African Dust over the Northern Tropical Atlantic: 1955–2008

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ABSTRACT

African dust outbreaks are the result of complex interactions between the land, atmosphere, and oceans, and only recently has a large body of work begun to emerge that aims to understand the controls on—and impacts of—African dust. At the same time, long-term records of dust outbreaks are either inferred from visibility data from weather stations or confined to a few in situ observational sites. Satellites provide the best opportunity for studying the large-scale characteristics of dust storms, but reliable records of dust are generally on the scale of a decade or less. Here the authors develop a simple model for using modern and historical data from meteorological satellites, in conjunction with a proxy record for atmospheric dust, to extend satellite-retrieved dust optical depth over the northern tropical Atlantic Ocean from 1955 to 2008. The resultant 54-yr record of dust has a spatial resolution of 1° and a monthly temporal resolution. From analysis of the historical dust data, monthly tropical northern Atlantic dust cover is bimodal, has a strong annual cycle, peaked in the early 1980s, and shows minimums in dustiness during the beginning and end of the record. These dust optical depth estimates are used to calculate radiative forcing and heating rates from the surface through the top of the atmosphere over the last half century. Radiative transfer simulations show a large net negative dust forcing from the surface through the top of the atmosphere, also with a distinct annual cycle, and mean tropical Atlantic monthly values of the surface forcing range from \(-300 \text{ to } -9 \text{ W m}^{-2}\). Since the surface forcing is roughly a factor of 3 larger in magnitude than the top-of-the-atmosphere forcing, there is also a positive heating rate of the midtroposphere by dust.

1. Introduction

The first unambiguous account of an African dust outbreak over the Atlantic Ocean appeared in the 1721 proceedings of the French Academy of Sciences by P. Feuillée, “Observation sur une pluie de sable dans la mer Atlantique précédée d’une aurore boréale” (Observation on a sand rain in the Atlantic Ocean preceding an aurora borealis) (Feuillée 1721). This brief account describes the encounter of a ship with a “sand rain” on 6 April 1719 at 45°N and 38°W that lasted 3 h. Since that time there has been a steady increase in the number of publications related to dust outbreaks (Stout et al. 2009), and since the dawn of the satellite era the rate of publications on Saharan dust has grown exponentially (Kaufman et al. 2005b).

One interesting subset of study into aeolian dust focuses on the relationship between climate variability and African dust outbreaks. Much focus has been placed on demonstrating that Atlantic dust storm frequency and intensity changes are forced by variability in Sahelian precipitation (e.g., Chiapello et al. 2005; Prospero and Nees 1977), and others have conversely examined the ability of atmospheric dust to locally alter precipitation (e.g., Miller et al. 2004; Rosenfeld et al. 2001; Yoshioka et al. 2007) and cloud formation (Kaufman et al. 2005a), and the development of tropical cyclones on seasonal time scales (Evan et al. 2006a). More recently, model-based (Yoshioka et al. 2007) and observation-based (Evan et al. 2009; Foltz and McPhaden 2008a) studies have identified a possible important role African dust outbreaks play in modulating surface temperatures of the northern tropical Atlantic, although it is also possible that Atlantic temperatures alter regional circulation patterns in such a way as to change Atlantic dust cover (Wong et al. 2008).

One impediment to understanding the nature of the relationship between African dust outbreaks and the climate of the northern tropical Atlantic is the relatively short observational record of Atlantic dustiness (historical
satellite records that retrieve aerosol optical depth only go back to the late 1970s or early 1980s). Since there is decadal-scale variability in tropical Atlantic Ocean temperatures (Goldenberg et al. 2001) and Atlantic dust cover (Prospero and Lamb 2003), a longer time series of dust is needed to more clearly establish causal relationships between the two and to identify more robust correlations between dust and atmospheric phenomenola like tropical cyclones (Evan et al. 2006a).

Recently Mukhopadhyay and Kreyzik (2008) created a proxy record of atmospheric dust over Cape Verde that is annually resolved and extends from 1955 to 2004, and here we describe a simple model to exploit this dust data to extend back in time to 1955 satellite-based estimated of Atlantic dust optical depth. We then characterize the radiative effect of dust over the northern tropical Atlantic on these time scales. The remainder of this paper is organized as follows. In the next section we describe the data and models used in this study. In section 3 we develop a simple statistical model for using the dust proxy record to extend satellite-based estimates of dust optical depth over the northern tropical Atlantic back in time to 1955 and present results from the model output. In the fourth section we use this reconstructed dust optical depth data to model the magnitude of the dust aerosol direct effect, describing the technique used to do so and then presenting results from the radiative transfer model. We conclude with interpretation of the output from the dust optical depth reconstruction and the radiative transfer model, summarize weaknesses with our methodologies, and discuss future efforts that can improve the accuracy of these results.

2. Data and models

We use an annually resolved historical proxy record for atmospheric dust based on measurements of crustal helium-4 ($^4$He) flux from a Porites coral at a water depth of 5 m near Pedra de Lume on Sal Island (16°45′44″N, 22°53′23″W), part of the Cape Verde archipelago, for the period of 1955–94 (Mukhopadhyay and Kreyzik 2008). $^4$He is produced from radioactive decay of U and Th, elements that are abundant in dust from continental sources (Patterson et al. 1999). The concentration of $^4$He in the coral therefore, reflects the abundance of continental dust incorporated into the coral skeleton structure (Mukhopadhyay and Kreyzik 2008). Minerals in rocks accumulate $^4$He until they are ground down to a size of roughly 20–30 μm, which is the alpha stopping distance. Once particles are broken down to sizes that can be carried by wind, there is no additional accumulation of $^4$He as the alpha particles are ejected out of the dust. Therefore in geologic archives the total amount of $^4$He is directly proportional to the total amount of dust and thus there is no time lag between dust in the archive and $^4$He contents.

