The Climatological Effect of Perturbations in Atmospheric Burden and Optical Properties of Saharan Dust

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Abstract

Mineral dust aerosols are a diverse set of atmospheric constituents which provide some of the largest natural direct radiative forcing and have the propensity to affect several human-relevant issues. This dissertation investigates the regional and global climatic response to aerosol radiative forcing from dust using simulations with a suite of fully coupled climate models. An idealized perturbation to global dust climatology, with changes in Saharan-born dust comparable to the observed changes between the 1960s and 1990s, and an ensemble of realistic dust optical properties are utilized to study the climatological effect of perturbations in atmospheric burden and optical regime of Saharan dust.

Changes in dust atmospheric concentration lead to direct radiative responses from the top of the atmosphere (ToA) through to the surface along with regional hydrologic and thermodynamic responses, depending crucially on the amount of aerosol absorption versus scattering. There are large anomalies in the West African monsoon due to moist enthalpy changes throughout the atmospheric column over West Africa. In the tropical North Atlantic, there are significant responses in the upper ocean heat budget arising from the wind stress curl response to a shift in the Atlantic Intertropical Convergence Zone and associated mixed layer depth anomalies.

Simultaneously, there are changes in tropical cyclone activity across the North Atlantic Ocean with the largest response occurring in the most absorbing and scattering optical regimes. There are also non-negligible anomalies in the North Pacific and Indian Oceans. A relationship between accumulated cyclone energy and ToA radiative flux anomalies is used to explain the North Atlantic anomalies while several known climate variations are theorized to explain the far-field response to the dust forcing.

Changing the optical regime of dust alone is found to lead to radiative anomalies larger than simply adding dust. As dust becomes more scattering, there is a net global cooling focused in the Northern Hemisphere and a general equatorward shift of tropical precipitation and the mid-latitude atmospheric jets. This leads to a preferential negative phase of the
North Atlantic Oscillation, a decrease in the Atlantic Meridional Overturning Circulation, and associated changes to the global meridional heat transport.
Acknowledgements

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Finally, my deepest appreciation goes to my family. Mom and Dad, thank you for always supporting me in whatever paths I choose and for offering me guidance when I knew not which path to take. My sisters Melanie and Marilyn, thank you both for your care and focus as bigger sisters, even if I may be bigger than both of you. And thank you all for your unconditional love for all my life.
“Though wise men at their end know dark is right,
Because their words had forked no lightning they
Do not go gentle into that good night.”

–Dylan Thomas
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Chapter 1

Introduction

Mineral dust is a natural aerosol of the atmosphere which primarily originates in arid or semi-arid regions of the world and preferentially from topographic depressions (Prospero et al. 2002), from where surface winds uplift soil particles into the atmosphere (Gillette et al. 1980; Knippertz and Todd 2012). These particles may extend out of the atmospheric boundary layer into the mid- or upper-troposphere where they can be advected thousands of kilometers (Grousset et al. 2003). There is a lack of consensus on global dust emission, as was shown by Huneeus et al. (2011) who diagnosed a large spread between models of 500 – 4300 Tg yr$^{-1}$ global annual dust emission. 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. 2012b).

North Africa, and in particular the Sahara desert, is the world’s largest source of aeolian mineral dust, contributing to over 75% of the world’s annual budget (Ginoux et al. 2012b). Dust emissions from North Africa have a distinct seasonal cycle with the greatest areal extent occurring in Boreal summer and autumn (Adams et al. 2012). There is also evidence for significant interannual (Ridley et al. 2014) and multi-decadal scale variability (Mahowald et al. 2010) in dust transport from North Africa. For example, dust concentrations as measured at Barbados increased almost 5x between the 1960s and 1990s (Prospero and
Lamb 2003). Interestingly, this increase in dust cover occurred at a similar time as a decrease in tropical cyclone activity across the North Atlantic, particularly in the Western North Atlantic (Vecchi and Knutson 2011).

Dust is primarily emitted from distinct, highly localized sources, and these regions can vary in mineralogical composition (Claquin et al. 1999; Nickovic et al. 2012; Journet et al. 2014). Such variations are then translated to the type of mineral dust emitted and advected through the atmosphere (Caquineau et al. 2002) and lead to substantially different optical properties of the aeolian dust (Kim et al. 2011). There is significant spread among the observations of aeolian dusts’ single scattering albedos (SSAs) ranging from particles which are strong radiative absorbers to near perfect scatterers (Highwood and Ryder 2014). The choice of these optical properties have been shown to affect the simulated interactions between dust aerosols and radiation (Sokolik and Toon 1999; Balkanski et al. 2007). However, the vast majority of modeling studies are limited in that they only use one set of optical properties in their experiment and no study has even thought about the importance of first quantifying dust absorption versus scattering despite the range of mineralogical composition between sources.

Dust increases the shortwave attenuation of the atmosphere due to both scattering and absorption which leads to a decrease in net shortwave radiative flux at the surface (Tegen and Lacis 1996). Simultaneously, mineral dust absorbs and re-emits longwave radiation in the atmosphere (Volz 1973). These effects combine and lead to a redistribution of radiative heating of the atmospheric column (Miller et al. 2014). This effect can be further modulated by interactions of dust with other chemical aerosols (Ginoux et al. 2012a). In addition, mineral dust has been shown to affect cloud microphysics (Levin et al. 1996; DeMott et al. 2003) and biogeochemistry of both the land (Swap et al. 1992) and ocean (Jickells et al. 2005). These climatologically important changes can further modulate the impact of dust on such human-important phenomena as air pollution (Liu et al. 2009).
There have been a number of modeling studies attempting to quantify the impact of dust on climate. The earliest simulations with general circulation models (GCMs) showed that dust’s direct radiative impact was essentially focused in the local region, but hinted at far reaching effects through teleconnections (Miller and Tegen 1998). With more sophisticated parameterizations, the next generation of modeling studies began analyzing the effect of dust on the local hydrologic cycle. These results were mixed, with some studies showing a decrease in regional precipitation (Miller et al. 2004; Yoshioka et al. 2007; Yue et al. 2010, 2011; Mahajan et al. 2012) and others showing an increase in regional precipitation (Miller et al. 2004; Solmon et al. 2008; Lau et al. 2009; Gu et al. 2012; Solmon et al. 2012). Simultaneously, all of these models have described a decrease in surface temperature between the atmosphere and the ocean from dust. These models all had relatively simple representations of the oceans, most with passive (“prescribed sea surface temperature (SST)”) oceans, and the most sophisticated used a “slab” ocean model that responded only to local heat fluxes. None of these studies used a fully resolved ocean model, although significant perturbations to the physical ocean by dust are expected (Evan et al. 2012; Serra et al. 2014).

Additionally, there is a growing number of studies suggesting dust impacts tropical cyclone genesis and development, based on observations (Evan et al. 2006; Sun et al. 2008; Braun 2010) and models (Karyampudi and Pierce 2002; Zhang et al. 2007, 2009; Reale et al. 2014; Bretl et al. 2015). However, there are still a large number of uncertainties and thus no consensus on the sign of this impact. For instance, the development of thermally-driven easterly jets can act to increase the genesis frequency of tropical cyclones along the jet’s southern border (Braun 2010). Conversely, the significant drying and heating aloft, surface cooling, and increased wind shear can dampen the probability of tropical cyclone genesis (Dunion and Velden 2004). Some studies have attempted to link atmospheric dust load and tropical cyclone frequency on climatological timescales (Wang et al. 2012). However, none of the dynamical models used in these efforts resolved all of these processes in a consistent way
with fully-coupled atmosphere-land-ocean-ice models, nor did the global models explicitly resolve tropical cyclones.

The current literature covering the impact of dust on climate is severely hampered by a lack of model simulations exploring the coupled atmosphere-ocean response to dust as well as the sensitivity of this response to the chosen optical properties of dust. This dissertation presents several original contributions to the existing literature through its use of high-resolution, fully-coupled simulations of dust’s direct radiative impact on climate across a sensitivity range of dust optical properties. In Chapter 2, we conduct novel experiments utilizing a state of the art, fully-coupled GCM forced by realistic perturbations of the Saharan dust layer, meant to idealize the changes between the 1960s and the 1990s as observed at Barbados, and covering two ends of the spectrum of dust optical properties. We focus our attention to the impact of these perturbations on North Africa and the North Atlantic Ocean, where the majority of past studies have similarly focused. We show how dust, through direct radiative perturbations, influences the West African monsoon and circulations across the North Atlantic Ocean. The results of Chapter 2 are largely drawn from the works of Strong et al. (2014, 2015b).

In view of this potentially very important factor that dust could have on hurricane genesis and development, and the lack of prior consistent analysis, we present the first coherent study of dust radiative impact on tropical cyclones. In Chapter 3, we use a high resolution GCM with the same realistic dust forcing climatology as Chapter 2 alongside an expanded range of optical properties. At the same time, we extend our analysis to the entire globe. This leads to several discoveries of Saharan dust’s influence on the tropical precipitation around the world. A key aspect is the model resolution is fine enough for us to track tropical cyclone like vorticies (TCs) and simulate the impact of Saharan dust on the North Atlantic TC climatology, among other basins. The majority of Chapter 3 is described by Strong et al. (2015a, 2016a,b).
Finally, to understand how the choice of dust optical regime affects the mean climate state we present an innovative sensitivity analysis of differing dust optical properties in a fully-coupled modeling framework. In Chapter 4, we conduct a suite of experiments within a single fully-coupled GCM to explain the climate simulation’s sensitivity to various optical properties of dust. In explaining the differences between simulated model states, we find a significant impact of the choice of optical properties on the global atmospheric and oceanic circulation patterns and transports. Crucially, these changes can be on the order of or larger than those simulated simply by changing dust atmospheric burden. Most of this work has been presented in Strong et al. (2017).
Chapter 2

Effect of Aerosol Burden:
Low-Resolution Simulations

2.1 Preliminaries

Dust is one of the most abundant aerosols in the atmosphere with over 75% of global emissions arising from the Sahara desert (Ginoux et al. 2012b). There is a lack of consensus on global dust emission Huneeus et al. (2011), however it is well known that Saharan dust concentrations over the Atlantic have significant interannual and decadal scale variations (Chiapello et al. 2005; Evan and Mukhopadhyay 2010; Ridley et al. 2014). For instance, atmospheric dust concentrations as observed at Barbados increased nearly 5-fold 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 dissertation. Due to the variability of its mineralogy, the impact of dust on radiation (Boucher et al. 2013) may vary greatly in time and space. The choice of dust mineralogy dataset to use, 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). Unfortunately, there is still uncertainty about the mineral-
ogy dependent optical properties of dust and little information on variable dust mineralogy is incorporated into modeling experiments. Instead of solving for dust mineralogy, we will perform a sensitivity analysis covering a range of dust absorption and scattering.

A number of modeling studies have attempted to define the impact of dust on climate by using everything from simple radiative column models (Miller 2012) through single-component atmosphere (Miller and Tegen 1998) or ocean GCMs (Serra et al. 2014). Due to the large contribution of Saharan dust to the global aerosol budget, many studies focused on analyzing the effect of dust on the hydrologic cycle of North Africa with some studies showing a decrease in regional precipitation (Miller et al. 2004; Yoshioka et al. 2007; Yue et al. 2010, 2011; Mahajan et al. 2012) and others showing an increase 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 2.1 and plotted in Figure 2.1 alongside the effects of specifically Saharan-borne dust which are calculated in this chapter. There is a large spread between models, both in terms of the projected African monsoon response and the SSAs used. A significant limitation to these modeling studies is the majority of them are limited in the use of only one set of optical properties. In addition, no modeling studies have been conducted using a full ocean model coupled to an atmospheric GCM.

The purpose of this chapter 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 multi-centenial multi-member ensemble runs with a 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
Table 2.1: 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 $\mu$m dust particles at 550 nm used in each model, derived from either in-text tabulation when provided or through model paper references. The precipitation anomalies and associated bounds (mm day$^{-1}$) of each model result are derived either from in-text tabulation when provided (italicized) or from averaging plotted results and calculating the regional range of response. The results of this chapter 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>$\delta P$</th>
<th>$\delta P_{\text{upper}}$</th>
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<tr>
<td>Coakley and Cess (1985) [1x Dust Climo.]</td>
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<td>0.5</td>
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<td>0.3</td>
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<td>Yoshioka et al. (2007) [AMIP.SL]</td>
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<td>-0.15</td>
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<tr>
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<td>-0.22</td>
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<td>Solmon et al. (2008) [0.95*SSA]</td>
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<td>-0.1</td>
<td>0.275</td>
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<tr>
<td>Solmon et al. (2008) [1.0*SSA]</td>
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<td>-0.4</td>
<td>0.05</td>
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<td>Solmon et al. (2008) [1.05*SSA]</td>
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<td>-0.1</td>
<td>0.275</td>
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<td>-0.6</td>
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<td>0.521</td>
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<tr>
<td>SCT-Dust</td>
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<td>0.99261</td>
<td>-0.75</td>
<td>-0.4</td>
<td>-0.06</td>
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</table>
Figure 2.1: The response of the West African monsoon to global dust in several modeling studies using various ocean configurations. The squares and associated error bars represent the precipitation anomalies and confidence intervals, respectively, listed in Table 2.1 and progress from oldest to newest study from left to right in each column. The color of each square and color bar at right indicate the single scattering albedo (SSA) of 1 \( \mu \text{m} \) dust particles at 550 nm used in each model, derived from either in-text tabulation when provided or through model paper references. The bold circles represent the results of the two experiments run in this chapter, ABS-dust (lower SSA) and SCT-dust (higher SSA), and are representative of 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. The bar in each column indicates the model average for each ocean model configuration.
ocean influences the African monsoon. In addition, we will seek to understand the coupled atmosphere-ocean response in the tropical North Atlantic.

This chapter is organized as follows. In Section 2.2 we describe the model and datasets used in this chapter as well as the experimental design. In Section 2.3 we analyze the direct radiative effect of dust. In Section 2.4 we explore several regional hydrologic and thermodynamic effects of dust on the climate system. Part 2.4.1 covers the impact of cloud and precipitation anomalies over North Africa and the tropical Atlantic and how those changes affect the local energy balance. Part 2.4.2 focuses on the West African monsoon system and associated local circulations. Part 2.4.3 describes the effect in the tropical North Atlantic Ocean and the coupled atmospheric-oceanic response. Section 2.5 gives a discussion of the presented results and concluding remarks. These results have all been pulled from Strong et al. (2014, 2015b).

