The Response of the Tropical Atlantic and West African Climate to Saharan Dust in a Fully Coupled GCM

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ABSTRACT

This study examines the climate response in West Africa and the tropical Atlantic to an idealized aerosol radiative forcing from Saharan mineral dust, comparable to the observed changes between the 1960s and 1990s, using simulations with the fully coupled GFDL Climate Model, version 2.1 (CM2.1), for two optical property regimes: more absorbing (ABS) and more scattering (SCT) dust. For both regimes dust induces significant regional reductions in radiative flux at the surface (approximately $-230$ W m$^{-2}$). At the top of the atmosphere (TOA) dust in the two simulations produces a radiative flux anomaly of opposite sign ($+30$ W m$^{-2}$ in the ABS case and $-20$ W m$^{-2}$ in the SCT case). These differences result in opposing regional hydrologic and thermodynamic effects of dust. The ABS-forced simulations show an increase in the West African monsoon resulting from dust, whereas in the SCT-forced simulations dust causes a decrease in the monsoon. This is due to moist enthalpy changes throughout the atmospheric column over West Africa creating either horizontal divergence or convergence near the surface, respectively. In the tropical North Atlantic, dust acts to cool the ocean surface. However, in the subsurface the ABS-forced simulations show a decrease in upper-ocean heat content, while the SCT-forced simulations show an increase in upper-ocean heat content. The peak differences primarily arise from the wind stress curl response to a shift in the Atlantic ITCZ and associated mixed layer depth anomalies. Changes to upper-ocean currents are also found to be important in transporting energy across the equator.

1. Introduction

Dust is one of the most abundant aerosols in the atmosphere. It originates in arid or semi-arid regions, preferentially from topographic depressions (Prospero et al. 2002). Surface winds uplift soil particles into the atmosphere (Gillette et al. 1980), and these particles may reach the middle or upper troposphere where they can be carried thousands of kilometers (Grousset et al. 2003; Adams et al. 2012). Because of the variability of its mineralogy, the impact of dust on radiation (Boucher et al. 2013), cloud microphysics (Levin et al. 1996; DeMott et al. 2003), and biogeochemistry (Swap et al. 1992; Jickells et al. 2005) may vary greatly in time and space. This could further modulate the impact of dust on such phenomena as air pollution (Liu et al. 2009) and hurricane frequency in the Atlantic basin (Dunion and Velden 2004; Braun 2010; Wang et al. 2012). Unfortunately, there is still uncertainty about the mineralogy-dependent optical properties of dust, and little information on variable dust mineralogy is incorporated into modeling experiments. Thus, a sensitivity study is required.

There is also a lack of consensus on global dust emission, as was shown by Huneeus et al. (2011) who diagnosed a large spread between models of 6–30 Tg global annual dust load. This spread arises even without considering dust from agriculture or vegetation changes due to climate change, which may account for 25% of global dust emissions (Tegen et al. 2004; Ginoux et al. 2012). In addition, atmospheric dust concentrations show significant multidecadal-scale variability (Mahowald et al. 2010). For instance, atmospheric dust concentrations as observed at Barbados increased nearly fivefold
between the 1960s and 1990s (Prospero and Lamb 2003). Based on the objective analysis of different dust inventories by Cakmur et al. (2006), we use the topography-based dust source inventory of Ginoux et al. (2001) in this study. The choice of dust mineralogy dataset, and in particular its absorption and scattering characteristics, has been shown to considerably affect the interactions with radiation (Sokolik and Toon 1999; Miller et al. 2004). Instead of solving for dust mineralogy, we will perform a sensitivity study covering a range of dust absorption and scattering.

There have been a number of studies attempting to quantify the impact of dust on climate through modeling. The earliest simulations with general circulation models (GCMs) showed that dust’s direct radiative impact was felt very strongly in the local region and hinted at far-reaching effects, outside of the main dust cloud, through clouds and precipitation (Miller and Tegen 1998). With increasing resolution, later models began analyzing the effect of dust on the local hydrologic cycle, with some studies showing a decrease in regional precipitation (Miller et al. 2004; Solmon et al. 2008; Lau et al. 2009; Gu et al. 2012; Solmon et al. 2012). The responses in the West African monsoon to global dust concentrations for several of these models are listed in Table 1 and plotted in Fig. 1 alongside the effects of specifically Saharan dust calculated in this study. All of these models also showed a decrease in the surface temperature under the dust cloud and, for the ones that were coupled to a slab ocean model, showed a decrease over the North Atlantic as well. However, the majority of modeling studies are limited in that they use only one set of optical properties in their experiment. In addition, no modeling studies have been done using a full ocean model coupled to an atmospheric GCM, despite results showing significant perturbations to the physical ocean in the local region (Evan et al. 2012).

The purpose of our study, then, is to expand beyond the past GCM simulations of the climatological effect of dust by using a fully coupled GCM in conjunction with a range of dust optical properties. Our hypothesis is that the optical properties of dust can significantly affect the regional climate response to dust of North Africa and the tropical North Atlantic Ocean, with significantly different results depending on the amount of absorption versus scattering of the mineral aerosols. Utilizing multicentennial multimember ensemble runs with a

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### Table 1. The simulated JJA precipitation response in the area of the West African monsoon to global dust in previous studies. The configuration for each study’s ocean model is presented, and those studies with multiple experiments are noted. The single scattering albedo (SSA) is that of 1-μm dust particles at 550 nm used in each model, derived from either in-text tabulation when provided or through model paper references. The precipitation (P) anomalies and associated bounds (mm day⁻¹) of each model result are derived either from in-text tabulation when provided (denoted by an asterisk) or from averaging plotted results and calculating the regional range of response. The results of this study are presented last and represent the effect of changes in Saharan dust comparable to the 1960s to the 1990s in comparison to the global changes in dust presented in prior studies.

<table>
<thead>
<tr>
<th>Study (Experiment notes)</th>
<th>Ocean model</th>
<th>SSA</th>
<th>(\Delta P_{\text{lower}})</th>
<th>(\bar{\Delta P})</th>
<th>(\Delta P_{\text{upper}})</th>
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<td>SCT dust</td>
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range of dust optical regimes, we will analyze the full climate effect of dust perturbations by allowing the ocean to respond in its full capacity and produce a sensitivity analysis of the climatic response to dust’s radiative properties. For instance, we will focus on the extent to which the ocean influences the African monsoon. In addition, we will seek to understand the coupled atmosphere–ocean response in the tropical North Atlantic.

The paper is organized as follows. In section 2 we describe the model and datasets used in this study as well as the experimental design. In section 3 we analyze the direct radiative effect of dust. In section 4 we explore several regional hydrologic and thermodynamic effects of dust on the climate system. Section 4a covers the impact of cloud and precipitation anomalies over North Africa and the tropical Atlantic and how those changes affect the local energy balance. Section 4b focuses on the West African monsoon system and associated local circulations. Section 4c describes the effect in the tropical North Atlantic Ocean and the coupled atmospheric–oceanic response. Section 5 gives a discussion of the presented results and concluding remarks.