We employ historical satellite retrievals of aerosol optical thickness (AOT) from the Advanced Very High Resolution Radiometer (AVHRR) Pathfinder Extended dataset (PATMOS-x). The monthly-mean PATMOS-x 670-nm AOT retrievals (Stowe et al. 1997) span the period of 1982–2008 and are available at a 0.5° horizontal resolution. AVHRR AOT retrievals in dust regions benefit from a dust detection algorithm and have been shown to compare very well against optical depth measurements from Aerosol Robotic Network (AERONET) instruments in the presence of dust (Evan et al. 2006b).

We use monthly-mean 550-nm AOT retrievals and so-called fine-mode-fraction (FMF) estimates from the Moderate Resolution Imaging Spectroradiometer (MODIS) on board the Aqua satellite (Remer et al. 2005), which cover the period of 2002–08 and are available at a 1° horizontal resolution (MODIS products are all level 3 collection 5). Random errors in the MODIS overwater aerosol optical depths are estimated to be ±0.05 (Remer et al. 2005). FMF is a ratio of cloud-free radiances normalized by scattering angle and is proportional to the size of the scattering aerosol in the column (Remer et al. 2005). We convert AOT values into an equivalent dust optical depth $\tau_{dust}$ in a manner consistent with Kaufman et al. (2005b) by using surface winds from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996) to estimate marine aerosol optical depths. We also accounts for stratospheric aerosols with stratospheric optical depth values from Sato et al. (1993), assuming stratospheric aerosol optical depths are zero post-1999.

To calculate all-sky atmospheric radiative forcing by dust we estimate the clear-sky fraction using long-term 3-hourly monthly-mean cloud cover amounts from the International Satellite Cloud Climatology Project (ISCCP) (Rossow and Schiffer 1999). All radiative transfer calculations are made using dust optical properties from Myhre et al. (2003) and dust size distributions from Haywood et al. (2003). We estimate the vertical distribution of $\tau_{dust}$ by scaling the column total $\tau_{dust}$ with profiles of the Saharan air layer from Dunion and Marron (2008). All surface through top-of-the-atmosphere (TOA) radiative flux and heating rate calculations are done using the Streamer radiative transfer model (Key and Schweiger 1998).

3. Methods

First we estimate $\tau_{dust}$ for the period of 1982–2008 using PATMOS-x AOT retrievals (670 nm). Here we
adopt the following equation to derive $\tau_{\text{dust}}$ from MODIS AOT (550 nm) values (Kaufman et al. 2005b):

$$\tau_{\text{dust}} = \frac{\text{AOT}(0.9 - \text{FMF}) - 0.6\tau_{\text{marine}}}{0.4}, \quad (1)$$

where

$$\tau_{\text{marine}} = 0.007W_{10m} + 0.02, \quad (2)$$

and $W_{10m}$ is the mean wind speed (m s$^{-1}$) at 10 m from NCEP reanalysis. To calculate monthly-mean $\tau_{\text{dust}}$, we use monthly-mean values of AOT and FMF and long-term monthly-mean values of $W_{10m}$. We therefore calculate monthly-mean fields of $\tau_{\text{dust}}$ for MODIS using monthly-mean FMF and long-term monthly-mean $W_{10m}$ fields for the period of 2003–08 using AOT data from the 

Aqua satellite. AOT from the MODIS instruments on board the Aqua and Terra satellites exhibit a high level of agreement (Ichoku et al. 2005; Remer et al. 2006), making our choice of Aqua over Terra data somewhat arbitrary. However, Aqua is in an afternoon orbit and Terra is in a morning orbit, and we prefer to use the Aqua data since the Aqua AOT retrievals are made for lower solar zenith angles than those for Terra.

a. Estimating AVHRR $\tau_{\text{dust}}$

We use monthly-mean FMV values for the period 2002–08. As we intend to estimate $\tau_{\text{dust}}$ going back to 1982, it is necessary to calculate and use long-term monthly-mean values of the FMF. Here we assume that using long-term mean FMF values does not introduce large bias or errors into the estimates of $\tau_{\text{dust}}$. A scatterplot of 1° horizontally resolved monthly versus long-term mean monthly MODIS FMF values, over the region of $0^\circ–30^\circ$N, $10^\circ–65^\circ$W, suggests that interannual variability in FMF is relatively small when compared with the seasonal cycle over this region (Fig. 1). The linear least squares best-fit line through the data in Fig. 1 is $y = 0.98x + 0.01$. The bias of climatological FMF when compared with the monthly-mean values is 0.00, and the RMSE is 0.08. Assuming a FMF value of 0.5 (the average over this region), the bias in $\tau_{\text{dust}}$ estimates, based on the RMSE in FMF, would be $\pm 0.05$ for an AOT of 0.2, $\pm 0.10$ for an AOT of 0.5, and $\pm 0.20$ for an AOT of 1.0. Considering that the majority of AOT retrievals are well below 0.5, the error introduced in using an FMF climatology is likely to be smaller than the error in the AVHRR AOT retrievals themselves.

We also calculate MODIS $\tau_{\text{dust}}$ using both monthly and long-term monthly-mean FMF values [Eq. (1)] at a 1° horizontal resolution and over the region of $0^\circ–30^\circ$N, $10^\circ–65^\circ$W. A scatterplot of the monthly and long-term monthly-mean $\tau_{\text{dust}}$ estimates is very similar (Fig. 2), suggesting that year-to-year changes in $\tau_{\text{dust}}$ are dominated by variability in AOT rather than variability in the FMF [Eq. (1)]. The linear least squares best-fit line through the data in Fig. 2 is $y = 1.02x + 0.00$, the bias is 0.00, and the RMSE is 0.05. At very high values of $\tau_{\text{dust}}$ (greater than 1.0) there is a slight bias toward underestimating $\tau_{\text{dust}}$ when using climatological FMF values, but this bias is small ($\pm 0.06$), especially when compared with $\tau_{\text{dust}}$ values greater than 1.

