African dust outbreaks: A satellite perspective of temporal and spatial variability over the tropical Atlantic Ocean

Jingfeng Huang,¹ Chidong Zhang,¹ and Joseph M. Prospero¹

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[1] We describe the temporal evolution and spatial structure of extreme African dust outbreaks and their associated meteorological fields over West Africa and the tropical Atlantic using A–Train data and a global reanalysis product. We used Aqua–Moderate Resolution Imaging Spectroradiometer daily aerosol optical depth (AOD) to identify major dust outbreaks, defined as AOD events one standard deviation above the background along the African coast. Dry air outbreaks were defined using water vapor data of the Aqua Atmospheric Infrared Sounder. Dry air outbreaks do not always coincide with dust outbreaks. Most boreal summer outbreaks reached the West Indies between 10⁰N and 20⁰N, some traveling on to the southeastern United States; winter outbreaks moved to South America between 0⁰ and 10⁰N. Outbreaks travel westward at an average speed of 1000 km d⁻¹, reaching the Caribbean or South America in a week’s time. The advance of a dust front is associated with decreases in water vapor (up to −1.0 g kg⁻¹) and increases in temperature (up to 1.0 K) and, behind the fronts, an anticyclonic circulation. We used Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) data to characterize dust altitude distributions. The vertical distribution of warm dry air is similar to that of dust observed in CALIPSO. The dust layer altitude decreases during transport across the Atlantic and is significantly lower in boreal winter than summer. The study highlights the temporal and spatial variability of African dust outbreaks, which are important to improving our understanding of climate impacts of African dust and Atlantic climate variability in general.


1. Introduction

[2] African dust has attracted increasing attention because of its potential importance on the global energy and hydrological cycles, human health, marine and continental ecosystems, and climate. African dust may affect the environment through its radiative effects [e.g., Carlson and Benjamin, 1980; Miller et al., 2004; Ramanathan et al., 2001], microphysical effects as cloud condensation nuclei [e.g., Johnson, 1982; Wurzler et al., 2000; Lohmann and Feichter, 2005] or ice nuclei [DeMott et al., 2003; Sassen et al., 2003]. Moreover, dust plays an important role in biogeochemical processes. Deposition to land surfaces contributes to soil development [e.g., Muhs et al., 2007] and as a source of nutrients [e.g., Swap et al., 1992]. Deposition to the ocean supplies iron [e.g., Jickells et al., 2005; Mahowald et al., 2005] and phosphorous [Mahowald et al., 2008], elements that are essential to phytoplankton growth and, hence, to the oceanic and global carbon cycle. North Africa is the largest dust source in the world. Each year more than 1000 × 10¹² g of African dust is transported over the Atlantic to reach southeastern United States, Caribbean, South America, and sometimes even eastern Pacific [e.g., Prospero et al., 1981, 1996, 1999a, 1999b; Prospero and Lamb, 2003; Torres et al., 2002; Kaufman et al., 2005b; Perry et al., 1997; Chin et al., 2007]. Large quantities are also transported northward to Mediterranean and Europe [e.g., Mattsson and Nihlen, 1996; Prospero, 1996].

[3] Dust radiative forcing has been simulated in global climate models (GCMs) under different climates [e.g., Claquin et al., 2003; Mahowald et al., 2006]. Dust radiative forcing to the atmosphere can be both positive and negative [e.g., Myhre and Stordal, 2001]. Increased dust emissions could result in decreased surface heating and reduced global evaporation and precipitation [e.g., Mahowald et al., 2006; Foltz and McPhaden, 2008], which in return could lead to increased global dust emissions [e.g., Miller et al., 2004]. But there are substantial differences among the outputs of dust models [Textor et al., 2006]. There are few field observations that might serve to constrain and validate models. Improved large-scale observations are needed to assist the progress in GCM simulations of dust climatic effects.
[4] African dust outbreaks are known to be associated with dry and warm air and anticyclonic circulation [e.g., Carlson, 1979; Carlson and Prospero, 1972; Prospero and Carlson, 1972; Karyampudi and Carlson, 1988; Karyampudi et al., 1999]. It has been suggested that African dust may have substantial effects on cloud and precipitation [Kaufman et al., 2005a; Koren et al., 2005; Rosenfeld et al., 2001] and possibly on the formation/intensification of tropical cyclones [e.g., Karyampudi and Pierce, 2002; Dunion and Velden, 2004]. But it is difficult to quantify the extent to which such effects come directly from dust or from its associated meteorological variables. However, much of our knowledge about African dust outbreak is based on case studies of individual events. Thus, there is no coherent statistically based description of these outbreaks and the associated synoptic meteorological fields.

[5] Over the past few years, a suite of satellite products have become available that present us with an unprecedented opportunity to study in detail African dust, including its spatial distributions and three-dimensional structures, its transport patterns across the Atlantic Ocean, and its accompanying meteorological properties. These products include the Moderate Resolution Imaging Spectroradiometer (MODIS) daily aerosol optical depth (AOD), the Atmospheric Infrared Sounder (AIRS) water vapor and temperature profiles, and Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) aerosol profiles. The purpose of this study is to document satellite-observed statistics of African dust transport across the tropical Atlantic Ocean and to contribute to our understanding of the role of dust in weather and climate by providing observational validation basis for model simulations.

[6] The data used in this study are described in section 2. Climatology of AOD and water vapor over the tropical Atlantic are summarized in section 3. Statistical results on propagation features of African dust outbreaks are discussed in section 4. Section 5 describes the vertical structures of dust layers. Associated meteorological fields across dust fronts are introduced in section 6. Discussions are given in section 7 where the dominant transport patterns and vertical structures of African dust outbreaks are summarized into schematic diagrams.

2. Data

2.1. Aqua-MODIS AOD

[7] To study the annual cycle of African dust, we used the 5 year (2003–2007) data of level-3 version 5 Aqua-MODIS daily aerosol optical depth (AOD) at 550 nm (MYD08) [Kaufman et al., 1997; Kaufman and Tanré, 1998]; over bright surfaces such as deserts, we used the MODIS Deep Blue AOD product [Hsu et al., 2004]. The MODIS Deep Blue algorithm uses sophisticated dust and smoke models that discriminate between fine and coarse mode aerosols [Hsu et al., 2006]. To avoid complications from model assumptions and considering dust and smoke are the only two major aerosol types over the study region, we use a simple fine mode fraction (FMF) criterion to separate dust and smoke. To qualify as “dust” the FMF must be less than 0.7. This threshold value lies between dust FMF, 0.5, and smoke FMF, 0.9 [Kaufman et al., 2005b; Yu et al., 2009]. However, we note this simple FMF criterion may not completely remove the complication from smoke. This should not present a problem during boreal summer because burning activity is located far to the south of the dust sources, but it could be a problem in boreal winter when dust and smoke are often mixed [e.g., Torres et al., 2002; Haywood et al., 2008].