2. Methodology
a. Coupled model

For this study, we used the Geophysical Fluid Dynamics Laboratory (GFDL) Climate Model, version 2.1 (CM2.1), (Delworth et al. 2006) and perturbed the concentration of mineral dust atmospheric burden, exploring the sensitivity based on two of the optical regimes of mineral dust aerosols. The GFDL CM2.1 is described in detail by Delworth et al. (2006), Gnanadesikan et al. (2006), Wittenberg et al. (2006), and Stouffer et al. (2006) and has been shown to have one of the best simulations of the West African monsoon among all the models from phase 3 of CMIP (CMIP3) (Cook and Vizy 2006). CM2.1 is a fully coupled general circulation model, combining atmospheric, oceanic, land, and ice models.

The atmosphere model (AM2.1) in CM2.1 runs on a finite volume dynamical core (Lin 2004) with a horizontal resolution of 2° latitude by 2.5° longitude and 24 vertical levels. The specific parameterizations used in AM2.1, including the treatment of scattering and absorption of radiation by aerosols, are described by Anderson et al. (2004) and Delworth et al. (2006). Aerosols are calculated offline and introduced into the atmospheric model as a monthly climatological average (Horowitz 2006), and interactions with radiation have been validated (Ginoux et al. 2006). One shortfall of the model is that only the aerosol direct effect is considered.

The ocean model (OM3.1) in CM2.1 is based on the Modular Ocean Model, version 4 (MOM4) (Griffies et al. 2003). The horizontal domain uses a tripolar grid with a 1° latitude by 1° longitude horizontal resolution,
which enhances to $\frac{1}{3}^\circ$ within 10$^\circ$ of the equator. The vertical domain uses 50 grid levels, 22 of which are evenly spaced within the top 220 m alone, and retains the ability to somewhat follow rough bottom topography using partial grid cells. The parameterizations employed by OM3.1 are described by Griffies et al. (2005) and Gnanadesikan et al. (2006).

The land model used in CM2.1 is a variant of the Land Dynamics (LaD) model (Milly and Shmakin 2002). It uses the same grid as the atmospheric model, has 18 vertical levels for heat storage, and uses a two-layer bucket scheme for water storage. The final component is the sea ice simulator (SIS), which uses a modified Semtner three-layer thermodynamic scheme (Winton 2000) and the elastic–viscous–plastic technique for ice internal stresses (Hunke and Dukowicz 1997). It uses the same horizontal grid as the ocean and two vertical layers for ice and snow. The coupled model is initialized by forcing the climate with constant 1990 insolation, gas and aerosol concentrations, and land cover for 300 years as described by Delworth et al. (2006).

b. Dust forcing

The climatological annual cycle of monthly, global mineral dust aerosol burden (Horowitz 2006) was calculated using the Model of Ozone and Related Chemical Tracers, version 2 (MOZART-2), (Horowitz et al. 2003; Tie et al. 2005) forced by the NCEP–NCAR reanalysis of 1990 (Kalnay et al. 1996) (Fig. 2a). This climatology, called the full or base dust climatology, is relatively well validated in CM2.1, particularly over the North Atlantic Ocean where it is generally within one standard deviation of observations (Ginoux et al. 2006). One shortfall is that this climatology lags the observed maximum dust concentration in June by one month over the North Atlantic. To assess the impact of variations in dust concentrations, we utilized an additional MOZART-2 simulation forced by the NCEP reanalysis of 1990 wherein the dust emission over North Africa is reduced to 20% of its 1990 value (Fig. 2b). This climatology will be called the reduced-dust climatology. These two datasets are representative of the high and low dust concentrations of the 1990s and the 1960s, respectively (Prospero and Lamb 2003). The difference between the base and reduced concentration can be seen in Fig. 2c.

We explore the sensitivity of the climate response to dust concentrations to the optical properties of dust by performing both the control and perturbation experiments using two estimates of dust optical properties. The first optical property of absorbing (ABS) dust is derived from a combination of the in situ observations of Volz (1973) and Patterson et al. (1977). Of the two optical property regimes, this combination has the highest imaginary part of the refractive index and thus is more absorbing. This combination has been used as the standard in previous GFDL climate models (Anderson et al. 2004), but observational studies have shown the results of Patterson et al. (1977) to be overly absorbing (Sinyuk et al. 2003). Thus, the second optical property explored is chosen to be more scattering (SCT) and is calculated using Mie theory for highly scattering dust. The values of the refractive index are taken from Balkanski et al.
coefficient, single scattering albedo, and the asymmetry parameter for both optical regimes for fine (0–1 μm) and coarse (1–10 μm) dust modes. As can be seen, the ABS dust is much more absorbing, particularly in the longwave, whereas the SCT dust is much more scattering across all wavelengths.

c. Experimental design

We form two control runs by forcing the model with an annually invariant 1990 radiative forcing. Each control run uses the full (base) dust climatology for the annual cycle of atmospheric dust load and either the ABS-dust optical properties or the SCT-dust optical properties. These are run for 155 years each. Then, a snapshot at the end of every 5 years of each control run is used as the initialization state for a perturbed case wherein the atmospheric dust load climatology is switched to the reduced-dust climatology (corresponding to 5 times reduced dust load). The optical properties are kept consistent with the parent control runs, and the perturbation cases are run for 50 years in parallel with the control runs. We repeat these perturbation simulations out to year 105 of the control runs. This produces two ensemble sets of simulations with 21 members wherein we can observe the effect of varying dust concentrations in different optical regimes. To assess the effect of dust decadal variation, we subtract the reduced dust concentration simulations from their associated base dust concentration simulation years and then average across all members. This process creates two sets of 50-yr ensemble average anomalies due to variations in dust concentrations. Significance of the anomalies is then determined by a Student’s t test at 95% confidence.

3. Direct radiative effects

a. Clear-sky top-of-atmosphere radiative fluxes

We begin by analyzing the direct radiative impact of both ABS and SCT dust to changes in climatological dust concentration, first at the top of the atmosphere and then at the surface. Each of the following results is averaged over the summer months [June–August (JJA)], when the climatological extent of dust is largest and the West African monsoon is strongest, and are clear-sky calculations, by which we mean that they neglect the radiative forcing of clouds. In addition, each anomaly map is calculated so that it represents the effect of increasing dust.

