Aerosol-Induced Large-Scale Variability in Precipitation over the Tropical Atlantic

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ABSTRACT

Multiyear satellite observations are used to document a relationship between the large-scale variability in precipitation over the tropical Atlantic and aerosol traced to African sources. During boreal winter and spring there is a significant reduction in precipitation south of the Atlantic marine intertropical convergence zone (ITCZ) during months when aerosol concentrations are anomalously high over a large domain of the tropical Atlantic Ocean. This reduction cannot be linearly attributed to known climate factors such as El Niño–Southern Oscillation, the North Atlantic Oscillation, and zonal and meridional modes of tropical Atlantic sea surface temperature or to meteorological factors such as water vapor. The fractional variance in precipitation related to aerosol is about 12% of the total interannual variance, which is of the same order of magnitude as that related to each of the known climate and weather factors. A backward trajectory analysis confirms the African origin of aerosols that directly affect the changes in precipitation. The reduction in mean precipitation mainly comes from decreases in moderate rain rates (10–20 mm day$^{-1}$), while light rain (<10 mm day$^{-1}$) can fluctuate in the opposite direction. The results cannot be readily explained in terms of wet deposition or uncertainties in satellite retrievals, and suggest that the observations demonstrate clearly identifiable effects of African aerosol on large-scale variability in precipitation in the tropical Atlantic region.

1. Introduction

The possibility that aerosol may influence precipitation and thereby modulate the global hydrological cycle has been the focus of much research (e.g., Haywood and Boucher 2000; Ramanathan et al. 2001; Rosenfeld et al. 2001; Lohmann and Feichter 2005). Aerosol influences on precipitation may derive from radiative effects (Carlson and Benjamin 1980; Miller and Tegen 1998; Diaz et al. 2001; Ramanathan et al. 2001; Yoshioka et al. 2007) and/or its role as cloud condensation nuclei (CCN) or ice nuclei (IN) (Twomey et al. 1984; Albrecht 1989; Wurzler et al. 2000; Sassen et al. 2003; Lohmann and Feichter 2005). Aerosol radiative processes (absorption and scattering) can modify the atmospheric and surface conditions to affect precipitation; meanwhile, precipitation efficiency and amount sensitively depend on aerosols serving as CCN and IN. Intuitively, aerosol modulation of large-scale variability in precipitation should be identifiable. In reality, however, convincing evidence of such modulation has not been easy to obtain because of the complexity of the processes. Precipitation variability depends on many factors ranging from the large-scale circulation, cloud dynamics and microphysics, and the complex interrelationships among these factors, most of which are poorly understood. Many observational and modeling studies of aerosol effects on precipitation have yielded inconsistent results (Menon 2004; Takemura et al. 2005; Denman et al. 2007; Menon and Del Genio 2007; Tao et al. 2007). The inconsistency mostly likely comes from the fact that aerosol effects on precipitation sensitively depend on cloud types and environmental conditions, such as water vapor supply and atmospheric instability (Lohmann and Lesins 2002; Matsui et al. 2004; Menon 2004; Segal et al. 2004; Koren et al. 2005; Tao et al. 2007), as well as on the characteristics (e.g., size distribution, concentration, chemical composition, and surface physical properties) of different aerosol types such as smoke (Kaufman et al. 2005b; Rosenfeld 1999), urban and industrial air pollution (Rosenfeld 2000; Menon et al. 2002; Kaufman et al. 2005b; Huang et al. 2007), and mineral dust (Rosenfeld et al. 2001; Mahowald and Kiehl 2003; Kaufman et al. 2005b). Aerosol radiative forcing on the surface and atmosphere also influence
precipitation at remote locations (e.g., Rotstayn et al. 2007; Wang 2007).

Investigations on aerosol–precipitation interaction on large or climatic scales face difficult challenges (Denman et al. 2007). There are uncertainties in satellite retrievals of aerosol and precipitation. Numerical models suffer from limitations in their parameterizations of aerosol effects and precipitation processes. In situ observations from field campaigns and high-resolution model simulations are often limited to certain specific environments and specific types of precipitating clouds; consequently it is impossible to generalize their results to climatic scales. Given the complications of precipitation mechanisms and aerosol effects on precipitation, we consider a different approach and ask whether coherent large-scale variability in precipitation can be induced by aerosols and, if it can, whether it can be detected by satellite observations. In this study, we address this question by testing a null hypothesis that there is no coherent large-scale variability between aerosol and precipitation in the tropics. Rejecting null hypothesis should be a necessary step to justify other studies, observational and modeling alike, on climatic effects of aerosol on precipitation. Observations of large-scale covariability between aerosol and precipitation are also needed to validate numerical models for the study of climatic effects of aerosol on precipitation.