Confident that the use of long-term monthly-mean FMV values does not introduce large errors when calculating $\tau_{\text{dust}}$, we next create an AVHRR $\tau_{\text{dust}}$ by using monthly PATMOS-x AOT values, that have been re-binned to a 1° horizontal resolution, using Eq. (1). We compare 1° horizontally resolved MODIS and AVHRR $\tau_{\text{dust}}$ data, where PATMOS-x $\tau_{\text{dust}}$ uses climatological monthly-mean FMV values and the MODIS $\tau_{\text{dust}}$ uses monthly FMF (Fig. 3). There is considerably less agreement in $\tau_{\text{dust}}$ between sensors, which have different observation times and cloud-clearing techniques. For $\tau_{\text{dust}}$ less than 1.0, where the vast majority of all AVHRR $\tau_{\text{dust}}$ values reside, AVHRR is biased low (0.06) when compared with MODIS. Although MODIS $\tau_{\text{dust}}$ values could be overestimates (Remer et al. 2005), there is no reason to believe that the PATMOS-x retrievals are more accurate. For $\tau_{\text{dust}}$ greater than unity, the AVHRR retrievals are
on average biased very high, although this is not obvious from Fig. 3. It is possible that this high bias is an artifact in the PATMOS-x AOT, or that the application of a dust detection algorithm for the AVHRR (Evan et al. 2006b) has identified dusty pixels that were not included in the MODIS AOT retrievals. However, since very few (less than 0.3%) \( \tau_{dust} \) values are greater than one, this is an acceptable result. The overall bias in AVHRR and MODIS \( \tau_{dust} \) is \(-0.02\) (AVHRR underestimating \( \tau_{dust} \) when compared to MODIS), the RMSE is 0.13, and the linear least squares best-fit line is \( y = 1.04x + 0.02 \) (Fig. 3).

Similar scatterplots of AVHRR and MODIS \( \tau_{dust} \) using monthly-mean FMF for both satellite datasets, and then climatological monthly FMF values for both, provided essentially the same results, as differences in AOT retrievals between MODIS and AVHRR dominate over the error introduced by using a climatological monthly FMF in Eq. (1). Using monthly-mean AVHRR AOT retrievals, and climatological monthly-mean fields of FMF and \( \tau_{marine} \), we create fields of \( \tau_{dust} \) over the region of \( 0^\circ - 30^\circ \) N, \( 0^\circ - 65^\circ \) W and from 1982 to 2008 from Eqs. (1) and (2). Since AOT values are also sensitive to stratospheric aerosols associated with volcanic eruptions, before calculating \( \tau_{dust} \) with AVHRR AOT values we subtract zonally averaged values of volcanic stratospheric aerosol optical depths (Sato et al. 1993), which are dominated by the eruptions of El Chichón and Mount Pinatubo, in a manner consistent with Evan et al. (2009). Missing data are filled in using linear interpolation in time. Having developed a climatology of northern tropical Atlantic \( \tau_{dust} \), we next use the coral proxy record to extend this climatology back to 1955.

**b. Reconstructing \( \tau_{dust} \) from 1955 to 2008**

The coral proxy record for atmospheric dust has been compared to AOT retrievals from the Total Ozone Mapping Spectrometer (TOMS) and dust mass concentrations at Barbados (Mukhopadhyay and Kreyck 2008). Here we compare this same time series of \(^4\)He flux with the time series of AVHRR \( \tau_{dust} \) for the \( 1^\circ \) horizontally resolved region that contains Sal Island. The two time series shows excellent agreement for the common period of 1982–94 (Fig. 4). In particular, comparisons from Mukhopadhyay and Kreyck (2008) show an underestimation of dust from the \(^4\)He flux when compared with Barbados and TOMS data for the period post-1988. However, here we show excellent agreement for the entire span of the record. It is possible that the TOMS data are reflecting stratospheric aerosols from the 1991 Mount Pinatubo eruption, and that this is resulting in the discrepancy between the two. There is no simple explanation for the difference between the coral proxy and Barbados records post-1988; however, a similar divergence has been noted in a comparison using dust fraction data from the AVHRR (Evan et al. 2007).
Since there are only 13 common data points between the $^4$He and AVHRR Sal Island time series, and as there is a strong linear downward trend over most of that period (1982–94), it is difficult to fit a robust model to the data to express $\tau_{dust}$ as a function of $^4$He. Although the proxy data are annually resolved, it is possible that the method for estimating $^4$He underestimates large mineral aerosol particles, and therefore it is necessary to filter the data with a 3-yr running mean to reduce higher-frequency uncertainty in the coral proxy data (Mukhopadhyay and Kreycik 2008). The correlation between the $^4$He and AVHRR Sal Island time series is 0.40 (80% significance level based on the $t$ score of a two-tailed $t$ test) for the annual time series and 0.83 for the smoothed time series (significant at the 99% level). Therefore, we assume that

$$\frac{\tau_{dust} - \overline{\tau_{dust}}}{\sigma_{\tau_{dust}}} = \frac{^4\text{He} - \overline{^4\text{He}}}{\sigma_{^4\text{He}}}$$  \hspace{1cm} (3)$$

