model interpolates 6 hour NCEP reanalysis data to this time step. The following procedure was used to calculate the spectrum: (1) the seasonal mean and linear trend were removed; (2) the resulting time series was tapered with a split cosine bell function (5% at each end); (3) fast Fourier transform was used to obtain raw spectral

Table 2. Annual Mean Contribution (Percent) of Synoptic (s) and Diurnal (d) Variances for MATCH/DEAD/NCEP (BASE)a

<table>
<thead>
<tr>
<th></th>
<th>West Sahara</th>
<th>East Sahara</th>
<th>Sahara Peninsula</th>
<th>Australia</th>
<th>East Asia</th>
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<tbody>
<tr>
<td>mbl (s)</td>
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<td>17.8</td>
<td>10.0</td>
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<td>43.6</td>
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<td>72.9</td>
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<tr>
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<td>37.9</td>
<td>50.9</td>
<td>52.5</td>
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<td>44.2</td>
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<tr>
<td>wnd (s)</td>
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<td>75.4</td>
<td>74.9</td>
<td>67.5</td>
<td>42.4</td>
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</table>

aData averaged from spring, summer, fall, and winter.

Table 3. Seasonal Contribution (Percent) of Synoptic (s) and Diurnal (d) Variances in West Sahara for MATCH/DEAD/NCEP (BASE)

<table>
<thead>
<tr>
<th></th>
<th>Spring</th>
<th>Summer</th>
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<th>Winter</th>
<th>Average</th>
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<tr>
<td>dst (s)</td>
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<td>wnd (s)</td>
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<td>wnd (d)</td>
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<td>22.4</td>
<td>30.8</td>
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<td>wndf (s)</td>
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<td>16.6</td>
<td>23.1</td>
<td>19.3</td>
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<tr>
<td>wndf (d)</td>
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<td>74.9</td>
<td>69.2</td>
<td>63.4</td>
<td>66.6</td>
</tr>
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</table>
Figure 4
estimates; (4) the raw spectral estimates were then smoothed with a running average of length $L = 3$. The degrees of freedom were estimated by

$$D = \frac{2LN_{ef}}{N} N_m = \frac{2L(0.873N)}{N} N_m = (2L)(0.873)N_m,$$

taking into account tapering of the time series [Madden and Julian, 1971; Jones et al., 2003], where $L$ is the length of running average, $N_{ef}$ is effective number, and $N$ is total number. The red noise background spectrum and 95% significance level were calculated following the methodology of Mitchell [1966], Madden and Julian [1971], and Jones et al. [2003].

[11] The contributions of variability from different timescales were calculated by integrating the relative variability spectral peak area divided by total spectral area. For example, the diurnal variability contribution was calculated by dividing the diurnal variance peak area ($\sim 0.45–1.2$ days) divided by total spectral area, and synoptic variability contribution was calculated by the synoptic peak area ($\sim 3–10$ days) divided by total spectral area.

3. Dust Mobilization Variability

[12] Figure 2 shows the dust mobilization and dust surface concentration time series simulated by the BASE and GDD cases in May 1995 in the west Sahara (23.8°N, 11.25°W) source region. Figure 2 shows the strong diurnal cycle and some synoptic timescale variations. In order to identify the diurnal variability and synoptic variability of dust mobilization and dust concentration we begin by examining the time series with spectral variance method from each season in 1995 for North Africa.