We chose the tropical Atlantic as our focus. In this region, the concentration and variability of aerosol from Africa are among the largest, if not the largest, in the world (Anderson et al. 1996; Prospero 1996, 1999; Prospero and Lamb 2003; Torres et al. 2002; Kaufman et al. 2002). It is well known that mineral dust from West Africa is often transported westward into the tropical Atlantic region north of the ITCZ within a layer of the lower troposphere known as the Saharan air layer (SAL) (Carlson and Prospero 1972; Karyampudi et al. 1999; Kaufman et al. 2005c). African dust can reach the Caribbean (Prospero and Lamb 2003) and the southeastern United States (Prospero 1999; DeMott et al. 2003), most strongly in boreal summer. In boreal winter, mineral dust from Africa can be transported southward across the Guinean coast and westward into the tropical Atlantic south of the ITCZ, even into South America (Prospero et al. 1981; Prospero and Lamb 2003; Chiapello et al. 2005). At the microphysical level, mineral dust may serve both as giant CCN to facilitate warm rain formation (Johnson 1982; Rosenfeld and Farberstein 1992) and as IN to facilitate cold rain formation (DeMott et al. 2003). However, precipitation suppression due to desert dust is also possible (Rosenfeld et al. 2001). In addition, carbonaceous aerosol from biomass burning in Africa is also transported westward into the tropical Atlantic, especially in boreal winter and south of the ITCZ (Hao and Liu 1994; Torres et al. 2002; Kaufman et al. 2005c). It has been reported that carbonaceous aerosols tend to suppress warm rain in the tropics (Kaufman and Fraser 1997; Rosenfeld 1999; Wang 2007). Numerical simulations have suggested that anthropogenic aerosol forcing can potentially move the ITCZ southward by cooling the Northern Hemisphere more strongly than the Southern Hemisphere (e.g., Rotstayn and Lohmann 2002; Biasutti and Giannini 2006; Cox et al. 2008). There is also evidence suggesting that Saharan dust may modulate the interannual variability of Atlantic hurricanes (Dunion and Velden 2004; Evan et al. 2006; Lau and Kim 2007). These and other previous results notwithstanding, observational evidence of climatic effects by African aerosol on precipitation over the tropical Atlantic has yet to be found, and hence the motivation of this study.

There are both pros and cons in choosing the tropical Atlantic as a test ground for detecting climatic effects of aerosol on precipitation. The processes that produce dynamic forcing to precipitation in the tropical Atlantic include local sea surface temperature (Zebiak 1993) and synoptic-scale waves (e.g., Thompson et al. 1979; Gu and Zhang 2002) and remote influences by the El Niño–Southern Oscillation (ENSO) (e.g., Alexander and Scott 2002; Xie and Carton 2004), the North Atlantic Oscillation (NAO) (e.g., Wu and Liu 2002; Kyte et al. 2006), and perhaps the West African monsoon (Hagos and Cook 2005). But, arguably, these forcing mechanisms are relatively weak in comparison to those in other tropical regions such as the eastern Pacific, where ENSO signals dominate, and the western Pacific and Indian Oceans, where the Madden–Julian oscillation prevails (Madden and Julian 1971). The high aerosol concentrations and their great temporal and spatial variability combined with a relatively weak dynamic forcing on precipitation make the tropical Atlantic an ideal region to detect climatic effects of aerosol on precipitation. On the other hand, the mixture of mineral dust and smoke aerosol in the tropical Atlantic region complicates the interpretation of any observed large-scale variability in precipitation associated with aerosols. Satellite aerosol data, especially those with long-term records that we used in this study (see section 2a), cannot readily distinguish between dust and smoke. We are therefore forced to settle with a lower expectation: to find out whether there is coherent large-scale variability in precipitation associated with anomalous aerosols regardless of their types.

Our analysis strategy is the following: within the limitation of linear data analysis, we first remove influences of known climate and weather factors (e.g., ENSO, NAO, tropical Atlantic SST, and water vapor) from long-term satellite data of aerosol and precipitation. Then we look for large-scale covariability between any
remaining anomalies in aerosol and precipitation data using statistical significance tests. If significant linkages are found, we employ more recent high-resolution and high-accuracy satellite data to investigate physical significance of the linkage by comparing the magnitudes of precipitation fluctuations related to aerosols and to the other climate and weather factors. We then use a backward trajectory analysis to identify the source regions of aerosols that may be directly involved with the observed variability in precipitation. The null hypothesis $H_0$ will be rejected only if all these steps are successfully accomplished. Finally, we employ several other approaches to help interpret the observed covariability between African aerosol and precipitation.

The data used in this study and our analysis method are described in section 2. Results are presented in section 3. A summary and discussions are provided in section 4.

2. Data

a. Main datasets

Our analyses cover three periods. The longest is from 1979 through 2000 with a data gap over 1993–96. In this period, the Global Precipitation Climatology Project (GPCP) (Huffman et al. 1997) precipitation and the Total Ozone Mapping Spectrometer (TOMS) aerosol index (AI) are available. Our main statistics about aerosol and precipitation are derived from this period. The second period is from 1988 through 2000 during which the Special Sensor Microwave Imager (SSM/I) precipitable water data are available. The objective of the analysis for this period is to isolate water vapor effects on precipitation from aerosol effects. The third period covers the most recent years, 2000–06, when the Tropical Rainfall Measuring Mission (TRMM) precipitation and the Moderate Resolution Imaging Spectroradiometer (MODIS) aerosol data are available. These two datasets are of higher resolution and higher quality than the earlier products; they were used to confirm the results based on the GPCP and TOMS data.

The GPCP version 2 monthly and pentad (five day) precipitation data (mm day$^{-1}$, 2.5° × 2.5°) incorporate precipitation estimates from low-orbit satellite microwave data, geosynchronous-orbit satellite infrared data, and surface rain gauge observations but there were no microwave data before mid-1987 (Huffman et al. 1997; Adler et al. 2003; Yin et al. 2004). The GPCP pentad precipitation data is used to examine variability in precipitation at different rain rates.

The TOMS aerosol index (AI) version 8 (1° × 1.25°) monthly data are available from two platforms: Nimbus-7 (November 1978–April 1993) and Earth Probe (August 1996–December 2000). Thus there is a data gap from 1993 to 1996. Also, the Earth Probe TOMS AI data after December 2000 were not used because of their known calibration drift and biases (see http://toms.gsfc.nasa.gov/news/news.html). TOMS AI can be taken as a qualitative indicator of UV-absorbing aerosols of which dust and carbonaceous aerosol are the dominant types (Herman et al. 1997; Torres et al. 1998, 2002). We used 0.5 as threshold values for TOMS AI because low AI values may contain ground noise or cloud contamination (Herman et al. 1997). TOMS AI is more sensitive to aerosols at high altitudes than those in the low-lying layer but, nonetheless, Barbados surface dust concentration and TOMS AI are closely correlated (Chiapello et al. 1999).