2.2 Methodology

2.2.1 Coupled Model

For this chapter, we used the Geophysical Fluid Dynamics Laboratory’s (GFDL) Climate Model v2.1 (CM2.1) (Delworth et al. 2006) and perturb the concentration of mineral dust atmospheric burden, exploring the sensitivity based on two of the optical regimes of mineral dust aerosols. The GFDL’s 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 CMIP3 models (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) 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 only the aerosol direct effect is considered.

The ocean model (OM3.1) is based on the Modular Ocean Model v4 (MOM4) code (Griffies et al. 2003). The horizontal domain uses a tripolar grid with a 1° latitude by 1° longitude horizontal resolution which enhances to 1/3° within 10° 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 on 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 a 2-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).

### 2.2.2 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 (MOZART) v2 (Horowitz et al. 2003; Tie et al. 2005) forced by NCEP Reanalysis of 1990 (Jickells et al. 1996) (Fig. 2.2a). 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. In order to assess the impact of variations in dust concentrations, we utilized an additional MOZARTv2 simulation forced by NCEP Reanalysis of 1990 wherein the dust emission over North Africa is reduced to 20% of its 1990 value (Fig. 2.2b). 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 Figure 2.2c.

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 regime 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. For simplicity, we shall henceforth refer to this regime as the ABS-dust regime. 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 less absorbing and is calculated using Mie theory for highly scattering dust. The values of the refractive index are taken from Balkanski et al. (2007) as their methodology provides values for the full solar spectrum compared to Sinyuk et al. (2003) which only provides values 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 and set the longwave refractive index to a constant value. As this regime is less absorbing and far more scattering than the ABS-dust regime, we shall henceforth refer to it as the SCT-dust regime. Figure 2.3 shows the extinction coefficient, single scattering albedo, and the asymmetry parameter for both optical regimes for fine
Figure 2.2: The climatology of dust aerosol optical thickness (AOT) used in this experiment. The column integrated dust AOT is calculated with the ABS-dust optical properties at 550 nm (Fig. 2.3) and averaged over the months June, July, and August. The top panel (a) shows the atmospheric AOT used in the base run. The middle panel (b) shows the atmospheric AOT using the perturbed (reduced) emission dataset. The bottom panel (c) shows the difference between these two climatologies.
(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.

2.2.3 Experimental Design

We form two control runs by forcing the model with an 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. In order 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 year 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.

2.3 Direct Radiative Effects

2.3.1 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 (JJA), when the
Figure 2.3: 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 µm, 0.2 µm, 0.4 µm, and 0.8 µm; dashed lines) and the coarse (large) dust bin sizes (1 µm, 2 µm, 4 µm, and 8 µm; solid lines) of the ABS-dust regime (red lines) and the SCT-dust regime (blue lines). The top panel (a) shows the extinction coefficient (m² g⁻¹) as a function of wavelength. The middle panel (b) shows the single scattering albedo of the dust as a function of wavelength. The bottom panel (c) shows the asymmetry parameter of the dust as a function of wavelength.
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 they represent 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-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. 2.4a). 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. 2.4b). 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 the same 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, the eastern and equatorial tropical North Atlantic with a peak around 10 W m\(^{-2}\) along the southern Sahara (Fig. 2.4c). 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. 2.4c). 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. 2.4d). 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
Figure 2.4: The top of atmosphere, clear-sky radiative anomalies due to dust with differing optical properties (Positive downwards; $W \text{ m}^{-2}$). The values are averaged over JJA and only those values that pass a Student’s t-test to 95% are shaded. The left column shows the ABS-dust simulation results while the right column shows the SCT-dust simulation results. The top row (Panels a & b) shows the total shortwave flux anomaly at the top of the atmosphere. The middle row (Panels c & d) shows the total longwave flux anomaly at the top of the atmosphere. The bottom row (Panels e & f) shows the total radiative flux anomaly at the top of the atmosphere.
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 2.4.1.

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 $30 \text{ W m}^{-2}$ (Fig. 2.4e), and a general decrease in the SCT-dust simulation of up to $-20 \text{ W m}^{-2}$ (Fig. 2.4f). Both simulations result in a general decrease of clear-sky total radiative flux across the tropical North Atlantic of the order $-5 \text{ W m}^{-2}$. 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).

2.3.2 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. 2.5) 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 $-45 \text{ W m}^{-2}$ along the West coast of North Africa (Fig. 2.5a). The same effect occurs to the clear-sky SCT-dust SW radiative fluxes which decrease by a more modest $-10 \text{ W m}^{-2}$ across North Africa and the North Atlantic, particularly near the West African coastline (Fig. 2.5b) due 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 $25 \text{ W m}^{-2}$ (Fig. 2.5c). 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 $-15 \text{ W m}^{-2}$ along the Southern Sahara and Sahel (Fig. 2.5d). This is again due to the increased surface temperature in this region as will be discussed in Section 2.4.1.
Figure 2.5: Same as in Figure 2.4, but for the surface.
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 focus in the eastern tropical North Atlantic with decreases up to $-30 \text{ W m}^{-2}$ (Fig. 2.5e). Similarly, in the SCT-dust simulation there are negative anomalies in total radiative flux of almost $-20 \text{ W m}^{-2}$ focused primarily along the West coast of North Africa and along the Sahel (Fig. 2.5f). 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).

2.4 Regional Hydrologic and Thermodynamic Effects

2.4.1 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. 2.6). 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 $-10\%$ (Fig. 2.6a). There is a generally more muted change in mid-level clouds with decreases across the western Atlantic and Sahara of around $-5\%$ and a strong increase over the region between $5^\circ\text{N}-15^\circ\text{N}$ and $15^\circ\text{W}-15^\circ\text{E}$ of near $15\%$ (Fig. 2.6b). Because of it’s importance in later analyses, we will call this region ($5^\circ\text{N}-15^\circ\text{N}$ and $15^\circ\text{W}-15^\circ\text{E}$) the West African Monsoon region (WAM). The low cloud anomalies are much more heterogeneous over the Atlantic, but over the western WAM region increase by almost $15\%$ (Fig. 2.6c). In combination, the total cloud amount increases over much of North Africa and equatorial Atlantic, by as much as 15%, and decreases across the central and western North Atlantic by about $-10\%$, and appears to be a strong function of the high cloud changes (Figs. 2.6a &
2.6d). 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. 2.7). 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. 2.7a). Again, the response in mid-level 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. 2.7b). 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. 2.7c). 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. 2.7d). These anomalies indicate a significant decrease in deep convection over North Africa and 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 mm day\(^{-1}\) and 3 mm day\(^{-1}\), respectively (Fig. 2.6e). 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. 2.6e). 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 \(10^\circ\)N with peaks of \(-2.5\) mm day\(^{-1}\), \(-3\) mm day\(^{-1}\), and \(-2\) mm day\(^{-1}\), respectively (Fig. 2.7e). These changes
Figure 2.6: The cloud field (%), precipitation (mm day\(^{-1}\)), and soil moisture (kg m\(^{-2}\)) response due to 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. The top left panel (a) shows the high-level cloud amount anomaly. The top right panel (b) shows the mid-level cloud amount anomaly. The middle left panel (c) shows the low-level cloud amount anomaly. The middle right panel (d) shows the total cloud amount anomaly. The bottom left panel (e) shows the total precipitation anomaly. The bottom right panel (f) shows the soil moisture anomaly.
Figure 2.7: Same as in Figure 2.6, but for the SCT-dust simulation.
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. 2.7e). These changes have the potential to be a non-negligible feedback on the amount of dust aerosols in the atmosphere due 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 non-negligible 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. 2.6f). Meanwhile, the decreases in precipitation over the Americas leads to reductions in soil moisture of order $-30$ kg m\(^{-2}\) over North America and order $-10$ kg m\(^{-2}\) over South America (Fig. 2.6f). Conversely, in the SCT-dust simulation there are decreases primarily in the WAM region of up to $-20$ kg m\(^{-2}\) due to the strong reductions in precipitation in that area (Fig. 2.7f). Again, the Americas show the opposite picture with strong increases of soil moisture content nearing 20 kg m\(^{-2}\) in accordance with the increases in precipitation (Fig. 2.7f). 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 this subsection.

**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 2.3, 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 (Fig. 2.8a&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. 2.8a) whereas in the SCT-dust simulation the anomaly is of the order $15 \text{ W m}^{-2}$ (Fig. 2.8b). 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 due to 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 (Fig. 2.8a). The changes north of $10^\circ\text{N}$ 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 $10^\circ\text{N}$ match the total cloud amount anomalies better 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. 2.8b), but again matches decently with the observed cloud amount anomalies.

Now, considering the all-sky longwave (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. 2.8c). 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 allow more LW radiation to escape to space (Fig. 2.8d).
Figure 2.8: The top of atmosphere, all-sky radiative anomalies due to dust with differing optical properties (Positive downwards; W m$^{-2}$). The values are averaged over JJA and only those values that pass a Student’s t-test to 95% are shaded. The left column shows the ABS-dust simulation results while the right column shows the SCT-dust simulation results. The top row (Panels a & b) shows the total shortwave flux anomaly at the top of the atmosphere. The middle row (Panels c & d) shows the total longwave flux anomaly at the top of the atmosphere. The bottom row (Panels e & f) shows the total radiative flux anomaly at the top of the atmosphere.
Across the tropical North Atlantic, just South of 10°N, there is a significant decrease in all-sky LW radiative flux of about $-10 \text{ W m}^{-2}$ (Fig. 2.8c). 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 which 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. 2.8e) and SCT-dust (Fig. 2.8f) 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. 2.8e) leads to a more heterogeneous response than the SCT-dust simulation (Fig. 2.8f) with the latter having a band of decreased top of atmosphere radiative flux across much of the central tropical North Atlantic.

All-Sky Surface Radiative Fluxes

Looking now at the all-sky surface radiative flux (Fig. 2.9), 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 (Fig. 2.9a). 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. 2.9a). In the SCT-dust simulation, we notice a constricting of the SW radiative flux anomaly over the tropical North Atlantic with positive flux anomalies near the equator and 30°N (Fig. 2.9b). 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\(^\circ\)N-30\(^\circ\)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. 2.9c). Conversely, in the SCT-dust simulation there is a modest decrease in line with the loss of cloud amount in that region (Fig. 2.9d). There are also small changes to the anomaly in the WAM region due to the increase (decrease) of cloudiness in that area in the ABS-dust (SCT-dust) simulation.

Figure 2.9: Same as in Figure 2.8, but for the surface.
When combined, the SW term dominates in both the ABS-dust and the SCT-dust simulations and we observe all-sky total radiative fluxes which are more heterogeneous versions of the clear-sky radiative fluxes (Fig. 2.5e&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.

**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 $-40 \text{ W m}^{-2}$ along the Sahel and an increase in sensible heat flux across the Americas (Fig. 2.10a). 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. 2.10a&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 $35 \text{ W m}^{-2}$ while over the WAM region there are decreases of $-30 \text{ W m}^{-2}$ which extend out over the tropical North Atlantic and Gulf of Guinea under the dust plume and across the Americas (Fig. 2.10b). South of the equator there is a significant increase in latent heat flux of up to $25 \text{ W m}^{-2}$ focused on the eastern portion of the tropical South Atlantic (Fig. 2.10b).

These anomalies over North Africa appear to be associated with the precipitation (Fig. 2.6e) and soil moisture anomalies (Fig. 2.6f). 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. 29).
Figure 2.10: The surface turbulent flux balance (Positive upwards; W m$^{-2}$) and surface temperature (K) change due to 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. The top left panel (a) shows the sensible heat flux anomaly at the surface. The top right panel (b) shows the latent heat flux anomaly at the surface. The bottom left panel (c) shows the total energy flux anomaly at the surface. The bottom right panel (d) shows the surface temperature anomaly.
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. 2.9e) 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 a half a degree (Fig. 2.10d).

These turbulent flux terms appear to balance the all-sky direct radiative anomalies over the Sahara (Fig. 2.10c). However, over much of the North Atlantic there is a rather heterogeneous all-sky net energy flux anomaly (Fig. 2.10c). 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. 2.10c). This is also associated with a significant warming in the southeastern tropical South Atlantic of up to $1$ K (Fig. 2.10d). 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 the Sahara and increases peaking at $20$ W m$^{-2}$ in the WAM region (Fig. 2.11a). There is also a significant negative sensible heat flux anomaly across the Americas (Fig. 2.11a). 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 a positive anomaly over the WAM region of $10$ W m$^{-2}$ (Fig. 2.11b). 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. 2.11b).

Again, these anomalies are associated with decreases in precipitation (Fig. 2.7e) and soil moisture content (Fig. 2.7f) 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. 2.11d). 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. 2.9f) 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 2.11c). However, there is a strong positive anomaly along the equatorial Atlantic peaking around 30 W m$^{-2}$ (Fig. 2.11c). 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.

Figure 2.11: Same as in Figure 2.10, but for the SCT-dust simulation.
2.4.2 West African Monsoon Response

We will now turn our focus towards 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, which has been cooled by the absorption of radiation above, and thus enhancing precipitation. However, Lau et al. (2009) state this effect becomes minimized for reflecting dust with SSA at 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.

Zonal and Meridional Diagnostics

To understand the local hydrologic and circulation anomalies we plot similar figures to Fig.5 and Fig.7 of Lau et al. (2009) to describe changes to the West African Monsoonal system (Figs. 2.12 & 2.13). 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 over the African continent between 20°W and 20°E (Fig. 2.12a). In the ABS-dust simulation there is a westerly wind anomaly through much of the dust layer, reaching over 2 m s⁻¹ above the coast of the WAM region (Fig. 2.12a). 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 the coast of West Africa where there is a sharp increase in the precipitation amount (Fig. 2.12c). The dust plume is also associated
with a significant positive temperature anomaly throughout much of the troposphere (Fig. 2.12e). 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. 2.12e). 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 2 separate jets, one to the north and one to the south of the main dust loading (Fig. 2.12b). These are accompanied by a significant westerly anomaly at the surface beneath the dust plume. 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. 2.12d). This is supported by the strong area of lift along the southern flank of the dust plume (Fig. 2.12f). There is a significant warming trend through and to the North of the dust plume (Fig. 2.12f). 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 aerosol optical thickness (AOT) through the plume (Fig. 2.13a). The zonal wind
Figure 2.12: The effect of dust over the West African Monsoon 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. The left column presents a zonal view of the meridional average between 5°N-15°N. The right column presents a meridional view of the zonal average between 10°W-10°E. The top row (Panels a & b) shows the dust optical thickness (shading) and the zonal wind anomaly (m s⁻¹; contoured every 1 m s⁻¹). The middle row (Panels c & d) shows the precipitation anomaly (mm day⁻¹). The bottom row (Panels e & f) shows the wind streamlines and potential temperature anomaly (K, shading and contoured every 0.5 K).
response is primarily easterly through much of the lower and mid-troposphere, in contrast to the ABS-dust case (Fig. 2.13a). 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. 2.13c). Analyzing the temperature response, there is significant cooling throughout the troposphere over West Africa and the tropical North Atlantic while a slight surface warming over the WAM region (Fig. 2.13e). 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. 2.13e). This subsidence corresponds with the decrease in precipitation over West Africa. There are also areas of rising motion around 80°W and 60°W-40°W which align with the areas of slight positive anomalous precipitation over the tropical North Atlantic (Figs. 2.13c & 2.13e).