2.2. CALIPSO Lidar Data

[8] The CALIPSO lidar level-2 vertical feature mask (VFM) product contains the column properties of cloud and aerosol layers [e.g., Vaughan et al., 2004] with coverage available from June 2006. Vertical profile data are available from ground to 8.2 km with 30 m vertical resolution and 333 m horizontal resolution. The CALIPSO VFM data discriminate dust from smoke on the basis of their distinctive depolarization ratios and the wavelength dependence of the signal attenuation. In comparison to dust layers, smoke layers have much lower depolarization ratios and a stronger wavelength dependence of the signal attenuation within the layer [Liu et al., 2008a].

2.3. Aqua-AIRS Water Vapor and Temperature

[9] Level-3 version 5 Aqua-AIRS (2003–2007) daily gridded (1° × 1°) atmospheric water vapor and temperature data of 12 layers from 1000 to 100 hPa [Chahine et al., 2006] are available concurrent with the MODIS AOD data. They have been validated over both land and ocean for a broad range of geographic conditions [Divakarla et al., 2006, Tobin et al., 2006]. Direct comparisons between profiles of water vapor from Aqua-AIRS and atmospheric rawinsondes suggest that Aqua-AIRS may underestimate extreme dryness associated with the Saharan air layer (SAL) (P. Minnette, personal communication, 2009).

2.4. NCEP-DOE Reanalysis

[10] The main variables from the NCEP-DOE Reanalysis or NCEP II [Kanamitsu et al., 2002] used in this study are air temperature and the three-dimensional wind fields. They are all daily data with 2.5° × 2.5° resolution and 17 vertical layers from 1000 to 10 hPa.

3. Climatology

[11] Figure 1 shows the climatological means and their respective standard deviations of dust AOD and columnar water vapor over the tropical Atlantic and adjacent regions in boreal summer (June–July–August or JJA) and winter (December–January–February or DJF). In boreal summer, there is a clear dust corridor centered roughly at 17°N from the west coast of North Africa to the Caribbean in both the mean (Figure 1a) and variance (Figure 1b). The structural features of the summer dust outbreaks have been illustrated in the conceptual model of the SAL by Karyampudi et al. [1999].

[12] The dust content along the center of the corridor decreases westward, with its oceanic maximum located immediately off the West African coast. From the coast eastward over land, the highest AOD belt extends to the Bodélé Depression (17°N, 18°E) northeast of Lake Chad. There, the AOD mean and standard deviation are both much higher than over the surrounding areas. The Bodélé Depression was flooded in the Pleistocene and Holocene
when it accumulated vast sedimentary deposits [Prospero et al., 2002], which now serve as the largest and most persistently active dust source on Earth [Washington and Todd, 2005; Koren et al., 2006].

In contrast to dust, the mean distribution of water vapor over the tropical Atlantic Ocean in boreal summer is nearly uniform in longitude and decreases both northward and southward from its zonally elongated maximum (>50 kg m\(^{-2}\)) between 5°N and 10°N, along the axis of the intertropical convergence zone (ITCZ) (Figure 1a). There are signs of dry air intrusion from the northeastern Atlantic southwestward into the tropics and moist air intrusion from the western tropical Atlantic northeastward into the midlatitudes. This pattern is consistent with the anticyclonic flow associated with the North Atlantic Subtropical High. Other than these, the variability of water vapor is zonally uniform and its variability maximum coincides roughly with that of dust (Figure 1b). This suggests possible association between the two parameters, which will be discussed later in this study.

The mean and the variability of dust AOD and water vapor in boreal winter are qualitatively the same as in summer except that the entire spatial patterns shift southward (Figures 1c and 1d). The dust corridor, along with its associated variability maximum in both dust and water vapor, is now centered at 5°N–10°N and it extends to South America instead of the Caribbean. The Bodélé Depression, however, remains as the location of the highest mean and variability in dust in summer. Because of intensive biomass burning activities in the Sahel region in boreal winter, aerosols over the tropical Atlantic will most likely contain more smoke than in boreal summer [Kaufman et al., 2005b]. Although in winter the main fire sources lie to the south of the major dust sources, dust and smoke become mixed over West Africa and in the ITCZ over the Gulf of Guinea [Haywood et al., 2008]. This results in a very complex mixing structure both in the vertical and the horizontal extents but typically with smoke overlying the dust layer [Haywood et al., 2008]. This mixing results in aerosols that span a wide range of physical and optical properties and thereby complicates the interpretation of satellite retrievals [Osborne et al., 2008].

Figure 1 suggests that there are annual latitudinal migrations in both dust and water vapor. Such annual migrations are better illustrated in Figure 2, which plots means and standard deviations of AOD and water vapor zonally averaged over 50°W–15°W and over each calendar month. The largest AOD mean and variability are found in June and July at its northernmost position between 20°N and 30°N. In Figure 2a, there is evidence of increased AOD extending to the high latitudes in March; this transport is seen most clearly in Figure 2b, which shows a pronounced spike in standard deviation at this time. This feature is attributed to the infrequent but spectacular dust outbreaks that occur along the northwest coast of West Africa in the late boreal winter and early spring. Satellite images often show dust fronts moving rapidly away from the coast and into the midlatitudes of the central and eastern North Atlantic [e.g., cover photo, Dunion and Velden, 2004]. It is interesting to note that, while the largest mean and variability of
Figure 2. Annual cycle of (a) monthly mean dust AOD (shading) and water vapor (contour, kg m\(^{-2}\)) and (b) monthly averaged daily standard deviations of dust AOD (shading) and water vapor (contour, kg m\(^{-2}\)). All variables were averaged over 15°W–50°W.

dust roughly coincide in latitude, the largest variability in water vapor lies north of its maximum in the mean (Figure 2b). But the largest water vapor variability in the dust region is in November when the variability of dust is relatively small. This is the first sign that fluctuations in water vapor and dust over the tropical Atlantic Ocean can be independent of each other. We will further demonstrate this later.

4. Propagation

[16] We used the MODIS daily dust AOD and its anomalies (with the mean annual cycle removed) to document dust events propagating across the tropical Atlantic region. We first demonstrate a special case to show the typical behavior of such dust events. Then we show statistics of origins, pathways, and other properties of dust events including their associated water vapor anomalies.