Monthly SSM/I atmospheric water vapor data (mm, 0.25° × 0.25°) (Jackson and Stephens 1995; Wentz 1997) were used to examine water vapor effects on precipitation in the second analysis period (1988–2000). The monthly TRMM precipitation 3B43 version 6 product (mm day$^{-1}$, 0.25° × 0.25°) (Fisher 2004) and the monthly Terra MODIS aerosol optical depth (AOD) (1° × 1°) (Kaufman et al. 1997), both available for the third analysis period (2000–06), were used to validate GPCP–TOMS results.

Major climate factors influencing Atlantic precipitation include ENSO (Philander 1990), NAO (Hurrell 1995), and the zonal and meridional modes of Atlantic SST (Zebiak 1993). Indices used for these climate factors are the ENSO Niño-3.0 (5°S–5°N, 90°–150°W) SST index, the NAO index, the tropical Atlantic zonal mode Atl.3.0 (3°S–3°N, 20°W–0°) SST index, and the tropical Atlantic meridional mode SST index (SST anomaly difference between 5° and 25°N, 55° and 15°W and between 0° and 20°S, 30°W and 10°E) (Servain 1991; Wang 2002).

b. Analysis domains

We focus mainly on precipitation variability in the Atlantic marine ITCZ (AMI), which has a strong influence on the climate of adjacent continents, especially Northeast Brazil and West Africa (Hastenrath and Heller 1977; Lamb 1978; Janicot et al. 1998; Robertson et al. 2004). The analysis domains for precipitation and aerosol are selected to reflect both the spatial and the temporal (monthly and interannual) variability of precipitation and aerosol. Land effects are minimized by excluding continental regions from the analysis domain. A domain for precipitation extends over the range that encompasses the seasonal shifting of the AMI from its southernmost position in boreal spring to its northernmost position in boreal summer. This domain will hereafter be referred to as the AMI domain (Fig. 1). The analysis domain for aerosol, denoted as Atlantic aerosol (AA) domain in Fig. 1, covers a much larger region than the AMI precipitation area at any time. We assume that the variability of aerosol directly affecting AMI precipitation is part of
the basin-scale coherent variability that can be measured by aerosol data averaged over the larger AA domain. The assumption is supported by daily satellite observations of aerosol and large-scale synoptic patterns responsible for transporting aerosol from Africa westward to the tropical Atlantic region (Prospero and Lamb 2003; Chiapello et al. 2005). With this two-domain approach, we minimize the possible contaminations of the detected variability in aerosol averaged over the AA domain by the complications of wet deposition and uncertainties in satellite aerosol retrieval in cloudy areas. We took additional steps to distinguish between possible aerosol effects on precipitation and those of wet deposition. For example, we further investigated the sensitivity of the observed aerosol effect to rain rates and also compared the spatial and seasonal variability of the observed aerosol effect to that of the model-simulated wet deposition.

c. Data processing

Aerosol and precipitation anomalies are calculated by removing their seasonal cycles. Effects on aerosol and precipitation by known climate factors, such as the ENSO, NAO, and the zonal and meridional modes of Atlantic SST, were removed through multivariable linear regression. This is to reduce the ambiguity that covariability between aerosol and precipitation might be caused simultaneously by these climate factors. The resulting aerosol and precipitation anomalies still exhibit residual seasonal cycles in their variance. To treat all months equally so that the total sample size is increased, the anomalies are normalized by their interannual standard deviation in each month averaged over the respective domains. We will use the term “normalized anomalies” when referring to these latter time series.

The data series are not sufficiently long to explicitly address issues related to any long-term trend. To avoid uncertainties in this and to keep the focus within the scope of this study, the domain means of normalized anomalies were detrended. We subsequently obtained similar results using data that were not detrended.

3. Large-scale covariability

We begin by examining climatological features of precipitation (GPCP) and aerosol (TOMS AI). Figures 1, 2, and 3 show, respectively, their climatological means, seasonal cycles, and interannual variability in the
tropical Atlantic. A prominent feature of the climatology is that the AMI, represented by the narrow band of concentrated precipitation in Figs. 1 and 2, is flanked most of the time by high concentration of aerosol on both its north and south sides. The seasonal cycles in aerosol and precipitation are clearly evident in both the latitudinal positions of the activity and their amplitudes (Fig. 2a). They also undergo obvious interannual and decadal variability (Fig. 3) (Cakmur et al. 2001; Prospero and Lamb 2003; Gu and Adler 2006). The domain-averaged Nimbus-7 and Earth Probe TOMS AI are comparable in their seasonal cycles. The TOMS AI values are relatively higher in boreal spring and summer and lower in autumn and winter. There is an increasing trend in the early years of the Nimbus-7 TOMS AI; the increase is consistent with long-term dust measurements made on Barbados (Prospero and Lamb 2003) and is linked to the onset of severe drought in the Sudano–Sahel region of North Africa. Beyond that period, the TOMS AI varies around a mean of about 0.75 from year to year. The interannual variability of precipitation is steady with an averaged annual mean of ~3 mm day$^{-1}$. Its seasonal cycle varies from 2 mm day$^{-1}$ in boreal winter to as high as 5 mm day$^{-1}$ in boreal summer. Given the intensity and scale of aerosol transport that takes place over the tropical Atlantic for so much of the year, we might expect that, if aerosols have an impact on precipitation, the impact should be most readily detectable over this region.