where the overbar represents the mean and $\sigma$ the standard deviation, and the series of $^4$He and $\tau_{dust}$ have been smoothed by a 3-yr running mean filter (with first and last values of the annual series removed to avoid edge effects from the filter). The assumption in Eq. (3) is equivalent to expressing both time series in units of standard deviation and assuming that the best-fit line between them is $y = 1.0x + 0.0$ (Fig. 5), putting the $^4$He time series in units of dust optical depth. The 54-yr time series of reconstructed $\tau_{dust}$ at Cape Verde therefore consists of the following: a linear transform of the $^4$He flux from 1955 to 1981, an average of the $^4$He flux and AVHRR $\tau_{dust}$ data from 1982 to 1994, and AVHRR $\tau_{dust}$ only from 1994 to 2008 (Fig. 6). The annual time series of Cape Verde reconstructed $\tau_{dust}$ follows that of the $^4$He and AVHRR $\tau_{dust}$ series (Fig. 5), and a smoothed version of the series shows minimums at the beginning and end of the record and a maximum in the early 1980s (Fig. 6).
We next estimate annual fields of $\tau_{dust}$ spanning the region of 0°–30°N, 10°–65°W for the period of 1955–2008 using the reconstructed 54-yr annual time series of $\tau_{dust}$ from Cape Verde. There is a very high correlation between the annual time series of AVHRR $\tau_{dust}$ at Cape Verde with $\tau_{dust}$ over the northern tropical Atlantic (Fig. 7), in particular over the eastern half of the basin where aerosol loadings are largest (e.g., Kaufman et al. 2005b). We use linear least squares regression to predict $\tau_{dust}$ from 0° to 30°N and from 10° to 65°W from annual-mean AVHRR $\tau_{dust}$ values for the period 1982–2008, and then apply these regression coefficients (Fig. 8) to extend the reconstructed Cape Verde annual $\tau_{dust}$ time series across the basin and for the period of 1955–2008. In other words,

$$\tau_{lat,lon} = a_{lat,lon} \tau_{CV} + b_{lat,lon},$$

where $\tau_{lat,lon}$ is the annual time series of $\tau_{dust}$ at some location over the northern tropical Atlantic, $a_{lat,lon}$ and $b_{lat,lon}$ are the regression coefficients for the same location, and $\tau_{CV}$ is the annual time series of $\tau_{dust}$ at Cape Verde. West of 50°W the correlation coefficients (Fig. 7) are on the order of 0.5 or smaller, and therefore there may be substantial errors in the linear regression. We estimate the error in our reconstructed dust product section on model results.

Then next step is to convert the 54 years of annual-mean fields of northern tropical Atlantic $\tau_{dust}$ to monthly resolved ones. While we experimented with several linear methods to accomplish this, we decided to essentially resample the monthly AVHRR $\tau_{dust}$ values. For each 1° horizontally resolved location, and for each year in the reconstructed time series (1955–2008) we found the annual-mean AVHRR $\tau_{dust}$ value (over the period 1982–2008) that was the closest to the reconstructed annual-mean $\tau_{dust}$ value. We then assumed that the annual cycle of $\tau_{dust}$ could be described as the annual cycle of $\tau_{dust}$ from closest (with respect to the annual mean) AVHRR year, plus a small offset applied equally to each month to ensure that the annual-mean value from the reconstructed time series was preserved.

We test this assumption by randomly splitting the 27-yr AVHRR dataset into a 12-yr training and 13-yr testing dataset. We then apply the resampling technique to the testing dataset annual values of $\tau_{dust}$, using the training data annual cycles, and then compare results against the withheld (testing) AVHRR monthly data. We repeat this test 100 times for the northern tropical Atlantic to generate a distribution of mean monthly RMSE and bias statistics. When averaging over the distribution of test cases, this method for estimating the annual cycle of $\tau_{dust}$
exhibits no bias, and monthly-mean RMSE values fall between 0.05 and 0.15 units of optical depth, with the highest RMSE of the 100 test cases being less than 0.20 (Fig. 9). We apply this technique to the 54-yr reconstructed $\tau_{\text{dust}}$ dataset using all 27 years of AVHRR data as the final step in converting the Cape Verde proxy record of atmospheric dust into monthly fields of $\tau_{\text{dust}}$ for the period of 1955–2008, and over the region of $0^\circ–30^\circ$N, $0^\circ–65^\circ$W.

c. Model results

Comparison of the new monthly, reconstructed $\tau_{\text{dust}}$ with AVHRR (1982–2008) and MODIS (2002–08) shows excellent agreement (Fig. 10). Comparison with AVHRR data gives a near-zero bias (0.002) and a RMSE of 0.11, with the monthly distribution of the RMSE closely following that in Fig. 9. Comparison with MODIS data also shows excellent agreement, with a mean bias of 0.006 and a RMSE of 0.13, where the mean monthly RMSE changes very little from month to month (range of 0.11–0.15). Therefore, a rough error estimate for this method of historical reconstruction of $\tau_{\text{dust}}$ is to assume error bars of ±0.1 units of optical depth. We note that the scatterplots in Fig. 10 indicate uncertainty in reconstructed $\tau_{\text{dust}}$ that results from the statistical methods used in the reconstruction’s statistical methods.

Spatially, monthly-mean reconstructed $\tau_{\text{dust}}$ is biased high when compared to monthly-mean MODIS $\tau_{\text{dust}}$ west of $45^\circ$W and is biased low east of this longitude, but the magnitude of the bias is generally less that 0.1 (Fig. 11). When compared with monthly-mean MODIS $\tau_{\text{dust}}$, the RMSE of reconstructed $\tau_{\text{dust}}$ is less than 0.2 over most of the northern tropical Atlantic, although close to the coast of West Africa the RMSE values are greater than 0.2 (Fig. 11). The spatial patterns of the errors in $\tau_{\text{dust}}$ in general follow the field of long-term mean $\tau_{\text{dust}}$, where
Considering the number of assumptions made in our model of reconstructed $\tau_{dust}$, these error statistics demonstrate that over our region of interest the errors in our estimation of $\tau_{dust}$ are of a similar magnitude to errors in the aerosol retrievals themselves. However, the west–east gradient in the bias suggests that future improvements in the reconstructed $\tau_{dust}$ dataset may be made by replacing the linear regression [Eq. (4)] with a nonlinear method for estimating the annually resolved fields of $\tau_{dust}$. We note that our error analysis only considers errors related to our statistical methods for creating a historical $\tau_{dust}$ product and do not take into account errors in AOT retrievals from the AVHRR or MODIS satellites. We refer the reader to Evan et al. (2006b) and Remer et al. (2005) for evaluation of the AVHRR and MODIS AOT products, respectively.

