Response of the eastern subtropical Atlantic SST to Saharan dust: A modeling and observational study

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Retrievals of aerosol optical depth (AOD) from MODIS, and of sea surface temperature (SST) from TMI are analyzed jointly with the output of a numerical model for the period 2000–2006 to determine the impact of Saharan dust on the eastern subtropical North Atlantic SST. Simultaneously with, or shortly after strong dust outbreaks, a decrease in SST of 0.2° to 0.4°C can be observed in the microwave SST data set, which is consistent with an independent estimate of SST decrease simulated here by a local mixed layer model. However, low wind conditions and a shallow mixed layer are required to reach this response, and it is therefore unlikely that a clear response of SST to dust lasting more than a few days can be seen in the microwave SST observations. An inspection of microwave SST observations suggests that about 30% of SST variance could be explained by dust-induced cooling in our study region that is not represented in existing AVHRR SST fields nor represented in reanalysis centers–provided surface heat fluxes. On longer time scales, a comparison between observed SST fields and simulated SST, using an eddy-permitting model of the North Atlantic, suggests a cooling of about 0.5°C on the local SST on sub–seasonal to interannual time scales which is significantly correlated and consistent with a dust-induced cooling. However, while supportive of the hypothesis that Saharan dust lead to a reduction in SST, the eddy–resolving model results are not by themselves conclusive. Moreover, the effects of dust–induced cooling on simulations of the ocean circulation, on atmospheric forecasts and on climate simulations remains to be investigated in future studies.


1. Introduction

Mineral dust and aerosols originating from wind-induced erosion of soil in arid regions can impact the climate system in many ways, e.g., by altering weather or by affecting atmospheric chemistry. As an example, mineral dust and the very dry Saharan Air Layer (SAL), with which it is associated, have been shown to affect the development of clouds and precipitation, as well as modulating thunderstorm activities and tropical cyclogenesis [Sassen et al., 2003; Dunion and Velden, 2004; Yoshioka et al., 2007; Wu, 2007; Evan et al., 2008]. Mineral dust, in particular, by increasing the attenuation (backward scattering and absorption) of solar radiation in the atmosphere, can lead to a redistribution of radiative heating from the surface upward into the dust layer [e.g., Miller and Tegen, 1999] and to a decreased shortwave irradiance at sea level. By changing the atmospheric opacity, mineral dust can thus alter the shortwave radiative forcing at the surface of the ocean [Jickells et al., 2005; Evan et al., 2009]; in addition, thermal emissions from dust aerosols can increase the surface longwave forcing [Vogelmann et al., 2003; Zhu et al., 2007]. Both processes play a role in the ocean mixed layer heat budget and can therefore affect the climatologically important sea surface temperature (SST). In principle, the ocean circulation and the transport of properties can also change in response to enhanced mineral dust or aerosol concentrations. This would be caused through dust-induced differential heating of the upper ocean and resulting thermal wind changes which can in turn affect the mixed layer heat budget and circulation [Marzeion et al., 2005]. Moreover, once settled on the oceans surface, mineral dust can act as a fertilizer (either directly or via stimulation of nitrogen fixation) potentially enhancing biological productivity, and can thereby change the composition of phytoplankton [Erickson et al., 2003; Coale et al., 2004; Boyd, 2007; Blain et al., 2007].

As compared to the rest of the World, the erosion of the Saharan soil is by far the largest annual source of mineral dust aerosols [Prospero et al., 2002; Washington et al., 2003], resulting in a deposition of more than 40% (200–260 Tg yr⁻¹) of the global dust into the North Atlantic Ocean [Kaufman et al., 2005; Mahowald et al., 2005, and references therein].

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The SAL can be traced far over the western Atlantic [Prospero and Carlson, 1981; Colarco et al., 2003] and associated with this layer are often enhanced concentrations of mineral dust. As an example, Lau and Kim [2007] reported a significant increase in Saharan dust over the western North Atlantic during 2006 and argued that this enhanced dust concentration might have been responsible for a cooling of the ocean’s surface there, during the same year, by scattering more solar radiation to space and depriving the surface of some solar heating. Knowing the impact of Saharan dust on the circulation of the eastern subtropical Atlantic and on associated transports of nutrients is therefore specifically important for understanding the climatic relevance of Saharan dust in terms of ocean circulation, air–sea interaction, and biological productivity.

Figure 1. MODIS Rapid Response System 2-km resolution true color image off the west coast of Africa, acquired on March 4, 2004 at 11:55 UTC on board the EOS Terra platform.

\[5\] An example of a Saharan dust outbreak is shown in Figure 1, illustrating an intense dust storm lasting 5 days during early March 2004, as seen by the Moderate Resolution Imaging Spectroradiometer (MODIS) on board of the US NASA Earth Observing System (EOS) Terra satellite. Figure 1 clearly reveals the large spatial extent of a dust layer, which typically can cover an area of about 550 km in the zonal direction and 1750 km meridionally between the African coast and the Canary Islands. Radiometric measurements taken quasi-simultaneously underneath this dust layer during the trans-Atlantic Aerosol and Ocean Science Expeditions (AEROSE) suggest surface radiative forcing anomalies of approximately \(-100 \text{ W m}^{-2}\) [Morris et al., 2006], equivalent to a 40% reduction of shortwave heating at the surface. Given these strong negative shortwave heating anomalies, a strong thermal response of the upper ocean to Saharan dust should be expected (a radiative forcing anomaly of \(-100 \text{ W m}^{-2}\) over 15 days would cool a 30-m thick mixed layer by approximately 1°C).