At the top of the atmosphere, we observe a largely opposite response between the addition of ABS and SCT dust to the direct radiative balance over the Sahara Desert (Fig. 4). Over the Sahara, the addition of ABS

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**Figure 3.** The optical properties of both regimes of dust used in this experiment. The values are averaged over the fine (small) dust bin sizes (0.1, 0.2, 0.4, and 0.8 μm; dashed lines) and the coarse (large) dust bin sizes (1, 2, 4, and 8 μm; solid lines) of the ABS-dust regime (red lines) and the SCT-dust regime (blue lines). (a) The extinction coefficient (m² g⁻¹) as a function of wavelength. (b) The SSA of the dust as a function of wavelength. (c) The asymmetry parameter of the dust as a function of wavelength.

(2007) as their methodology provides values for the full solar spectrum compared to Sinyuk et al. (2003), which provides values only at near-UV wavelengths. We consider their 0.9% hematite case and artificially enhance the scattering effect by multiplying the imaginary part of the refractive index by 0.1. Figure 3 shows the extinction

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Dust causes a significant increase in incoming clear-sky shortwave (SW) radiative flux at the top of the atmosphere, up to 20 W m\(^{-2}\) in the southwestern Sahara (Fig. 4a). In contrast, the addition of SCT dust causes a modest decrease in incoming clear-sky SW radiative flux across North Africa, peaking around −15 W m\(^{-2}\) near the west coast of North Africa (Fig. 4b). This difference is due to the dust reducing (increasing) the net column albedo over the very reflective Sahara and decreasing (increasing) the amount of upward-reflected SW radiation in the ABS-dust (SCT dust) simulation. Over the tropical North Atlantic, both simulations result in a negative clear-sky SW radiative flux anomaly of order −10 W m\(^{-2}\). This is due to the slightly increased backscattering over the relatively dark ocean in both simulations.

Similarly, there is a nearly opposite difference in clear-sky longwave (LW) radiative flux at the top of the atmosphere between the ABS-dust and SCT-dust simulations. In the ABS-dust simulation, the clear-sky LW flux increases under much of the dust plume, including North Africa and the eastern and equatorial tropical North Atlantic, with a peak around 10 W m\(^{-2}\) along the southern Sahara (Fig. 4c). Conversely, the far field away from the dust plume, including the Americas and the northern tropical North Atlantic, has predominantly negative LW flux anomalies of the order −5 W m\(^{-2}\) (Fig. 4c). This is largely opposite to the anomalies in the SCT-dust simulation, which include a decrease of −5 W m\(^{-2}\) along the Sahel and increases of up to 5 W m\(^{-2}\) across the tropical North Atlantic and Americas (Fig. 4d). The addition of ABS dust leads to a positive downward LW radiative flux anomaly at the top of the atmosphere because dust is absorbing terrestrial radiation and emitting at lower temperatures than the surface. Meanwhile, the addition of SCT dust leads to a weaker negative downward LW radiative flux anomaly because of the increased surface temperature in the Sahel, as will be discussed in section 4a.

These SW and LW radiative anomalies combine to cause a strong increase in clear-sky total radiative flux...
across the top of the atmosphere over North Africa in the ABS-dust simulation, reaching up to 30 W m$^{-2}$ (Fig. 4e), and a general decrease in the SCT-dust simulation of up to $-20$ W m$^{-2}$ (Fig. 4f). Both simulations result in a general decrease of clear-sky total radiative flux across the tropical North Atlantic of the order $-5$ W m$^{-2}$. These results are generally in line with the current understanding of dust-induced anomalies in top-of-atmosphere radiative fluxes and are of relatively similar value to previous studies (Miller et al. 2014).

b. Clear-sky surface radiative fluxes

Turning now to the clear-sky radiative effect at the surface, we note that the response per radiative component is generally stronger for the ABS-dust experiments at the surface (Fig. 5) than at the top of the atmosphere. Whereas at the top of the atmosphere there are competing effects of absorption and scattering, the addition of ABS dust dims the surface beneath the aerosol cloud by up to $-45$ W m$^{-2}$ along the west coast of North Africa (Fig. 5a). The same effect occurs with the clear-sky SCT-dust SW radiative fluxes, which decrease by a more modest $-10$ W m$^{-2}$ across North Africa and the North Atlantic, particularly near the West African coastline (Fig. 5b) owing to the reduced absorption of SW radiation by SCT dust.

The clear-sky LW radiative flux increases across North Africa in the ABS-dust simulation by up to 25 W m$^{-2}$ (Fig. 5c). This is due to the emission of LW radiation by the dust plume. However, in the SCT-dust simulation, there is a clear-sky LW radiative anomaly of $-15$ W m$^{-2}$ along the southern Sahara and Sahel (Fig. 5d). This is again due to the increased surface temperature in this region, as will be discussed in section 4a.

When we combine the SW and LW clear-sky radiative flux anomalies and the surface we observe a generally negative total radiative flux in both optical regimes. In the ABS-dust simulation, this anomaly focuses in the eastern tropical North Atlantic with decreases up to $-30$ W m$^{-2}$ (Fig. 5e). Similarly, in the SCT-dust simulation there are negative anomalies in total radiative flux of almost $-20$ W m$^{-2}$ focused primarily along the west coast of North Africa and along the Sahel (Fig. 5f). These anomalies generally corroborate the results of previous studies in that dust causes a negative surface radiative flux anomaly underneath the plume and there is a stronger response for stronger absorption (Miller et al. 2014).
4. Regional hydrologic and thermodynamic effects

a. Clouds and precipitation

Switching away from the clear-sky radiative effect of dust, we will explore the changes to the climatological cloud fields and associated precipitation patterns. In the ABS-dust experiment, we observe that an increase in dust causes significant changes to clouds over North Africa and the North Atlantic (Fig. 6). The addition of ABS dust causes a marked increase in high-level clouds over all of North Africa and the Gulf of Guinea, by as much as 15%, while simultaneously decreasing cloud cover over much of the North Atlantic and Americas, locally as high as 2×10% (Fig. 6a). There is a generally more muted change in midlevel clouds with decreases across the western Atlantic and Sahara of around −5% and a strong increase over the region between 5°–15°N and 15°W–15°E of near 15% (Fig. 6b). Because of its importance in later analyses, we will call this region (5°–15°N and 15°W–15°E) the West African monsoon (WAM) region. The low-cloud anomalies are much more heterogeneous over the Atlantic, but over the western WAM region increase by almost 15% (Fig. 6c).

In combination, the total cloud amount increases over much of North Africa and the equatorial Atlantic, by as much as 15%, and decreases across the central and western North Atlantic by about −10%, and this appears to be a strong function of the high-cloud changes (Figs. 6a and 6d). These changes indicate an increase in deep convective systems over the WAM region and eastern equatorial Atlantic Ocean.