In the remainder of this paper, we refer to our reconstructed monthly fields of dust optical depth as $\tau_{dust}$. The field of long-term mean $\tau_{dust}$ follows the pattern of Atlantic dust cover (Kaufman et al. 2005b), and exhibits the highest values (greater than 0.3) in the eastern half of the basin and between the latitudes of 5° and 20°N, with maximums (greater than 0.5) near the coast of West Africa (Fig. 12). There are also two small regions with large $\tau_{dust}$ values that are off the eastern coast of Brazil, which result from satellite-retrieved AOT artifacts due to increased sediments associated with river discharge.

The annual time series of $\tau_{dust}$, averaged over the area where mean values are the highest (5°–20°N and east of 50°W) (Fig. 13) closely follows the series of $^4$He and AVHRR $\tau_{dust}$ at Cape Verde (Fig. 4). The annual-mean minimum of 0.24 occurs in 2004, and the years of 1955–57, 1964, 1976, and 2005 all show annual-mean $\tau_{dust}$ values that are lower than 0.25. The maximum annual-mean $\tau_{dust}$ of 0.37 occurs in 1984, and the years of 1962, 1983–85, and 1987 all have values greater than 0.32. With the maximum in annual-mean $\tau_{dust}$ found in the middle of the record, and

FIG. 11. Maps of bias and RMSE in reconstructed $\tau_{dust}$. Error statistics are based on comparisons with monthly-mean $\tau_{dust}$ values from MODIS for the period 2002–08.

FIG. 12. Map of the long-term mean $\tau_{dust}$ for the period 1955–2008. Climatology is based on output from the statistical model for extending satellite estimates of $\tau_{dust}$ back in time using the Cape Verde $^4$He proxy record for atmospheric dust.
minimums occurring at the beginning and ends, there is an apparent increasing (1955–84) and then decreasing (1984–2008) trend in Atlantic dustiness. The range of annual-mean $\tau_{\text{dust}}$ is relatively small, but monthly values (after the annual cycle is removed) span nearly 0.5 units of optical depth (Fig. 13). Annually, minimums in $\tau_{\text{dust}}$ change little from year to year, but maximums span 0.3–0.6 units of optical depth, and the years of minimums and maximums in monthly $\tau_{\text{dust}}$ follows those of the annual series.

The distribution of monthly-mean $\tau_{\text{dust}}$ ($5^\circ$–20$^\circ$N and east of 50$^\circ$W) is bimodal, with an absolute maximum occurring at 0.30 units of optical depth and a relative maximum near 0.15 (Fig. 14). The secondary maximum at 0.15 results from variability in wintertime [January–March (JFM)] $\tau_{\text{dust}}$ (not shown). There is also a very small local maximum near 0.55 that results from the very high spring and summer $\tau_{\text{dust}}$ values during the early 1980s. The fact that the histogram of monthly-mean $\tau_{\text{dust}}$ is bimodal suggests that a simple model assuming normally distributed monthly-mean dust values will not be able to effectively represent the processes responsible for year-to-year and month-to-month variability in Atlantic dust cover.

4. Dust radiative forcing

We next explore the magnitude and variability of the dust-aerosol direct effect by estimating the radiative impact of suspended dust across the northern tropical Atlantic. To do this we follow the methods of Evan et al. (2009) except that we exclusively use the Streamer radiative transfer model (Key and Schweiger 1998) for all radiative transfer calculations, employing 4 streams for long- and shortwave calculations and 24 phase function Legendre coefficients. There are errors associated with any attempt to estimate the direct aerosol effect that result from assumptions about aerosol optical properties, the vertical distribution of aerosols and water vapor, fractional cloud cover, and spherical assumptions for the dust particles. We have attempted to be clear about our methodology and the potential sources of error, and given the data currently available we believe the radiative estimates presented here represent the best attempt to estimate this process over these long time scales.

Here we provide error estimates for the radiative forcing calculations that are based on climatological monthly maps of the RMSE between $\tau_{\text{dust}}$ and MODIS $\tau_{\text{dust}}$ (not shown). We perform all radiative forcing calculations using $\tau_{\text{dust}}$, $\tau_{\text{dust}}$ plus the RMSE, and $\tau_{\text{dust}}$ minus the RMSE, and present the spread of those estimates as one way of quantifying uncertainty. It is important to point out that our methods for estimating the dust direct radiative effect provide results that are in agreement with independent estimates of dust radiative forcing from models (Evan et al. 2008; Yoshioka et al. 2007) and satellites (Foltz and McPhaden 2008b; Wong et al. 2009).

a. Radiative transfer setup

As was done in Evan et al. (2009), all column-integrated values of $\tau_{\text{dust}}$ are distributed vertically according to mean...
profiles of the Saharan air layer (Fig. 15) as identified by Dunion and Marron (2008). Here dust concentrations are highest between 700 and 500 hPa and fall away quickly below and above this altitude. It is possible that model-based profiles of atmospheric dust are more accurate than the profile used here, although our dust profile does show many similarities to those from the Model of Atmospheric Transport and Chemistry (Wong et al. 2009). Data from the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation platform can also be used to create an observational climatology of Atlantic dust profiles, and future efforts will explore this possibility.