a. Effects from known climate factors

We apply a multivariable regression to the anomalous time series and compare the aerosol-related precipitation variance (averaged over the AMI domain) to that related to each of the known climate factors, namely, ENSO, NAO, and Atlantic SST zonal and meridional modes (Fig. 4a). The highest aerosol-related precipitation variance is about 30% of the total in January and about 3%–15% in other months. The largest fractional variance in precipitation anomalies related to ENSO occurs in January (~43%). The Atlantic zonal SST mode dominates the precipitation variability in boreal summer (July and August) and the meridional mode dominates in boreal winter and spring (March and April), each of them accounts for 25%–40% of the variability. The largest fractional variance linked to the NAO is in February (~15%); the NAO is not a dominant factor in the ITCZ precipitation variability in any month. Although aerosol is not a dominant factor in precipitation variability in any single month, its influence is comparable to those by ENSO, NAO, and tropical Atlantic SST, especially in boreal winter and spring. On average, the fractional interannual variance in precipitation related to aerosol is about 12%—the same order of magnitude as those related to each of the other known climate factors.

These climate factors also affect the aerosol variability (e.g., Prospero and Ness 1986; Moulin et al. 1997; Ginoux et al. 2004). Similar multivariable regression was
conducted on aerosol variance and results are shown in Fig. 4b. ENSO explains as high as 38% of aerosol variance in January. Effects from the tropical Atlantic SST on aerosol are at a maximum in April for the zonal mode (35%) and in June for the meridional mode (40%).

The NAO, with an average less than 10% of the aerosol interannual variance, is not a dominant factor for the aerosol variability in any month.

The effects from these climate factors on both precipitation and aerosol may complicate the interpretation of observed aerosol–precipitation relationships. To avoid such possible complication, their effects were removed from the anomalous time series of both precipitation and aerosol. The analyses described hereafter were therefore made using data that are linearly independent of ENSO, NAO, and tropical Atlantic SST.

b. Correlation

Figure 5 compares time series of monthly normalized anomalies in GPCP precipitation averaged over the AMI domain and TOMS AI averaged over the AA domain. A visual inspection gives an impression that anomalously low (high) precipitation tends to occur in months of anomalously high (low) aerosol concentration. A scatter diagram for the two variables (Fig. 6) confirms this casual observation. The inverse relationship between aerosol and precipitation normalized anomalies is significant at the 99% confidence level with a correlation coefficient ($R$) of $0.324$ (a two-tailed nondirectional $t$ test is used to determine confidence levels of the correlation significance). The inverse relation implies that precipitation averaged in the AMI domain is reduced when aerosol concentration averaged in the AA domain is anomalously high.

To test the sensitivity of the correlation to the domain selection, we list in Table 1 seven cases in which correlation coefficients were calculated for the two variables averaged in different domains. The only insignificant correlation for precipitation and aerosol is found for both in the north AA (NAA) domain (case E). The most interesting is case G in which the precipitation and aerosol domains are mutually exclusive but the correlation between their normalized anomalies remain significant.

The correlation between aerosol and precipitation varies through the year (Fig. 6). It is largest in January ($R = 0.626$) and February ($R = 0.649$), both significant at the 99% confidence level. The correlations are significant in boreal winter (December–February, $R = 0.51$, 99% confidence level), summer (June–August, $R = 0.31$, 95% confidence level), and spring (March–May, $R = 0.27$, 95% confidence level). The most significant correlation between the two occurs in months of the highest aerosol concentration south of the ITCZ (e.g., January and February), while north of the ITCZ the months with the highest aerosol concentrations are June and July. This seasonality in correlation is hardly explainable in terms of wet deposition or uncertainties in satellite retrievals.

To demonstrate the spatial distribution of the precipitation variability associated with aerosol, we calculated correlation between the normalized anomalous aerosols averaged over the AA domain and the normalized anomalous precipitation at each grid point. Figure 7 shows the correlation coefficients. Areas with correlation statistically significant at the 95% and 99% confidence levels are outlined. The inverse correlation is highest (at the 99% confidence level) south of the
ITCZ in the western tropical Atlantic. Evident in Fig. 2, December–May is the season when absorbing aerosol south of the equator is relatively high and the ITCZ approaches its southernmost position. The high values of TOMS AI are possibly the combined effect of carbonaceous particulate produced by the intense fire season in central Africa and the southward flow of mineral dust from Lake Chad, the Sudano–Sahel region, and other significant sources of mineral dust in West Africa (Torres 2002). The aerosol sources will be identified in section 3h. This preference of negative correlation south of the AMI rainband is, again, unexplainable in terms of wet deposition or uncertainties in satellite retrievals.

### c. Composites

To further explore the covariability between aerosol and precipitation, we compare precipitation and aerosol between months of extreme high and low aerosol concentrations. The top and bottom thirds (terciles) of all

![Fig. 5. Time series (1979–2000) of AMI domain-averaged precipitation normalized anomalies (dashed line) and AA domain-averaged aerosol normalized anomalies (solid line).](image1)

![Fig. 6. Scatter diagram of monthly AA domain-averaged aerosol normalized anomalies and AMI domain-averaged precipitation normalized anomalies. Solid line indicates a linear fit (correlation coefficient $R = 0.324$, significant at the 99% confidence level).](image2)
months ranked by normalized anomalies in aerosol averaged over the AA domain were selected to represent such extreme months of aerosol concentrations. Composites of the normalized anomalies in precipitation and aerosol for the two terciles were made at each grid point. The differences between the composites of the two terciles highlight the variability between months of high and low anomalous aerosol concentrations. They are shown in Figs. 8a,b for precipitation and in Figs. 8c,d for aerosol. The spatial distribution of the difference in precipitation shows a coherent pattern over the western tropical Atlantic (Fig. 8a). The largest difference is south of the ITCZ, consistent with the correlation pattern in Fig. 7. Its magnitude is about $-1.5$, corresponding to a reduction in monthly precipitation of 55 mm, roughly 40% of monthly mean precipitation in the region (Fig. 1). When the composite procedure is repeated for each calendar month, a distinct seasonal cycle in the difference emerges (Fig. 8b). The reduction in precipitation remains south of the ITCZ center and migrates in latitude along with the ITCZ. Its largest amplitudes are in boreal winter and spring.