4.1. Example

[17] Figure 3 shows the propagation of two African dust events during 13–18 July 2007. As seen in Figure 3a, one event propagated westward from 20°W on 13 July (first panel) to 60°W on 16 July (fourth panel) at a speed of \(\sim 8°–10°\) longitude per day. Dust concentrations on Barbados (13.17°N, −59.43°W), measured as part of a continuing program [Prospero and Lamb, 2003], sharply increased from 3.5 \(\mu g\) m\(^{-3}\) on 15 July to 41 \(\mu g\) m\(^{-3}\) on 16 July, consistent with the continuing movement of the dust plume. The second event emerged from the coast of West Africa on 15 July (third panel); on 18 July (sixth panel) it shows a large tail extending from 50°W to 20°W. Consistent with the satellite depiction, Barbados dust concentrations rose to 16.5 \(\mu g\) m\(^{-3}\) on 19 July and 51 \(\mu g\) m\(^{-3}\) on 20 July, and they remained high through 27 July, roughly in the range of 36–66 \(\mu g\) m\(^{-3}\). Dry air associated with both dust events can be seen as large negative water vapor anomalies in Figure 3b. In both cases, the pathways of dust and dry air events overlapped. The aerosol vertical structure along the CALIPSO track is shown in Figure 3c; the CALIPSO lidar track is shown in Figure 3a. Initially, on 14 July, the main part of the dust layer (in red) concentrated between the altitudes of 2–5 km from the surface. In the daily sequence in Figure 3c, we note a clear reduction in the thickness and cross-sectional area of the dust cloud. We would expect that some of the reduction is due to the removal of aerosol during transport although this cannot be quantified because the CALIPSO track does not necessarily cross the same air mass on succeeding days. Information from Figures 3a and 3c indicates that initially the dust layer was approximately 15° wide in longitude and 5° in latitude, and 3 km thick. There is also a systematic decrease in the altitude of the top of the SAL. The vertical structure of African dust layers is discussed more fully in section 5.

4.2. Outbreak Identification

[18] To help objectively identify and characterize the temporal and spatial evolution of dust outbreaks, we selected three Saharan air layer (SAL) boxes oriented from north to south off the West African coast. They are North-SAL (N-SAL hereafter) (20°N–30°N, 15°W–25°W), Middle-SAL (M-SAL hereafter) (10°N–20°N, 15°W–25°W) and South-SAL (S-SAL hereafter) (0°–10°N, 15°W–25°W) (see Figure 4a). We identified the presence of a dust outbreak on a daily basis using a simple criterion: The daily AOD averaged over a SAL box exceeds the long-term “background” AOD of that box by one standard deviation. At each box, the background AOD was defined as the AOD of the PDF of AOD within the box. We chose this definition for the background because the PDF of AOD is highly non-Gaussian with a long tail of high AOD, as seen in Figure 4b as the PDF of AOD within the three reference boxes, respectively. By this definition, intense African dust outbreaks are relatively infrequent extreme events. However, the “background” condition is by no means dust free. In Figure 4a, to demonstrate its spatial variability, the background AOD is calculated by confining the AOD data distribution to each grid. The background AOD is relatively high in a latitude belt from equator to 20°N along the tropical North Atlantic, corresponding to the dust corridor seen from the mean (Figure 1a). The highest background
AOD (>0.5) is over the Bodélé Depression in northern Chad. In Figure 4b, the background AOD values corresponding to the peaks of the PDFs are 0.17, 0.29, and 0.24 for the N-SAL, M-SAL, and S-SAL boxes, respectively. These background AODs are lower than the mean AODs, which are 0.29, 0.48, and 0.37 for the N-SAL, M-SAL, and S-SAL boxes, respectively. The AOD thresholds for identifying dust outbreaks (i.e., the background AOD plus one standard deviation) are 0.35 for N-SAL, 0.57 for M-SAL, and 0.46 for S-SAL. For all seasons in the 5 year period, 69, 63, and 59 days per year were classified as dust outbreak days in the N-SAL, M-SAL, and S-SAL boxes, respectively, a total of 191 per year. However, it should be noted that on many occasions a large dust event could trigger an “outbreak” classification in more than one box. On the basis of a simple average, 64 dust outbreak days per year is equivalent to slightly above one dust outbreak day per week. But a single dust outbreak can result in AOD values that exceed the threshold criterion for more than 1 day. Thus, the number of discrete dust outbreak events is considerably less than 64 per year. Using the high AOD tails in the PDF to measure the strength of dust outbreaks, one sees that dust outbreaks that go through the M-SAL box appear to be statistically stronger (i.e., larger PDFs of AOD that are above the background AOD) than those passing through the other two boxes (Figure 4b); in contrast, the N-SAL box experiences the weakest dust outbreaks.

[20] Background water vapor, defined as the water vapor value corresponding to the peak PDF, is calculated at each grid and shown in Figure 4c to demonstrate its spatial variability. The highest background water vapor is associated with the Atlantic marine ITCZ, and the lowest is over the Sahara desert. In contrast to AOD, the PDFs of water vapor in the three boxes are very different from each other (Figure 4d): Air in the N-SAL box is almost always much drier than in the S-SAL box. These differences are consis-
tent with the mean distribution of water vapor (Figure 1a). Note also that the water vapor in the M-SAL box undergoes much larger variability (wider PDF) than that in both N- and S-SAL boxes. 

Similarly to dust outbreaks, we defined a dry-air outbreak as an abrupt decrease of water vapor that is lower than its background by one standard deviation in each of the SAL boxes. In the 5 year period, 6, 22, and 46 days per year were classified as dry-air outbreak days for the N-, M-, and S-SAL boxes, respectively. They are much less frequent than dust outbreaks, especially in the N-SAL box. There are more dry-air outbreak days through S- than N-SAL boxes because the background water vapor content in N-SAL is lower than that in S-SAL (Figure 4d) due to its distance from the ITCZ. Air from Africa, which may bear similar water vapor content across all three boxes, would be classified as much drier than the background in the S-SAL box than in N-SAL and M-SAL boxes. Over the 5 year period a total of 87 events were identified as being both dust outbreak day and dry-air outbreak day by our definitions, which is about 9% of dust outbreaks and 23% of dry-air outbreak days. Meanwhile, 54% of dust outbreak days are associated with air drier than the background, 46% of dry-air outbreak days are associated with AOD higher than the background. These statistics imply that extremely dry dust outbreaks and extremely dusty dry-air outbreaks do not always coincide [Zhang and Pennington, 2004], and that the water-vapor content of dust outbreaks and the dust concentration of dry-air outbreaks both vary substantially among events. Different definitions of dust and dry-air outbreaks (e.g., based on means instead of maximum PDF) may lead to quantitatively different statistics.