The response to the addition of SCT dust appears to be largely opposite to the results of the ABS-dust experiment (Fig. 7). An increase in SCT-dust causes a significant decrease in high cloud cover over North Africa and the Gulf of Guinea, by as much as −15%, while simultaneously causing an increase over much of the North Atlantic and Americas, by as much as 10% (Fig. 7a). Again, the response in midlevel clouds is more muted than at high levels and primarily consists of a decrease over the WAM region and western Sahara, by as much as −10% (Fig. 7b). At low levels, the cloud field shows an anomalous decrease over the WAM region, equatorial Atlantic, and North Atlantic of up to −10%
while conversely resulting in an increase over the central tropical North Atlantic and Americas of about 5% (Fig. 7c). When combined, the total cloud field shows negative anomalies over much of North Africa and the eastern equatorial Atlantic of up to −15%, while there are positive anomalies over the eastern North Atlantic and Americas of almost 5% (Fig. 7d). These anomalies indicate a significant decrease in deep convection over North Africa and the eastern equatorial Atlantic.

The changes expressed in the cloud fields are associated with precipitation patterns over North Africa and the North Atlantic. The most striking response occurs over the Sahel and tropical North Atlantic in both the ABS- and SCT-dust ensemble simulations. We see that the addition of ABS dust causes an increase in precipitation over the WAM region and the Gulf of Guinea with maximum values reaching 2.5 and 3 mm day$^{-1}$, respectively (Fig. 6e). We note that these same areas had increases in cloud cover at almost every level corroborating the development of deep convection over these regions. Simultaneously, there is a significant decrease across the tropical North Atlantic of up to −4 mm day$^{-1}$ (Fig. 6e). The addition of SCT dust has much the opposite response with a significant decrease over the WAM region, Gulf of Guinea, and the tropical North Atlantic north of 10°N, with peaks of −2.5, −3, and −2 mm day$^{-1}$, respectively (Fig. 7e). These changes are concurrent with decreased cloudiness at all levels, which is in line with our suggestion of a reduction of the deep convection associated with the West African monsoon. Meanwhile, there is an increase of precipitation over the western tropical North Atlantic reaching up to 2 mm day$^{-1}$ (Fig. 7e). These changes have the potential to be a nonnegligible feedback on the amount of dust aerosols in the atmosphere owing to the associated changes in the wet deposition rate. Such feedbacks onto the climatological effect of dust have been shown to be significant for Africa (Miller et al. 2004) and North America (Cook et al. 2009).

Because of the moisture-starved nature of North Africa, these changes in precipitation patterns have nonnegligible effects on the soil moisture balance of the region. The increases in precipitation over North Africa in the ABS-dust simulation generally lead to increases in soil moisture content of up to 25 kg m$^{-2}$ over the regions south of the Sahel (Fig. 6f). Meanwhile, the decreases in precipitation over the Americas lead to reductions in soil moisture of order −30 kg m$^{-2}$ over North America.

FIG. 7. As in Fig. 6, but for the SCT-dust simulation.
and order $-10 \text{ kg m}^{-2}$ over South America (Fig. 6f). Conversely, in the SCT-dust simulation there are decreases primarily in the WAM region of up to $-20 \text{ kg m}^{-2}$ resulting from the strong reductions in precipitation in that area (Fig. 7f). Again, the Americas show the opposite picture with strong increases of soil moisture content nearing $20 \text{ kg m}^{-2}$ in accordance with the increases in precipitation (Fig. 7f). The results over the Americas are surprisingly large and could have implications for long-term drought conditions in those areas. These significant hydrologic changes raise the question of how the regional climate energy budget is perturbed and are the topic of the rest of section 4a.

1) ALL-SKY TOP-OF-ATMOSPHERE RADIATIVE FLUXES

We will be analyzing the energetic budget of the region by working through the all-sky radiative flux anomalies, by which we mean that they include the radiative forcing of clouds (in contrast to section 3), for each simulation. At the top of the atmosphere we observe that including clouds in the calculation of SW radiative flux leads to the development of a strong dipole feature between the Sahara and the WAM region (Figs. 8a,b). The response over North Africa is the same as the clear-sky calculations because of the lack of moisture in this region. However, the addition of ABS dust (SCT dust) causes a significant decrease (increase) in incoming all-sky SW radiative flux over the WAM region. In the ABS-dust simulation this amounts to almost $-25 \text{ W m}^{-2}$ (Fig. 8a), whereas in the SCT-dust simulation the anomaly is of the order $15 \text{ W m}^{-2}$ (Fig. 8b). These changes are associated with the anomalies in cloud amount over the WAM region. With the ABS-dust simulation, an increase in cloud cover reflects more incoming SW radiation to space. Meanwhile, in the SCT-dust simulation a decrease in cloud cover reduces the amount of backscattered radiation resulting from clouds.

Over the tropical North Atlantic there is a more heterogeneous response in the ABS-dust simulation, with significant decreases in the central and far-eastern portions and a strong decrease over the Gulf of Guinea.

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**Fig. 8.** The TOA all-sky radiative anomalies resulting from dust with differing optical properties (positive downward; W m$^{-2}$). The values are averaged over JJA, and only those values that pass a Student’s $t$ test to 95% are shaded. (left) The ABS-dust simulation results and (right) the SCT-dust simulation results. (a),(b) The total SW flux anomaly at the TOA. (c),(d) The total LW flux anomaly at the TOA. (e),(f) The total radiative flux anomaly at the TOA.
(Fig. 8a). The changes north of 10°N match the low-cloud amount anomalies where more (less) low clouds reflect more (less) incoming SW radiation leading to a negative (positive) all-sky SW radiative flux anomaly at the top of the atmosphere. The changes south of 10°N better match the total cloud amount anomalies, implying a change in the large-scale atmospheric circulation. Conversely, the all-sky SW radiative anomaly over the tropical North Atlantic in the SCT-dust simulation is less heterogeneous than its ABS-dust counterpart (Fig. 8b) but again matches decently with the observed cloud amount anomalies.

Now, considering the all-sky LW radiative flux we note that it is effectively the same map as the clear-sky flux across North Africa, albeit with higher-magnitude anomalies. In the ABS-dust simulation, this is associated with the large increase in high-level clouds, which trap LW radiation in the atmosphere reducing the outgoing LW radiative flux (Fig. 8c). Conversely, in the SCT-dust simulation there is an enhanced decrease in LW radiative flux because of the significant decrease in high-level clouds across North Africa, which then allows more LW radiation to escape to space (Fig. 8d).

Across the tropical North Atlantic, just south of 10°N, there is a significant decrease in all-sky LW radiative flux of about −10 W m\(^{-2}\) (Fig. 8c). The location of this anomaly coincides with the climatological position of the Atlantic intertropical convergence zone (ITCZ) in summer. Therefore, the significant decrease in cloud amount in this region suggests a reduction or a movement of this large-scale circulation feature. In addition, both simulations show significant changes in the Gulf of Guinea that coincide with changes in cloud amount. In the ABS-dust simulation, the increase in LW radiative flux is associated with an increase in total cloud amount, particularly high-cloud amount in this region. Conversely, in the SCT-dust simulation, the decrease in LW radiative flux coincides with a strong decrease in total cloud amount over the Gulf of Guinea.