The same composites and their differences were made for aerosol (Figs. 8c,d). The most striking feature is that largest differences in aerosol between the top and bottom terciles are not associated with the regions where the largest differences in precipitation are found. This is partially understandable. The largest aerosol variability should be near the aerosol sources, over Africa in this case, where there might not be much precipitation at all. What directly matters to ITCZ precipitation processes is not so much the amount of aerosol present but the availability of aerosols to interact with precipitating clouds in the ITCZ in cloud microphysics or the effectiveness of aerosols on changing large-scale circulation through their radiative forcing. Meanwhile, Fig. 8 also justifies our approach of using the AA domain to represent the large-scale variability of aerosols that may affect ITCZ precipitation. Selecting an aerosol domain with precipitation domain mutually exclusive (like case G in Table 1) does not change the composites significantly. The composite variability in precipitation and aerosol in both space and time cannot be explained by wet deposition or uncertainties in satellite retrievals.

The significance of the difference shown in Fig. 8a was tested from two standpoints. First, we calculated the probability that the difference pattern could be produced by randomly selecting the same number of months for the composites. This was done by comparing the probability distribution function (PDF) of the composite difference within the AMI domain (bars in Fig. 9) to a Gaussian distribution with the same standard deviation but zero mean, which represent the PDF from random sampling (solid curve in Fig. 9). A Kolmogorov–Smirnov (K–S) test indicates that the two distributions are different at the 99% confidence level. In other words, the probability that the difference pattern in Fig. 9a might be reproduced by random sampling is only 1%. Second, we calculated the probability that this difference pattern might be the same as the interannual variability of precipitation without considering any influences from aerosol. The PDF of interannual variability in precipitation was calculated using the entire normalized anomalies in the same domain (dotted curve in Fig. 9). Again, a

### Table 1. Testing cases for the selection of aerosol and precipitation domains.

<table>
<thead>
<tr>
<th>Case</th>
<th>Aerosol domain</th>
<th>Precipitation domain</th>
<th>Confidence level</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>AA</td>
<td>AMI</td>
<td>&gt;99%</td>
</tr>
<tr>
<td>B</td>
<td>AA</td>
<td>AA</td>
<td>&gt;99%</td>
</tr>
<tr>
<td>C</td>
<td>NAA</td>
<td>AMI</td>
<td>&gt;99%</td>
</tr>
<tr>
<td>D</td>
<td>SAA</td>
<td>AMI</td>
<td>&gt;99%</td>
</tr>
<tr>
<td>E</td>
<td>NAA</td>
<td>NAA</td>
<td>&lt;95%</td>
</tr>
<tr>
<td>F</td>
<td>SAA</td>
<td>SAA</td>
<td>&gt;99%</td>
</tr>
<tr>
<td>G</td>
<td>Within AA, but with precipitation domain exclusive</td>
<td>10° latitudes to the north and south of the AMI centerline (see Fig. 2)</td>
<td>&gt;99%</td>
</tr>
</tbody>
</table>
K–S test indicates that this PDF is different from that of composite difference at the 99% confidence level. We thus conclude that the difference pattern in Fig. 8a indeed represents a significant change in precipitation between months of anomalously high and low aerosol concentrations.

d. Variability at different rain rates

The previous analyses focus on changes in mean precipitation. It is possible that a change in the mean is not uniformly contributed through the entire range of the rain rate. In other words, it is reasonable to expect that the observed reduction in mean precipitation associated with aerosol might be larger in a certain range of the rain rates than in others. On the other hand, when small, statistically insignificant changes in mean precipitation are found, there could be opposite fluctuations in different precipitating systems (rain rates) with the result that they might tend to cancel each other. To address this issue, we used the pentad GPCP precipitation

![Fig. 8. Difference composites of precipitation and aerosol normalized anomalies between the top and bottom aerosol tercile months: (a) the annual mean and (b) seasonal cycle of precipitation; (c) the annual mean and (d) seasonal cycle of aerosol; (e),(f) the same as (a) and (b), but for December–April only. Dashed lines in (b) and (d) mark the center of the ITCZ rainband.](image-url)
data to compare monthly precipitation as a function of the rain rate for the top and bottom tercile months of aerosol concentrations. Their differences are shown in Fig. 10 for several subdomains in the analysis region (Fig. 10f).

In subdomain A, the precipitation reduction related to aerosol, seen in the mean (Figs. 8a and 10f), mainly occurs in the rain rate range of 5–15 mm day\(^{-1}\) with a maximum reduction in monthly precipitation of 120 mm at the rain rate of 11 mm day\(^{-1}\) (Fig. 10a). There is a sign of precipitation enhancement for very light rain (<3 mm day\(^{-1}\)) and there is no obvious change in heavy rain (>25 mm day\(^{-1}\)). In the center of the ITCZ (subdomain B in Fig. 10f), the precipitation reduction seems to shift toward higher rain rates (9–20 mm day\(^{-1}\) and the light rain (<7 mm day\(^{-1}\)) enhancement becomes more obvious in comparison to subdomain A. Their combined contribution gives a weak reduction in the mean (Figs. 8a and 10f). The tendency of the aerosol-related reduction to shift toward higher rain rates and increased enhancement at light rain rates becomes more obvious along the northern edge of the ITCZ (subdomains C and D). Comparing the aerosol-related changes to subdomain A, the total amount of reduction is about the same north and south of the ITCZ, but the enhancement in light rain is much greater to the north. This is especially true in subdomain C, where the net effect of the reduction at moderate rain rates and the enhancement in light rain yields a negligible change in the mean (Figs. 8a and 10f). To the south of the ITCZ in subdomain E (Fig. 11e), we find a different feature: light rain (<10 mm day\(^{-1}\)) is reduced as opposed to being enhanced north of the ITCZ in subdomain D. While the complication in aerosol-related changes in precipitation at different rain rates will be discussed more in section 4, an important message here is that effects of aerosol on precipitation cannot always be measured solely in terms of changes in mean precipitation. Again, we emphasize that these contrasting fluctuations at different rain rates and at different locations cannot be explained in terms of wet deposition and uncertainties in satellite retrievals.