4.3. Pathways and Composites

To better understand the spatial pattern of all dust events propagating across the tropical Atlantic, we calculated their pathways both forward and backward in time starting from the SAL reference boxes. The pathway analysis was conducted using a “moving window” approach. The procedure is demonstrated in Figure 5 that shows a series of MODIS AOD maps spanning a 16 day period (5–20 July 2007). On 11 July (day D, Figure 5b), the latitude and longitude of the AOD maximum in the M-SAL box were designated as the core of the dust outbreak and it becomes the reference point of the analysis. From this point, we draw a new box extending across ±5° latitude. Considering the traveling speed of a typical dust event (see Figure 3), the west boundary of the box is placed 15° west of that reference point for the forward window for day D+1 to avoid missing the current track of a dust event or overlapping with a previous dust event. The location of the maximum AOD in this box becomes the new reference point for the construction of a new box on the following day D+2. This process is repeated in a similar way for forward
tracking on day D+2, D+3, etc. Following these steps, we identified the locations of maximum AOD on succeeding days (for example, D+4 and D+9 in Figures 5c and 5d) until the dust event could no longer be identified using our AOD criterion. Using the same procedure used for constructing the westward boxes and using the same initial reference point, we draw a box over Africa with the east boundary at 15° to the east for D-1 and we construct a series of eastward moving boxes on succeeding days. The maximum AOD points were then linked to form a pathway. The moving windows on D-6, D, D+4, and D+9 are shown in Figures 5a–5d, respectively. The pathway for the entire life history of the dust outbreak is shown in all four parts.

[23] By using the moving window approach, we objectively calculated pathways solely on the basis of the daily AOD observations. The merit of this approach is particularly valuable for statistical studies based on long-term satellite observations, especially when satellite observations are providing better and better spatial coverage over the whole region without missing data strips, for example, a better data coverage by jointly using MODIS and MISR data sets [Kalashnikova and Kuhn, 2008]. This procedure yielded relatively consistent and unambiguous results for the forward tracking of dust over the ocean. Ideally, we should be able to use the same procedure to backtrack dust outbreaks over land and, thus, to identify the source regions. However, over land the results are not so clear because of the possible effects of emissions from the many sources in West Africa. Thus, the maximum AOD identified in a moving window on (for example) day D-1 might indeed represent the center of the dust cloud produced by a source or sources to the east (e.g., the Bodélé Depression) or it might be due in part or in whole to a new dust outbreak that occurs within the day D-1 box. As a result, the moving window approach may lead to uncertainties in identifying major dust sources.

[24] Figure 6 plots all identified pathways of dust outbreaks that pass through the three SAL boxes. The pathways cover a broad region of north tropical Atlantic from the equator to 30°N. The Sahel region and the Bodélé Depression in particular appear to be the most prominent sources of most dust outbreaks. The Bodélé Depression stands out in this analysis partly because of the limitations of the moving window approach and partly because it is one of the strongest and most persistently active African dust sources. To indicate a general propagating pattern of dust outbreaks, we defined a central path for each 5° longitude interval from 75°W to 20°E. This central path is at the latitude of the most concentrated pathways. The central path is flanked by two thinner dashed lines at both sides, which define a zone embracing 75% of the total pathways.

[25] Most N-SAL events travel to the West Indies and North America (Figure 6a), which is consistent with the
mean AOD distribution in boreal summer (Figure 1a). The central path hooks to the north over the Caribbean and suggests a tendency to transport to the Gulf of Mexico and the southern and eastern United States. In contrast, most S-SAL events travel to the northeast coast of South America (Figure 6c), consistent with the mean AOD distribution in boreal winter (Figure 1c). A few N-SAL or M-SAL dust outbreaks moved northward to the U.S. east coast or even Canada.

[26] It is notable that central path of N-SAL events lies to the south of the reference box and the central path of S-SAL events lies to the north of the reference box. This indicates that while the three reference boxes may help identify some interesting features of dust events they do not necessarily always distinguish them from each other; that is, according to our definition of an event, a single large dust event could simultaneously trigger individual “events” within more than one box. It also suggests that the real tendency of the dust outbreaks is through the M-SAL box and that the other two boxes catch the spillover from these events. Therefore, because of the synoptic scale of dust events, the number and pathways of outbreaks shown in Figure 6 must be viewed in reference to their predefined reference box.

[27] Figures 7a–7e show statistics of these dust outbreaks identified in the three reference boxes. The traveling time of a dust event over the Atlantic Ocean is defined as the time between the day when it is observed in a reference box and the day when its AOD no longer satisfies the dust outbreak criterion (i.e., the last day its AOD exceeds the background AOD by one standard deviation). Most dust events retain their identities for 7 days, some last as long as 12 days (Figure 7a). Comparing the life spans of dust events passing through the three reference boxes and over the Atlantic, we see no major differences, except that there seems to be more short-lived events coming out of the S-SAL box. The longitude at the end of the pathways of a dust outbreak is defined as its western termination. Most dust events propagate to 70°W–80°W; a few can travel to as far as 100°W and beyond (Figure 7b). The statistics on dust strength in terms of dust optical depth are shown as the PDF of the box average AOD deviation from the background AOD in the three SAL boxes (Figure 7c). There is evidence that M-SAL events are on average stronger than those from the other two reference boxes, a conclusion similar to that from Figure 4b.

[28] The seasonal cycle of dust outbreaks in number per month is shown in Figure 7d. As previously discussed, because each outbreak is defined by referring to the background AOD in its specific reference box, and also because the synoptic scale of dust events, a dust outbreak in one reference box cannot be well separated from a dust outbreak

Figure 6. Pathways of all dust outbreaks passing through reference box of (a) N-SAL, (b) M-SAL, and (c) S-SAL, which are shown as white-lined boxes. Thick gray lines define central paths, the latitudes of highest pathway concentrations. Dashed lines flanking the central paths define a zone embracing 75% of the total pathways. The eastern starting points of gray lines are approximately located around the Bodélé Depression (17°N, 18°E).
in another reference box. So the seasonal variability of dust outbreaks should be discussed only in the context of a specific reference box. The peak season for N-SAL events is boreal summer. S-SAL events appear to undergo a semi-annual cycle with peaks in both boreal spring and autumn. M-SAL events do not show an obvious seasonal cycle.