3.1. North African Source Areas

[13] Here we look at the temporal variability in three regions of North Africa (west Sahara, east Sahara, and the Sahel). We calculated the spectral variances of dust mobilization, dust concentration, surface wind, and wind friction velocity, which are essential to the dust mobilization and dust concentration. Figure 3 shows the spectral variability of dust mobilization (mbl), dust surface concentration (dst), surface wind (wnd), and surface wind friction velocity (wndf) during the summer in the west Saharan area (20°W–5°E, 20°–35°N). It can be seen that there are peaks of spectra which exceed the red noise background values from 0.45 to 1.4 days in the mbl, dst, wnd, and wndf associated with diurnal variability. There are also other peaks associated with synoptic timescales which do not exceed the red noise background from 3 to 9 days in dust mobilization and wind friction velocity spectral variance and which exceed the red noise background from 3 to 9 days for surface wind and surface dust concentration. Figure 3 shows that spectral variances of diurnal variability are smaller for dust surface concentration and surface wind velocity than for dust mobilization and wind friction velocity, also shown quantitatively in Table 2. Table 2 shows that the synoptic timescale variability (such as the AEW) and diurnal variability of surface wind and dust surface concentration are similar in the North African region. The variability is similar during spring, fall, and winter in North Africa. The diurnal variance is larger, and synoptic variance is smaller in summer (Table 3), consistent with more diurnal variability during the summer time. Seasonal contributions are similar in other dust source areas (not shown). Since easterly waves are strongest in the summer, we expected easterly waves to be more important in the summer [Diedhiou et al., 1999; Jones et al., 2003]. The dominance of the diurnal cycle in the summer appears to be due to the AEW being more important at 700 mbar, while surface processes are more important for dust mobilization (Table 3). The observation that the dust concentration has more synoptic-scale variability than the mobilization suggests that transport in this region is driven more by synoptic-scale variability.

[14] Results for the east Sahara region (5°–30°E, 20°–35°N) are similar to the west Sahara, except that the synoptic-scale variability is larger in the west Sahara than the east Sahara (Table 2). This result is consistent with the synoptic-scale variability, presumably driven by AEW, being stronger in west Sahara than in east Sahara [e.g., Jones et al., 2003].

[15] Similar to the other North African sources, for the Sahel region (15°W–20°E, 12°–20°N), diurnal variability dominates the mobilization and wind friction velocity, while dust concentration and wind speed have more variability in the synoptic timescale. The Sahel region has smaller amounts of synoptic variability in the mobilization or wind friction velocity than either the west or east Sahara (Table 2).

3.2. Arabian Peninsula, Australia, and East Asia Dust Source Areas

[16] Spectral analyses show that the Arabian Peninsula source (43°–56°E, 20°–30°N) has variability similar to the West African region in terms of the relative roles of diurnal and synoptic variability in the mobilization (Table 2), while the dust concentrations tend have more synoptic variability (and less diurnal variability) than in North Africa. In the Arabian Peninsula dust concentrations the synoptic variability occurs between about 3 and 9 days, similar to the Sahel, and is likely to be associated with frontal depressions [e.g., Pye, 1987].

[17] Figure 4 illustrates the spectral variance of dust mobilization, concentration, surface wind, and wind friction velocity in Australian (27°–31°S, 118°–138°E) dust source region during spring. Spectral analysis suggests that synoptic systems in Australia are more important than in the North Africa or the Arabian Peninsula source regions (especially for dust mobilization) (Figure 4 and Table 2). Synoptic variability is statistically significant for the dust mobilization and concentration between 2 and 10 days, and diurnal variability is less peaked than in Figure 3 for North Africa.

Figure 4. Spectral variance of (a) dust mobilization (mbl), (b) dust surface concentration (dst), (c) surface wind (wnd), and (d) wind friction velocity (wndf) in spring in Australian source region for BASE case. Smoothed solid line represents the background red noise spectrum, and dashed line represents the 95% significance level.
Figure 5. Same as Figure 4 except for the east Asian source region.
Africa. Previous studies shows that many major dust storms in south Australia are associated with nonprecipitation cold fronts, which are driven by the synoptic systems [e.g., Garratt, 1984].

Table 4. Annual Mean Contribution (Percent) of Synoptic (s) and Diurnal (d) Variances for MATCH/GOCART/DAO (GDD) Sensitivity Analysesa

<table>
<thead>
<tr>
<th>Region</th>
<th>West Sahara</th>
<th>East Sahara</th>
<th>Sahel</th>
<th>Arabian</th>
<th>Australia</th>
<th>East Asia</th>
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</thead>
<tbody>
<tr>
<td>mbl (s)</td>
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<td>28.36</td>
<td>24.66</td>
<td>20.36</td>
<td>39.05</td>
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<tr>
<td>mbl (d)</td>
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<td>67.08</td>
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<td>wnd (s)</td>
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<td>41.71</td>
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<td>39.67</td>
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<tr>
<td>wndrst (^{2}) (d)</td>
<td>59.26</td>
<td>61.54</td>
<td>73.84</td>
<td>71.42</td>
<td>42.81</td>
<td>55.57</td>
</tr>
</tbody>
</table>

aData averaged from spring, summer, and fall.

bWind at reference height.