\[6\] Previously, atmospheric model simulations had suggested that underneath dense atmospheric mineral dust layers the SST may decrease by as much as 1°C [Miller and Tegen, 1998]. In addition, recent works from Yoshioka et al. [2007], Foltz and McPhaden [2008] and Evan et al. [2008] observe that changes in aerosol cover have a significant impact on the Atlantic SST. Yet, using the existing observational database, a quantitative estimation of how much of the observed ocean surface cooling is dust-induced is still missing.

\[7\] The overall goal of this study is accordingly to perform a simultaneous analysis of several satellite data sets to investigate whether Saharan mineral dust leads to a cooling of SST of the eastern North Atlantic, both during strong storms and at seasonal time scales. At the same time, we attempt to quantify how much of the observed SST changes are dust-induced, using for that purpose a one-dimensional (1D) local mixed layer model of the upper ocean as well as a full three-dimensional (3D) eddy-permitting circulation model of the North Atlantic. While the eddy-permitting model will give us the full perspective of the dynamical changes in SST that can be expected in our study area, the 1D perspective will be used to provide a measure of a local change in SST that can be expected in response to a local dust-induced perturbation.

\[8\] The remainder of the paper is organized as follows. In Section 2 we describe the data used in the present work. Aerosol radiative forcing anomalies will be shown in Section 3. The relation between the observed SST and Aerosol Optical Depth (AOD) anomalies will be discussed in Section 4 and numerical simulations of SST anomalies as they should result from observed dust loads are presented in Section 5. A joint interpretation of observed AOD and SST anomalies and model simulations of SST is provided in Section 6. A discussion and concluding remarks are given in Section 7.

2. Data and Methodology

\[9\] Our study is based on a suite of parameters measured from satellites between 2000 and 2006, such as all-weather microwave SST measurements and retrievals of AOD of the atmosphere taken in the visible. In addition, we used several parameters measured during the AEROSE expedition.
conducted from aboard the NOAA Ship Ronald H. Brown in the eastern subtropical North Atlantic.

2.1. MODIS AOD

[10] AOD time series provide a measure of Saharan mineral dust concentrations and can be used to compute the anomaly of solar shortwave irradiance at sea level underneath the dust layer. Daily estimates of AOD are available from the MODIS instrument on board the Terra and Aqua satellites [Remer et al., 2005, 2006]. These data are used here to determine the frequency and intensity of Saharan dust outbreaks over the eastern subtropical North Atlantic, as well as to estimate the associated anomalies in solar forcing of the ocean. For our study we used the daily gridded AOD Level-3 MODIS Terra Collection 5 product, available on a 1° × 1° spatial resolution for the period February 2000 to December 2006 (ftp://ladsweb.nascom.nasa.gov). During the beginning of the dust outbreak event shown in Figure 1, AOD values as large as 4.5 were observed; during its further evolution the main axis of the plume traveled less than a few degrees westward, while its optical depths decreased to 0.5 on March 9, suggesting primarily dry deposition of dust onto the ocean surface [Kaufman et al., 2005; Schepanski et al., 2009].

[11] Daily global fields of AOD contain gaps that arise due to various reasons (e.g., sun glint, clouds and/or bright underlying surfaces). Dealing with those gaps in our analysis is difficult and to avoid them we produced weekly composites, which still preserve the major Saharan dust outbreaks. The composites were computed by area-weighted averaging of all observations within each of the 1° × 1° grid cells.

[12] We identified Saharan dust outbreaks from the daily MODIS AOD product and from the computed weekly AOD anomalies (relative to a climatological seasonal cycle), when values exceeded the AOD time-mean plus twice the local standard deviation \((\langle AOD \rangle + 2\sigma) > 0.85\) and \(>0.30\), for daily and weekly fields, respectively. From Figure 2, several strong dust outbreaks can be detected during the period 2000–2006 between Cape Verde and the Canary Islands (4 in summer, 7 in winter, 1 in spring and 2 in autumn). It is also obvious from Figure 2 that daily data would lead to much higher AOD values during dust storm events and that some of those are eliminated in the weekly averages. However, because our focus here is primarily on the ocean’s long-term response to dust, the outbreaks with longer duration are the most relevant events and these will still be captured in the weekly averages.

2.2. TMI SST

[13] While AVHRR (Advanced Very High Resolution Radiometer) and MODIS infrared (IR) SST retrievals are significantly impacted by the presence of mineral dust in the atmosphere, SST retrievals from microwave (MW) radiometry are not [May et al., 1992; Wentz et al., 2000; Chelton and Wentz, 2005]. For our study, microwave SST retrievals from the TRMM Microwave Imager (TMI) radiometer on board the Tropical Rainfall Measuring Mission (TRMM) satellite were used as weekly SST fields at a spatial resolution of 0.25° over the period January 1998 to December 2006. The data set is described in detail by Wentz [1998]. In addition to the TMI data set, and partly to test results from TMI, we used the version 5 of the SST data available from the Advanced Microwave Scanning Radiometer-EOS (AMSR-E) [Kawanishi et al., 2003] on board of the EOS-Aqua satellite.