Figure 7. (a) Life span of dust events over the tropical Atlantic, (b) west end longitudes of dust events, (c) PDF of AOD deviation above background AOD for all dust events, (d) annual cycles of dust event days per month, and (e) west end latitude distributions of all events in three reference boxes.
The average of \( \sim 6 \) dust outbreak days in a month (Figure 7d) is consistent with the AEW periodicity of 3–5 days [Burpee, 1972] considering that our study focuses on very strong dust events.

The latitudinal destinations of the dust outbreaks are further summarized in Figure 7e, which compares the west end latitudes of dust pathways from the three boxes. The \( 10^\circ \) N–20\(^\circ\)N latitude belt in the West Indies is the top destination for all dust outbreaks. But about a half of the N-SAL outbreaks reach latitudes higher than 20\(^\circ\)N, such as the U.S. east coast as observed by Prospero [1999a, 1999b]. There are also considerable numbers (\( \sim 10\% \)) of S-SAL outbreaks that terminate below 5\(^\circ\)N and into the southern hemisphere in South America. This transport scenario is consistent with dust measurements made in French Guiana [Prospero et al., 1981] and in the Amazon Basin [Swap et al., 1996]. The association of strong dust transport to the Amazon is supported by modeling studies [Koren et al., 2006], which identify the Bodélé Depression as the major source of dust over the Amazon.

The seasonal character of the pathways is shown in Figure 8. Their central path and boundaries of 75\% of pathways are also plotted. Transport to South America is strongest in boreal spring and somewhat less in winter (Figures 8a and 8d) as supported by measurements in French Guiana [Prospero et al., 1981]. In contrast, they reach the West Indies more frequently in boreal summer and autumn and the U.S. east coast in boreal summer (Figures 8b and 8c) [Prospero, 1999a, 1999b; Prospero and Lamb, 2003]. Boreal autumn shows the most diverse destinations with pathways ranging from the U.S. east coast to South America.

To better illustrate their spatial patterns, we show in Figure 9 the calculated composites for AOD anomalies of dust outbreaks (Figure 9a) and water vapor anomalies of dry air outbreaks (Figure 9b) for the four seasons using the 75\% of the events that are adjacent to the seasonal central paths depicted in Figure 8. In addition to composites on day T when a dust (or dry air) outbreak is identified in the reference box, the composites on D-3, D, D+2, and D+4 days are also shown to provide additional information on the propagation of dust event across the tropical Atlantic in terms of changes in AOD and water vapor anomalies at different stages of the dust event lifetime and the associated variability between different seasons. The spatial patterns of AOD and water vapor composites in boreal spring and winter are substantially different from those of summer and autumn (Figures 9a and 9b) [Husar et al., 1997]. Autumn stands out as the least dusty period; however, dry air outbreaks are still quite evident (Figure 9b), more so than in summer. In boreal winter-spring, there is a persistent dusty and dry period in the Sahel region, which can last for a week.
and extend for more than 20° in longitude. Biomass burning, which also occurs in these seasons, can complicate the issue, as discussed in the following paragraph. In contrast, in summer-autumn both AOD and water vapor anomalies are more isolated in longitude and short lived in time at a given location.

This seasonal contrast in dust outbreaks can be related to the seasonal position and strength of the Atlantic marine ITCZ, the West African Monsoon, and AEW. In the boreal spring and winter, the southern extent of the dust distribution is clearly bounded by the ITCZ; its most intense precipitation, however, is located further to the south and away from the dust sources and pathways. In winter and spring, AEWs are generally weak [Gu and Zhang, 2002], allowing dusty and dry conditions to persist over the ocean. In boreal summer and autumn, while the spatially isolated dust events might be related to AEW [Jones et al., 2004], stronger rainfall associated with AEW may modulate the frequency and intensity of the dust and dry events, and shortening their lifetime as defined “events.” Another possible reason is that biomass burning peaks in the Sahel-Soudan region in winter. Our use of coarse mode AOD may not sufficiently remove these smoke events from our record, particularly over land where the confidence level of FMF values deteriorates (R. Levy, personal communication, 2009). But the CALIPSO results (to be discussed in a later section) indicated that smoke layers are suspended in the atmosphere at much higher levels than the dust layers; thus, they are likely to have a longer lifetime [e.g., Haywood et al., 2008; Heese and Wiegner, 2008].

Seasonal characteristics of all dust outbreaks are summarized in Figure 10. The PDFs of the west end latitude

Figure 9. (a) Composites of AOD anomalies for outbreaks within the zone of 75% of pathways (see the text and Figure 8) in four seasons and (b) composites of water vapor anomalies (kg m⁻²) for the same outbreaks as in Figure 9a in four seasons.
of pathways for the four boreal seasons are plotted in Figure 10a. The top destination latitude for boreal winter dust is in the range of 0°N–10°N while it is 10°N–20°N in the other three seasons, as seen in Figure 7e. In general, dust is transported to more northern destinations in boreal summer and autumn than boreal spring and winter. The strength of the dust outbreaks in terms of AOD deviation from its background or AOD anomaly from its seasonal mean is significantly lower in boreal autumn than in the other three seasons (Figures 10b and 10c). This is especially notable in the absence of any appreciable autumn events with AOD deviation greater than about 0.4; in contrast, the other seasons yield events with AOD as high as two standard deviations above background (Figure 10b).

As discussed previously, the Bodélé Depression is the most outstanding African dust source [e.g., Prospero et al., 2002; Washington and Todd, 2005]; it exhibits the largest mean and variance (Figure 1) and background AOD (Figure 4a). To illustrate the path and seasonality of dust events from the Bodélé Depression, we conducted 10 day forward trajectory analysis using the NOAA Hybrid Single Particle Lagrangian Integrated Trajectory (HYSPLIT) 4 model (R. R. Draxler and G. D. Rolph, HYSPLIT model access via READY Web site (http://www.arl.noaa.gov/ready/hysplit4.html), NOAA Air Resources Laboratory, Silver Spring, Maryland, 2003) and the NCEP reanalysis meteorological data set. Instead of tracking dust signals on a horizontal plane as in the preceding discussion, this trajectory analysis tracks the three-dimensional movement of air mass. The starting point was set on the surface at 17°N and 18°E, the center of the depression, and the location of the AOD maximum in Figure 1a. Only trajectories in four seasons of 2007 are shown in Figure 11 for illustration purposes.