From the meridionally averaged perspective, there are significant easterly wind anomalies South, 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. 2.13d). In fact, the strongest decreases in precipitation align with major areas of subsidence in the meridional plane (Fig. 2.13f). The dust causes a significant cooling aloft with a strong, but constrained warming at the surface (Fig. 2.13f). 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 precipitation over the region. Interestingly, the SCT-dust simulation shows a comparable magnitude anomaly to the ABS-dust simulation, which is inconsistent with the mechanism proposed by Lau et al. (2009).

**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
Figure 2.13: Same as in Figure 2.12, but for the SCT-dust simulation.
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:

\[ F^{\text{net}} = SW^{\downarrow}_{\text{ToA}} - SW^{\uparrow}_{\text{ToA}} - SW^{\downarrow}_{\text{Sfc}} + SW^{\uparrow}_{\text{Sfc}} - LW^{\uparrow}_{\text{ToA}} - LW^{\downarrow}_{\text{Sfc}} + LW^{\uparrow}_{\text{Sfc}} + SH + LH \]

where the superscripts 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. 2.14a). Simultaneously, there is an anomalous net divergence over the WAM region and eastern tropical North Atlantic, dropping to \(-20\) W m\(^{-2}\) (Fig. 2.14a). 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 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. 2.14b). 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. 2.14b). The anomalies over North Africa are again due to the top of the atmosphere radiative flux anomalies adjusting slower 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. 2.6e & 2.7e) to the respective net energy convergence figures (Fig. 2.14) 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. 2.14a) is coincident with rising motion to compensate for the increased column diabatic heating (Fig. 2.12f) and an increase in precipitation (Fig. 2.12d). Conversely, in the SCT-dust simulation, the negative net column energy convergence over these same regions (Fig. 2.14b) is associated with subsidence to compensate for the column diabatic cooling (Fig. 2.13f) and a modest decrease in precipitation (Fig. 2.13d). However the precipitation anomaly is limited in these cases possibly due to 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. 2.6e) and negative precipitation anomalies (Fig. 2.7e) 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 non-steady 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.

2.4.3 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. 2.10d), whereas the SCT-dust simulation showed a thinner plume of cold surface water, again extending from the central West African coast (Fig. 2.11d). 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 chapter we integrate from the surface down to 400 m depth. We plot this quantity across the Atlantic basin in Figure 2.15 (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. 2.15a). 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. 2.15a). Conversely, the SCT-dust simulation
Figure 2.15: The tropical North Atlantic Ocean upper ocean heat content response due to dust with differing optical properties. The values are averaged over JJA and only those values that pass a Student’s t-test to 95% are contoured. The top row (Panels a & b) shows the upper ocean heat content anomaly (kJ cm$^{-2}$; shading) for the ABS-dust optical regime (Global average of $-4.36$ kJ cm$^{-2}$) and the bottom row (Panels c & d) shows the same for the SCT-dust optical regime (Global average of $-11.71$ kJ cm$^{-2}$). The left column (Panels a & c) shows the total surface energy flux response (Positive downwards; W m$^{-2}$; contours every 5 W m$^{-2}$) and the right column (Panels b & d) shows the wind stress curl response (N km$^{-3}$; contours every 10 N km$^{-3}$).
shows significant heating across the central tropical North Atlantic, with heating reaching up to 80 kJ cm$^{-2}$ (Fig. 2.15c). 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 towards the equator (Fig. 2.15c).

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 Figures 2.15a & 2.15c (black contours; positive downwards). 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 doesn’t 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 Figures 2.15b & 2.15d (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 we see a strong decrease in UOHC (Fig 2.15b). 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. 2.15d). 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. 2.16). We observe that in the summer months, the addition of ABS-dust causes the ITCZ to shift poleward by almost a degree, inducing a positive wind stress curl anomaly further North. Conversely, the addition of
Figure 2.16: The anomaly in the location of the Atlantic Intertropical Convergence Zone, as defined by the location of zero meridional wind stress, due to dust with differing optical properties. Only those values that pass a Student’s t-test to 95% are contoured. The red line is the anomalous monthly shift of the ITCZ from the ABS-dust forced experiment base run climatology. The blue line is the anomalous monthly shift of the ITCZ from the SCT-dust forced experiment base run climatology.

SCT-dust causes the ITCZ to shift equatorward by more than a half of a degree 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. 2.6e). This could indicate a weakening of the Atlantic ITCZ with an apparent shift to the North, potentially due to the increase in low level entropy as indicated in Figure 2.12e. 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 further South (Fig. 2.7e). This may be due to the strong decrease in entropy throughout the column over the Atlantic (Fig. 2.13e) which would act to push the center of convection further South.
In line with the observed changes in UOHC, we plot the changes to the ocean mixed layer depth in Figure 2.17. 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 Figures 2.15b&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°N-20°N signifying a shallowing of the mixed layer in this region (Fig. 2.17a). This is in line with the positive wind stress curl anomaly over this same band. Conversely, there is a strong positive anomaly between 20°N-30°N of over 4m 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°N-20°N of up to 3 m and a negative anomaly between 20°N-30°N of around −3 m (Fig. 2.17b). These changes translate to a deepening and a shallowing of the mixed layer, respectively. Such anomalies are significant to note as often 1-dimensional models of the ocean temperature’s response to dust perturbations prescribe the mixed layer depth (Miller 2012).

Also in Figure 2.17 are plotted the surface current anomaly (streamlines) averaged over the top 100 m of the ocean. For both optical regimes, we note an anomalous anticyclonic circulation develop around the positive mixed layer depth anomalies and a cyclonic circulation develop 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. 2.17a). 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 equa-
Figure 2.17: The tropical North Atlantic Ocean mixed layer depth and surface current response due to dust with differing optical properties. The values are averaged over JJA and only those values that pass a Student’s t-test to 95% are contoured. The left panel (a) shows the mixed layer depth anomaly (m; shading) and anomalous current streamlines in the top 100 m of the ocean for the ABS-dust optical regime. The right panel (b) shows the same for the SCT-dust optical regime.

tor bringing surface waters North in eastern portion of the basin and South in the western tropical Atlantic (Fig. 2.17b).

2.5 Chapter Summary

Saharan dust aerosols can produce one of the largest regional aerosol radiative forcing in the world. Previous studies have attempted to quantify the impact of this forcing on climate, but have so far only explored the effect in single components of the coupled Earth system. Using the fully coupled GFDL CM2.1, we have defined 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 the atmosphere clear-sky radiative fluxes over North Africa between the more absorbing and more scattering dust simulations. At the surface, any radiative anomalies are dominated by the SW response to dust, despite differences in LW flux, leading to a general local dimming. We also noted opposing responses in cloud fields and precipitation patterns between absorbing and scattering dust 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 corresponded 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 more absorbing dust causes a significant increase in precipitation whereas more scattering 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, respectively. However, the column energy convergence criteria is not as precise over the region between the Sahel and the Gulf of Guinea and could be due to a number of factors including non-steady circulations and the oceanic transport of energy to offset the aerosol induced heating or cooling.

Over the tropical North Atlantic ocean, our simulations again show opposing results. When using more absorbing dust, the central tropical North Atlantic UOHC decreases significantly whereas the UOHC increases markedly when using more scattering dust. The main contribution to this impact is the dynamical ocean response to a shifting ITCZ. The largely absorbing 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 primarily scattering dust cloud causes the ITCZ to shift southward inducing a negative wind stress curl anomaly, downwelling of the warm surface ocean, and a deepening of the mixed layer. In addition, the varying optical regimes lead to non-negligible surface current anomalies which 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.

These results have enormous implications for the meteorology of North Africa and the tropical North Atlantic. The simulated changes in the West African monsoon as well as the
heat budget of the North Atlantic could lead to anomalies in such features as hurricanes and other non-steady state circulations. In the next chapter, we expand upon these results by running similar experiments in a higher resolution GCM. In addition, we will increase the range of optical properties used to see if we can pinpoint the crossover regime between our aforementioned opposing results. This will hopefully shed light on Saharan dust’s impact on more transitory features of the climate system.
Chapter 3

Effect of Aerosol Burden:
High-Resolution Simulations

3.1 Preliminaries

In the previous chapter, we have seen that Saharan dust, the single largest contributor to global aeolian dust (Ginoux et al. 2012b), can significantly impact the climate when the atmospheric burden of dust aerosols is varied the match the natural interannual (Ridley et al. 2014) and multi-decadal (Mahowald et al. 2010) scale variations. In addition, Saharan-born dust is commonly not of a single homogeneous composition, but is instead a combination of multiple regional mineral deposits (Caquineau et al. 2006), and we have also seen that the choice of mineralogy dataset can considerably affect the calculated interaction of dust with climate (Strong et al. 2014, 2015b). This is compounded by results which show mineral dust can affect such human-important phenomena such as air pollution (Liu et al. 2009) and tropical cyclone frequency (Dunion and Velden 2004).

Several observational based studies have attempted to come up with a concrete answer to the question of whether dry, dusty air associated with the Saharan Air Layer, an air-mass which advects westward from North Africa, positively or negatively impacts tropical
cyclone genesis and development across the North Atlantic (Evan et al. 2006; Sun et al. 2008; Braun 2010; Wang et al. 2012). Unfortunately, no consensus has been reached due in part to observational constraints or conflicting impacts on potential genesis parameters. For instance, it is hard to directly cite dust as the cause for apparent changes in tropical cyclone evolution when only associating variations in hurricanes with the proximity of nearby dusty air. Additionally, the competition between such features as enhanced mid-level warming and drying from dust as well as increased wind shear and positive vorticity anomalies from an enhanced African Easterly Jet make it hard to balance the probability of tropical development. Some have tried to resolve these issues with model simulations of dust’s impact on hurricane development (Karyampudi and Pierce 2002; Zhang et al. 2007, 2009; Reale et al. 2014; Bretl et al. 2015), but all have been limited to individual tropical cyclones or single seasons. Wang et al. (2012) attempted to link increases in atmospheric dust load to decreases in North Atlantic tropical cyclone frequency on decadal timescales, but were unable to fully explore a coherent cause for the relationship. There has been little to no work in modeling these relationships in a fully-coupled framework on climatologically relevant timescales, nor with global modeling frameworks that resolve tropical cyclones.

The purpose of this chapter then is to investigate the coupled atmospheric and oceanic climatological effect of Saharan dust on tropical cyclones. To this end, we will use a state of the art general circulation model run at a sufficiently high resolution to adequately resolve tropical cyclone like vorticies (TCs). Our hypothesis is that Saharan dust will cause a significant decrease in tropical cyclone frequency across the North Atlantic Ocean for the full range of investigated optical regimes and with diminishing significant results farther from the main dust cloud.

The chapter is organized as follows. In Section 3.2 we describe the model, computations, and datasets used in this chapter as well as the experimental design. In Section 3.3 we discuss the global radiation, surface temperature, and precipitation response to perturbations in Saharan-born dust, principally for comparison with other studies. In Section 3.4 we ana-
lyze the global changes to tropical cyclone density and compare them with the observational record as well as several common genesis forecast indicies. In Section 3.5 we explore changes to several regional factors which could influence the modeled anomalies in tropical cyclogenesis and development and craft several hypotheses for the modeled changes. In Section 3.6 we present a discussion of the results and our concluding remarks. The majority of the results in this chapter are from Strong et al. (2015a, 2016a,b).

3.2 Methodology

The methodology presented next generally follows the same experimental protocol as in Chapter 2. It features improvements to both the modeling framework and sensitivity range of optical properties used for modeling dust’s radiative forcing as well as streamlining the calculation of dust’s impact on the climate.

3.2.1 Coupled Model

For this chapter, we used the GFDL Coupled Model CM2.5 Forecast-oriented Low Ocean Resolution version (CM2.5-FLOR) (Vecchi et al. 2014) to calculate the effect of perturbations to the atmospheric aerosol burden of dust under various optical regimes. GFDL’s CM2.5-FLOR is an offshoot of the previous Coupled Model v2.5 (CM2.5) (Delworth et al. 2012) developed at GFDL and uses many of the same model configurations. The atmosphere and land components have a horizontal resolution of 0.5 by 0.5 degrees using a cubed sphere, finite volume dynamical core. The ocean and ice components use a lower 1 degree horizontal resolution which enhances to 1/3 degree near the equator. This model framework shows substantial improvement to its regional hydroclimate simulations relative to the previous CM2.1 and marginal improvements relative to CM2.5 (e.g., Jia et al. (2015)).
3.2.2 Dust Forcing

We use the same prescribed climatological annual cycles of monthly, global mineral dust aerosol burden as in Figure 2.2. As a reminder, both climatological cycles were calculated using the Model of Ozone and Related Chemical Tracers, version 2 (MOZART-2) (Horowitz et al. 2003; Tie et al. 2005; Horowitz 2006) forced by the NCEP-NCAR reanalysis of 1990 (Jickells et al. 1996) and are meant to represent the high and low Saharan dust concentrations of the 1990s (Base Emission) and 1960s (Reduced Emission).

We increased the range of optical properties used to calculate the modeled radiative forcing of dust from the prior ABS-dust and SCT-dust regimes to now incorporate six optical regimes, the properties of which are detailed in Figure 3.1. The first regime (V&P) is derived from a combination of the in-situ observations of Volz (1973) and Patterson et al. (1977). This combination has been used in previous GFDL climate models (Anderson et al. 2004) and is the same as the prior ABS-dust of Chapter 2. The remaining optical regimes are calculated using Mie theory with the refractive indices given by Balkanski et al. (2007). We consider the cases of 2.7hem, 1.5hem, and 0.9hem. In addition, we create two other artificial optical regimes aimed at replicating extremely-scattering mineral dust. For the first regime, we multiply the imaginary part of the refractive index of the 0.9hem case by 0.1 (0.1x0.9hem) and for the second regime we set the longwave refractive index of the 0.1x0.9hem case to a constant value (0.1x0.9hem [Fixed LW]). This latter optical regime is the same as the prior SCT-dust case of Chapter 2.