When combined, the SW and LW radiative anomalies in both the ABS-dust (Fig. 8e) and SCT-dust (Fig. 8f) simulations lead to a significant dipole in total radiative flux across the Sahara and the WAM region. Over the tropical North Atlantic the ABS-dust simulation (Fig. 8e) leads to a more heterogeneous response than the SCT-dust simulation (Fig. 8f), with the latter having a band of decreased top-of-atmosphere radiative flux across much of the central tropical North Atlantic.

2) ALL-SKY SURFACE RADIATIVE FLUXES

Looking now at the all-sky surface radiative flux (Fig. 9), we note that again the response is generally of the same magnitude as the clear-sky results. For the all-sky SW flux response to ABS dust we observe a broadening of the negative anomaly over the tropical North Atlantic to include the entire region between the equator and 30°N. The anomaly is less heterogeneous than its ABS-dust counterpart (Fig. 9a). This is likely associated with the increase in cloud amount in these areas, which would block the incoming SW flux to the surface. There is also a significant decrease in the WAM region, again due to the increase in cloudiness throughout the column (Fig. 9a). In the SCT-dust simulation, we notice a constriction of the SW radiative flux anomaly over the tropical North Atlantic with positive flux anomalies near the equator and 30°N (Fig. 9b). In addition, a dipole structure develops between the WAM region and the Sahara with a local maximum anomaly of almost 20 W m\(^{-2}\) in the WAM region. These anomalies can largely be associated with the significant decreases in total cloud amount over these areas.

The all-sky LW radiative flux anomalies are not much different from their clear-sky counterparts. The most apparent difference arises between 20° and 30°N in the eastern North Atlantic. In the ABS-dust case there is a modest increase in LW flux at the surface associated with the increase in cloudiness in that region trapping LW radiation near the surface (Fig. 9c). Conversely, in the SCT-dust simulation there is a modest decrease in line with the loss of cloud amount in that region (Fig. 9d). There are also small changes to the anomaly in the WAM region as a result of the increase (decrease) of cloudiness in that area in the ABS-dust (SCT dust) simulation.

When combined, the SW term dominates in both the ABS-dust and SCT-dust simulations, and we observe all-sky total radiative fluxes that are more heterogeneous versions of the clear-sky radiative fluxes (Figs. 5e,f). The largest difference arising from adding clouds to the radiative flux calculations is the development of a dipole between the Gulf of Guinea–WAM region and the Sahara in the SCT-dust simulation, which is due to the relative changes of increased surface dimming from dust and decreased cloudiness.

3) SURFACE TURBULENT FLUXES AND TEMPERATURE

Radiative anomalies at the surface are quickly equilibrated by the turbulent flux of energy through sensible and latent heating terms. Focusing first on the addition of ABS dust, we see a decrease in sensible heat flux across North Africa, with maxima around −40 W m\(^{-2}\) along the Sahel, and an increase in sensible heat flux across the Americas (Fig. 10a). As is to be expected, sensible heat flux anomalies are small in comparison to the latent heat flux anomalies over the majority of the tropical Atlantic Ocean (Figs. 10a,b). Simultaneously,
we see a strong north–south dipole of latent heat flux anomalies across North Africa and the eastern tropical Atlantic. Over the southern Sahara there are increases of latent heat flux of 35 W m\(^{-2}\), while over the WAM region there are decreases of −30 W m\(^{-2}\) that extend out over the tropical North Atlantic and Gulf of Guinea under the dust plume and across the Americas (Fig. 10b). South of the equator there is a significant increase in latent heat flux of up to 25 W m\(^{-2}\) focused on the eastern portion of the tropical South Atlantic (Fig. 10b).

These anomalies over North Africa appear to be associated with the precipitation (Fig. 6e) and soil moisture anomalies (Fig. 6f). In the warm and relatively moisture-starved region of the Sahara and the Sahel, any increase in precipitation will fuel an increase in evaporation (and therefore latent heat flux) while leading to a decreased sensible heat flux. This increase of evaporation at the expense of the sensible heat flux is then associated with cooler temperatures, so an increase in soil moisture causes a cooling effect on the surface (Fig. 10d). This leads to strong cooling along the Sahel–Sahara boundary reaching up to −3.5 K. Meanwhile over the less moisture-burdened WAM region and tropical North Atlantic, the negative surface radiative flux anomalies (Fig. 9c) will lead to a decrease in surface turbulent fluxes and a cooling effect. This is visible in the general cooling across the tropical North Atlantic basin of about 0.5 K (Fig. 10d).

These turbulent flux terms appear to balance the all-sky direct radiative anomalies over the Sahara (Fig. 10c). However, over much of the North Atlantic there is a rather heterogeneous all-sky net energy flux anomaly (Fig. 10c). In the Gulf of Guinea, there is an interesting dipole around the equator with a positive all-sky net energy flux anomaly in the eastern tropical North Atlantic peaking around 25 W m\(^{-2}\) and a negative anomaly in the eastern tropical South Atlantic reaching −35 W m\(^{-2}\) (Fig. 10c). This is also associated with a significant warming in the southeastern tropical South Atlantic of up to 1 K (Fig. 10d). These combined would suggest the ocean has a cross-equatorial transport of heat from the Northern Hemisphere into the Southern Hemisphere.

The impact of the addition of SCT dust on the surface turbulent fluxes is nearly opposite from adding the ABS dust. The sensible heat term shows an anomalous dipole structure with decreases reaching −10 W m\(^{-2}\) in

![Fig. 9. As in Fig. 8, but for the surface.](image-url)
the Sahara and increases peaking at 20 W m$^{-2}$ in the WAM region (Fig. 11a). There is also a significant negative sensible heat flux anomaly across the Americas (Fig. 11a). Again, the sensible heat flux anomalies are small compared to the latent heat flux terms over the open ocean. The latent heat flux term is predominantly negative over much of North Africa, with maxima around $-30$ W m$^{-2}$ along the Sahel, and has a positive anomaly over the WAM region of 10 W m$^{-2}$ (Fig. 11b). There is a much less coherent negative anomaly across the tropical North Atlantic, with a peak along the equator of about $-20$ W m$^{-2}$, and a strong positive anomaly across much of the Americas (Fig. 11b).

Again, these anomalies are associated with decreases in precipitation (Fig. 7e) and soil moisture content (Fig. 7f) in these regions. In contrast to the ABS-dust simulation, the decrease of precipitation in the Sahel is associated with a decrease in soil moisture content available for evaporation. This corresponds to a reduction in the Bowen ratio and a general increase in surface temperature (Fig. 11d). This leads to a strong warming along the Sahel reaching up to 1.5 K. Meanwhile, the negative surface radiative flux anomalies over the dry Sahara (Fig. 9f) are then associated with a general cooling and a reduction in surface turbulent fluxes.