### e. Variance

The importance of the aerosol to precipitation needs to be understood in the context of total variability in precipitation. This can be addressed from two perspectives. First, the variance in precipitation explainable by aerosol effects should be compared to those by other known climate factors such as ENSO, tropical Atlantic SST, and NAO. This was discussed in section 3b. In certain months of the year, the aerosol-related variance in monthly mean precipitation is at the same order of magnitude as those related to the other climate factors (Fig. 4). Similarly, we may ask how much of the variance in precipitation that is linearly independent of the other climate factors can be explained by aerosol. To quantify this, we calculated the “residual variance” of precipitation, which is the variance of precipitation after the influences by ENSO, tropical Atlantic SST, and NAO were linearly removed by the multivariable regression as described in section 3b. The aerosol-related fraction of this residual variance was then calculated at each grid point using aerosol anomalies averaged over the AA domain. As expected, the large fractional variance due to aerosol is mainly south of the ITCZ (Fig. 11) where the largest aerosol-related reduction in precipitation is found. Aerosol-related variance explains up to 15% of the residue variance. This aerosol-related fractional variance can be as high as 45% in certain months (e.g., January). All these observations suggest that precipitation variability related to aerosol is a nonnegligible component of its total variability.


An outstanding problem not yet addressed is the possibility that the observed precipitation change associated with aerosol is caused not by aerosol but rather...
by the presence of anomalously low water vapor in air masses bearing anomalously high aerosol. It is known that during much of the year dusty air masses emerge from North Africa and pass over the tropical Atlantic north of the equator and into the Caribbean. During the summer, these outbreaks are characterized by the presence of an elevated layer of hot, dry, dust-laden air—the Saharan air layer; because of these distinguishing characteristics it is commonly referred to as the SAL (Carlson and Prospero 1972; Karyampudi et al. 1999). South of the equator, dry air outbreaks from Africa have also been observed (Zhang and Pennington 2004). Some of them may have high aerosol concentrations. To quantitatively isolate effects of aerosol from those of water vapor, we made two similar sets of composites of precipitation between top and bottom aerosol tercile months for the shorter second period (1988–2000) when SSM/I water vapor data are available. In one set of composites, only the known climate factors are removed through a multivariable regression as done previously for the longer first period (1979–2000). In another, the SSM/I water vapor averaged over the AA domain is included in the multivariable regression and its linear association with precipitation is removed along

Fig. 10. (a)–(e) Differences in monthly precipitation between the top and bottom aerosol tercile months as a function of the rain rate averaged in (f) five subdomains (A–E). Total difference in monthly precipitation (mm) is given for each subdomain. The thick solid lines in (a)–(e) are 2 mm d$^{-1}$ moving averages.
with the other climate factors. The composite difference for the second period, with or without water vapor effect removed, is almost identical (cf. Fig. 12 to Fig. 8a). The domain-averaged reduction in mean precipitation attributable to water vapor is only about 4% of that due to aerosol. The fractional variance of monthly precipitation that is explainable by water vapor variability is about 1% in comparison to 9% by aerosol. We can therefore confidently say that the aerosol-related changes in precipitation observed here cannot be attributed to accompanying water vapor variability.

g. Analysis for the third period (2000–06): Validation by high-resolution data

The long-term satellite data used thus far in this study (GPCP and TOMS) may suffer from inaccuracies for various reasons. To validate their results we perform the same analysis using more recent satellite products from TRMM and MODIS that provide more accurate data with higher resolutions. On the other hand, when data of higher temporal (e.g., daily) resolution are used, some of the assumptions underlining our analysis method (e.g., that the AA domain mean aerosol represents the variability that directly affects ITCZ precipitation) may not be valid. Consequently the distinction between changes resulting from wet removal of aerosol and those resulting from aerosol-induced changes in precipitation may not be clearly resolved solely from satellite observations.

Despite these limitations, we ask whether the same analysis method applied to monthly mean TRMM and MODIS data would yield results consistent with those obtained from the GPCP and TOMS data even though the results may not be identical. Following this line of thinking, we repeated the composites for the top and bottom tercile months of aerosol using MODIS aerosol optical depth (AOD) data and TRMM precipitation data for the period 2000–06. Despite the limited temporal coverage of data and detailed discrepancies, we obtain a relatively consistent result for the main regions where precipitation reduction is found (Figs. 8a and 13a). When aerosol is anomalously high, precipitation is reduced in the same region as found from the GPCP and TOMS data: south of the ITCZ in the western tropical Atlantic. The strongest reduction there also occurs during boreal winter and spring (Fig. 13b). We find, however, that TRMM precipitation is obviously enhanced north of the ITCZ in association with increases in the MODIS aerosol concentration (Fig. 13a). This enhancement, very weak in the GPCP–TOMS data, is worthy of further scrutiny.