[18] Figure 5 shows the spectral analysis for the spring for east Asia (35°–45°N, 75°–90°E, 37°–45°N, 100°–110°E). The diurnal variability in east Asia is different than in other regions in that the mobilization has no distinct diurnal peak but is more “noisy” and the variability associated with the diurnal cycle in concentration is much smaller. Figure 5 shows that synoptic systems are more important in dust mobilization and dust concentration in east Asia dust source area, consistent with observations of strong cold fronts and cyclonic motions driving dust events [e.g., Qian et al., 1997; Zhou et al., 2002].

3.3. Impact of Meteorology and Source Parameterization on Variability

[19] Also shown in Figure 2 are time series of dust mobilization and concentrations at the same location using the MATCH transport model, GOCART’s mobilization scheme [Ginoux et al., 2001] and NASA DAO’s meteorological data set (case GDD from Luo et al. [2003]). Notice the very different variability seen in the time series. Spectral analyses were also calculated for this case (referred to as case GDD) shown in Table 4 and Figure 6. Table 4 shows the mean synoptic and diurnal variances of mobilization, wind at reference height (10 m) averaged from spring, summer, and fall for the GDD case. The results for dust mobilization in North Africa suggest that there is more dust mobilization associated with synoptic systems in GDD compared with the BASE case. Figure 6 shows the spectral analyses of dust mobilization, dust surface concentration, and surface wind (magnitude) for MATCH/GOCART/DAO in west Sahara in summer. Figure 6 illustrates that for the summer season, synoptic variance in mobilization is larger in GDD than in the BASE case (42.2 versus 15.4%) and diurnal variance in mobilization for GDD is less than for BASE (40.6 versus 68.4%). However, the variability in dust surface concentrations shows a different signal in that synoptic variance of dust surface concentration is much smaller for GDD than for BASE (7.2 versus 26.0%) and diurnal variance of dust surface concentration is much larger for GDD than for BASE (86.6 versus 59.2%) in summer. A comparison of Tables 2 and 4 suggests that these signals are seen throughout the year (Table 4 is averaged from spring, summer, and fall since we only have 11 months of DAO data for 1995, but this does not qualitatively change our remarks, since the seasonal cycle in relative variability is not as large as the differences between BASE and GDD (see Table 3)). From Tables 2 and 4 it can be seen that diurnal variability of DAO surface winds (60–70%) is larger than synoptic variability of surface wind (15–30%) in Sahara, Sahel, and Arabian Peninsula dust source areas, while in the NCEP reanalysed the synoptic variability in surface winds is similar to the diurnal variability (30–50%). This suggests that the explanation for the increase in mobilization associated with synoptic systems in the GDD is not simply associated with surface winds.

[20] Understanding the differences becomes even more complicated when we look also at the concentration variability in the source regions of North Africa. Whereas the BASE case has more synoptic variability in the concentrations than in the mobilization (30–50% versus 10–20%), the GDD case has less synoptic variability in concentrations than mobilization (6–12% versus 25–40%) in North Africa. We conducted further sensitivity studies where we used the DEAD source scheme with the DAO winds (base case with DAO winds for source and transport or BDD) and the GOCART source scheme (DAO winds for mobilization, n NCEP winds for transport (called GDN)), and the results are shown in Figure 7. In order to compare different cases, four cases in Figure 7 are averaged from spring, summer, and fall. These results show that changing from NCEP winds (BASE) to DAO winds (BDD) does not systematically change the synoptic variability in mobilization in North Africa (6–20% versus 10–20%) or the daily variability (55–74% versus 63–72%) but increases the diurnal variability in concentration (50–55% versus 40–50%) and decreases the synoptic variability in concentration (17–27% versus 30–40%). The differences in behavior between the GDD and BASE appear to be due to differences in the atmospheric stability and the model’s sensitivity to atmospheric stability, as seen in the negative correlations of GDD to planetary boundary layer height (PBLH) and as discussed in section 3.4. In the other regions, especially the Arabian Peninsula and east Asia, we see similar differences between the BASE and GDD cases, with differences in the relative importance of diurnal and synoptic variability in mobilization and concentration. (There are small differences in mobilization variability between GDN and GDD, due differences in calculations using slightly different configurations in our model). The importance of the change in transport winds as well as mobilization (due to changes in wind and source parameterization) is shown in the GDN case, where the diurnal variability in dust concentrations is intermediate between case BDD and case GDD. Thus the differences between our BASE case and the GOCART-like case (GDD) are due to differences in the source mobilization, source area winds, and transported winds. It is unclear which of these reanalyses is more similar to the real atmosphere, but these results show the importance of both diurnal variability and synoptic variability in dust source regions in dust models.