We estimate clear-sky (cloud free) dust radiative forcing and heating rates by differencing Streamer model output from runs with and without dust. Although dust does emit and absorb longwave radiation, shortwave scattering is by far the dominate effect at the surface. When the sun is below the horizon there is weak positive surface forcing (on the order of 10 W m\(^{-2}\) for \(\tau_{\text{dust}}\) of 1.0), and for small solar zenith angles there is a very strong negative surface forcing (\(-200\) W m\(^{-2}\) for the same \(\tau_{\text{dust}}\)) (Fig. 16). Therefore, dust has an overwater net negative (downward) forcing at the surface (e.g., Evan et al. 2008; Myhre et al. 2003). TOA forcing has a larger nighttime positive forcing, relative to the daytime negative forcing (20 to \(-60\) W m\(^{-2}\) for \(\tau_{\text{dust}}\) of 1.0), but the diurnally averaged effect for all \(\tau_{\text{dust}}\) values is negative TOA forcing (Fig. 16).

To estimate all-sky dust forcing (i.e., accounting for cloud cover), we scale all forcing values by the cloud-free fraction of sky at each location, every 3 h, assuming that radiative forcing in the presence of both dust and clouds is dominated by the meteorological clouds. Therefore, the magnitude of the aerosol direct effect is inversely proportional to the total cloud cover. Since we use long-term monthly-mean 3-hourly cloud cover data from ISCCP (Rossow and Schiffer 1999), we have no way to account for interannual changes in cloudiness and the effects on year-to-year variability of the aerosols direct
effect. It is possible to use 4-hourly cloud cover from NCEP reanalysis to account for cloud cover changes from one year to the next. However, the climatological spatial pattern of cloudiness over the Atlantic from NCEP data is very dissimilar to that from ISCCP (Fig. 17). In particular, the NCEP cloud data do not resolve the intertropical convergence zone well and have a maximum in cloud cover near the coast of West Africa and at 15°N (Fig. 12). The cause of discrepancies between NCEP and ISCCP cloud cover over the northern tropical Atlantic may be related to the strength of the Hadley circulation in the reanalysis dataset (Trenberth and Guillemot 1998).

It is important to note that Atlantic dust outbreaks are accompanied by a midlevel dry air and warm anomaly (e.g., Dunion and Marron 2008; Karyampudi et al. 1999) and that these environmental features also contribute to change in radiative fluxes at the surface and throughout the column (Wong et al. 2009). However, here we are only considering the aerosol direct effect and are not accounting for the radiative effects of the Saharan air layer as a whole.

b. Dust forcing results

We apply our method for estimating the dust radiative forcing from monthly \( \tau_{\text{dust}} \) values to our 54-yr dataset. The long-term mean spatial pattern of surface radiative forcing (Fig. 18) generally follows that of \( \tau_{\text{dust}} \) (Fig. 12), except that the forcing values are weak closer to the equator, where cloud cover is pervasive (Fig. 17). The strongest forcing values of less than \(-10\ \text{W m}^{-2}\) are located close to the coast of West Africa and from 8° to 25°N, and in general forcing values at these latitudes are on the order from \(-4\) to \(-8\ \text{W m}^{-2}\) westward to the Caribbean. The spatial patterns of surface forcing do not change when considering errors in \( \tau_{\text{dust}} \), but the magnitude of the direct effect does vary considerably. Dust surface forcing is less than \(-10\ \text{W m}^{-2}\) across much of the northern tropical Atlantic for the case of \( \tau_{\text{dust}} \) minus the RMSE, and between \(-2\) and \(-6\ \text{W m}^{-2}\) for the case of \( \tau_{\text{dust}} \) minus the RMSE (Fig. 18).

A map of TOA forcing shows a very similar spatial pattern to the surface forcing, but TOA values are about one-quarter to one-third the magnitude of those at the surface (Fig. 19). TOA forcing estimates are negative, even when considering errors in \( \tau_{\text{dust}} \), with forcing values ranging from \(-1\) to \(-3\ \text{W m}^{-2}\) across much of our region of interest for \( \tau_{\text{dust}} \) minus or plus the RMSE, respectively (Fig. 19).

Annual-mean values of dust surface forcing, averaged over 5°–20°N, 10°–50°W, range from \(-5.0\) to \(-6.5\ \text{W m}^{-2}\), with the strongest forcing occurring in the early 1980s and the weakest forcing in the beginning and ends of the 54-yr record (Fig. 20). Errors in \( \tau_{\text{dust}} \) only slightly change the temporal variability in the forcing time series but do add offsets of \(\pm2.5\ \text{W m}^{-2}\). Monthly surface forcing values, with the annual cycle removed, range from \(-3.0\) to \(-9.0\ \text{W m}^{-2}\), where values near \(-9.0\ \text{W m}^{-2}\) occur frequently throughout the period of 1970–95. A similar time series of TOA forcing shows less year-to-year variability in the annual-mean values, with annual values ranging from \(-1.5\) to \(-2.0\ \text{W m}^{-2}\) and monthly values (also with the annual cycle removed) from \(-1.0\) to \(-2.5\ \text{W m}^{-2}\) (Fig. 20). However, overall the TOA and
FIG. 18. Long-term mean (1955–2008) all-sky surface radiative forcing (downward) by dust. (top) The climatological average, (middle) the upper estimate based on RMSE in reconstructed $\tau_{\text{dust}}$, and (bottom) the lower estimate based on RMSE in reconstructed $\tau_{\text{dust}}$.

FIG. 19. As in Fig. 18, but for long-term mean (1955–2008) all-sky TOA radiative forcing (downward) by dust.
surface forcing series follow the same general pattern of dust forcing throughout the 54-yr record. As with the surface forcing series, errors in $\tau_{\text{dust}}$ tend to result in an offset in the magnitude of dust forcing ($\pm 0.7 \text{ W m}^{-2}$) but do not alter the year-to-year variability in TOA forcing.