h. Case study: Back trajectory analysis

A remaining question from the previous analysis is, What is the source of the aerosol that may directly relate to the precipitation reduction? To identify the aerosol origin, we conducted backward trajectory analyses using the Hybrid Single-Particle Lagrangian Integrated Trajectory version 4 (HYSPLIT4) model applied to the National Center for Atmospheric Research (NCAR) reanalysis data (Draxler and Rolph 2003; Rolph 2003). The starting points of the back tracking are set at three levels (500, 1500, and 3000 m) at the meridional grid of 0° and 35°W, where the aerosol-related precipitation reduction is largest. This point will be referred to as the target point of the backward trajectory analysis. Because the precipitation reduction is most significant in boreal winter and spring (section 3c), two cases were selected for the backward trajectory tracking: January–February
of 1997 and April–May of 1987. In the time series of normalized anomalies (Fig. 5), these two seasons are observed with highest anomalous aerosol concentration and significant precipitation reductions. The backward trajectories are shown in Fig. 14. It is clear that aerosols reaching the target area are from the east, with the most persistent transport occurring at the 1500-m and 3000-m levels; the 500-m trajectories arrive mostly from oceanic regions. In January and February of 1997, the most possible aerosol sources are in central Africa where at this time of year the concentrations of dust and smoke are often very high (Torres 2002; Husar et al. 1997). In April and May 1987, there appear to be two separate cluster paths of aerosol from Africa, instead of one as in January and February. The dominant paths pass over southern West Africa along the coast of the Gulf of Guinea, as in January–February, and the second, much weaker, origi- nates from south-central Africa. The details of sources and trajectories need to be further explored using a more systematic approach and extending over a sufficiently long period to develop more reliable statistics. Nevertheless, these two cases give us confidence that aerosols directly related to the precipitation reduction south of the ITCZ are from African sources.

\[\text{(a)}\]

\[\text{(b)}\]

**Fig. 13.** Difference composite of TRMM precipitation normalized anomalies between top and bottom tercile months of MODIS AOD normalized anomalies: (a) annual mean; (b) seasonal cycle. The center of the ITCZ rainband is marked by black dashed lines.

\[\text{4. Summary and conclusions}\]

In an attempt to explore the effects of aerosol on large-scale variability in tropical precipitation relevant to climate time scales, we have documented statistically significant covariability between African aerosol and precipitation in the tropical Atlantic using long-term multiyear satellite data of aerosol (TOMS A1) and precipitation (GPCP). The strongest signature of this covariability lies south of the ITCZ over the western tropical Atlantic Ocean where precipitation is reduced in those months when aerosol concentrations are anomalously high over a large area of the tropical Atlantic. This aerosol-related reduction in precipitation is most prominent during boreal winter and spring. The aerosol-related fluctuations in precipitation are linearly independent of known climate and weather factors (e.g., ENSO, NAO, tropical Atlantic SST, and water vapor) and are of the same order of magnitudes as those associated with these factors. The aerosol-related changes in precipitation vary with rain rate; the effect is greatest at lower rain rates. Under some conditions modest increases are found. In addition to monthly results, a separate study

\[\text{i. Wet deposition}\]

We have pointed out that the observed aerosol-related precipitation reduction cannot be explained by wet deposition. To further substantiate our argument, we compare the observed spatial distribution and seasonality of aerosol-related precipitation change to that of the model-simulated wet deposition [see Wang (2007) for model description]. The patterns of the wet deposition can be illustrated by the grid-to-grid correlation between black carbon and precipitation when the black carbon effects on precipitation are turned off in the model and the wet deposition is the sole interaction (Fig. 15). The outstanding features from Fig. 15 are that signals of wet deposition are robust nearly everywhere in the AMI domain and in every calendar month, closely following the seasonal migration of the ITCZ. Interestingly, the largest wet deposition signals are south of the ITCZ but they occur in May–July. In general, the spatial distribution and seasonality of wet deposition are unmistakably different from those of the observed aerosol-related precipitation reduction in Fig. 8.
(Huang et al. 2009b) using daily TRMM precipitation and MODIS AOD data and back trajectory analysis also yielded significant reduction in precipitation anomalies attributable to elevated AOD. Based on these results, we confidently reject the null hypothesis $H_0$: there is no coherent large-scale variability between aerosol and precipitation in the tropics. Furthermore, using several different approaches we presented evidence that our results cannot be readily explained by invoking biases introduced by the wet deposition of aerosols or by uncertainties in satellite retrievals. Having ruled out the possible interpretations of our results in terms of the climate and weather factors, wet deposition, and uncertainties in satellite retrievals, the remaining plausible mechanism would be the effect of aerosol on precipitation.

Although our data support this interpretation, more observational and modeling studies are needed to further quantify the aerosol effect on tropical precipitation and to understand its mechanisms. The following issues need to be addressed:

1) As discussed in section 1, aerosols can affect precipitation through their radiative effects or their roles as CCN or IN. However, our observations, although statistically significant, provide no information on the mechanisms of aerosol–precipitation interactions. It is notable, however, that global climate models that incorporate the radiative effects of black carbon aerosol have produced patterns of reduction in precipitation similar to those found in our analysis (Wang 2007; Chung and Seinfeld 2005). The consistent results between modeling and observations lend confidence to the modeling results and provide possible mechanisms to explain the observations.