3.2.3 Experimental Design

The model is initialized in a spun-up state following 100 years of forcing with an invariant 1990 climatology of insolation, gas and aerosol concentrations, and land cover as described by Vecchi et al. (2014). We create a control state for each of the six optical regimes using the base emission climatology and run each individually for 100 years to allow the climate system to further equilibrate. We then run each control simulation for a further 200 years. Using the
Figure 3.1: The optical properties of all regimes of dust used in this experiment. The values are averaged over the fine (small) dust bin sizes (0.1 $\mu$m, 0.2 $\mu$m, 0.4 $\mu$m, and 0.8 $\mu$m; dashed lines) and the coarse (large) dust bin sizes (1 $\mu$m, 2 $\mu$m, 4 $\mu$m, and 8 $\mu$m; solid lines). The top panel (a) shows the extinction coefficient (m$^2$ g$^{-1}$) as a function of wavelength. The middle panel (b) shows the single scattering albedo of the dust as a function of wavelength. The bottom panel (c) shows the asymmetry parameter of the dust as a function of wavelength.
climate state at years 100 and 200 as initialization we create two perturbation simulations for each optical regime where we keep the optical properties the same, but change the dust climatology for the annual cycle of atmospheric dust load to the reduced emission state. Each of these perturbations is allowed to run for 100 years in parallel to their respective control simulations. To calculate the impact of dust, we align the control and perturbation simulations by equivalent model year and then subtract the perturbation simulations from their respective control simulations to arrive at a value representative of the impact of adding a realistic amount of Saharan dust aerosol burden to the atmosphere.

### 3.2.4 Tropical Cyclone Tracker

Using the model’s six-hourly output, we track simulated tropical-cyclone-like vorticies (TCs) offline with the method described by Zhao et al. (2009) and the parameter settings of Kim et al. (2014). In calculating the density of TCs, we define the TC track density as the number of days with a tropical-cyclone-like vortex present in a box 10 by 10 degrees centered on each 1 degree grid box. This is for several reasons as described by Vecchi et al. (2014): The size of the 10x10 degree grid is much smaller than the scale of the basins, on a scale comparable to the average diameter of observed TCs (Chavas and Emanuel 2010), and is large enough to include most of the areas where impacts of individual TCs in models are evident (Lin et al. 2010). The choice of grid size also minimizes the impact of the edges of larger discrete boxes in computing density, effectively smoothing the field of TC density. Similarly, we define the TC genesis density as the number of storms spawned in the same 10 by 10 degree box.

To focus on the mean response, we implement a bootstrap, without replacement, style calculation to determine the dust-induced changes in TC density and genesis. For each model grid point, we construct a time series of the annual sum of TC density or genesis in both the base and reduced dust cycle experiments. The order of these time series is randomized and then the reduced dust series is subtracted from the base dust series and the result is averaged. This process is repeated 1,000,000 times on the original time series and the resultant set
of 1,000,000 values is averaged to produce a single value for that grid point. This extreme repetition guarantees statistical convergence of the final solution and while the results are not fundamentally different from doing a strict anomaly calculation between the base and reduced dust climatologies, they do highlight the most significant responses.

For clarity, we define several basins across the globe. The North Atlantic is all ocean grid points from the Americas east to Africa/Europe, north of the Equator. The Eastern North Pacific is all ocean grid points from 160W east to the Americas, north of the Equator. The Western North Pacific is all ocean grid points from Asia east to 200E, north of the Equator. The North Indian is all ocean grid points in the Indian Ocean north of the Equator. The South Atlantic is all ocean grid points from the Americas east to Africa, south of the Equator. The South Pacific is all ocean grid points from 165E east to South America, south of the Equator. The Australian Basin is all ocean grid points from 105E east to 165E, south of the Equator. The South Indian is all ocean grid points from Africa east to 105E, south of the Equator. And finally, the Indian Ocean basin is all ocean grid points from Africa east to the maritime continent.

### 3.3 Mean Climate State Anomalies

For comparison with other studies, we begin by analyzing the Saharan dust induced changes to the global climate as detailed in several common radiative and thermodynamic variables. The following results are all averaged over the Boreal summer and autumn seasons (JJASON) as this is the time of peak areal extent of the Saharan dust plume (Adams et al. 2012) as well as the climatological peak of Northern Hemisphere tropical cyclone season. The general results do not change significantly when considering the annual average.
3.3.1 ToA Net Clear-Sky Radiative Flux Anomalies

We observe in our model simulations that Saharan-born dust causes significant anomalies in the net clear-sky radiative flux at the Top of the Atmosphere (ToA) (Fig. 3.2, first column). As would be expected from clear-sky calculations, which neglect the effects of changes in clouds, these anomalies are primarily focused in the regions directly influenced by the changes in atmospheric dust burden. The largest anomalies, and simultaneous largest discrepancies between optical regimes, occurs over much of North and West Africa. In the most absorbing cases (V&P, 2.7hem, 1.5hem, 0.9hem) we note a strong positive anomaly along the western Sahara/Sahel boundary, peaking to the north of Lake Chad at over 20 W m$^{-2}$ in the V&P regime. This is largely due to the shortwave component of the net radiative flux as the aeolian, absorbing dust decreases the net column albedo over the relatively high albedo Sahara desert, thus allowing less shortwave radiation to be reflected back to space and leading to a positive ToA anomaly.

This anomaly begins to slowly transition as the optical regime becomes more scattering. Beginning in the 1.5hem and 0.9hem dust cases, we note a negative ToA radiative flux anomaly develop in the region between 5°N-15°N and 15°W-15°E, referred to as the West African Monsoon (WAM) region. As we continue to increase the scattering ability of the dust to the 0.1x0.9hem and 0.1x0.9hem (Fixed LW) cases, we develop a strong negative anomaly across much of North and West Africa, peaking around $-20$ W m$^{-2}$ along the southern Sahara desert. This negative anomaly is primarily due to the shortwave scattering of dust increasing the net column albedo over these regions, beginning with the relatively low albedo WAM region in the 1.5hem and 0.9hem cases and extending to much higher albedo Sahara desert in the most scattering cases.

In all optical regimes, we observe a strong negative anomaly across much of the tropical North Atlantic Ocean, even stretching across the Caribbean Sea and portions of Central America. The largest anomalies are focused in the eastern Tropical North Atlantic, just off the coast of West Africa, and fade both westward and poleward/equatorward as the total
Figure 3.2: The global anomalies due to an increase in Saharan dust, comparable to the observed changes between the 1960s and 1990s, with differing optical properties. The values are averaged over JJASON and only those values that pass a Student’s t-test to 95% are shaded. The columns, from left to right, show the net radiative flux anomalies at the top of the atmosphere (Positive downwards; W m$^{-2}$), the net radiative flux anomalies at the surface (Positive downwards; W m$^{-2}$), the 2m air temperature anomalies (K), and the precipitation anomalies (mm day$^{-1}$). The rows, from top to bottom, show the response in the V&P, 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem [Fixed LW] optical regimes, and are ordered by increased scattering of dust.
atmospheric dust burden becomes smaller. This is again due to the increased backscatter of solar radiation by dust over the relatively low albedo ocean surface, which is evident from the fact that the anomalies become larger as the dust becomes more scattering from the V&P case to the 0.1x0.9hem (Fixed LW) case.

There are also far-field ToA radiative flux anomalies of note, particularly across the Himalayas and in the polar regions. In the more absorbing V&P, 2.7hem, and 1.5hem cases we observe very strong positive anomalies across the Himalayas and portions of the Indian subcontinent while in the more scattering 0.9hem, 0.1x0.9hem, and 0.1x0.9hem (Fixed LW) cases we observe negative anomalies across these regions. The mountainous response is due to the significant warming/cooling occurring in the Himalayas (Fig. 3.2, third column) that decreases/increases the snowpack and associated albedo leading to an increased/decreased net radiative flux at the ToA. The response over the lowlands of the Indian subcontinent is largely explained by the warming/cooling over the region leading to a increase/decrease in outgoing longwave radiation by the Stefan-Boltzmann Law, which for the observed changes in surface temperature give a similar result of the same order of magnitude. Meanwhile, the polar responses in Baffin Bay (V&P case) as well as the Barents (V&P, 2.7hem, 0.1x0.9hem cases), Greenland (0.1x0.9hem, 0.1x0.9hem [Fixed LW] cases), Norwegian (V&P, 2.7hem, 1.5hem cases), Weddell (V&P, 2.7hem, 0.9hem cases), and Bellingshausen Seas (0.9hem, 0.1x0.9hem cases) are described by changes in sea ice concentration in those areas. An increased/decreased sea ice concentration will lead to an increased/decreased surface albedo which will then lead to a decreased/increased net radiative flux at the ToA. These last two regional anomalies are beyond the scope of this dissertation, but deserve a more detailed analysis in later work.

3.3.2 Surface Net Clear-Sky Radiative Flux Anomalies

Transitioning now to the surface radiative balance (Fig. 3.2, second column), we note the similarity in major spatial patterns between the net radiative flux at the surface and that at
the ToA. The largest and most homogeneous response is negative and occurs across much of North and West Africa and stretches out across the Tropical North Atlantic Ocean into Central America. This negative response is strongest in the most absorbing (V&P) and the most scattering (0.1x0.9hem [Fixed LW]) optical regimes with anomalies just off the West African coastline reaching over $-30 \text{ W m}^{-2}$. The cause of these negative anomalies is the increased atmospheric attenuation of radiation provided by the aeolian dust particles. Despite the optical regime and no matter the surface albedo, dust will absorb and scatter shortwave radiation leading to a decreased radiative flux at the surface. This effect is amplified when either the absorption or scattering properties of dust are maximized, leading to the largest anomalies occurring in the V&P and 0.1x0.9hem (Fixed LW) cases.

We observe the same far-field anomalies in radiative flux at the surface as we did in the radiative flux response at the ToA. This includes the increased radiative flux over the Himalayas and Indian subcontinent in the most absorbing regimes, the decreased radiative flux over the Himalayas and Indian subcontinent in the most scattering regimes, and the assortment of polar anomalies listed in the previous subsection. These anomalies appear in the surface budget as well as the ToA budget because they rely on surface changing and not any associated radiative effects in the atmosphere.

### 3.3.3 Surface Temperature Anomalies

The 2m surface temperature anomalies due to changes in Saharan-born dust (Fig. 3.2, third column) extend across much more of the globe than the previous radiative anomalies. On average, these anomalies lead to a global warming of around 0.02K in the most absorbing, V&P case or lead to a global cooling of around $-0.07K$ in the most scattering, 0.1x0.9hem (Fixed LW) case. However, some of the largest surface air temperature anomalies are still focused on North Africa and the Tropical Atlantic Ocean. Looking first at North Africa, we observe a tri-pole like feature where the WAM region and much of Mediterranean coastline are out of phase with the Central Sahara desert. In the most absorbing optical regime
(V&P), this takes the form of negative anomalies over $-1K$ across the WAM region, Sahel, and Mediterranean basin with a weaker positive anomaly up to $0.25K$ covering the rest of North Africa. As we increase the amount of scattering, this anomaly pattern slowly changes sign until the 0.1x0.9hem and 0.1x0.9hem (Fixed LW) regimes where there is now a positive anomaly of around $0.25K$ in the WAM region and a negative anomaly, peaking near $-1K$, across much of the Sahara desert. These responses were investigated in Chapter 2 and explained in part by changes to the evaporative flux over this region from associated dust-induced precipitation anomalies.

Across the Tropical North Atlantic Ocean, we see a negative surface air temperature anomaly in all optical regimes, reaching up to $-1K$ in the V&P and 0.1x0.9hem (Fixed LW) optical regimes. This is due in large part to the decreased radiative flux from the attenuation by dust, as emphasized by the fact that the temperature anomalies are strongest in the most absorbing and most scattering regimes. However, in nearly all cases there is also a slight warming of the surface air temperature to the east of the North American continent in addition to a similar pattern across the Western North Pacific, albeit of opposite sign. These anomalies could be indicative of larger scale circulation features such as a shifting of the jet stream in these principal storm tracks.

In addition, we notice the surface temperature anomalies which corroborate our earlier statements about changes to the cryosphere leading to shifts in surface albedo. In the most absorbing regimes, the simulations show a significant warming in the Himalayas exceeding $1K$ which was associated with decreased snowpack and increased downward radiative flux. This transitions to a broader cooling effect in the most scattering regimes which would be associated with the converse growth of mountainous snowpack. Similarly, there are warm anomalies in polar areas increased downward radiative flux and cold anomalies in areas of decreased downward radiative flux which are associated with changes in sea ice cover.
3.3.4 Precipitation Anomalies

The simulated precipitation anomalies due to an increase in Saharan-born dust (Fig. 3.2, fourth column) are again much more heterogeneous than the previous anomalies. While there is no large global change in precipitation, it is interesting that most of the precipitation anomalies appear to be focused in the tropics across the globe. When taken as a whole, it appears that more absorbing dust causes an increase in tropical precipitation in the Northern Hemisphere and a decrease in the Southern Hemisphere while more scattering dust causes a decrease in Northern Hemisphere tropical precipitation and an increase in Southern Hemisphere precipitation.

The largest changes occur in the WAM region and have been the subject of multiple studies (Miller et al. 2004; Yoshioka et al. 2007; Mahajan et al. 2012; Solmon et al. 2012; Strong et al. 2015b). Here, we observe in the most absorbing V&P, 2.7hem, and 1.5hem optical regimes that we get a general increase in the precipitation across the WAM region, up to 2.5 mm day$^{-1}$ in the V&P case. However, as we increase the scattering properties of the dust, we drastically change the sign of this response to negative in the 0.1x0.9hem and 0.1x0.9hem (Fixed LW) regimes, exceeding $-1.5$ mm day$^{-1}$. These anomalies are associated with a strengthening or a weakening, respectively, of the West African monsoon as was shown in Chapter 2.

Moving to the Tropical Atlantic Ocean, we observe a significantly negative precipitation anomaly in every optical regime, with various spatial patterns. In the most absorbing V&P and 2.7hem cases there is a strong negative anomaly focused on the Western Atlantic Ocean and Caribbean Sea with anomalies up to $-4$ mm day$^{-1}$ across Central America. This negative anomaly stretches north into the Central Atlantic basin and slightly south of the Equator, enveloping the weaker positive Eastern Atlantic and Gulf of Guinea precipitation anomalies. As the scattering of dust increases, the pattern begins to shift such that there is a negative anomaly along the southern border of the dust plume, extending from West Africa to the Lesser Antilles, abutting a positive anomaly extending from South America.
These Equator-centric anomalies could be due to changes in the Intertropical Convergence Zone (ITCZ) (Wilcox et al. 2010; Woodage and Woodward 2014; Strong et al. 2015b).