The climatological annual cycle of dust radiative forcing at the surface and over the region of $5^\circ$–$20^\circ$N, $10^\circ$–$50^\circ$W varies from the November minimum (in magnitude) of $-0.5 \text{ W m}^{-2}$ to the July maximum of $-15 \text{ W m}^{-2}$ (Fig. 21). Climatological monthly values of dust surface forcing are small (from $-1$ to $-5 \text{ W m}^{-2}$) in the months of January–April and October–December, and large (from $-12$ to $-15 \text{ W m}^{-2}$) from June–August. Uncertainties in the climatological mean surface forcing that result from errors in $\tau_{\text{dust}}$ are $\pm 5 \text{ W m}^{-2}$ during the summer months and $\pm 1 \text{ W m}^{-2}$ during the winter and fall months (Fig. 21). The range of monthly values is on the order of $\pm 1$–$4 \text{ W m}^{-2}$ above or below the climatological mean. The climatological annual cycle of TOA forcing follows the same pattern as the surface forcing, except that monthly values are near $0 \text{ W m}^{-2}$ during January–April and October–December and from $-4$ to $-5 \text{ W m}^{-2}$ during

**Fig. 20.** Time series of (left) surface and (right) TOA forcing by dust. Monthly- (gray line) and annual-mean (black line) time series of dust surface forcing are for the period 1955–2008 and averaged over the region of $5^\circ$–$20^\circ$N and east of $50^\circ$W. Monthly-mean values are shown without the seasonal cycle. The dashed lines indicate uncertainties in the annual-mean time series associated with errors in $\tau_{\text{dust}}$.

**Fig. 21.** Climatological seasonal cycle of dust (left) surface and (right) TOA forcing. Climatological mean values are based on the period 1955–2008 and all data are averaged over the region of $5^\circ$–$20^\circ$N and east of $50^\circ$W. The gray shaded region represents the range of monthly-mean values, and the black line is the long-term mean. The dashed lines indicate uncertainties in the climatological forcing associated with errors in $\tau_{\text{dust}}$. 
the summer months. Month mean values are on the order of $\pm 1$–2 W m$^{-2}$, as are the forcing uncertainty associated with errors in $\tau_{\text{dust}}$ (Fig. 21).

The long-term mean profile of dust radiative forcing (also averaged over the region of $5^\circ$–$20^\circ$N, $10^\circ$–$50^\circ$W) shows the strongest negative forcing value of $-5.5$ W m$^{-2}$ at the surface, which decreases to $-2.0$ W m$^{-2}$ at 200 hPa and remains at this value through the top of the atmosphere (Fig. 22), consistent with recent, independent estimates of the dust direct effect (Wong et al. 2009). All monthly-mean forcing profiles follow this same pattern throughout the atmosphere, owing to the profile of dust we are using in our radiative transfer calculations (Fig. 15). The values of dust forcing decrease (in magnitude) from the surface to 150 hPa, remain constant through 50 hPa, and slightly increase (in magnitude) through the top of the atmosphere (Fig. 22). Uncertainties in the climatological forcing profile that are associated with errors in $\tau_{\text{dust}}$ are $\pm 2.5$ W m$^{-2}$ at the surface and $\pm 1$ W m$^{-2}$ at 200 hPa to the top of the atmosphere.

The seasonal cycle of dust forcing profiles (Fig. 23) are largely predictable based on TOA and surface forcing annual cycles (Fig. 21); surface forcing values are strongest during the summer months and weakest during the boreal autumn and winter, and TOA forcing is between zero and $-6.0$ W m$^{-2}$. Less apparent, though, is that the mean annual cycle of dust forcing does not begin to noticeably deviate from that at the surface until a height of 2 km; at a height of 8 km the range of the mean annual cycle is less than 8 W m$^{-2}$, and at 15 km it is less than 6 W m$^{-2}$.

c. Atmospheric heating rates

We also briefly describe estimates of the dust-forced atmospheric heating rate based on output from our radiative transfer model. Monthly-mean heating rates are constructed in the same manner as the all-sky forcing estimates. It is important to note that these all-sky heating rates are a regional ($5^\circ$–$20^\circ$N and east of $50^\circ$W) average based on the assumption that the dust-heating rate is zero when water clouds are present.

The mean profile of the climatological dust-forced atmospheric heating rate is roughly 0 K day$^{-1}$ just above the surface, has a maximum of nearly 0.05 K day$^{-1}$ at heights between 350 and 500 hPa, is 0 K day$^{-1}$ at roughly 60 hPa (where the heating rate changes sign), and is a minimum of $-0.01$ K day$^{-1}$ at 25 hPa (Fig. 24). Monthly profiles deviate little from the general shape of the climatological profile, owing to the single atmospheric profile of dust and standard Streamer “tropical” atmospheric profile of temperature and moisture that

![Image of dust forcing profiles](image-url)
we use for all radiative transfer calculations. The range of the monthly-mean heating rates is from 0.17 K day$^{-1}$ in the midtroposphere to $-0.03$ K day$^{-1}$ near the top of the atmosphere. The patterns of the monthly values also reveal the bimodal nature of Atlantic dust outbreaks (Fig. 14), with two distinct groupings of heating rates visible in the troposphere.

The magnitude our heating rate profiles are consistent with those of Wong et al. (2009), except that the heating rate profiles in Wong et al. (2009) are a maximum below 800 hPa, and go to 0 K day$^{-1}$ much lower in the atmosphere. However, the Wong et al. (2009) results do consider a smaller portion of the Atlantic and look at dust-heating rates over a much shorter period of time, so this level of agreement is encouraging. Uncertainty in the climatological heating rate profile, associated with errors in $\tau_{\text{dust}}$, is between ±0.01 and 0.02 K day$^{-1}$ (Fig. 24).