2) One intriguing result from this study is the opposite effects of aerosol on precipitation at light and moderate rain rates. The observed reduction in precipitation at moderate rain rates is consistent with many previous studies reporting suppression of warm rain (see discussion in section 1). But the accompanying enhancement of light rain by aerosol is unexpected. Sometimes it is this enhancement of light rain that

**Fig. 14.** 10-day back trajectory analysis using HYSPLIT4 model applied to the NOAA reanalysis data for (a) January and February 1997 and (b) April and May 1987. Starting point is $0^\circ$, $35^\circ W$ with three height levels: 500 m (in gray dotted lines), 1500 m (in gray dashed lines), and 3000 m (in black solid lines).
partially offsets the reduction in moderate rain and leads to a trivial change in the total (or mean) precipitation. This has two related implications. First, the effects of aerosol on precipitation cannot be solely judged by changes in the mean. While mean precipitation is a main concern for the hydrological cycle at the surface, how precipitation at various rain rates may vary is directly relevant to dynamics. Second, if changes in different rain rates are associated with different precipitating cloud systems, then aerosol-induced fluctuations in precipitation would also produce fluctuations in the vertical latent heating structure. It is well known that responses of the large-scale circulation to tropical diabatic heating sensitively depend on its vertical profiles (Hartmann et al. 1984; Bergman and Hendon 2000; Schumacher et al. 2004). TRMM retrievals of tropical latent heating profiles (Tao et al. 2006) may provide insights to this problem.

3) It is clear from this study that the largest fluctuations in precipitation associated with aerosol are not in regions where the largest aerosol concentrations are found. One reason for this is simply that, north of the ITCZ where aerosol concentrations can be very high, there is not much precipitation. The other reason might be related to background aerosol. Wang (2005) found in a cloud-resolving model that higher CCN concentrations do not necessarily lead to increased precipitation efficiency; the aerosol effect on precipitation is more readily seen in a pristine environment than in a polluted one. Other large-scale background environments may also play a role in determining the pattern of precipitation response to aerosol effects. For example, one puzzling result from this study is that the largest reduction in precipitation by aerosol is located in the western tropical Atlantic, far from the sources of African aerosol. Even though the significant correlation pattern links the upstream sources to this response area (Fig. 7), the great distance between the largest precipitation response and the aerosol source region needs to be explained. One possible explanation is that the dynamic forcing of precipitation (e.g., SST meridional gradient) in the eastern part of the AMI domain is much stronger than in the west part; thus the precipitation in the eastern region may be less susceptible to external perturbations such as those from aerosol. Figure 16 superposes the precipitation difference composite (as seen in Fig. 8a) onto the climatological mean of tropical Atlantic SST. The largest SST meridional gradient is found where the aerosol effect on mean precipitation change is relatively weak.

4) The asymmetric aerosol effects on precipitation north and south of the ITCZ found in this study may reflect the differential response of precipitation to mineral dust and biomass burning smoke. Even though the aerosol data that we have used cannot distinguish these two types of aerosol, there are reasons to believe that biomass burning aerosol from Africa might have played a major role in the observed reduction of precipitation. Our backward trajectory analysis suggests that the aerosol that impacts on the areas of the largest aerosol-related precipitation reduction is carried from source in equatorial Africa, where biomass burning is most intense in boreal winter and spring (Hoelzemann et al. 2004)—the same season of our observed largest
reduction in precipitation. Other studies have shown that aerosol released from biomass burning over equatorial Africa can spread over the entire tropical Atlantic (Duncan et al. 2003; Hoelzemann et al. 2004; Ito and Penner 2004; Ito and Penner 2005). Previous studies seem to agree that the biomass burning smoke reduces coalescence efficiency and consequently suppresses warm rain (Rosenfeld 1999; Kaufman et al. 2005b). The aerosol effect on warm rain, which typically yields moderate rain rates (<20 mm day$^{-1}$), might be the main reason for the observed reduction in precipitation in this study.

A model simulation of black carbon radiative forcing also suggested a northward shifting of the AMI (Wang 2007), as opposed to some simulations that show a southward shifting of the ITCZ attributed to the asymmetric surface cooling across the equator caused by anthropogenic aerosol forcing (e.g., Rotstain and Lohmann 2002; Biasutti and Giannini 2006; Cox et al. 2008). Nonetheless, simulations confirmed that aerosols can effectively redistribute precipitation on the large scale.

A role of mineral dust in the observed reduction of precipitation cannot be ruled out. During boreal winter and spring large quantities of mineral dust are carried from sources in the Sahel–Sudano region of Africa into the tropical Atlantic (Washington and Todd 2005) and to South America (Prospero et al. 1981; Swap et al. 1996). Mineral aerosols can suppress precipitation in thin low-altitude clouds and changing cloud amounts in ice phase clouds (Mahowald and Kiehl 2003). The reduction of precipitation by desert dust in other parts of the world has been reported (Kaufman et al. 2005b; Rosenfeld et al. 2001; Yin and Chen 2007). Human activities are known to be influential to biomass burning (Dwyer et al. 2000; Kaufman et al. 2005a) and mineral dust mobilization (Moulin and Chiapello 2006). Thus, our results point to a possible role of human activity in modulating precipitation even in remote areas.

The observed precipitation reduction possibly induced by aerosol is not unique over the western tropical Atlantic Ocean. Following the same analysis procedure as that used here, we have obtained similar results in the West African monsoon region (Huang et al. 2009a). It has yet to be seen if similar results can also be found in other tropical regions with aerosol from different sources. With a given temperature distribution, aerosol should not produce a net increase or decrease in global precipitation without altering total surface evaporation. Regionally, aerosol may help redistribute precipitation and change rain characteristics (types, rate, duration, etc.). Therefore, an effect of aerosol on the global hydrological cycle, if any, must be manifested in terms of its regional signatures. Studies of aerosol effects on the global hydrological cycle must therefore be conducted with regional focuses (Haywood and Boucher 2000; Lohmann and Feichter 2005). Our study provides a basis for a regionally focused exercise aimed at understanding the possible climatic effects of aerosol on the global hydrological cycle.

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