Of final note, we return to the Indian subcontinent which has significant changes to its simulated precipitation. In the most absorbing regimes there is a net increase in precipitation of almost 2 mm day$^{-1}$ while in the most scattering cases there is a net decrease of up to $-1$ mm day$^{-1}$. These anomalies could be indicative of changes to the Indian monsoon system and would have significant impact on any circulation features in the region.

### 3.4 Tropical Cyclone Anomalies

#### 3.4.1 Simulated Anomalies

The global tropical cyclone genesis and track density anomalies due to an increase in Saharan-born dust, as determined by our TC tracking routine, are plotted in Figure 3.3 where the contours represent anomalies in genesis or track density while the shading represents the percent change as given by the formula:

$$\text{Percent Change} = \frac{2 \times (\text{Density}_{\text{Full Dust}} - \text{Density}_{\text{Reduced Dust}})}{\text{Density}_{\text{Full Dust}} + \text{Density}_{\text{Reduced Dust}}} \times 100\%$$

We use this specific formulation because we are comparing model climatologies and neither can be said to be a true reference state for the world. Unlike other formulas, we keep a sign convention to match the calculation of differences between high and low atmospheric dust concentration states. A value of 100% (-100%) means dust causes an increase (decrease) equivalent to adding (subtracting) the average of the full and reduced dust states. A value of 0% means there is no change between the full and reduced dust states. Similarly, the anomalous accumulated cyclone energy (ACE) for various basins is provided in Table 3.1. ACE is a useful measure of seasonal tropical cyclone activity and is defined as the time
Table 3.1: The simulated accumulated cyclone energy ($10^4 \text{kts}^2$) anomalies due to an increase in Saharan dust, comparable to the observed changes between the 1960s and 1990s, with differing optical properties averaged across several basins.

<table>
<thead>
<tr>
<th>Basin</th>
<th>Volz &amp; Patterson</th>
<th>2.7hem</th>
<th>1.5hem</th>
<th>0.9hem</th>
<th>0.1x0.9hem</th>
<th>0.1x0.9hem (Fixed LW)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Global</td>
<td>−6.05</td>
<td>9.95</td>
<td>8.68</td>
<td>2.04</td>
<td>7.99</td>
<td>10.95</td>
</tr>
<tr>
<td>Northern Hemisphere</td>
<td>−2.24</td>
<td>16.36</td>
<td>9.44</td>
<td>0.76</td>
<td>0.60</td>
<td>1.60</td>
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<tr>
<td>North Indian</td>
<td>−0.68</td>
<td>0.69</td>
<td>0.68</td>
<td>−1.52</td>
<td>−1.18</td>
<td>1.62</td>
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<tr>
<td>West Pacific</td>
<td>3.19</td>
<td>15.37</td>
<td>5.27</td>
<td>1.45</td>
<td>3.32</td>
<td>4.26</td>
</tr>
<tr>
<td>East Pacific</td>
<td>1.61</td>
<td>3.96</td>
<td>3.95</td>
<td>2.43</td>
<td>1.40</td>
<td>0.93</td>
</tr>
<tr>
<td>North Atlantic</td>
<td>−6.36</td>
<td>−3.65</td>
<td>−0.48</td>
<td>−1.59</td>
<td>−2.94</td>
<td>−5.20</td>
</tr>
<tr>
<td>Southern Hemisphere</td>
<td>−3.81</td>
<td>−6.41</td>
<td>−0.76</td>
<td>1.26</td>
<td>7.39</td>
<td>9.35</td>
</tr>
<tr>
<td>South Indian</td>
<td>−4.75</td>
<td>−3.84</td>
<td>−0.53</td>
<td>0.81</td>
<td>2.32</td>
<td>7.86</td>
</tr>
<tr>
<td>Australia</td>
<td>−1.09</td>
<td>−0.02</td>
<td>0.00</td>
<td>−1.04</td>
<td>1.48</td>
<td>0.41</td>
</tr>
<tr>
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<td>1.44</td>
<td>−2.57</td>
<td>0.22</td>
<td>1.63</td>
<td>3.24</td>
<td>1.19</td>
</tr>
<tr>
<td>South Atlantic</td>
<td>0.59</td>
<td>0.01</td>
<td>−0.45</td>
<td>−0.13</td>
<td>0.36</td>
<td>−0.11</td>
</tr>
</tbody>
</table>

sum of the squared maximum sustained wind velocity of every active tropical cyclone using six-hourly data.

We observe that in our simulations an increase of Saharan-born dust has substantial impacts to the global TC climatology. In every optical regime, there is a significant decrease in North Atlantic TC genesis events with the strongest response in the most absorbing and most scattering cases. In the V&P optical regime dust causes over 0.5 less storms per year, which is over 75% of the modeled climatology, and in the 0.1x0.9hem (Fixed LW) optical regime dust causes a little over 0.25 less storms per year, which is around 50% of the model’s climatology. This effect is weakened in the more moderate optical regimes of 1.5hem and 0.9hem where there are less than 0.25 less storms per year and only about a 20% change from the model’s climatology. These anomalies are primarily focused in the main development region (MDR) of the North Atlantic, between $10^\circ$N-$25^\circ$N and $80^\circ$W-$20^\circ$W, with an emphasis towards its easternmost boundary.

The genesis anomalies translate to decreased TC activity across the North Atlantic where again we have the largest changes in the most absorbing and most scattering optical regimes. There are anomalies up to $-2$ storm days per year in the MDR for the V&P regime, which
Figure 3.3: The TC genesis and lifetime-track density anomalies due to an increase in Saharan dust, comparable to the observed changes between the 1960s and 1990s, with differing optical properties. The shading denotes the percent change while the contours represent the absolute anomaly with a contour interval of 0.25 storm year\(^{-1}\) for TC genesis density and 0.5 day year\(^{-1}\) for TC track density. The values are indicative of changes over a 10x10 degree grid centered on each grid cell and only those values with absolute magnitude larger than 0.1 storm year\(^{-1}\) for TC genesis density or 0.25 day year\(^{-1}\) for TC track density are shaded. The left column shows the TC genesis density anomalies (storms spawned per year) while the right column shows the TC lifetime-track density anomalies (days with storm present per year). The rows, from top to bottom, show the response in the V&P, 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem [Fixed LW] optical regimes, and are ordered by increased scattering of dust.
is over 75% of the climatology in that region, and additional negative anomalies which overlay the traditional northwestward track of most Cape Verde style storms up to the Southeastern United States of America. Interestingly, while the negative anomalies in the 0.1x0.9hem (Fixed LW) regime are of comparable magnitude, the peak anomalies appear to be shifted farther west from North Africa than the previous example but still maintain a similar westward extent. And again, the weakened impact of the 1.5hem and 0.9hem dust means there is a weakened response in the TC track density. These results are corroborated by the anomalies in ACE which show a significant decrease in tropical cyclone activity for each optical regime and with maximum anomalies occurring in the V&P regime.

Surprisingly, there are also significant anomalies in other basins around the globe, particularly in the Northern Pacific and Indian Oceans. Focusing first on the Northeast Pacific basin, we see a strong east/west dipole develop in the V&P regime with decreased activity directly off the west coast of Central America amounting to over 0.25 less storms per year and over 1.5 less storm days per year. These anomalies are balanced by those just west of 240°E and into the Central North Pacific basin where we see anomalies exceeding 0.25 more storms per year and over 1.5 more storm days per year, in a band focused tightly around 15°N. However, the spatial scale of the positive anomalies far outweighs that of the negative anomalies, leading to a net increase in TC activity as measured by ACE. The density anomalies slowly decrease in magnitude as we transition to more scattering optical regimes with the far eastern negative anomalies decreasing by the 1.5hem optical regime, allowing for the largest increase in basin-wide TC activity as measured by ACE. Unlike in the North Atlantic, the largest and most homogeneous anomalies exist in only the most absorbing dust cases so while we get positive track density anomalies above 1 more storm day per year in the 0.1x0.9hem (Fixed LW) case, the overall pattern is much more heterogeneous than in the V&P, 2.7hem, and 1.5hem cases.

In the Northwestern Pacific basin our simulations show an interesting trend in TC activity. Despite TC genesis density anomalies rarely exceeding ±0.5 storms per year, we note
some of the largest global anomalies in TC track density with values exceeding 2 more storm
days per year in both the V&P and 2.7hem optical regimes. There is also an interesting
spatial pattern which develops as we transition to more scattering regimes. In the V&P
dust case, there is a large north/south gradient in TC track density anomalies with negative
anomalies towards the equator and positive anomalies toward the pole, signifying a potential
shift in TC activity out of the deep tropics towards the north. As we increase the scattering
we get largely basin-wide positive anomalies in the 2.7hem and 1.5hem cases which match
the largest positive anomalies in ACE. This spatial pattern flips signs by the 0.1x0.9hem and
0.1x0.9hem (Fixed LW) cases where there are now positive anomalies towards the equator
and negative anomalies towards the poles, potentially implying a shift of TC activity into the
deep tropics and towards the South. These patterns move in tandem with similar anomalies
off the east coast of Australia.

Lastly, in the Indian Ocean we note relatively large changes in TC genesis density as
well as TC track density, focused primarily in the region between 0°S-15°S and 60°E-100°E.
Again there is an interesting transition as we increase the scattering of dust. For the most
absorbing dust cases we get a significant decrease of −0.5 storms per year and upwards of −2
storm days per year. But as we move to the 0.1x0.9hem and 0.1x0.9hem (Fixed LW) regimes
we get significant increases of over 0.5 storms per year and well over 3.5 storm days per year
in these same areas. When taken together with the anomalies in the Western Pacific basin,
it appears as if there’s a transition from shifting TC genesis and development to the North
in the most absorbing cases (V&P and 2.7hem) to the South in the most scattering cases
(0.1x0.9hem and 0.1x0.9hem [Fixed LW]). However, while some of these absolute anomalies
in the Northern Pacific and Indian Oceans are on the order of or larger than those in the
North Atlantic, they represent only a small percentage change from the model’s climatology,
usually at most around 25%.
3.4.2 Comparison to Observations

As was stated in the methodology section, our choice of dust atmospheric loading climatology was influenced by the period of low dust transport away from the Sahara in the late 1960s compared with the period of high dust transport in the mid-1980s through the early 1990s as measured at Barbados (Prospero and Lamb 2003). These two periods represent the extremes in the direct observational record and as such presented a rich opportunity for a sensitivity study. While not an exact comparison, our simulations can be viewed as idealizations of the climate state in these two periods and thus lead us to compare our modeled TC anomalies with those of the real world. We collect the observed TC record from the IBTrACS dataset (Schreck et al. 2014) and calculate the percent change from the 5 year periods 1965-1970 and 1983-1988, periods roughly centered on the maximum and minimum dust transport over Barbados in the observational record (Prospero and Lamb 2003). These results are compared with the basin total percent change between the 200 year full and reduced dust simulations for each optical regime and presented in Figure 3.4.

The most striking comparison occurs in the North Atlantic where our simulated TC anomalies explain between 10-150% of the observed trend of 30% fewer storm days per year. Interestingly, the most similar percent difference occurs in our most artificial optical regime of 0.1x0.9hem (Fixed LW) while the optical regimes regarded as most accurate of Saharan dust, V&P and 2.7hem, either over or under estimate the observed trend, respectively. As to be expected, the more middle-of-the-road optical regimes of 1.5hem and 0.9hem display the smallest changes as they were also the ones with the smallest climate perturbations. On the whole, however, every optical regime simulates a negative trend in TC activity across the North Atlantic as was observed between the late 1960s to mid-1980s.

Other basins around the world are not as coherent in their relationship between our simulations and the observed trend. In the Northeast Pacific basin, our simulations show an increase in TC activity anywhere between 3-7%, which is of the same sign as the observations but dwarfed by the substantial 40% increase in observed TC activity. This may indicate that
Figure 3.4: A comparison between the observed anomalies in total TC activity and the simulated anomalies due to an increase in Saharan dust, comparable to the observed changes between the 1960s and 1990s, with differing optical properties across several basins. The observations are drawn from the IBTrACS dataset (Schreck et al. 2014) and calculated as the percent difference between the periods 1960-1965 and 1983-1988 for each basin. Each bar represents the percent difference in total TC activity across individual basins with the shading representing the single scattering albedo of 1µm dust at 550nm of each optical regime. The bars, from left to right, show the response in the observations and V&P, 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem [Fixed LW] optical regimes, and are ordered by increased scattering of dust.
while Saharan dust has a non-negligible impact on Northeastern Pacific TCs, it is not the primary driving mechanism. Interestingly, the largest changes occur in the most intermediate of optical properties, 2.7hem, 1.5hem, and 0.9hem. Meanwhile, in the Northwestern Pacific our simulations have the complete opposite sign from the observed trend in every optical range which was originally expected as the region of forcing is not in near-proximity to the region of response.

Lastly, in the Southern Hemisphere the observations of both the South Pacific and Indian Oceans show a general increase in TC activity. Meanwhile, our simulations show a common pattern of decreasing TC activity for the most absorbing optical regimes and increasing TC activity for the most scattering optical regimes, with the obvious exclusion of the V&P dust case in the South Pacific. This means that the closest our simulations get to representing the observed trend in Southern Hemisphere TC activity is in the 0.1x0.9hem and 0.1x0.9hem (Fixed LW) regimes. In fact, the 0.1x0.9hem case in the Indian Ocean is the closest to resolving the actual observed trend. This is again surprising as these are our most artificial dust optical regimes.

3.4.3 Comparison to Predictive Indicies

To further corroborate our results, we examine how Saharan dust impacts several common genesis potential indicies (GPIs). These indicies are empirically-derived formulations of the likelihood for a tropical storm to develop given the local conditions. However, note that there is evidence that GPIs can be problematic in resolving simulated TC climatologies under various climate forcings (Camargo et al. 2014). We use the GPIs developed by Emanuel and Nolan (2004), Emanuel (2010), Tippett et al. (2011), Gray (1979), and Royer et al. (1998). Hereafter, these will be referred to as EN04, E10, T11, G79, and R98, respectively. To focus on the regions in which our model simulates tropical cyclogenesis, we follow the protocol of
Figure 3.5: A comparison between the simulated anomalies in total TC genesis events and the simulated anomalies in genesis potential indices due to an increase in Saharan dust, comparable to the observed changes between the 1960s and 1990s, with differing optical properties across several basins. Each bar represents the basin-averaged anomalies in TC genesis events from the tropical cyclone tracking scheme or as predicted from several genesis-weighted genesis potential indices. The bars, from left to right, show the response in the TC tracking scheme and as predicted from the genesis potential indices of Emanuel and Nolan (2004), Emanuel (2010), Tippett et al. (2011), Gray (1979), and Royer et al. (1998).