Finally, these terms appear to balance the anomalous radiative perturbations over the majority of landmasses as seen in the all-sky net energy flux at the surface (Fig. 11c). However, there is a strong positive anomaly along the equatorial Atlantic peaking around 30 W m$^{-2}$ (Fig. 11c). This, in comparison to the ABS-dust simulation, would suggest a cross-equatorial transport of heat by the ocean from the Southern Hemisphere into the Northern Hemisphere.

b. West African monsoon response

We will now turn our focus toward a specific hydrologic cycle in Africa: the West African monsoon. On short time scales following a forcing event, before the system reaches equilibrium, the effect of dust can be described by the elevated heat pump (EHP) mechanism of Lau et al. (2009). The EHP for absorbing particles begins with the dust absorbing radiation and warming the atmosphere while cooling the surface. This warming leads to vertical motion, which in turn leads to divergence aloft and convergence over a boundary layer that has been cooled by the absorption of radiation above, thus enhancing precipitation. However, Lau et al. (2009) state that this effect becomes minimized for reflecting dust with a SSA of 0.95 or higher. Additionally, on longer time scales after the surface has had a chance to adjust to the imposed forcing, the general column diabatic heating is controlled primarily through the all-sky top-of-atmosphere radiative flux anomalies (Miller et al. 2014). These anomalies adjust at a slower rate compared to those at the surface.

Fig. 10. The surface turbulent flux balance (positive upward; W m$^{-2}$) and surface temperature (K) change resulting from dust for the ABS-dust simulation. The values are averaged over JJA, and only those values that pass a Student’s $t$ test to 95% are shaded. (a) The sensible heat flux anomaly at the surface. (b) The latent heat flux anomaly at the surface. (c) The total energy flux anomaly at the surface. (d) The surface temperature anomaly.
1) ZONAL AND MERIDIONAL DIAGNOSTICS

To understand the local hydrologic and circulation anomalies, we plot similar figures to Figs. 5 and 7 of Lau et al. (2009) to describe changes to the West African monsoonal system (Figs. 12 and 13). For both simulations we see that in a meridional average the dust extends through much of the troposphere but weakens away from the source region. The strongest concentration is between 800 and 700 mb (1 mb = 1 hPa) over the African continent between 20°W and 20°E (Fig. 12a). In the ABS-dust simulation there is a westerly wind anomaly through much of the dust layer, reaching over 2 ms⁻¹ above the coast of the WAM region (Fig. 12a). The westerly wind anomalies reach up to the tropopause over much of West Africa but switch to a weaker easterly wind anomaly above the tropical North Atlantic and Caribbean. The low-level westerly flow across much of the tropical North Atlantic basin is coincident with a reduction of precipitation until reaching the coast of West Africa where there is a sharp increase in the precipitation amount (Fig. 12c). The dust plume is also associated with a significant positive temperature anomaly throughout much of the troposphere (Fig. 12e). This anomaly peaks above the main dust plume over the WAM region between 400 and 300 mb. Over this same region we see a general lifting motion in the atmosphere, which corresponds well to the area of increased precipitation (Fig. 12e). Meanwhile at the surface there is significant cooling over the West African landmass and west out over the tropical North Atlantic. The lift over the WAM region is somewhat balanced by subsidence over the majority of the central tropical North Atlantic.

Viewing the effects from the zonally averaged direction, in the ABS-dust simulation we are better able to see that the lower-tropospheric westerly wind anomalies are actually composed of two separate jets: one to the north and one to the south of the main dust loading (Fig. 12b). These are accompanied by a significant westerly anomaly at the surface beneath the dust cloud. All of this surrounds a weaker easterly anomaly above the center of the climatological dust plume. The westerly wind anomalies allow for more moisture transport over West Africa, enhancing the precipitation in this region (Fig. 12d). This is supported by the strong area of lift along the southern flank of the dust plume (Fig. 12f). There is a significant warming trend through and to the north of the dust plume (Fig. 12f). This induces rising motion over the Sahara and aids in the small increase in precipitation over that region. We see that this strong heating is associated with transport of moist air from the Gulf of Guinea over the cooler boundary layer before recirculating it southward aloft. These results largely concur with the results of Lau et al. (2009) in terms of sign and location, especially for the larger-scale features such as the lower-level cool tongue and upper-level warming, strong coastal precipitation asymmetry, and a vertical dipole in zonal wind anomalies. The one exception is the precipitation response, which is about twice as high in our experiments.

Considering the SCT-dust simulation, we use the same dust climatology between the ABS-dust and SCT-dust...
cases, but because of the reduced absorption of the dust there is a lower AOT through the plume (Fig. 13a). The zonal wind response is primarily easterly through much of the lower and middle troposphere, in contrast to the ABS-dust case (Fig. 13a). This easterly anomaly dries out the WAM region and weakens the West African monsoon; however, there is little impact to the tropical North Atlantic and Caribbean (Fig. 13c). Analyzing the temperature response, there is significant cooling throughout the troposphere over West Africa and the tropical North Atlantic, while there is a slight surface warming over the WAM region (Fig. 13e). There is little vertical motion in the zonal cross section except over the WAM region between 500 and 300 mb where there is a region of strong subsidence between the upper-level easterly jet and lower-level westerly jet (Fig. 13e). This subsidence corresponds with the decrease in precipitation over West Africa. There are also areas of rising motion around 80°W and 60°–40°W, which align with the areas of slight positive anomalous precipitation over the tropical North Atlantic (Figs. 13c,e).

From the meridionally averaged perspective, there are significant easterly wind anomalies south of, below, and north of the dust plume, while there is a weaker westerly wind anomaly above the dust. These easterly wind anomalies bring dry air from off the continent and drive the strong decrease in precipitation (Fig. 13d). In fact, the strongest decreases in precipitation align with major areas of subsidence in the meridional plane (Fig. 13f). The dust causes a significant cooling aloft with a strong but constrained warming at the surface (Fig. 13f). Despite the presence of this anomalous inversion, this cooling is associated with general subsiding motion over much of West Africa and the Gulf of Guinea, decreasing

Fig. 12. The effect of dust over the WAM region for the ABS-dust simulation. The values are averaged over JJA, and only those values that pass a Student’s t test to 95% are shaded/contoured. (left) A zonal view of the meridional average between 5° and 15°N and (right) a meridional view of the zonal average between 10°W and 10°E. (a),(b) The dust optical thickness (shading) and the zonal wind anomaly (m s⁻¹; contoured every 1 m s⁻¹). (c),(d) The precipitation anomaly (mm day⁻¹). (e),(f) The wind streamlines and potential temperature anomaly (K; shading and contoured every 0.5 K).
precipitation over the region. Interestingly, the SCT-dust simulation shows an anomaly of comparable magnitude to that in the ABS-dust simulation, which is inconsistent with the mechanism proposed by Lau et al. (2009).