Consistent with our previous results on the pathways and seasonality (see Figure 8), the trajectory analysis shows that most air masses from the Bodélé Depression ended in the West Indies and South America in 2007. They tend to move southward to the Atlantic in boreal spring and winter and northward in boreal summer and autumn. In some cases, air mass from the Bodélé Depression reached as far as the U.S. east coast, the eastern Mediterranean, and southern Europe. It is notable that in spring the Bodélé Depression conceivably is a strong source of dust transported in a northeasterly direction toward the eastern Mediterranean.

Altitudes of the Bodélé Depression air masses during their transport are shown in Figure 11b. Along the trajectories, there is a clear tendency for rising motions during the early transport phase. There are interesting seasonal contrasts. Transport altitudes in boreal autumn and winter remain tightly grouped below the 700 hPa level, lower than those in boreal spring and summer. This is especially notable in the winter where many trajectories do not lift until about 200 h or later. In contrast, in summer most trajectories start to lift within the first 24–48 h presumably by upward motion associated with the strongly heated land surface and convective activity. In spring, strong lifting does not become apparent until about 100 h. Although the majority of the trajectories go above 700 hPa (Figure 11), there are two types of air mass motions in spring and summer. One type
rises very quickly from the ground to the mid-troposphere, above about 600–700 hPa. The other type, relatively less frequent, tends to remain below these levels relatively longer before they reach 700 hPa. This altitude range is consistent with the observed top of the SAL as it emerges from the coast of West Africa and starts its travel across the Atlantic.

5. Vertical Structure

We use CALIPSO data to investigate the vertical structure of African dust outbreaks. To discriminate between dust and smoke, we use the CALIPSO vertical feature mask data. Because of the high noise level of CALIPSO at day time [Liu et al., 2008a; Liu et al., 2008b], we only use nighttime data for quality assurance purposes. We separately analyzed the VFM data of boreal summer and winter. We first compiled the dust altitude information for three summers (2006, 2007, and 2008) and two winters (2006, 2007) and regidded the VFM data of each track onto a 1° × 1° grid over the region. During the regridding, we statistically calculated the maximum and minimum dust altitudes of each profile, which allows us to identify the top and bottom altitudes of dust layers sensed by CALIPSO. Then at each 1° × 1° grid we separately calculated for boreal summer and winter the climatological top and bottom altitudes of dust layers by averaging all the tracks that overpass the grid.

[39] In boreal summer, the top altitude of the dust layer over the African continent and eastern tropical Atlantic extends to about 5 km (Figures 12a and 12d). This is consistent with previous studies which show that the intense sensible heating in the Sahara produces a surface heat low and low-level convergence; these processes produce strong lifting of dust from the surface [e.g., Karyampudi et al., 1999] and leads to dry convection which extends to 5–6 km [Carlson and Prospero, 1972]. As the dusty air mass moves off the coast, the top begins to sink steadily. The
The decrease in the SAL top altitude during the transit of the Atlantic is most likely related to tropospheric clear-sky subsidence (\(~0.5\) km d\(^{-1}\)) due to longwave radiative cooling (\(~1\)°C d\(^{-1}\)), which would lead to an altitude decrease of about 2.5–3 km over a period of 5–7 days. The exact rate of the subsidence depends on the water vapor content in the atmosphere and therefore is highly variable. With dust, the net cooling rate might be reduced because of shortwave absorption, which would result in a reduced subsidence rate (Figures 12d and 12h). The mean altitude of the base of the dust layers is at about 1–2 km over Africa and along the coast; over the open Atlantic it remains mostly at about 0.5 km (Figures 12c and 12d).

In boreal winter, the dust corridor is significantly shifted southward to 5°S–15°N in comparison to 5°N–25°N in boreal summer (Figure 12e versus Figure 12a). Over the African continent the tops of the dust layers are in the range of about 3–4 km, about 1–2 km lower than those in boreal summer; over the Atlantic, the top is mostly between 2 and 3 km, substantially lower than during the summer (Figures 12e and 12h versus Figures 12a and 12d). The tops of winter dust layers decrease westward across the Atlantic but the rate of decrease is much less than that during summer (Figure 12h versus Figure 12d). The bottom altitudes also tend to be somewhat lower in winter, dipping to 200 m over the central ocean (Figures 12g and 12h versus Figures 12c and 12d). The entire layer begins to rise at about 40°W. This lifting is probably associated with convection over South America. The greatest lifting occurs between 60°W and 80°W, over the Amazon Basin. The HYSPLIT trajectory analysis on air mass (Figure 11) and the CALIPSO results on dust (Figure 12) both show descending altitudes in the westward transport over the tropical Atlantic. Both also show much lower altitude transport paths of air mass and dust in boreal winter relative to boreal summer. The results on the top altitudes of dust layers over West Africa are consistent with another independent CALIPSO data analysis [Liu et al., 2008b], which shows the variability of dust height over land in more detail.

The seasonal variability of the altitude of dust layers is intriguing. The relatively lower altitude in boreal winter than in boreal summer might be related to the following two factors. The first factor could be the closeness to convective activities. In boreal summer, over the tropical Atlantic, close to the ambient Atlantic Marine ITCZ (AMI) to its south, summer dust layers can be uplifted higher by stronger updrafts. In contrast in boreal winter, the AMI retreats to central Atlantic with relatively weaker strength. The departure of dust corridor from the AMI is relatively larger. Thus, the convective force in the tropical Atlantic climate system that uplifts dust layers could be significantly weakened. Similarly, over the West Africa, the continental monsoon activity is located northernmost close to dust sources in boreal summer. Stronger convective activity enhances gust occurrence in the ambient desert area which is directly responsible for uplifting dust into the air. In boreal
winter, however, the Monsoon is off to the Gulf of Guinea with weaker convective strength, away from the African dust sources. Consequently the gust force becomes weaker for dust uplifting; thus the altitude of winter dust layers starts at lower altitude than their summer counterparts in general. The second factor is the difference of the subsidence rates in boreal summer and winter. In boreal winter in comparison to summer, air is drier over the dust transport route. In addition, smoke particles from winter biomass burning activities are more actively hygroscopic than dust particles, and they are more shortwave absorbing than dust particles. Drier environment and stronger shortwave absorption in the air result in relatively lower subsidence rate in boreal winter than in summer.