[21] In order to explore the interannual variability in temporal variability we compare spectral variances between 1995 and 1998. Results are quite similar between the 2 years.
in terms of percent variability in the different regions, and they suggest that our results are not highly sensitive to interannual variability. These results show important differences depending on the meteorological data set and source parameterization used in the model. Unfortunately, high-resolution dust concentration data are not available for these remote regions, and thus verifying our results with observations is not tenable. Subgrid-scale processes which may be responsible for dust mobilization in these regions (such as dust devils or gust fronts associated with thunderstorms) are neglected in our modeling study, and thus the real atmosphere may have more dependence on small-scale impacts [e.g., Pye, 1987].
3.4. Correlation Analysis of Dust Mobilization and Concentration

[23] In order to estimate the relative roles of different processes on mobilization and dust concentration a correlation analysis is conducted. We calculated the correlations between dust mobilization and dust concentration with (1) the magnitude of the surface wind, (2) wind friction velocity, (3) soil moisture, and (4) boundary layer height, in which analysis values every 2400 s are correlated over the year for BASE and GDD cases. Significant correlations of dust mobilization with surface wind and wind friction velocity have been obtained in all dust source areas for BASE case. The correlations of mobilization versus surface wind are 0.43, 0.27, 0.50, 0.19, 0.49, and 0.23, and correlations of mobilization versus wind friction velocity are 0.68, 0.70, 0.75, 0.66, 0.60, and 0.34 in west and east Sahara, Sahel, Arabian Peninsula, Australia, and east Asia, respectively. High correlations between mobilization and wind friction velocity are consistent with the mobilization scheme used in the simulations. Lower correlations are observed with surface winds, since wind friction velocities incorporate a dependence on atmospheric stability. The correlations between surface wind and wind friction velocity are 0.50, 0.34, 0.66, 0.29, 0.76, and 0.77 in west and east Sahara, Sahel, Arabian Peninsula, Australia, and east Asia areas, respectively. However, there is no significant correlation between mobilization and soil moisture (volume water content) or PBLH except in Sahel. The reason for lower correlation between mobilization and soil moisture could be that the correlation is calculated using data from every time step, and variations in soil moisture occur on longer timescales. Additionally, soil moisture is not usually limiting dust mobilization in these regions. As expected, there are moderate to high correlations between dust mobilization and dust concentration in source regions (0.85, 0.86, 0.67, 0.78, 0.95, and 0.66 in west and east Sahara, Sahel, Arabian Peninsula, Australia, and east Asia, respectively). Correlations suggest that the rest of the variability in concentration in dust source regions is not clearly related to transport by wind components $u$ or $v$ or by the vertical mixing inherent in a higher PBLH, but most are due to a combination of these transport parameters.