2) **RESIDUAL ATMOSPHERIC COLUMN RADIATIVE FLUXES**

It has been posited that precipitation anomalies in the West African monsoon can be related to changes in the net energy convergence in the atmospheric column over the region. We have seen that there are discrepancies between the total energy flux anomalies at the surface and the radiative perturbations at the top of the atmosphere, and so we can calculate the convergence of energy into the atmospheric column. This term is defined as

\[
F_{\text{net}} = \text{SW}^\uparrow_{\text{TOA}} - \text{SW}^\downarrow_{\text{TOA}} - \text{SW}^\downarrow_{\text{Sfc}} + \text{SW}^\downarrow_{\text{Sfc}} - \text{LW}^\downarrow_{\text{TOA}} - \text{LW}^\downarrow_{\text{Sfc}} + \text{LW}^\uparrow_{\text{Sfc}} + \text{SH} + \text{LH},
\]

where the superscript arrows denote the direction of the flux and subscripts denote top of atmosphere (TOA) and surface (Sfc) for the SW and LW radiative components added to the turbulent sensible (SH) and latent (LH) heat fluxes. Chou and Neelin (2003) explain that in the tropics the net energy convergence must be largely balanced by divergence of moist static energy. Because of the small horizontal gradients of temperature and moisture in the tropics, positive net energy convergence is balanced by rising motion. This convective motion leads to enhanced precipitation over the regions of positive net energy convergence.

In both optical regimes, this parameter shows a strong north–south dipole over North Africa and similarly over the eastern half of the equatorial Atlantic while being mostly heterogeneous over the remainder of the tropical North Atlantic basin. Following the addition of ABS dust, there is net convergence of energy into the atmospheric column over the Sahara and eastern tropical South Atlantic, reaching up to 40 W m\(^{-2}\) (Fig. 14a). Simultaneously, there is an anomalous net divergence over the WAM region and eastern tropical North

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**Fig. 13.** As in Fig. 12, but for the SCT-dust simulation.
Atlantic, dropping to $-20 \text{ W m}^{-2}$ (Fig. 14a). This dipole over North Africa is due principally to the all-sky TOA radiative flux anomalies bringing energy into the atmospheric column over the Sahara and removing energy from the column over the WAM region while the surface radiative and turbulent flux anomalies adjust rapidly to the anomalous forcing. The equatorial dipole is largely a result of the anomalous surface energy flux and could imply that the ocean is exporting some of the net column energy into the Southern Hemisphere.

The addition of SCT dust also causes a dipole but in the opposite sense to the ABS-dust perturbation. The Sahara and eastern tropical North Atlantic show an anomalous net energy divergence through the column, up to $-35 \text{ W m}^{-2}$ (Fig. 14b). There is also an anomalously strong net convergence of energy over the WAM region and eastern tropical North Atlantic, reaching $15 \text{ W m}^{-2}$ (Fig. 14b). The anomalies over North Africa are again a result of the top-of-atmosphere radiative flux anomalies adjusting more slowly than the surface radiative flux anomalies while the surface energy flux anomalies are the root of the equatorial dipole. This latter effect is indicative of a cross-equatorial ocean transport of energy into the Northern Hemisphere.

Comparing the precipitation figures (Figs. 6e and 7e) to the respective net energy convergence figures (Fig. 14) we see that the correspondence is imprecise in many regions. In the ABS-dust simulation over the Sahara and Sahel, the positive net column energy convergence (Fig. 14a) is coincident with rising motion to compensate for the increased column diabatic heating (Fig. 12f) and an increase in precipitation (Fig. 12d). Conversely, in the SCT-dust simulation, the negative net column energy convergence over these same regions (Fig. 14b) is associated with subsidence to compensate for the column diabatic cooling (Fig. 13f) and a modest decrease in precipitation (Fig. 13d). However, the precipitation anomaly is limited in these cases possibly because of the air being too dry outside the WAM region of North Africa.

When we look at the WAM region itself, the theory of net column energy convergence controlling precipitation appears to be less precise. Instead, we have a negative anomaly in net column energy convergence in the ABS-dust simulation and a slight positive anomaly in the SCT-dust simulation. These are coincident with positive (Fig. 6e) and negative (Fig. 7e) precipitation anomalies in their respective simulations. In general, the heating only accounts for the regionally averaged sign of the precipitation anomaly over North Africa. This could be due to some form of nonsteady circulation, such as African easterly waves, because one of the assumptions of Chou and Neelin (2003) is the compensation of heating only by the direct circulation. In addition, the ocean may be transporting some energy away from the column as evidenced by the previously mentioned dipole structure along the equator. To examine the possibility of this last source of error, we will next examine the effect of dust on the tropical North Atlantic.

c. Tropical North Atlantic response

Turning now to the effect on the ocean, we recall that in the ABS-dust simulation, the tropical North Atlantic significantly cooled, primarily near the West African coastline (Fig. 10d), whereas the SCT-dust simulation showed a thinner plume of cold surface water, again extending from the central West African coast (Fig. 11d). A measure of the effect to the tropical North Atlantic Ocean is the upper-ocean heat content (UOHC), calculated as the depth-integrated heat content. In this study we integrate from the surface down to 400-m depth. We plot this quantity across the Atlantic basin in Fig. 15 (shading).

In the ABS-dust case we see a strong cooling through the subsurface across much of the central tropical North Atlantic, reaching up to $-105 \text{ KJ cm}^{-2}$, slowly becoming more positive poleward and equatorward (Fig. 15a). This cooling is equivalent to a decrease of roughly $-0.6 \text{ K}$ in the average temperature of the layer between the surface and 400-m depth. This parameter.
becomes increasingly banded in the North Atlantic and also has a weak east–west dipole along the equator (Fig. 15a). Conversely, the SCT-dust simulation shows significant heating across the central tropical North Atlantic, with heating reaching up to 80 kJ cm$^{-2}$ (Fig. 15c). This heating is equivalent to an increase of roughly 0.5 K in the average temperature of the layer between the surface and 400-m depth. There is again a banded structure to the north and an overall weak cooling toward the equator (Fig. 15c).

One possible explanation for these patterns is the direct input of energy through the total energy flux at the surface. This quantity is plotted in Figs. 15a and 15c (black contours; positive downward). We observe that this quantity does not match many of the overall patterns in the central tropical North Atlantic and in fact is counter to many of the most significant features. If the local diabatic heating does not explain the changes in UOHC, then we need to consider the adiabatic response to wind changes in order to explain the UOHC anomalies.