Held and Zhao (2011) and genesis weight the GPIs using the formula:

\[
[GPI(x,y)]_G = \frac{G(x,y)GPI(x,y)}{G(x,y)}
\]

where \(GPI(x,y)\) is the spatial map of monthly means of the GPI, \(G(x,y)\) is the spatial map of monthly means of the genesis density for the reduced dust simulation, and the overline is an average over the 12 months. We then average over each individual basin and compare them to the results from our TC tracking routine (Fig. 3.5).
Beginning in the North Atlantic, it becomes readily apparent that many of the GPI's do not adequately represent the simulated changes in TC genesis derived from the TC tracking scheme and often overestimate the modeled response. Interestingly, for the more absorbing cases, the older G79 and R98 GPIS come closest to the magnitude of the simulated TC genesis anomalies. This persists from the V&P case through to the 0.9hem case. For the most artificially scattering regimes of 0.1x0.9hem and 0.1x0.9hem (Fixed LW), all but the E10 GPI, which has an erroneously positive basin-wide response, do an acceptable prediction of the modeled TC genesis anomalies and some like the EN04 are almost exact matches.

Many of the same issues get exacerbated in the North Pacific Ocean. In the Northeast Pacific for the most absorbing regimes of dust, there are several close predictions. The EN04 GPI for V&P dust and the E10 and G79 GPIS for 2.7hem and 1.5hem dust all predict similar anomalies to those simulated. However, in the most scattering regimes, all but the R98 GPI in the 0.1x0.9hem (Fixed LW) regime predict an incorrect decrease in TC genesis events. In the Northwest Pacific, most of the GPI's overestimate the simulated trend in TC genesis for the most extreme optical regimes. However, in the 2.7hem, 1.5hem, and 0.9hem cases, many of the GPIS correctly predict the magnitude of the simulated trend and get a similar basin-wide average value, save for the R98 GPI in the 1.5hem case and the G79 and R98 GPIS in the 0.9hem case. This is at odds with the V&P, 0.1x0.9hem, and 0.1x0.9hem (Fixed LW) regimes where the GPI’s do not adequately represent a balancing effect as noted in the TC genesis spatial anomalies.

In the South Indian Ocean, we see that the EN04 GPI does a good job at predicting the simulated responses in the V&P and 0.1x0.9hem (Fixed LW) regimes as well as the E10 GPI in the V&P regime, but both predict a positive overestimate in every other regime. Meanwhile, the R98 does a better performance in the 2.7hem and 0.1x0.9hem regimes and the T11 appears best for the 1.5hem and 0.9hem regimes. While it appears no one GPI is best for all dust optical regimes and in all basins of the world, we may be able to glean the controls on the simulated TC genesis anomalies from their constituent parts.
3.5 Impact of Dust on Tropical Cyclone Related Parameters

Understanding what drives changes in individual tropical cyclones, let alone their basin-wide climatologies, is a monumental task and the subject of numerous prior studies. Much of this research has pointed towards the influencing effects of low-level vorticity, vertical wind shear, mid-level vertical velocity, mid-level moisture, and sea surface temperature anomalies. In addition, several other quantities have shown utility in TC development prediction such as the presence of convective precipitating systems as well as upper ocean heat content, maximum potential surface wind speed, surface to mid-level entropy deficit, and lower level atmospheric thermal structure anomalies. We will explore these quantities, among a few others, in our attempt to explain how perturbations in the atmospheric loading of Saharan dust impacts basin-wide TC climatologies.

3.5.1 North Atlantic

We begin in the North Atlantic by calculating the basin average of the genesis weighted TC development parameters mentioned above alongside the basin-averaged TC track density anomalies (Tab. 3.2). Our aim is to find a variable which not only is the correct sign to be in line with the simulated TC track density anomalies, but also get the relative magnitudes between optical regimes correct, namely the largest anomalies in the most absorbing and most scattering regimes and the smallest anomalies in the middle-of-the-road regimes. For the purely dynamics-related variables, we see that low level vorticity and vertical wind shear do not consistently get the correct sign of anomaly, but the mid-level vertical velocity does, as suggested by Held and Zhao (2011). However, the vertical velocity shows a monotonically decreasing anomaly with increasing scattering which is not in line with the pattern of TC density anomalies. These same issues with consistent trends also arise in the moisture and vertically-integrated quantities. Interestingly, the most consistent set of variables are the...
temperature related relative sea surface temperature and upper ocean heat content. Both get the correct sign of anomalies to match those expected from the TC density anomalies, but the relative magnitudes between the 0.9hem and 0.1x0.9hem cases cause an exception. This may suggest a strong ocean influence on Atlantic TC variability as suggested by Wang et al. (2012).

So if none of the common TC development parameters can consistently explain the anomalies we see in TC track density across the North Atlantic, we must search for another answer. Dust’s impact to climate is highly dependent on it’s ToA net radiative flux anomaly (Miller et al. 2014). This would suggest that if we compare dust’s ToA radiative perturbation with the basin-averaged TC statistics we may find a possible relationship. To narrow in on the predictive behavior of this relationship, we calculate dust’s LW and SW all-sky radiative flux anomaly over the North Atlantic MDR and plot it against the basin wide ACE anomalies for each optical regime (Fig. 3.6). What we observe is a strictly monotonic relationship between the net ToA radiative imbalance over the MDR and the ACE anomaly for the entire basin. The most scattering and most absorbing regimes cause the largest ToA radiative anomalies as well as have the largest ACE anomalies. Meanwhile the more moderate optical regimes of 1.5hem and 0.9hem have the smallest net ToA radiative flux anomalies in addition to the smallest ACE anomalies. This would appear to suggest that dust’s radiative effects over the main TC generating region of the North Atlantic affect the entire TC climatology of the basin.

3.5.2 Northeast and Central Pacific

Shifting across Central America to the Northeast Pacific, we again look at the aforementioned TC development parameters averaged across the basin and compare them to the basin-averaged TC track density anomalies (Tab. 3.3). In the Northeast Pacific, the TC track density anomalies are all positive and actually largest for the more moderate optical regimes of 2.7hem, 1.5hem, and 0.9hem. As was common in the North Atlantic, no one parameter
Table 3.2: The simulated genesis-weighted tropical cyclogenesis parameter anomalies due to an increase in Saharan dust, comparable to the observed changes between the 1960s and 1990s, with differing optical properties averaged across the North Atlantic Ocean basin. Values are italicized when the sign of the tropical cyclogenesis parameter does not match that expected from the simulated tropical storm activity anomalies.

<table>
<thead>
<tr>
<th></th>
<th>Volz &amp; Patterson</th>
<th>2.7hem</th>
<th>1.5hem</th>
<th>0.9hem</th>
<th>0.1x0.9hem</th>
<th>0.1x0.9hem (Fixed LW)</th>
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<tbody>
<tr>
<td><strong>Storm Anomalies (Days/Year)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
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<td>850mb Absolute Vorticity (10^{-6} s^{-1})</td>
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<td>-0.065</td>
<td>0.024</td>
<td>0.041</td>
<td>-0.121</td>
</tr>
<tr>
<td>200mb-850mb Vertical Wind Shear (m s^{-1})</td>
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<td>-0.061</td>
<td>-0.040</td>
<td>0.166</td>
<td>0.050</td>
<td>0.026</td>
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<tr>
<td>500mb Vertical Pressure Velocity (hPa day^{-1})</td>
<td>1.513</td>
<td>1.056</td>
<td>0.818</td>
<td>0.496</td>
<td>0.437</td>
<td>0.104</td>
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<td>600mb Relative Humidity (%)</td>
<td>-0.981</td>
<td>-0.491</td>
<td>-0.227</td>
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<td>0.010</td>
<td>-0.042</td>
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<tr>
<td>Convective Precipitation (mm day^{-1})</td>
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<td>-0.028</td>
<td>-0.038</td>
<td>-0.041</td>
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</tr>
<tr>
<td>Relative Sea Surface Temperature (K)</td>
<td>-0.114</td>
<td>-0.040</td>
<td>-0.024</td>
<td>-0.063</td>
<td>-0.018</td>
<td>-0.044</td>
</tr>
<tr>
<td>Upper 60m Ocean Heat Content (10^3 cal cm^{-2})</td>
<td>-0.617</td>
<td>-0.270</td>
<td>-0.198</td>
<td>-0.388</td>
<td>-0.307</td>
<td>-0.496</td>
</tr>
<tr>
<td><strong>Vertically-Integrated Quantities</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Maximum Potential Surface Wind Speed (m s^{-1})</td>
<td>-1.605</td>
<td>-0.742</td>
<td>-0.400</td>
<td>-0.477</td>
<td>0.116</td>
<td>-0.017</td>
</tr>
<tr>
<td>Entropy Deficit</td>
<td>0.044</td>
<td>-1.419</td>
<td>-0.288</td>
<td>-0.015</td>
<td>0.796</td>
<td>0.115</td>
</tr>
<tr>
<td>500mb-925mb Vertical Gradient of $\theta_E$ (K mb^{-1})</td>
<td>-0.088</td>
<td>-0.084</td>
<td>-0.063</td>
<td>-0.077</td>
<td>-0.010</td>
<td>-0.015</td>
</tr>
</tbody>
</table>
Figure 3.6: The accumulated cyclone energy anomalies across the North Atlantic due to an increase in Saharan dust, comparable to the observed changes between the 1960s and 1990s, with differing optical properties. The x-axis is the top of atmosphere shortwave anomaly (Positive downwards; W m\(^{-2}\)) averaged across the North Atlantic main development region while the y-axis is the top of atmosphere longwave anomaly (Positive downwards; W m\(^{-2}\)) averaged across the North Atlantic main development region. The size of each circle represents the North Atlantic basin-averaged accumulated cyclone energy anomaly for each optical regime with each number being the exact value (10\(^4\) kts\(^{-2}\)). The shading of each circle represents the single scattering albedo of 1\(\mu\)m dust at 550nm of each optical regime.
consistently explains the simulated TC track density anomalies, and in fact each variable is wrong on the expected sign at least once for each optical regime. However, it appears that the mid-level vertical velocity and relative sea surface temperature come closest to the simulated TC track density anomalies, perhaps suggesting a atmospheric bridge type response reminiscent of Ham et al. (2013) and Kucharski et al. (2015).

In our simulations, the significant cooling of the North Atlantic could potentially lead to a central Pacific El-Niño-like state. As the North Atlantic develops anomalously cool sea surface temperatures, an anomalous easterly flow develops over the Eastern Pacific and an anomalous westerly flow over the Western Pacific (Ham et al. 2013). These lead to surface warming in the central Pacific and modest to little cooling in the Eastern Pacific. This pattern is reminiscent of a central Pacific El Niño whose action center for TC activity is focused more towards the central Pacific with little activity towards the Eastern Pacific. These anomalies match well to the TC track density anomalies found in the most absorbing regimes (Fig 3.3). However, this hypothesis does not seem to be as accurate for the most scattering regimes. This is possibly due to that fact that adding extremely scattering dust leads to a larger scale cooling of the Northern Hemisphere. Thus, even though the surface temperatures cool over the North Atlantic as shown in Chapter 2 and could potentially lead to the same central Pacific response, they are modulated by the increased global cooling from scattering dust which leads to a decreased potential for TC development.

### 3.5.3 Northwest Pacific and Indian

Finally, transitioning to the Indo-Pacific basins, we once again calculate individual basin-averaged TC parameters (Tabs. 3.4 & 3.5). Focusing first on the Northwest Pacific basin, there is even more disagreement between the listed parameters and the TC track density anomalies. No single parameter is able to capture the large increase in TC track density for the 2.7hem case and the lowest values for the 0.9hem and 0.1x0.9hem cases while simultaneously capturing the correct sign of the anomalies. Meanwhile, in the South Indian basin
Table 3.3: The simulated genesis-weighted tropical cyclogenesis parameter anomalies due to an increase in Saharan dust, comparable to the observed changes between the 1960s and 1990s, with differing optical properties averaged across the Northeast Pacific Ocean basin. Values are italicized when the sign of the tropical cyclogenesis parameter does not match that expected from the simulated tropical storm activity anomalies.

<table>
<thead>
<tr>
<th>Northeast Pacific</th>
<th>Volz &amp; Patterson</th>
<th>2.7hem</th>
<th>1.5hem</th>
<th>0.9hem</th>
<th>0.1x0.9hem</th>
<th>0.1x0.9hem (Fixed LW)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Storm Anomalies (Days/Year)</td>
<td>0.077</td>
<td>0.131</td>
<td>0.158</td>
<td>0.128</td>
<td>0.063</td>
<td>0.056</td>
</tr>
<tr>
<td><strong>Dynamics</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>850mb Absolute Vorticity (10(^{-6}) s(^{-1}))</td>
<td>(-0.058)</td>
<td>(-0.028)</td>
<td>(-0.045)</td>
<td>0.130</td>
<td>0.115</td>
<td>0.010</td>
</tr>
<tr>
<td>200mb-850mb Vertical Wind Shear (m s(^{-1}))</td>
<td>0.152</td>
<td>(-0.016)</td>
<td>(-0.002)</td>
<td>(-0.003)</td>
<td>(-0.018)</td>
<td>(-0.009)</td>
</tr>
<tr>
<td>500mb Vertical Pressure Velocity (hPa day(^{-1}))</td>
<td>(-0.661)</td>
<td>(-0.186)</td>
<td>(-0.686)</td>
<td>(-0.400)</td>
<td>(-0.201)</td>
<td>(0.028)</td>
</tr>
<tr>
<td><strong>Moisture</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>600mb Relative Humidity (%)</td>
<td>0.587</td>
<td>0.666</td>
<td>0.449</td>
<td>0.063</td>
<td>(-0.026)</td>
<td>0.006</td>
</tr>
<tr>
<td>Convective Precipitation (mm day(^{-1}))</td>
<td>0.002</td>
<td>0.038</td>
<td>0.026</td>
<td>(-0.007)</td>
<td>0.018</td>
<td>0.000</td>
</tr>
<tr>
<td><strong>Temperature</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Relative Sea Surface Temperature (K)</td>
<td>0.022</td>
<td>0.039</td>
<td>0.050</td>
<td>(-0.003)</td>
<td>(-0.009)</td>
<td>0.005</td>
</tr>
<tr>
<td>Upper 60m Ocean Heat Content (10(^3) cal cm(^{-2}))</td>
<td>0.352</td>
<td>0.326</td>
<td>0.222</td>
<td>(-0.102)</td>
<td>(-0.327)</td>
<td>(-0.429)</td>
</tr>
<tr>
<td><strong>Vertically-Integrated Quantities</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Maximum Potential Surface Wind Speed (m s(^{-1}))</td>
<td>(-0.111)</td>
<td>0.084</td>
<td>0.351</td>
<td>0.173</td>
<td>0.076</td>
<td>0.209</td>
</tr>
<tr>
<td>Entropy Deficit</td>
<td>(-0.026)</td>
<td>0.139</td>
<td>0.429</td>
<td>(-0.266)</td>
<td>(-0.058)</td>
<td>0.061</td>
</tr>
<tr>
<td>500mb-925mb Vertical Gradient of (\theta_E) (K mb(^{-1}))</td>
<td>0.039</td>
<td>0.058</td>
<td>0.030</td>
<td>0.025</td>
<td>(-0.007)</td>
<td>(-0.013)</td>
</tr>
</tbody>
</table>
there are two potential candidates, namely the vertical wind shear and the convective precipitation. While both do not perfectly match the pattern of a generally increasing level of TC activity, they do consistently get the signs correct and could suggest a atmospheric-based mechanism, perhaps dependent on the anomalies we saw in the Indian monsoon system.