[14] We are interested in the question whether Saharan dust (outbreaks and/or seasonal cycle) and its associated cooling can lead to cold SST anomalies in the eastern North Atlantic and to what extent they can be detected in existing satellite SST data sets. To answer both questions one needs a reliable satellite SST database that is not affected in its retrieval procedure by the dust. This holds for microwave SST observations and what follows is therefore mostly based on the TMI SST data set. Stammer et al. [2003] tested the quality of

Figure 2. (top) Daily retrievals and (bottom) anomalies of weekly composites with respect to its 2000–2006 climatological seasonal cycle of AOD (adimensional) from MODIS averaged over 21°–27°W and 19°–26.5°N, for the period 2000–2006. The horizontal lines represent the \((\langle AOD \rangle + 2\sigma)\) threshold value, respectively.
Vázquez are available [2004] who demonstrated that the aerosol surface forcing efficiency coefficient in –C08015 (Zhu et al. AOD Ramanathan et al. than 0.8 for weekly averages). Black arrows indicate strong dust events in winter (larger than 0.8 for weekly averages). Purple arrows in Figures 3a and 3b indicate seasonal high dust loads (centered in July 15) and black arrows indicate strong dust events in winter (larger than 0.8 for weekly averages).

Figure 3. (a) Time series of SST anomalies obtained from the TMI (red), AMSR-E (black), and AVHRR (blue) sensors, for the period 2002–2005. In all cases, the SST anomalies were calculated relative to a climatological seasonal cycle computed from TMI data over the 9 year period 1998–2006, and averaged subsequently over 25°–18°W; 15°–22°N. (b) Differences between TMI and AMSR-E data (black) and between TMI and AVHRR data (blue). Also shown (in green) is the time series of weekly averaged MODIS AOD data representing the same area and including the seasonal cycle. Purple arrows in Figures 3a and 3b indicate seasonal high dust loads (centered in July 15) and black arrows indicate strong dust events in winter (larger than 0.8 for weekly averages).

TMI data by comparing them against in situ measurements on a global scale and found a standard deviation difference of about 0.45°C.

To obtain further confidence into the quality of the TMI SST fields, we show in Figure 3a a comparison between SST anomalies as inferred from the TMI (red) and AMSR-E (black) microwave radiometers for the period 2002–2005 averaged over the region (18°–25°W; 15°–22°N). In both cases the SST anomalies were calculated with respect to a climatological seasonal cycle computed from 9 years (1998–2006) of TMI data. The MW SST anomalies are about the same albeit somewhat smaller in amplitude for AMSR-E. Arrows in Figure 3a mark seasonally high dust loads (in summer, purple arrows) and the strongest dust events in winter (black arrows), as identified from the AOD time series. A first visual comparison reveals that both MW SST time series decrease simultaneously to most of those events, as it would result from cooling.

Also shown in Figure 3a is a time series of SST anomalies from the AVHRR Pathfinder data [Kilpatrick et al., 2001], after subtracting the same TMI climatological seasonal cycle for comparison purposes. Differences between the SST from TMI and AMSRE and from TMI and AVHRR (shown in Figure 3b) indicate that the two MW SST products stay close together (within a 0.5°C range; rms(TMI-AMSRE) = 0.17°C). In contrast, the AVHRR SST estimates deviate substantially from both MW observations (rms(TMI-AVHRR) = 0.78°C), sometimes by as much as 2°C or more during strong dust events, with a clear seasonality also present in the difference. This suggests that AVHRR data is biased cold by as much as 1°C or more in this region during summer months.

It is known that both clouds and aerosol can lead to biases between IR and MW SST [Chelton and Wentz, 2005]. In our study region, clouds and Saharan dust have both seasonal cycles connected to the seasonal movement of the ITCZ and, in principle, the differences shown in Figure 3 could therefore result from either effect. However, Corlett et al. [2006] previously showed differences between SSTs from the dual view Advanced Along-Track Scanning Radiometer (AATSR) and SSTs from MODIS and AVHRR of about 1°C in our study area and argued that they result from dust impact on MODIS and AVHRR retrievals. This conclusion was backed up by Vázquez-Cuervo et al. [2004] who demonstrated that AATSR is less sensitive to aerosols due to the dual view. They also report negative correlations between aerosols and AATSR SST minus AVHRR SST (as well as with in-situ comparisons) in our studied region.

Those earlier results suggested to compare the SST differences shown in Figure 3 with the weekly averaged AOD (Figure 3b). A visual inspection shows a clear relationship between periods of enhanced AOD and associated dust concentrations and periods of large TMI minus AVHRR SST differences, suggesting that the impact of dust clouds on the AVHRR retrieval is the primary agent for causing the AVHRR SST to be biased cold. This applies for dust content on a seasonal cycle as well as for strong dust outbreaks on time scales of up to a week. These effects are consistent with the earlier finding from Vázquez-Cuervo et al. [2004] and are therefore the consequence of the AVHRR atmospheric correction algorithm failing to adequately compensate for the dust and the dry SAL. We conclude therefore that AVHRR data sets are not appropriate for the study intended here. At the same time it is obvious that microwave SST data have an important role to play in such studies of climate change, especially if there is an expected changing dust load of the atmosphere.