[24] The correlations between dust mobilizations with surface wind, wind at the reference height (10 m), soil moisture, and boundary layer height for the case using different winds and source parameterizations (GDD) were calculated. The high correlations between mobilization and surface wind and wind at 10 m have been obtained for the GDD case (correlations between mobilization and wind at 10 m are as follows: 0.91, 0.92, 0.85, 0.87, 0.72, and 0.72 in west and east Sahara, Sahel, Arabian Peninsula, Australia, and east Asia). The correlations between mobilization and wind at 10 m are almost the same as the correlation between mobilization and surface wind. The biggest difference between the BASE and GDD cases is the correlation between dust mobilization and dust concentration with boundary layer height in the North African regions, where there is no significant negative correlation between dust mobilization and dust surface concentration with PBLH in the BASE case but there are significant anticorrelations in the GDD case. In the GDD case, significant negative correlations between dust mobilization and boundary layer height in North Africa and Sahel areas (−0.42, −0.27, and −0.42 in west and east Sahara and Sahel source areas, respectively), could be partly caused by significant negative correlation between PBLH and wind at 10 m (−0.25). Additionally, significant negative correlations between dust surface concentration and PBLH are obtained for
GDD case (−0.69,−0.69, and −0.53 in west and east Sahara and Sahel source areas, respectively), while no significant correlations between dust surface concentration and PBLH have been found for BASE case. Compared with the BASE case, dust concentrations in GDD are more highly correlated with transport properties (winds and PBLH) and less with mobilization, which is consistent with the spectral analysis seen above (Table 2 versus Table 4).

4. Summary and Conclusions

[25] Summer transport of dust across the North Atlantic has long been associated with African Easterly Wave activity [e.g., Westphal et al., 1987; Jones et al., 2003], while east Asian dust has been associated with cold fronts during the spring [Qian et al., 1997; Zhou et al., 2002]. This is the first study to explicitly consider the spectra of temporal variability in mobilization and dust concentrations in source regions to understand what processes drive the dust mobilization. Because of a lack of available high-resolution observations in the source areas we use model simulations to estimate variability. Comparisons of dust modeled by this model and available measurements show that the model does a good job of simulating the real atmosphere’s variability [Mahowald et al., 2002; Mahowald et al., 2003; Zender et al., 2003; Jones et al., 2003; Luo et al., 2003], although these measurements are usually made at longer than 1 day timescales.

[26] Using our BASE model for the North African, Arabian Peninsula, Australian, and east Asian dust sources, 35–70% of dust mobilization variance, and ~20–50% of dust concentration variance, can be explained by diurnal variability. This is due to the strong impact of diurnal variability (and therefore the surface radiation budget) on the winds close to the surface which act to entrain the dust. About 10–40% of dust mobilization variance and 30–55% dust concentration variance are associated with synoptic systems in west and east Saharan, Sahel, Arabian Peninsula, Australian, and east Asian source regions. Dust concentrations in the source regions have a large component from the diurnal variability, but the synoptic-scale variabilities are much larger in the concentration than in the mobilization. This is due to the control of dust concentration in the source regions by a combination of local mobilization and transport out of the region. The Australian and east Asian source mobilization and concentration are relatively influenced more by synoptic-scale variability than the North African or Arabian Peninsula sources, consistent with cold fronts and cyclone movements playing a larger role in higher latitudes [e.g., Garratt, 1984; Qian et al., 1997; Zhou et al., 2002].

[27] Sensitivity studies with different meteorology and source parameterization suggest that these results are quite sensitive to changes in meteorology and source parameterization. The biggest differences are that the MATCH/DEAD/NCEP simulations have slightly more diurnal variability in the mobilization and substantially less diurnal variability in the concentration than the MATCH/GOCART/DAO–based simulations (with corresponding changes in the synoptic variability). This is partly due to differences in the mobilization scheme but also due to differences in the variability in the meteorological data sets. The differences seen in the variability analysis in the sensitivity analysis may suggest important differences as future predictions are made with these different model configurations. Other studies have suggested that dust feedbacks on dust [Perlwitz et al., 2001] are associated with changes in the surface heat budget, which is tied more tightly to the diurnal variability than synoptic variability. Thus changing models or source parameterization may substantially change the climate feedback associated with dust. Additionally, high-resolution concentration data in the source regions may help to tell which source parameterization is most accurate.

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References


Mahowald, N., and J. L. Duftresne (2004), Sensitivity of TOMS aerosol index to boundary layer height: Implications for detection of mineral