Looking for other explanations of the simulated TC track density anomalies, we again return to Figure 3.3 and note that while focusing on the Indo-Pacific region it appears that there is a general shift of TC activity from the North in the most absorbing cases to the South in the most scattering cases. For the V&P optical regime, this means an increased level of TC activity to the North in the Northern Hemisphere and decreased activity through the equator and into the Southern Hemisphere. However, as we increase the level of scattering, this trend reverses and instead we have increased convective activity in the Southern Hemisphere and decreased activity farther to the North. This may be due to the substantial heating or cooling we are imposing with our Saharan dust perturbations, respectively, affecting the general hemispheric energy balance. For our most absorbing cases we saw a general warming of the Northern Hemisphere while for our most scattering cases we saw a general cooling across the Northern Hemisphere, with both cases showing smaller changes across the Southern Hemisphere (Fig. 3.2, third column). This meridional asymmetry can lead to changes in the alterations in the zonal-mean circulation patterns of the tropics (Ming and Ramaswamy 2011), namely increased (decreased) convection in the Northern Hemisphere convective regions as the Northern Hemisphere warms (cools). This is balanced by decreased (increased) convection in the Southern Hemisphere convective regions as the northern Hemisphere warms (cools). These anomalies could lead to large-scale TC density anomalies similar to those we have simulated in the Indo-Pacific basin for our most absorbing and scattering regimes, respectively.
Table 3.4: The simulated genesis-weighted tropical cyclogenesis parameter anomalies due to an increase in Saharan dust, comparable to the observed changes between the 1960s and 1990s, with differing optical properties averaged across the Northwest Pacific Ocean basin. Values are italicized when the sign of the tropical cyclogenesis parameter does not match that expected from the simulated tropical storm activity anomalies.

<table>
<thead>
<tr>
<th></th>
<th>Volz &amp; Patterson</th>
<th>2.7hem</th>
<th>1.5hem</th>
<th>0.9hem</th>
<th>0.1x0.9hem</th>
<th>0.1x0.9hem (Fixed LW)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Storm Anomalies</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(Days/Year)</td>
<td>0.072</td>
<td>0.276</td>
<td>0.098</td>
<td>0.040</td>
<td>0.031</td>
<td>0.107</td>
</tr>
<tr>
<td><strong>Dynamics</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>850mb Absolute</td>
<td>−0.044</td>
<td>0.029</td>
<td>0.041</td>
<td>0.030</td>
<td>−0.063</td>
<td>0.132</td>
</tr>
<tr>
<td>Vorticity (10⁻⁶ s⁻¹)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>200mb-850mb Vertical</td>
<td>−0.081</td>
<td>0.032</td>
<td>−0.064</td>
<td>0.034</td>
<td>0.011</td>
<td>0.099</td>
</tr>
<tr>
<td>Wind Shear (m s⁻¹)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>500mb Vertical</td>
<td>0.349</td>
<td>0.027</td>
<td>0.274</td>
<td>0.178</td>
<td>0.431</td>
<td>0.042</td>
</tr>
<tr>
<td>Pressure Velocity</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(hPa day⁻¹)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Moisture</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>600mb Relative</td>
<td>0.190</td>
<td>0.058</td>
<td>−0.022</td>
<td>−0.117</td>
<td>−0.317</td>
<td>−0.150</td>
</tr>
<tr>
<td>Humidity (%)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Convective Precipitation</td>
<td>−0.014</td>
<td>0.004</td>
<td>−0.031</td>
<td>−0.019</td>
<td>−0.031</td>
<td>−0.006</td>
</tr>
<tr>
<td>(mm day⁻¹)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Temperature</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Relative Sea Surface</td>
<td>0.075</td>
<td>0.041</td>
<td>0.027</td>
<td>−0.036</td>
<td>−0.059</td>
<td>−0.073</td>
</tr>
<tr>
<td>Temperature (K)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper 60m Ocean Heat</td>
<td>0.464</td>
<td>0.244</td>
<td>0.065</td>
<td>−0.268</td>
<td>−0.506</td>
<td>−0.861</td>
</tr>
<tr>
<td>Content (10³ cal cm⁻²)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Vertically-Integrated Quantities</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Maximum Potential Surface Wind Speed (m s⁻¹)</td>
<td>0.301</td>
<td>0.106</td>
<td>0.119</td>
<td>−0.123</td>
<td>−0.185</td>
<td>−0.406</td>
</tr>
<tr>
<td>Entropy Deficit</td>
<td>−1.086</td>
<td>0.196</td>
<td>−0.324</td>
<td>0.346</td>
<td>−0.261</td>
<td>−0.370</td>
</tr>
<tr>
<td>500mb-925mb Vertical</td>
<td>0.055</td>
<td>0.026</td>
<td>0.030</td>
<td>−0.014</td>
<td>−0.063</td>
<td>−0.088</td>
</tr>
<tr>
<td>Gradient of θₑ (K mb⁻¹)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 3.5: The simulated genesis-weighted tropical cyclogenesis parameter anomalies due to an increase in Saharan dust, comparable to the observed changes between the 1960s and 1990s, with differing optical properties averaged across the Indian Ocean basin. Values are *italicized* when the sign of the tropical cyclogenesis parameter does not match that expected from the simulated tropical storm activity anomalies.

<table>
<thead>
<tr>
<th>Indian Storm Anomalies (Days/Year)</th>
<th>Volz &amp; Patterson</th>
<th>2.7hem</th>
<th>1.5hem</th>
<th>0.9hem</th>
<th>0.1x0.9hem</th>
<th>0.1x0.9hem (Fixed LW)</th>
</tr>
</thead>
<tbody>
<tr>
<td>850mb Absolute Vorticity (10⁻⁶ s⁻¹)</td>
<td></td>
<td>−0.009</td>
<td>0.054</td>
<td>−0.002</td>
<td>0.015</td>
<td>0.019</td>
</tr>
<tr>
<td>200mb-850mb Vertical Wind Shear (m s⁻¹)</td>
<td></td>
<td>0.028</td>
<td>0.038</td>
<td>0.077</td>
<td>−0.131</td>
<td>−0.132</td>
</tr>
<tr>
<td>500mb Vertical Pressure Velocity (hPa day⁻¹)</td>
<td></td>
<td>0.343</td>
<td>−0.032</td>
<td>−0.139</td>
<td>−0.224</td>
<td>0.200</td>
</tr>
<tr>
<td>Moisture</td>
<td></td>
<td>−0.046</td>
<td>0.171</td>
<td>−0.030</td>
<td>0.139</td>
<td>0.147</td>
</tr>
<tr>
<td>Convective Precipitation (mm day⁻¹)</td>
<td></td>
<td>−0.023</td>
<td>−0.011</td>
<td>−0.006</td>
<td>0.014</td>
<td>0.005</td>
</tr>
<tr>
<td>Temperature</td>
<td></td>
<td>0.012</td>
<td>0.004</td>
<td>0.011</td>
<td>0.027</td>
<td>0.019</td>
</tr>
<tr>
<td>Relative Sea Surface Temperature (K)</td>
<td></td>
<td>0.012</td>
<td>0.004</td>
<td>0.011</td>
<td>0.027</td>
<td>0.019</td>
</tr>
<tr>
<td>Upper 60m Ocean Heat Content (10³ cal cm⁻²)</td>
<td></td>
<td>0.108</td>
<td>0.078</td>
<td>0.076</td>
<td>−0.044</td>
<td>−0.233</td>
</tr>
<tr>
<td>Vertically-Integrated Quantities</td>
<td></td>
<td>−0.225</td>
<td>0.055</td>
<td>0.100</td>
<td>0.072</td>
<td>0.044</td>
</tr>
<tr>
<td>Maximum Potential Surface Wind Speed (m s⁻¹)</td>
<td></td>
<td>0.623</td>
<td>0.100</td>
<td>−7.114</td>
<td>−0.270</td>
<td>1.220</td>
</tr>
<tr>
<td>Entropy Deficit</td>
<td></td>
<td>−0.027</td>
<td>0.020</td>
<td>0.012</td>
<td>0.011</td>
<td>−0.020</td>
</tr>
<tr>
<td>500mb-925mb Vertical Gradient of θ_E (K mb⁻¹)</td>
<td></td>
<td>−0.027</td>
<td>0.020</td>
<td>0.012</td>
<td>0.011</td>
<td>−0.020</td>
</tr>
</tbody>
</table>
3.6 Chapter Summary

Due to Saharan-born dust’s climatological advection westward, it can induce substantive direct and semi-direct interactions with the climate across the Tropical North Atlantic. In particular, it can lead to significant anomalies in the North Atlantic’s main development region for tropical cyclones and thus has the potential to alter North Atlantic hurricane statistics. We have explored this impact from a climate perspective using the fully coupled GFDL CM2.5-FLOR model forced by realistic variations in atmospheric burden of Saharan dust across a wide range of aerosol optical regimes.

We found that Saharan-born dust affects the climate state in similar fashion to previous studies. There is a distinctly opposite response from our most absorbing regimes to our most scattering regimes in several common variables. The more absorbing dust leads to a net positive radiative flux anomaly at the ToA over North Africa whereas the more scattering dust causes a net negative radiative flux anomaly at the ToA over North Africa. Saharan-born dust in all optical regimes causes a significant negative radiative flux over the North Atlantic at the ToA. In addition, due to the solar dimming effect of dust, all cases see a negative net radiative flux at the surface underneath the dust plume anomaly. These radiative anomalies lead either directly or indirectly to other global anomalies in surface air temperature and precipitation. And while many of these anomalies are strongest in the area directly influenced by the dust plume, there are significantly large anomalies in other regions such as the equatorial Pacific and the Indian subcontinent.

We utilized a TC tracking scheme to determine how Saharan-born dust affects the climatological TC statistics in various basins of the world. As was predicted, we saw that Saharan-born dust causes a significant decrease in TC activity across the North Atlantic basin in nearly every optical regime, to varying magnitudes. In comparison to observations over the period idealized by our model simulations, we found that our variations in Saharan dust could explain nearly all of the observed trend depending on the optical regime used. Simultaneously, there were unexpectedly large TC differences in nearly every other basin
across the globe with some of the largest changes occurring in the West Pacific and Indian Oceans, albeit anomalies which only accounted for a small percentage of the model’s natural climatology. The agreement with observations for these differences was less striking than those in the North Atlantic.

To understand these TC anomalies, we examined several common predictive variables for TC activity. Many of the most common GPI’s failed to coherently predict TC changes in the North Atlantic, let alone in other basins. An examination of their constituent variables similarly led to inconclusive results. However, when considering other uncommon utilities we found that there was a monotonic relationship between North Atlantic basin-wide ACE anomalies and Saharan dust induced ToA radiative anomalies across the MDR. In other basins of the world, we put forth several theories to explain the observed TC anomalies. In the East and Central Pacific, we hypothesize that the simulated TC anomalies are due to the established-linkage between East Pacific circulations and the North Atlantic whereby certain El-Niño-like states are stimulated. Conversely, in the West Pacific and Indian Oceans, we theorize that the TC anomalies are due in part to large scale hemispheric shifts of the primary zones of convective activity.

These results have again shown the significant impact that Saharan dust has on not only on the local climate, but also in other regions around the globe. However, the previous two chapters have focused solely on the impact of changing the amount of dust atmospheric burden across several optical ranges. This drives the question of what are the effects of only changing the optical properties of dust? In the next chapter, we will explore how the mean climate state changes when the optical regime of dust varies over a range of optical properties.
Chapter 4

Optical Properties and Climatological Effect of Dust

4.1 Preliminaries

Mineral dust is emitted from point sources (Ginoux et al. 2012b) primarily from arid and semi-arid regions (Prospero et al. 2002). These source regions can have vastly different naturally occurring minerals which lead to substantially different mineral dust agglomerates (Claquin et al. 1999; Nickovic et al. 2012; Journet et al. 2014). These variations can persist as aeolian dust is advected through the atmosphere leading to mineralogically different aerosol plumes (Caquineau et al. 2002) and lead to many different optical properties of the dust plume (Tab. 4.1). When plotted against the optical properties from the previous chapter (Fig. 4.1), we see a large spread in single scattering albedos (SSAs) which covers both more absorbing and much more scattering regimes. For comparison, these values fall in between those commonly modeled for an internal mixture of very scattering sulfate and very absorbing black carbon aerosols (Donner et al. 2011). There is also significant spread even at single locations. When plotted against average location of the observation (Fig. 4.2) we see that these mineralogical variations can occur in quite close proximity to each other.
Table 4.1: Observed single scattering albedos of global aeolian dust in various studies. Each study has an ID used to reference Figures 4.1 and 4.2. The single scattering albedo (SSA) is that of 1 µm dust particles at 550 nm. Studies which do not present a range for observed SSAs are represented by a − in their respective columns.