3. Aerosol Radiative Forcing During AEROSE-I

To understand the impact of Saharan mineral dust on SST of the eastern Atlantic, an important quantity to know is the shortwave (SW) radiative forcing anomaly at sea level associated with a specific dust load of the atmosphere, referred to below as aerosol radiative forcing (ARF in W m⁻²) anomaly flux. ARF associated with dust in the SAL can be computed from AOD fields according to

$$ARF = f_e \cdot AOD,$$

where $f_e$ is the aerosol surface forcing efficiency coefficient in units of W m⁻²AOD⁻¹ [see Ramanathan et al., 2001] and AOD is the aerosol optical depth from MODIS discussed above. A few estimates of $f_e$ are available [Li et al., 2004; Yoon et al., 2005; Zhu et al., 2007]. However, a detailed
knowledge of \( f_e \) is missing and as part of this study, using data available from the AEROSE expedition, we revisit the question of what is the amplitude of \( f_e \) in our study region and whether there is a sensitivity of its estimate on the wavelength used to observe AOD.

[20] The AEROSE project consisted of a series of intensive field experiments conducted from aboard the U.S. National Oceanic and Atmospheric Administration (NOAA) research vessel Ronald H. Brown [Nalli et al., 2006; Morris et al., 2006; Hawkins et al., 2007]. Our analysis is based on data obtained during the first cruise (AEROSE-I), which took place during spring of 2004, departing from Bridgetown, Barbados, on February 29 and returning to San Juan, Puerto Rico on March 26, after crossing the Atlantic twice in zonal directions. During the cruise the ship encountered two significant Saharan dust events off the African coast in the period 9–18 March, and underneath it radiometric and ocean observations were obtained (Figure 4). The column-integrated optical depth observations were obtained with two upward looking, handheld, commercial Microtops sunphotometers operating in five different spectral bands. AOD estimates were derived from these measurements according to Knobelspiesse et al. [2004].

[21] In Figure 5 we show respective AOD measurements taken at sea level at wavelengths of 380 nm and 870 nm (Figure 5a, see Figure 4 for locations). Also shown in Figure 5 are SW radiation fluxes measured quasi-continuously at sea level \( (Q_{SL}, \text{Figure 5b}) \) from aboard the ship. Apparent from Figure 5 is the impact of clouds on the measurements, especially after March 11. In the solar SW radiation measurements, clouds usually lead to reduced radiation fluxes; however, in some instances, surface SW fluxes are also enhanced due to forward scattering from the broken clouds when the cloud fraction is small and the sun is in a cloud-free sky [e.g., Pfister et al., 2003]. The baselines of the AOD measurements are indicative of dust concentration changes under cloud-free conditions (Figure 5c). Cloud impacts lead to significant variability in the AOD observations. Because the ARF is defined as the difference between the incident radiative flux calculated under clear-sky conditions with and without aerosols [Won et al., 2004; Yoon et al., 2005], we have to eliminate the effect of cloud contamination on the radiation measurements, before \( f_e \) and ARF can be estimated from the AEROSE-I data.

[22] To identify cloud-free conditions, we use the amplitudes of the normalized SW atmospheric transmissivity, \( \tau \), which is the fraction of solar energy incident at the top of atmosphere (TOA) that would reach the surface with the sun overhead. Low values of \( \tau \) result from the presence of clouds, and anomalously high values occur when the sensor is in the direct beam of the sunlight with additional energy scattered by the clouds. To eliminate cloud-contaminated data, the first step was to discard all measurements for which \( \tau \) is higher than 0.85 and lower than 0.75. The high threshold was chosen to be somewhat lower than the clear-sky normalized atmospheric transmissivity measured in the clear polar atmospheres [Minnett, 1999; Hanafin and Minnett, 2001; Key and Minnett, 2006], and cases with values below the lower threshold were clearly influenced by the presence of clouds.

[23] As a second step of the cloud detection and elimination process, we fitted a sinusoidal function to the resulting SW radiation (Figure 5d). Because March 11 was essentially cloud-free, we used the radiation profile from that day to simulate the cloud-free daily cycle in solar radiation during the following days. But because AOD is varying with time, the cloud-free noon radiation also varies from day to day. To account for this we used a scaled version of the daily radiation cycle from March 11 and, superimposed on it, the measured shortwave radiation data of all other days.

[24] An estimate of the efficiency factor, \( f_e \), was obtained from the differences between the \( Q_{SL} \) and the shortwave fluxes at TOA \( (Q_{TOA}) \). These are plotted in Figure 6 as a function of AOD at 380 nm and 870 nm, considering all cloud-free values from 9 to 18 March 2004 and only the AEROSE-I AOD measurements between 0.1 and 0.75 (see Figure 5c). Values of \( Q_{TOA} \) were calculated for the position of the sun at the location of the ship at the times of the measurements using the equations given in the Astronomical Almanac [Astronomical Applications Department, 1990] and for the refraction of the atmosphere as given by Zimmerman [1981]. Using the results from the 380 nm channel, a linear regression between \( Q_{SL} - Q_{TOA} \) and AOD leads to values of \( f_e \) of

\[
  f_e = (Q_{SL} - Q_{TOA})/AOD = -73.5 \pm 7.1 \text{ W m}^{-2} \text{ AOD}^{-1},
\]