The curl of the wind stress is plotted in Figs. 15b and 15d (black contours) and corresponds more closely with the UOHC values in the central tropical North Atlantic, with regions of cyclonic curl associated with cooling and anticyclonic curl with warming. In the ABS-dust simulation we see a significant increase in positive wind stress curl over the same region where we see a strong decrease in UOHC (Fig. 15b). This is because the positive wind stress curl induces an upward Ekman pumping mechanism, which draws up cooler deep water to the surface. Conversely, in the SCT-dust simulation we see that there are significant negative wind stress curl anomalies over the region of significant UOHC.
increases (Fig. 15d). This is because the negative wind stress curl induces a downward Ekman pumping mechanism, which deepens the mixed layer and increases the UOHC of the region.

To test whether the ITCZ shifts discernibly enough to cause an impact to the wind stress curl, we run the ABS- and SCT-dust base simulations out for a total of 1000 years each and compare the average annual cycle of zero meridional wind stress over the Atlantic between the base simulations and ensemble perturbations (Fig. 16). We observe that in the summer months, the addition of ABS dust causes the ITCZ to shift poleward by almost 0.8°, inducing a positive wind stress curl anomaly farther north. Conversely, the addition of SCT dust causes the ITCZ to shift equatorward by more than 0.5°, inducing a negative wind stress curl anomaly. These changes could be due to a number of factors. In the ABS-dust simulation, there is a strong decrease in precipitation along the climatological position of the ITCZ without a similar increase to the north (Fig. 6e). This could indicate a weakening of the Atlantic ITCZ with an apparent shift to the north, potentially a result of the increase in low-level entropy as indicated in Fig. 12e. Conversely, in the SCT-dust simulation, there is a significant decrease in precipitation along the climatological position of the Atlantic ITCZ and a corresponding increase farther south (Fig. 7e). This may be due to the strong decrease in entropy throughout the column over the Atlantic (Fig. 13e), which would act to push the center of convection farther south.

In line with the observed changes in UOHC, we plot the changes to the ocean mixed layer depth in Fig. 17. In both instances, we observe that in the North Atlantic north of 10°N the anomaly in mixed layer depth matches well with the anomalous wind stress curl in Figs. 15b,d. However, near the equator the response is more complicated. In the ABS-dust simulation there is a strong negative anomaly in mixed layer depth of about −2 m stretching across the tropical North Atlantic between 10° and 20°N, signifying a shallowing of the mixed layer in this region (Fig. 17a). This is in line with the positive wind stress curl anomaly over this same band. Conversely, there is a strong positive anomaly between 20° and 30°N of over 4 m stretching across the basin, indicating a deepening of the mixed layer. This is in general alignment with the negative wind stress curl anomaly over the region. In the SCT-dust simulation, the results are generally of opposite sign north of 10°N, in accordance with the opposite changes in wind stress curl. We observe a positive mixed layer depth anomaly between 10° and 20°N of up to 3 m and a negative anomaly between 20° and 30°N of around −3 m (Fig. 17b). These changes translate to a deepening and a
shallowing of the mixed layer, respectively. Such anomalies are significant to note as often one-dimensional models of the ocean temperature’s response to dust perturbations prescribe the mixed layer depth (Miller 2012).

Also in Fig. 17 are plotted the surface current anomaly (streamlines) averaged over the top 100 m of the ocean. For both optical regimes, we note an anomalous anti-cyclonic circulation develops around the positive mixed layer depth anomalies and a cyclonic circulation develops around the negative mixed layer depth anomalies. In the ABS-dust simulation we can see a clockwise circulation develop around the equator, which transports surface waters north in the western portion of the tropical Atlantic and south in the eastern portion of the basin (Fig. 17a). This corroborates our earlier statement of heat transports within the tropical Atlantic Ocean as inferred by anomalous surface energy fluxes. Similarly, in the SCT-dust simulation we can see a counterclockwise circulation develop around the equator, bringing surface waters north in the eastern portion of the basin and south in the western tropical Atlantic (Fig. 17b).

5. Conclusions

Using the fully coupled GFDL CM2.1, we have explored the impact of realistic dust perturbations to the climate system across a range of viable optical properties. We have seen a largely opposite response in the top-of-atmosphere clear-sky radiative fluxes over North Africa between the ABS-dust and SCT-dust simulations. At the surface, both optical regimes are dominated by the SW response to dust, despite differences in LW flux, leading to a local dimming.

We also noted opposing responses in cloud fields and precipitation patterns between the two optical regimes. These lead to feedbacks on the vertical energy fluxes, leading to a dipole structure over North Africa, which modulates the column radiative budget, and a broadening of the radiative response at the surface. The changes in precipitation and soil moisture content correspond with significant changes in surface turbulent fluxes, particularly the latent heat flux, and surface temperature over North Africa. However, over the ocean the turbulent fluxes increase the heterogeneity of the surface energy balance.

When we inspect the influence of dust on the West African monsoon, we see that the ABS dust causes a significant increase in precipitation whereas the SCT dust leads to a significant decrease. Over the Sahara and Sahel, these changes are in accordance with the convergence or divergence of energy in the atmospheric column, fueling adiabatic ascent or descent. However, the column energy convergence criteria are not as precise over the region between the Sahel and the Gulf of Guinea; this could be due to a number of factors, including nonsteady circulations and the oceanic transport of energy to offset the aerosol-induced heating or cooling. These results have implications for the tropical meteorology of the North Atlantic because the West African monsoon often induces convective instabilities that travel west and become tropical cyclones.

Over the tropical North Atlantic Ocean, our simulations again show opposing results. The central tropical North Atlantic UOHC in the ABS-dust simulation decreases significantly, whereas in the SCT-dust simulation the UOHC increases markedly. The main contribution to this impact is the dynamical ocean response to a shifting ITCZ. In the ABS-dust ensemble, the dust cloud causes the ITCZ to shift northward, inducing a positive wind stress curl anomaly, an upwelling of cold water, and a shallowing of the mixed layer. Conversely, the SCT-dust cloud causes the ITCZ to shift southward, inducing a negative wind stress curl anomaly, a downwelling of the warm surface ocean, and a deepening of the mixed layer. In addition, the varying optical regimes lead to nonnegligible surface current anomalies that aid in the transport of energy across the basin. These results have enormous implications for the tropical meteorology of the region as the UOHC is what “fuels” many of the convective systems in the tropical North Atlantic.

Our results show a high sensitivity to the amount of SW absorption (i.e., the mineralogy), which varies considerably between sources (Claquin et al. 1999). These results expand our understanding of the coupled dust–climate system and pose several interesting implications for future work. Future simulations can be enhanced by including spatially varying mineralogy, conducting ensembles with more intermediate optical properties, and introducing several known functions of dust such as desert–vegetation feedback, source region wind changes and dust emission feedback, and the aerosol indirect effect. Simultaneously, with increased resolution, we can probe the effects of dust on tropical meteorology such as hurricanes. In addition, we could probe the global response to dust as well as explore the significantly large response to the hydrologic cycle across the Americas. Finally, with longer-running simulations we can examine the effect on and effect from various large-scale climate modes of variability such as the Atlantic multidecadal oscillation (AMO) and El Niño.

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