[43] The CALIPSO data report the bottom altitude of dust layers over the Atlantic as low as 0.5 km in summer and 0.2 km in winter. These altitudes would place dust within the nominal marine boundary layer. This raises several issues. First, measurements show that substantial quantities of dust are clearly present below the CALIPSO bottom altitude. The long dust record on Miami [Prospero, 1999b] and Barbados [Prospero and Lamb, 2003] are based on measurements made at surface-level sites. During the extensive dust measurements made aboard aircraft over a 3 month summer period during BOMEX [Prospero and Carlson, 1972], the mean dust concentration in the marine boundary layer was about 1/3 of that measured within the Saharan air layer. In the PRIDE study, Reid et al. [2002] experienced dust episodes where there was substantial dust in the boundary layer; on one occasion, dust was only observed in the boundary layer with none above at the altitudes typical of the SAL. Second, this inconsistency might be due to the CALIPSO VFM data uncertainty, perhaps a difficulty in resolving sea salt returns although we have no reason to believe that this is so. A relatively short-term aerosol study performed in the late 1970s at a surface site on coastal French Guiana [Prospero et al., 1981] yielded dust concentrations during boreal winter and spring that were as high as those measured on Barbados in summer along with moderately high concentrations of sea salt. It is not clear how CALIPSO might respond to a mixture of these two very different aerosol types. Finally, during intense dust episodes, the lidar signal can be attenuated to such a degree that it might not be possible to resolve the depth of the dust layer. In order to address these issues we would require a systematic investigation of the vertical profiles of various aerosol types (dust, smoke, sea salt, etc.) across the tropical Atlantic in conjunction with the CALIPSO measurements.

[44] As previously noted, the CALIPSO retrievals should enable us to discriminate between dust and smoke. This is a critical issue during boreal winter when both dust and smoke emissions are high in the low latitudes in West Africa. The emissions from these two types of sources are interlayered over this region in a complex way as observed during the winter AMMA field program in 2006 [Haywood et al., 2008]. In general, dust is advected southward (e.g., from the Sahel and Bodélé) toward the Gulf of Guinea; in contrast, smoke from biomass burning in the Sudan region is carried to the north by the monsoon circulation and lifted over the dusty air masses. The relatively low altitude distribution of dust that we show in Figure 12h is consistent with the AMMA findings [Haywood et al., 2008; Heese and Wiegner, 2008] and gives credence to the ability of CALIPSO to distinguish between dust and smoke in this way.

6. Dust Front

[45] We typically observe a sharp gradient in dust concentrations at the leading edge of a westward propagating dust event. For the purpose of this study we define a dust front by the location of the largest longitudinal AOD gradient within the region 60°W–20°W, 0°N–30°N. The daily AOD gradient can be as sharp as 0.5 per longitudinal degree. However, the dust “front” is not strictly analogous to a true front in the meteorological sense because we often observe substantial quantities of dust to the west of our defined dust fronts in the daily imagery. Nonetheless, we use the term “front” to signify the sharp transition in dust properties and we use the so-defined front as a reference point from which we can normalize the cross-sectional structure of dust outbreaks.

[46] To this end we examined the changes of the vertical profiles of the following meteorological factors across the dust fronts: Water vapor mixing ratio, atmospheric temperature, and wind. They were calculated for the longitudinal cross section that extended 10° to both east and west of the front, that is, across a box that is 20° wide centered at the dust front with maximum longitudinal AOD gradient. We previously noted that dust outbreaks normally move at roughly about 8°–10° longitude a day (Figure 3). Thus, the transitions that we observe across the box will be those that we might expect to observe over a 2 day period as a dust event advances toward our observing site in the box and passes over us. Most dust fronts were clearly identified between the equator and 20°N.

[47] Because dust outbreaks are normally drier than their ambient environments, their water vapor mixing ratio decreases dramatically from the west to east across the dust fronts, with magnitudes of reduction up to 1.0 g kg⁻¹ over less than 5° longitude (Figure 13a). Note that in Figure 13 the dust front is located at zero relative longitude. Thus in Figure 13a, the dry air “front” is about 3° west of the dust front. The dry air is located mostly below the 700 hPa level although there is some indication of dry air associated with the dust outbreak up to the 400 hPa level. The warm feature of dust outbreaks is shown by the significant temperature increase as high as 1.0°C in the dust layers. Warmest air is located immediately behind the dust front but without a sharp frontal transition. The warm layer is below the 700 hPa level (Figure 13b), consistent with the temperature inversion of the nominal SAL [e.g., Karyampudi and Carlson, 1988]. The SAL top (defined by the sharpest vertical gradient in water vapor and temperature) remains at a nearly constant level from 3° ahead of the dust front to the eastern edge of the composite box, 10° east of the front. Figures 12d and 12h show that the altitude of the dust layer declines as the dust outbreaks travel from east to west across the tropical Atlantic. The fact that this decrease in altitude is not observed in Figure 13 is related to the compositing technique which includes events close to the African coast where altitudes are higher (and which might explain the relatively dry air extending to 400 hPa in Figure 13a) and, at the other extreme, events in the western Atlantic where altitudes are lower. Thus, the averaging technique will
obscure the altitude trends; therefore, the 700 hPa level should be taken as an averaged altitude of these layers.

As expected, the easterly wind in the lower troposphere is strong at the dust front (Figure 13c), which is responsible for carrying dust over the tropical Atlantic. Because the leading edge of the dust front closely follows the 700 hPa synoptic-scale wave trough, the associated positive vorticity advection leads to a rising motion ahead of the front; meanwhile the negative vorticity advection associated with the 700 hPa ridge promotes sinking motions behind the front [e.g., Karyampudi and Carlson, 1988; Karyampudi et al., 1999]. The westerly wind in the upper troposphere around 150 hPa appeared also strong across dust fronts. These form a zonal overturning circulation extending about 10° in longitude and up to the tropopause in depth. Associated with the positive vorticity, the meridional wind is generally southerly south of the front and across it at 400 hPa (Figure 13d). There is no obvious meridional overturning circulation at the longitude of the front.

A notable large-scale circulation feature of dust events is the anticyclonic flow which is reflected as the enhanced northward and westward wind fields at, and to the west, of the dust front but opposite behind the front. When the SAL passes over the African coast, the pressure gradient force at midlevel flow weakens. This causes an excess of Coriolis force, resulting in the northward movement (anticyclonic rotation) of dust particle trajectories [Karyampudi and Carlson, 1988]. It is also possible that the diabatic heating in the SAL results in negative absolute vorticity in the form of the anticyclonic rotation of the SAL [Karyampudi and Carlson, 1988].

7. Summary

We used the five year (2003–2007) data of Aqua-MODIS AOD, CALIPSO VFM, Aqua-AIRS water vapor profiles and reanalysis data to characterize the properties of African dust outbreaks that propagate across the tropical Atlantic. We summarize the characteristics of the dust outbreaks in terms of their occurrence frequencies, their transport pathways, meteorological fields across their fronts, and their transport altitudes. To the best of our knowledge, this is the first comprehensive compilation of observational statistics of African dust events over the tropical Atlantic.