with a correlation coefficient of −0.41. The slope is smaller (−57.9 ± 8.9 W m\(^{-2}\) AOD\(^{-1}\)) when using the 870 nm channel (correlation of −0.27) and using an average of both channels would result in a value of \( f_e = -69.1 \pm 8.0 \text{ W m}^{-2} \text{ AOD}^{-1} \). We note, for a later interpretation, that the AOD measurements were taken between 13h and 19h local time, and that the \( f_e \) estimate therefore corresponds to the afternoon situation, rather than to a daily mean field. Compared to previous
estimates available from Li et al. [2004], Yoon et al. [2005] and Zhu et al. [2007] ($f_c \approx 80$ W m$^{-2}$ AOD$^{-1}$, during winter), our results are somewhat lower, although still close. Considering the offset for AOD = 0 as a further consistency check, the estimates lead to values of $-212.2 \pm 3.6$ W m$^{-2}$ and $-225.3 \pm 3.6$ W m$^{-2}$, for 380 and 870 nm, respectively. In comparison, the theoretical value is $-180$ W m$^{-2}$, which exist for the global ocean. Our estimates are larger, and this can partially be related to regional variations.

4. Observed SST and AOD Anomalies

[25] Examples of weekly-averaged AOD time series during strong dust events and contemporaneous weekly-averaged TMI SST anomalies (relative to a climatological seasonal cycle), are provided in Figure 7 for the years 2002 to 2005, both averaged over the region 27°–21°W, 19°–26.5°N. Figure 7 reveals several intense dust periods lasting several days. It appears that during or after some of those dust events, the SST shows downward tendencies (either as cooling or as reduced warming) as would be consistent with dust-related cooling.

[26] Figure 7a reveals an intense dust period during late April and beginning of May 2002. A second (although weaker) dust event occurs in the July/August time frame. After the first dust period, SST declines over a one and a half months period by as much as 1°C. The SST subsequently recovers until the end of June, however, consistent with the second dust period, the SST declines again during late July. Similarly, a drop in SST up to about 1.4°C is observed in mid-February 2003 (Figure 7b), after one strong dust event occurs in the preceding weeks. SST starts increasing during a subsequent low dust period but shows a new drop after a further strong event at the beginning of March. During 2004 (Figure 7c) two high AOD events occurred in March. Even though this seems to be a period with background positive
is the specific heat of seawater and (equation (2)). To be constant with time, then the difference between \( Q_{\text{SL}} - Q_{\text{TOA}} / C_1 \) and \( Q_{\text{SL}} \) in units of W m\(^{-2}\) would be a typical, de Boyer Montégut et al. typical density of the upper ocean, \( \rho = 0.55 \) when the anomalies of AOD lead those of SST by 1 week. The slope of the linear regression \((-0.27°C \text{ AOD}^{-1})\) lead to an estimate of the mean value of the MLD of about 22 m. This is somehow shallower than those values of de Boyer Montégut et al. [2004] and our findings in the next section. Nevertheless, the value is plausible and supports the validity of a relation between SST trends and AOD.

The results shown in Figure 8 suggest that about 30% (i.e., \( r^2 \)) of the SST variance could be explained by those of the accumulated AOD in our study region. However, the remaining 70% of SST variance occurs on all time scales because many processes lead to SST changes, most of which are not related to dust events at all. Accordingly, dust-induced SST anomalies will always be superimposed, or even masked, by SST changes associated with ocean dynamics or remote forcing. Separating local dust effects from dynamical SST variations (e.g., resulting from mechanical-wind stirring, upwelling and/or eddies and planetary waves) remains a challenge in any study based on (satellite) data alone, implying that more information from ocean dynamics is needed in any quantitative investigation.

We conclude that a quantitative and unambiguous connection between Saharan dust outbreaks and surface cooling as hypothesized by Miller and Tegen [1999] can not be made convincingly from the data at hand, but that additional information is necessary, e.g., as available from ocean dynamics embedded in circulation models. Nevertheless, a significant relation between SST changes and dust appears to be present.

5. Simulated SST Anomalies in the Mixed Layer

To simulate the response of SST to dust-induced cooling, an approach based on a simple bulk mixed layer model would be an oversimplification since the MLD, taken to be constant and known in equation (3), in reality is highly variable in space and time due to its dependence on many parameters, including surface wind stress. To further aid the investigation of dust effects on SST we use here a local mixed layer model to determine the amplitudes of SST anomalies that can be expected during individual dust events. In the next section, the use of a 3D circulation model will help to identify dynamical SST features.

The state-of-art 1D mixed layer model of the upper ocean is based on the coupled KPP mixed layer model [Large et al., 1994] (the model will be referred to as KPP-1D henceforth). The philosophy of this conceptual study is to...
neglect advective processes and thereby to investigate the magnitude of SST anomalies that can be forced locally by strong dust events. The central part of the KPP model consists of the 1D upper ocean mixing parameterization in which a bulk Richardson number is used to determine the depth of the oceanic boundary layer (OBL). A non-local “K profile parameterization” (KPP) is applied subsequently as the vertical mixing scheme in the OBL. In the runs performed, the vertical resolution is 1 m over the top 100 m. To run the KPP-1D model, initial temperature and salinity profiles were used as provided by CTD profiles collected during AEROSE-I at the location 19.0°W, 18.0°N (Station 5 in Figure 4).