The dust-forced atmospheric heating rate shows a distinct annual cycle in the atmospheric temperature response to African dust outbreaks over the Atlantic (Fig. 25). A maximum heating rate of greater than 0.12 K day$^{-1}$ is found during June–July at heights of 300–600 hPa. At this same altitude, the heating rate quickly falls off to less than 0.03 K day$^{-1}$ in September and does not begin to increase until April. Lower in the atmosphere the heating rate is less than 0.03 K day$^{-1}$, and above 50 hPa the heating rate is negative and greater than $-0.03$ K day$^{-1}$.

Although these all-sky dust-heating rates are small, we are averaging over a large region of the Atlantic and are assuming a zero heating rate throughout the column when meteorological clouds are present. Regionally we do show areas where more significant atmospheric heating is present. For example, the long-term August mean dust-heating rate at 6-km height is greater than 0.25 K day$^{-1}$ near the coast of West Africa and consistently greater than 0.10 K day$^{-1}$ spanning the Atlantic at latitudes of 12$^\circ$–22$^\circ$N (not shown).

5. Conclusions

Here we develop a methodology to use atmospheric dust proxy data from Cape Verde to estimate tropical northern Atlantic dust cover from 1955 to 2008. We demonstrated that a simple statistical model could well reproduce month-to-month dust optical depths across the oceanic region of 0$^\circ$–30$^\circ$N, 15$^\circ$–65$^\circ$W (Fig. 10), using as input long-term monthly-mean values of 10-m winds from reanalysis, satellite-based estimates of aerosol fine-mode fraction and optical depth, and annually resolved $^4$He flux from a Cape Verde coral record. This new 54-yr record presents an opportunity to study, at a high temporal and spatial resolution, the various roles of aeolian dust in the climate system on a time scale that is long enough to resolve decadal-scale variability.

Using this historical reconstruction of dust optical depths, we found that the spatial distribution of the
long-term mean dust optical depth had a maximum near 0.5 that was located off the coast of West Africa and from 10° to 20°N, and optical depths greater than 0.2 were located as far 50°W (Fig. 12). Time series analysis reveals that during the 1950s and the 2000s annual-mean northern tropical Atlantic dust optical depth was a minimum (less than 0.25) and dust optical depth was at a maximum during the early 1980s (greater than 0.35, Fig. 13). One interpretation of this pattern of dustiness is that there exists a 50-yr cycle to Atlantic dust cover. However, without a longer time series that can resolve more than one realization of this cycle, it is difficult to qualify the 50-yr cycle in Atlantic dustiness as a robust result of our analysis. From our reconstructed dust record, the histogram of monthly-mean Atlantic dust optical depth is bimodal, with a very small third maximum in dustiness at high values of dust optical depth (Fig. 14). This nearly trimodal distribution of dustiness suggests that simple linear models may not be useful for identifying drivers of—or predicting future variability in—Atlantic dust cover.

From 1955 to 2008 the range of monthly-mean dust optical depth averaged over the northern tropical Atlantic was 0.5 (Fig. 13), and such large swings in optical depth, especially averaged over a large region, suggest that dust variability may have an important effect of the regional energy budget. We explored the effect of dust on long and shortwave radiation using a radiative transfer model to calculate dust-forced changes in atmospheric and surface fluxes. Over the last 54 years dust surface forcing varied strongly in both space and time. Long-term mean dust surface forcing values were on the order of $-6 \text{ W m}^{-2}$ across the northern tropical Atlantic, with values stronger than $-10 \text{ W m}^{-2}$ within 15° of longitude of the coast of West Africa (Fig. 18). TOA forcing also exhibited the same spatial pattern, but with values roughly one-third of that at the surface: $-2.0 \text{ W m}^{-2}$ toward the Caribbean and less than $-2.5 \text{ W m}^{-2}$ off the coast of West Africa (Fig. 19). Monthly-mean values of dust surface forcing spanned from $-9$ to $-3 \text{ W m}^{-2}$ (with the seasonal cycle removed), but annually averaged values only exhibited a range of 1.5 $\text{ W m}^{-2}$ (Fig. 20). At the surface and TOA dust radiative forcing has a strong annual cycle, with maximums in magnitude occurring in June–August and near-zero minimums forcing during October–February (Fig. 21).

We also calculated the profiles of dust forcing and the resultant heating rates, although these results are less certain because of the single profile for dust optical depth used here (Fig. 15). From the surface through a height of 200 hPa dust forcing goes from a maximum (in magnitude) to the profile minimum (in magnitude), which is more or less maintained through to the TOA (Fig. 22). Seasonally the magnitude of the profile varies in a manner consistent with the annual cycle of regional mean surface and TOA forcing, with little change in the shape of the forcing profiles (Fig. 23). This pattern of forcing with height results in dust-forced heating of the troposphere, and a slight cooling above 50 hPa (Fig. 24). There is also a well-defined annual cycle of the profile of atmospheric heating by dust (Fig. 25), and local long-term averaged heating rates for the summer months can exhibit values on the order of 0.25 K day$^{-1}$ at some elevations.

Analysis of our long-term dust record and investigation into the characteristics of dust radiative forcing over the last 54 years show that mineral aerosols are an important part of the climate of the tropical North Atlantic, with variability on many different spatial and temporal scales. The study of aeolian dust is inherently interdisciplinary since dust storms are coupled to the atmosphere, land, and oceans. While future work will aim at improving this historical reconstruction of Atlantic dust optical depth, we anticipate that this new multidecadal dataset can be useful in investigating several facets of aeolian dust and its relationship to the climate of West Africa and the northern tropical Atlantic.

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