Dust and dry-air outbreaks were defined as extreme events with AOD and water-vapor mixing ratio deviating from their respective background (maximum PDF) by one standard deviation in one of the three reference boxes off the West African coast. By these definitions, extreme African dust outbreak day occurred almost once a week on average in these 5 years (roughly 65 events per year). Most dust outbreaks did not qualify as a dry air outbreak as defined here; only 9% of dust outbreaks qualified as dry-air outbreak days and only 23% of dry-air outbreaks as dust outbreaks). However, about 54% of identified dust outbreak days were associated with air drier than the background and about 46% of dry air outbreaks were associated with dust concentrations greater than the dust background level. This
indicates that there was substantial variability in dust and water-vapor contents among the identified events. It also suggests that dust and dry air outbreaks can be closely associated with each other sometimes but also that they can occur independently of each other.

Although there are many important dust sources in North Africa, the Bodélé Depression is clearly dominant in boreal winter although it is active throughout much of the year [Koren et al., 2006; Schepanski et al., 2007]. Therefore, in this paper, we presented the Bodélé Depression as the most representative dust source and we focused on transport from this source. From the Bodélé Depression, African dust may take different propagating pathways in different seasons. In boreal spring, summer, and autumn, the majority of dust outbreaks reach the West Indies. Some occasionally move northward toward the east coast of the United States in summer and autumn. In boreal spring and winter, dust outbreaks shift southward and most reach the northeast coast of South America. In spring, we find a singular pathway northward from the Bodélé Depression to Europe.

Our analysis shows that dust outbreaks travel westward at an approximate speed of 8°–10° longitude (~1000 km) a day. This speed is roughly consistent with that obtained in a previous study based on a comparison between Terra and Aqua-MODIS imagery [Kaufman et al.,

**Figure 14.** Schematic longitudinal vertical cross section along the Atlantic dust corridor with general characteristics of dust front and its associated changes in meteorological fields. Wind directions are illustrated as the solid arrows. The transport speed of the dust event is marked in the striped left arrow. The Bodélé Depression is used as a representative of African dust sources. The thick red line marks the ground level and denotes the centerline of a 20° latitude (10° to the north and 10° to the south) meridonal range used for the zonal vertical profile calculation.
which yielded 11 m/s, roughly 9° per day. From the coast of West Africa it takes about five days for outbreaks to reach Barbados and a week or so the Caribbean and Gulf of Mexico and the southeastern United States. In our analysis about half of the dust outbreaks retain their identities over the Atlantic after 7 days. Boreal summer is the peak season of African dust outbreaks in terms of their strengths although winter and spring transport is also quite strong as evidenced by dust measurements made along the coast of South America and in the Amazon.

The structure, especially the vertical extent, of African dust outbreaks is summarized in Figure 14. In boreal summer, the top altitude of aerosol layer observed by CALIPSO can be higher than 4 km over land, but it gradually decreases to 3–5 km over the northeastern Atlantic and to 2–3 km over the northwestern Atlantic (Figure 14a). In boreal winter, these levels are about 1 km lower and the gradual declining trend becomes milder (Figure 14b). These altitudes are consistent with the vertical structures of the dry and warm layers identified from the water vapor and temperature profiles, which are mainly in the lower troposphere (below the 700 hPa level). This implies that dust layers are one of the factors modulating the moisture content and static stability in the climate system over the north tropical Atlantic.

Our analysis was constrained by a number of factors. The use of satellite data only made the results vulnerable to retrieval errors. African dust is frequently transported above thick clouds and goes undetected by MODIS AOD products because of cloud screening [Kalashnikova and Kahn, 2008; R. Levy, personal communication, 2009]. Because of the different sample sizes of quality-passing profiles, the quality of AIRS level-3 water vapor profiles may not be the same at different levels (E. Olsen, personal communication, 2009). In the CALIPSO VFM product, misclassification may appear by labeling heavy dust aerosols as cloud possibly due to the intrinsic scattering properties of dust layers that are similar to those of cloud in certain conditions (CALIPSO VFM Version 2.01 data quality statement; Z. Liu, personal communication, 2009). The stringent definition of dust and dry-air outbreaks (peak PDF plus one standard deviation in AOD and water-vapor mixing ratio) selected only the extremely strong events; similarly this criterion can abruptly terminate a dust event when in reality dust concentrations remain high in the air mass. Many dust events continue to travel beyond their last day of identification by our definition. For example, high concentrations of dust are frequently measured in Miami every summer [Prospero, 1999a, 1999b] although our analysis scheme detects very few of these events. Similarly, substantial quantities of African dust are measured over the eastern United States as far north as New England [Perry et al., 1997] in dust events that would not be detected in our analysis. Our simple “moving window” method of tracking dust outbreaks might have left significant dust events undetected. Our focus on the Bodélé Depression as the dominant dust source is tempered by the fact that there are other large dust sources that contribute considerably to the distribution and seasonal variability of African dust [Prospero et al., 2002; Schepanski et al., 2007]. These limitations call for an improved approach in future work.

Despite these limitations our study provides an observational basis that can be used to test numerical simulations of African dust generation and transport. For a climate model to be useful in studies of African dust outbreaks and their impacts on weather and climate, it should produce not only realistic mean distributions of dust but also its transient features such as those identified and characterized in this study. The vertical structure of the dust layer, for example, is crucial to dust radiative effects.

Biomass burning aerosol is often another important constituent in African aerosols that we see transported across the tropical Atlantic on synoptic scales. Our analysis scheme could conceivably be used to study biomass burning smoke outbreaks which occur most frequently in the Sahel-Sudan region in boreal winter and spring and in southern Africa in boreal summer and autumn. While clearly separating mineral dust from biomass burning with satellite remote sensing products remains a challenge, it would be interesting and important to document what current satellite aerosol retrievals reveal in this regard.

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J. Huang, J. M. Prospero, and C. Zhang, Rosenstiel School of Marine and Atmospheric Science, University of Miami, 4600 Rickenbacker Cswy, Miami, FL 33149, USA. (jhuang@rsmas.miami.edu)

J. Huang, J. M. Prospero, and C. Zhang, Rosenstiel School of Marine and Atmospheric Science, University of Miami, 4600 Rickenbacker Cswy, Miami, FL 33149, USA. (jhuang@rsmas.miami.edu)