To simulate the effect of Saharan dust cooling of the sea surface, two simulations were performed: a control run and a perturbed run. The control run was driven by net forcing fields provided every 6 hours by the National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) Reanalysis-1 [Kalnay et al., 1996]. This version of the KPP-1D model can be forced either with the state of the atmosphere, or directly with the fluxes. In this case, we prescribed NCEP reanalysis surface fluxes which included wind stress, sensible heat flux, evaporation, radiative fluxes and precipitation. The evaporation was estimated with the latent heat fluxes from NCEP/NCAR Reanalysis-1 and the enthalpy of vaporization. Fields were linearly interpolated every 30 min. from the available 6-hourly values. Because the NCEP forcing does not include any impact of Saharan dust on the surface SW forcing [Lacis and Hansen, 1974], this run is not affected by any dust-related cooling.

The effect of dust loads was incorporated into the perturbed run through a reduction of the NCEP SW forcing by the ARF forcing. To this end, the following steps were performed:

The effect of dust loads was incorporated into the perturbed run through a reduction of the NCEP SW forcing by the ARF forcing. To this end, the following steps were performed:

Figure 7. Time series of weekly anomalies of MODIS AOD (triangles) and of TMI SST (stars), averaged over 27°–21°W; 19°–26.5°N, for the periods (a) April–August 2002 and January–May for years (b) 2003, (c) 2004 and (d) 2005.

Figure 8. Scatter plot between accumulated weekly anomalies of MODIS AOD and weekly anomalies of TMI SST, for the region 27°–21°W; 19°–26.5°N and for the period 2000–2006. The AOD is leading the SST for 1 week. The red line represents the fitted least-squares linear regression. The resulting coefficient of correlation and slope are 0.55 and −0.27, respectively.
where $ARF$ was provided every day, $SW(t)$ was available every 6h and $SW_{\text{max}}$ is the maximum noon radiation value. In this way, the night values are not altered and a maximum reduction of the SW is obtained at noon of every day.

3. No other dust effects on the surface radiative forcing were accounted for, such as enhanced incident longwave radiation. This step is justified by Vogelmann et al. [2003], who showed that the respective forcing would only be of the order of ±10W m$^{-2}$ which can be confidently neglected here.

The decrease in surface temperature resulting from the difference between the perturbed and unperturbed runs is therefore due only to the SW flux anomalies derived from the presence of dust in the atmosphere. Because what matters to our analysis here is the resulting ocean temperature at the uppermost level of the model, in the following we will only show the top level temperature anomaly, assuming it is equivalent to the SST anomaly. Results are shown in Figure 9, for strong dust outbreak periods during winter 2003 and 2004. Shown in Figure 9 (top) are SST anomalies as they result for the periods February 1 through March 31, 2003 (left) and February 1 through March 31, 2004 (right) with a temporal resolution of 30 min. Also shown are MODIS daily AOD observations at the same position. Because in our study dust impacts only the SW solar irradiance, SST anomalies show a clear daily cycle with maximum negative amplitudes occurring during mid-day and close to zero changes during night time. Moreover, an overall decrease due to the general presence of dust during the entire considered period is observed.

Focusing first on Figure 9 (left), two dust events are apparent from the AOD time series during early 2003, with peak AOD values of about 1.9 occurring during February 11. A second and somewhat longer event can be observed since the beginning of March (0.7 < AOD < 1.7). In both cases, SST decreases are simulated during and shortly after peak dust concentrations. This is specially obvious during the second event lasting for about 5 days in early March, when SST decreases by about 0.3°C. Even though a quantitative connection was not possible from our analysis of Figure 7, the decrease observed at the beginning of March 2003 (Figure 7b) agrees with the magnitude simulated here. We note that only weekly averages were shown in Figure 7, which inevitably lead to smaller amplitudes in the SST anomaly and temporal shifts of the maximum response. We also note that the first dust event, happening on February 11, has a smaller signature in SST, yet the forcing amplitude appears larger compared to the March event.

To understand why this is so we show, in Figure 9 (middle), anomalies of the SW forcing together with the absolute value of wind forcing. Moreover, shown in Figure 9 (bottom) are time series of the SST and MLD for the perturbed and unperturbed runs. The MLD is fairly deep during the beginning of February (around 40 m). A decline of MLD occurs during early March, when the wind stress suddenly drops to a minimum, leading to a MLD of less than 15 m. The same holds during 2004 (Figure 9 (right)): a shallow MLD is present only at the end of March (neglecting the two intermittent shoaling events), while the major dust event occurred 20 days before. Accordingly, the SST response simulated during 2004 is likewise moderate and only of the order of 0.1°C.

The experiments suggest that, if applied only over short periods, dust-induced cooling of the surface strongly depends on the structure of the upper ocean and especially the thickness of the surface mixed layer. To have a measurable impact on SST, a shallow mixed layer is required simultaneously to enhanced dust loads. Wind stirring tends to deepen the mixed layer through mixing and would thereby eliminate the signal. Amplitudes of SST anomalies, simulated with the KPP-1D model, are comparable to those observed in Figure 7 shortly after strong dust events. However, while observed changes are plausible, both in terms of amplitude and phase, a definitive proof of them being generated by dust can not be given from the 1D simulation; instead a full ocean circulation model is needed (next section). Nevertheless, the results provide the important information that a strong response of SST due to Saharan dust can be expected only during periods of low wind stirring (and associated low mixed layer depth) and that therefore more emphasis needs to be put on the impact of dust on SST on the seasonal time scale, discussed in the next section.

6. Isolating Dust-Induced SST Anomalies

To further discriminate dust-induced SST anomalies from dynamically-induced SST anomalies, the output from an ocean circulation model simulation of the North Atlantic, run over the period 1948–2008 is analyzed next. The run is described in detail by Serra et al. [2010] and is based on the coupled sea ice-ocean MIT general circulation model [Marshall et al., 2004] configured for the Atlantic Ocean north of 33°S including the Mediterranean Sea, the Nordic Seas and the Arctic Ocean. The model features a curvilinear grid with one pole situated over North America and a second pole over Europe, this way solving the singularity at the North Pole. The model is eddy-permitting with a horizontal resolution of about 15 km throughout the entire domain and 50 levels with vertical resolution varying from 10 m in the upper ocean to 550 m in the deep ocean. The model bottom topography derives from ETOPO2 and the initial temperature and salinity conditions from the WOCE Global Hydrographic Climatology [Gouretski and Koltermann, 2004]. The model is forced at the surface by fluxes of momentum, heat and freshwater computed using bulk formulae and the 6-hourly atmospheric state from the NCEP/NCAR Reanalysis-1 [Kalnay et al., 1996]. At the volume-balanced open northern and southern boundaries, the model is forced by a 1° resolution global solution of the MITgcm forced by the same NCEP data set. The model SST is relaxed to the monthly Extended Reconstructed SST V3 database [Smith et al., 2008] and the model sea surface salinity to the World Ocean Climatology 2005 monthly climatology [Boyer et al., 2005] with a relaxation time scale of 1 month. Vertical mixing is parameterized by the KPP formulation of Large et al. [1994]. Background coefficients of vertical diffusion and viscosity are $10^{-6}$ m$^2$ s$^{-1}$ and $10^{-5}$ m$^2$ s$^{-1}$, respectively.
and the coefficients of horizontal diffusion and viscosity are both $10^{11} \text{m}^4 \text{s}^{-1}$.

42 We use the model SST in the area of our study to see if the differences between the observed (with the microwave satellite sensor) and simulated (by the model) SST can be brought in closer relation with Saharan dust than we have been able to do with the KPP-1D model. The analyzed period of the SST is 1998–2006 (2000–2006 when the AOD is included) and the temporal resolution is 1 week. We recall that the NCEP reanalysis forcing does not include the effect of Saharan dust on the surface radiative forcing of the eastern North Atlantic, but otherwise leads to a realistic estimate of air–sea interaction, i.e., the model results can be used as an estimate of the local and non-local SST anomalies forced by ocean dynamics as well as by changes in the surface fluxes of heat, freshwater and momentum, all not related to dust.

43 Before using simulated SST fields to identify dynamical SST anomalies, we need to test the model’s skill against observed SST variations derived from microwave radiometry. To proceed in this direction we show in Figure 10 the differences in the root-mean-square (rms) SST variability between the TMI SST and the simulated SST for the period 2000–2006 (we show the difference in the rms variability rather than the rms SST difference, because any phase shift in...
SST variability, occurring close to Miller and Tegen. Distribution of the difference between the root mean square of TMI and model), suggesting differences quickly reduce to small values (<0.3°C, 20% of the SST vicinity, SST we will exclude it from the further intercomparison of the diagrams). As a consequence, that the model is well capable to simulate the observed SST but to model biases (Figure 10). As a consequence, that the model is well capable to simulate the observed SST but to model biases (Figure 10). A close inspection of the SST differences, which account for up to 0.5°C, reveals that, in the region enclosed in our box, they are not related to the SST differences evolution in the coastal region, suggesting again that the SST differences we are seeing in Figure 13 are not contaminated by model biases close to the coast but do provide a reasonable estimate of dust-induced SST cooling.

7. Concluding Remarks

For a comparison of observed and simulated SST fields, we show in Figure 11 longitude-time diagrams for the period 2002 through 2006 of the observed and simulated SST anomalies (i.e., relative to their respective climatological seasonal cycle) after averaging meridionally between 19°N and 26.5°N. From a visual inspection, we see a clear agreement between the simulated and observed SST anomalies over most parts of Figure 11. In particular, the model seems capable of simulating cold and warm events in quite some detail, suggesting that it is well capable of simulating the dynamical effects in the observed SST. The most obvious difference between Figures 11 (left) and 11 (right), is that the observed negative SST anomalies in the eastern (coast) part of the diagrams tend to be larger. This is consistent with what was discussed previously and displayed in Figure 10. We also note that this region shows a good correlation between the weekly means of AOD and the differences between the TMI SST and the simulated SST (Figure 12a), but this result is likely not related to dust-induced anomalies in SST but to model biases (Figure 10). As a consequence, we will exclude it from the further inter-comparisons involving model fields.

[45] The correlation between the SST differences and AOD is high offshore, showing significant negative correlations in the main dust deposition area (see box in Figure 12a). The positive correlations further south are likely due to other processes unrelated to dust. High correlations between SST differences and AOD time series can also be observed in the main dust area (indicated by absence of hatching in Figure 12b) after removing a climatological seasonal cycle from both time series, which suggests a statistically significant relationship between the AOD and the (observed-simulated) SST differences. Correlation coefficients larger than ±0.1 are significant at a 95% confidence level.

[46] To further highlight the data-model differences and their relation to dust, we show in Figure 13, as a function of time, the difference between the observed SST and those simulated by the model (black line) together with the weekly averages of MODIS AOD, after averaging both fields over 21°–27°W; 19°–26.5°N (black box in Figure 12a). We note that the SST difference in this box shows only a small regular seasonal cycle and primarily interannual variability superimposed to sub-seasonal changes. The same holds for the AOD time series and both are correlated (see Figure 12a). A close inspection of the SST differences, which account for up to 0.5°C, reveals that, in the region enclosed in our box, they are not related to the SST differences evolution in the coastal region, suggesting again that the SST differences we are seeing in Figure 13 are not contaminated by model biases close to the coast but do provide a reasonable estimate of dust-induced SST cooling.

In this study we used several satellite data sets and the output from numerical models to study the impact of aerosol layers of Sahara mineral dust on the eastern subtropical Atlantic with focus on the potential cooling affect of the airborne dust on the surface temperature. Such a cooling was postulated before [e.g., Miller and Tegen, 1999], but until now it has not been revealed in observations. Here we have identified several indications of Saharan dust impact on SST in the eastern subtropical North Atlantic.

Findings from our study suggest a reduction of SST simultaneously or shortly after enhanced aerosol loads. An inspection of microwave SST observations suggests that about 30% of SST variance could be explained by dust-induced cooling in our study region. To further support this finding, we used output from a numerical model to eliminate dynamical signals in the SST observations in an attempt to further isolate dust-induced SST cooling. The integration of the eddy-permitting model of the North Atlantic forced by the NCEP reanalysis (which does not include dust effects on surface heat fluxes) showed that most of the dynamical SST anomalies present in TMI observations can be simulated (see Figure 11) and that after removing them from the model output, remaining SST anomalies are consistent with dust-induced cooling not present in the simulations. From this result it appears that the model simulations are biased warm by about 0.5°C in the eastern subtropical North Atlantic due to the neglect of dust loading in the used NCEP atmospheric forcing fields. Moreover, because our model hindcast was relaxed towards the AVHRR SST fields [see also O’Neill...
Figure 11. (left) Longitude-time diagram of observed TMI weekly SST anomalies, shown for the years 2000 through 2006 after averaging over the latitude range 19°–26.5°N. SST anomalies are SST observations from which a climatological seasonal cycle calculated over the 1998–2006 period was removed. (right) Same as in Figure 11 (left), but obtained from the eddy-permitting model of the North Atlantic forced by the NCEP reanalysis.
et al., 2003, Figure 1] that also appear biased cold in our study area (due to dust impact on the retrieval), the still present warm model bias therefore has to be considered a lower bound. It can therefore be expected that the real cooling effect of dust on SST be even larger than estimated through the model-data comparison shown in Figure 12. In line with the conclusion drawn from the analysis of the observations and the specifically designed 1D experiments, the eddy-resolving model results provide additional support that a decrease in SST identified in the observations can be associated with the cooling effect of Saharan dust. However, while supportive of this hypothesis, the GCM experiments by themselves are not conclusive. To be more conclusive, separate model experiments would need to be performed in which a purely flux-driven model is being perturbed by dust-induced cooling anomalies, similar to what was performed here with the 1D model. While being beyond the scope of this paper, such an analysis in underway in a separate study [see also Martínez Avellaneda, 2010].

[46] Because AVHRR based SST retrievals are biased cold due to dust, we conclude that ocean models should use microwave SST fields as relaxation fields where possible, to avoid dust-induced biases in simulated SST (this calls for a long-term continuation of microwave SST measurements beyond current missions). In addition, ocean models should use surface radiative forcing that accounts for dust effects. This can be done by using surface-based measurements of satellite retrievals of AOD fields to correct the solar radiation according to equation (1). But this requires a better understanding of the forcing efficiency of dust, i.e., detailed and longer time series of AOD and downward solar radiation at sea level that can provide more reliable estimates of \( f_e \) than

Figure 12. Distribution of local correlations between: (a) AOD and TMI SST minus simulated SST and (b) anomalies of AOD (relative to AOD seasonal cycle) and anomalies of TMI SST (relative to TMI seasonal cycle) minus anomalies of simulated SST (relative to the simulated SST seasonal cycle). Correlation coefficients larger than ±0.1 are significant with 95% confidence level and denoted with the full and dashed black contours. The hatched area (Figure 12b) corresponds to a superposition of the following: pixels where the percentage of the MODIS AOD retrievals is less than 55%; pixels where the AOD anomalies is smaller than 0.5 more than 35 times (i.e., days); areas where the SST \( \text{rms} \) difference of Figure 10 is higher than 0.45°C. The black box in Figures 12a and 12b indicates the region where the spatial averages are performed.

Figure 13. Time series of weekly AOD from MODIS (blue) and TMI SST minus simulated SST (black), averaged in the box (27°–21°W; 19°–26.5°N) for the period 2000–2006.
those we provided here from short time series obtained during the AEROSE-I cruise. We anticipate that such a time-series of measurements will become available through the TENATSO (Tropical Eastern North Atlantic Time-Series Observatory) and SOPRAN (Surface Ocean Processes in the Anthropocene) programmes.

An open question remains also to what extent Saharan dust affects the quality of NCEP and ECMWF reanalyses and their estimates of surface fluxes, because no dust effects are included in the models that support the reanalysis and because AVHRR-based SST fields, by themselves biased low on a seasonal cycle, are used as lower boundary conditions. Moreover, using those SST fields in ocean models can have an impact on the simulated strength of the meridional overturning circulation and associated heat transport in the Atlantic. Also unclear is the impact that changes in dust load can have on air-sea feedbacks and climate simulations.

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