<table>
<thead>
<tr>
<th>ID #</th>
<th>Study</th>
<th>Location/Campaign</th>
<th>SSA&lt;sub&gt;lower&lt;/sub&gt;</th>
<th>SSA</th>
<th>SSA&lt;sub&gt;upper&lt;/sub&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Otto et al. (2009)</td>
<td>Morocco (SAMUM)</td>
<td>−</td>
<td>0.8</td>
<td>−</td>
</tr>
<tr>
<td>2</td>
<td>Müller et al. (2010)</td>
<td>SAMUM-I</td>
<td>−</td>
<td>0.82</td>
<td>−</td>
</tr>
<tr>
<td>3</td>
<td>Müller et al. (2012)</td>
<td>SAMUM-1</td>
<td>0.9</td>
<td>0.925</td>
<td>0.95</td>
</tr>
<tr>
<td>4</td>
<td>Müller et al. (2011)</td>
<td>SAMUM-2 Cape Verde</td>
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</tr>
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<td>9</td>
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<td>11</td>
<td>Chen et al. (2011)</td>
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<td>12</td>
<td>Meloni et al. (2006)</td>
<td>Mediterranean</td>
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<td>13</td>
<td>Johnson and Osborne (2011)</td>
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<td>14</td>
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<td>Sahara (Close to Source)</td>
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<td>15</td>
<td>Jung et al. (2010)</td>
<td>Asia – Long-Range Transport 0.85-0.975</td>
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<td>16</td>
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<td>17</td>
<td>Kim et al. (2011)</td>
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<td>18</td>
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<td>Banizoumbou (1995-2009)</td>
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<td>26</td>
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<td>Ilorin (1998-2009)</td>
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<td>27</td>
<td>Kim et al. (2011)</td>
<td>IER Cinzana (2004-2009)</td>
<td>0.844</td>
<td>0.943</td>
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<td>Agoufou (2003-2009)</td>
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**Figure 4.1:** A comparison of the single scattering albedos (SSAs) of all optical properties of dust used in this experiment with those from observations. The values are calculated for 1 µm dust particles at 550 nm wavelength. The shading represents the average of all reported values for each observation. The left column shows the SSAs for the dust optical properties used in this experiment. The right column shows the SSAs for dust collected from observational campaigns and are cross-listed in Table 4.1. The uncertainty bars in the right column represent the range of values reported for each observation. The boxes labeled “SULFATE” and “BLACK CARBON” represent the single scattering albedo of column totals of an internal mixture of sulfate and aged (hydrophilic) black carbon across the visible spectrum in the GFDL Coupled Model v3 (Donner et al. 2011)
Figure 4.2: A map of observed single scattering albedos (SSAs). These values are cross-listed listed in Table 4.1. Each circle is placed either over the documented observation location or over the geometric average of the spatial extent of the observational field campaign. The shading of each circle represents the average single scattering albedo of each observation, calculated for 1 μm dust particles at 550 nm wavelength.

It is well known that the choice of optical properties affects the simulated interactions between dust aerosols and radiation (Sokolik and Toon 1999; Balkanski et al. 2007). Some studies have attempted to explain the effect of mineralogically diverse dust aerosols on climate (Perlwitz et al. 2015a, b; Scanza et al. 2015), but they are still limited to atmosphere only simulations. For computational simplicity in running GCMs, dust is often limited to a single optical regime in climate simulations. Usually this choice is of a regime representative of Saharan dust as it is the largest single contributor to global aeolian dust. However, there is still significant disagreement about what optical properties best represent Saharan dust and this choice can lead to discrepancies in model simulations of the climate’s hydrologic and thermodynamic response to dust perturbations, sometimes even of opposite sign (Miller et al. 2004).

The goal of this chapter is to explore how the choice of optical properties affects the coupled mean climate state in model simulations with a GCM. To this end, we will use a high-resolution, fully-coupled GCM forced by a variety of realistic and artificial dust optical properties and run over climatologically relevant timescales. Our hypothesis is that the sensitivity of the climate to changes in optical properties of aeolian dust will be of similar magnitude to the sensitivity to perturbations in dust atmospheric burden alone.
This chapter is organized as follows. In Section 4.2 we cover the experimental methodology including the utilized GCM, dust forcing, and computational algorithms. In Section 4.3 we examine the anomalies in several climate state variables due to changing the optical properties of dust. In Section 4.4 we inspect in greater depth the anomalies in atmospheric circulation features between different optical regimes. In Section 4.5 we expand upon induced variations in oceanic circulations for our separate dust forcings. Finally, in Section 4.6 we present our concluding remarks.

4.2 Methodology

4.2.1 Coupled Model & Tropical Cyclone Tracker

To simulate the impact of dust of various optical properties on climate, we utilize the same CM2.5-FLOR as in Chapter 3 (Vecchi et al. 2014). As a quick reminder, the atmosphere and land components of CM2.5-FLOR have a horizontal resolution of 0.5 by 0.5 degrees using a cubed sphere, finite volume dynamical core. The ocean and ice components use a lower 1 degree horizontal resolution which enhances to 1/3 degree near the equator.

We use the same TC tracking methodology as in Chapter 3 and again we define the TC track density as the number of days with a tropical-cyclone-like vortex present in a box 10 by 10 degrees centered on each 1 degree grid box. To focus on the mean response, we implement the same bootstrap, without replacement, style calculation to determine the dust-induced changes in TC density. As a reminder, this consists of constructing a time series of the annual sum of TC density in both the base and reduced dust cycle experiments, randomizing the order of these time series, and then differencing the two states. This process is repeated 1,000,000 times and the resultant set is averaged to produce a single value for each grid point. Similarly, calculations of percent change are conducted using the same formula as in Chapter 3.
4.2.2 Dust Forcing

We use the same prescribed climatological annual cycle of monthly, global mineral dust aerosol burden as the Base Emission case of Figure 2.2. This climatology was constructed using the Model of Ozone and Related Chemical Tracers, version 2 (MOZART-2) (Horowitz et al. 2003; Tie et al. 2005; Horowitz 2006) forced by the NCEP-NCAR reanalysis of 1990 (Jickells et al. 1996) and is meant to represent the high dust concentrations of the 1990s (Prospero and Lamb 2003).

We also use the same six optical regimes for dust as in Figure 3.1. The first regime (V&P) is derived from a combination of the in-situ observations of Volz (1973) and Patterson et al. (1977). The remaining sets of optical properties are derived from the refractive indicies given by Balkanski et al. (2007), namely the cases of 2.7hem, 1.5hem, and 0.9hem. In addition, we craft two other artificial optical regimes aimed at replicating extremely-scattering mineral dust. For the first regime, we multiply the imaginary part of the refractive index of the 0.9hem case by 0.1 (0.1x0.9hem) and for the second regime we set the longwave refractive index of the 0.1x0.9hem case to a constant value (0.1x0.9hem [Fixed LW]). These choices for dust optical properties have SSAs at 550nm which compare well with those observed around the globe (Fig. 4.1).

4.2.3 Experimental Design

The model is initialized in a spun-up state following 100 years of forcing with an invariant 1990 climatology of insolation, gas and aerosol concentrations, and land cover as described by Vecchi et al. (2014). We subsequently spin-up each of the six optical regimes for 100 years using the aforementioned dust climatology, which are subsequently neglected in any later difference calculation. We then run each control simulation for a further 200 years. We treat the V&P simulation as the control state as it is generally the most absorbing optical regime as well as the most methodologically different from those based on Balkanski et al. (2007). To calculate the impact of changing optical properties of dust, we difference
the 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem (Fixed LW) simulations from the V&P simulation by equivalent model year and then calculate an average of the differenced time series. Each change is run through a Student’s t-test and only those which pass at 95% are kept.

4.3 Mean Climate State Differences

We begin by looking at several key variables describing dust’s impact on climate. All averages are annual averages, unless otherwise noted. While much of the analysis will be focusing in regions where dust forcing is the strongest (both at the top of atmosphere and the surface) climate in other regions will also be directly affected as optical properties are modified globally.

4.3.1 ToA Net Clear-Sky Radiative Flux Differences

It has been shown that dust’s effect on climate is principally determined by it’s perturbation to the top of atmosphere (ToA) radiative balance (Miller et al. 2014). Thus, we start our examination with the ToA net clear-sky radiative flux, by which we mean the sum of the shortwave and longwave radiative components passing through the ToA, neglecting the effects of clouds (Fig. 4.3, first column). Our control simulation with the V&P optical regime displays a ToA radiative balance with a net import of energy into the atmosphere in the tropics and subtropics and a net export out of the atmosphere in the high-latitudes and polar regions. In our annual average, these values range from a maximum of 107 W m$^{-2}$ to a minimum of $\mathbf{-136 \, W \, m^{-2}}$, respectively, and a global average of around $\mathbf{-8 \, W \, m^{-2}}$. As the all-sky ToA radiative balance is close to 0 W m$^{-2}$, this globally averaged balance must be due to clouds.

As we change the optical properties of dust in the model simulations, we note the most apparent difference from the control state occur over much of North Africa. While global
Figure 4.3: The global changes due to changes in dust optical regimes. The values are annually averaged and only those values that pass a Student’s t-test to 95% are shaded. The columns, from left to right, show the net radiative flux differences at the top of the atmosphere (Positive downwards; W m\(^{-2}\)), the net radiative flux changes at the surface (Positive downwards; W m\(^{-2}\)), the 2m air temperature differences (K), and the precipitation changes (mm day\(^{-1}\)). The rows, from top to bottom, show the annual average of the V&P regime and the differences from it for the 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem [Fixed LW] optical regimes, respectively, and are ordered by increased scattering of dust.
changes are consistently negative and produce global averages up to $-0.3 \text{ W m}^{-2}$, the maxima are located along the southern Sahara. They monotonically decrease to over $-40 \text{ W m}^{-2}$ locally as we increase the amount of scattering to the 0.1x0.9hem (Fixed LW) regime. We also see similar changes develop over the Arabian Peninsula and throughout much of central Asia. These areas are all under the main dust plume of the model’s climatology (Fig. 2.2) and as such when the amount of aerosol scattering increases there is an increased backscatter of solar radiation leading to a negative ToA radiative flux. The extreme flux changes are nearly double those shown in Chapter 3. Interestingly, the differences across the oceanic areas covered by the dust plume, such as the tropical North Atlantic, are not as large as their land counterparts, but this is likely due to the fact that in the annual average, most emitted dust aerosols stay over the land and only in the summer months do we see a large transport over the oceans.

We also note some interesting far field results across several polar regions, mountainous regions, and the Amazon basin. In the Greenland, Barents, Weddell and Bellingshausen Seas we observe a negative ToA net radiative flux which grows in magnitude with increasing scattering. This can be explained by the drastic decreases in surface temperature (Fig. 4.3, third column) which lead to an increase in sea ice and thus more reflected solar radiation to space. Similarly, there are marked negative differences in the Rocky Mountains and the Himalayas as we increase the scattering of dust. This is again due to the drastic cooling of the climate which leads to an increased snow pack in these regions. Finally, there is a weak, but positive net changes over the central Amazon basin in the most scattering 0.1x0.9hem (Fixed LW) regime, which is more pronounced in the seasonal averages of June-November. This can be partly explained by the cooling over the region leading to a decrease in outgoing longwave radiation by the Stefan-Boltzmann Law, which for the observed changes in surface temperature give a similar result of the same order of magnitude.
4.3.2 Surface Net Clear-Sky Radiative Flux Differences

Shifting down through the atmospheric column, we now look at the surface net clear-sky radiative flux differences (Fig. 4.3, second column). The annual average of the model experiment running with V&P dust has a peak positive response in the deep tropics of up to 230 W m\(^{-2}\), a general positive flux through the mid-latitudes, and a smaller negative flux in the polar regions of nearly \(-60\) W m\(^{-2}\). This leads to a global average around 90 W m\(^{-2}\). This large positive global average is offset by the effect of clouds and turbulent fluxes at the surface.

Over the dust belt we now see largely positive surface net radiative flux focused heavily on North Africa and the Arabian Peninsula. These changes generally increase in magnitude as we increase the scattering from the 2.7hem case to the 0.1x0.9hem case with peak differences over 25 W m\(^{-2}\) along the Atlantic coastline of North Africa and a global average of 0.4 W m\(^{-2}\). Though, as we transition to the 0.1x0.9hem (Fixed LW) case we see a modest decrease in the positive changes from less scattering regimes, particularly across the Sahel. The general net positive response we observe under much of the dust plume is due to the decreased absorption by the dust cloud and increased forward scattering by the more scattering dust. However, the decreased difference in the 0.1x0.9hem (Fixed LW) regime is due to the decreased extinction of longwave radiation by dust leading to less trapping of longwave radiation under the dust plume. While locally larger in some instances, these changes are of similar magnitude to those found by adding Saharan-born dust to the climate as was shown in Chapters 2 & 3.

Many of the same far field responses we noted in the ToA radiative differences also appear at the surface because of their dependence on changes at the surface. We again see negative net radiative fluxes in the polar seas due to changes in sea ice cover leading to increased shortwave reflection. Changes in the Rockies and Himalayas also appear for much the same reason except the driving force of the radiative flux is changing snow pack. Finally, we note there is a significant decrease in net radiative flux over the Indian subcontinent. Similar to
the mechanism described for the Amazon, these differences can be in part explained by the increase in surface temperature leading to an increase in emitted longwave radiation and a decrease in the net downward flux of radiation. So despite, being directly affected by the Arabian dust plume during the annual cycle, the radiative balance over India seems to be driven by the regional surface response to changes in global dust optical properties.

4.3.3 Surface Temperature Differences

As dust is radiatively cooling the surface, we analyze the surface air temperature (Fig. 4.3, third column). In the control regime using V&P dust we see a global average around 278K, peaking at 306K in the tropics and subtropics and generally decreasing with elevation and latitude towards the poles with a minimum of 219K. The transition to more scattering optical properties leads to a general cooling trend throughout the globe with a maximum global average difference around $-0.5K$ in the 0.1x0.9hem (Fixed LW) regime. This is in line with the current understanding of scattering aerosols effect on climate (Hansen et al. 2005).

However, the response is not homogeneous everywhere and shows some considerable regional variations. Most notable is a general warming trend throughout North Africa south of the Sahara desert with temperature increases of about 1K. This change has also been shown in Chapters 2 and 3 when looking at the impact of adding scattering dust to the climate. This warming can largely be explained by a reduction of latent heat flux or evaporation over this region from associated precipitation changes. Similarly, we see an increase in the surface air temperature over the lowlands of the Indian subcontinent which can be explained by the same argument. Interestingly, the strongest difference occurs in the Barents Sea where we see up to a 4K decrease in surface air temperature. This is likely associated with the radiative fluxes mentioned before. While this area is well outside the main dust belt, the changes in this region may be due in part to larger scale circulation changes.
4.3.4 Precipitation Differences

In addition to surface temperature, we inspect the differences in global precipitation induced by changing dust’s optical properties (Fig. 4.3, fourth column). Our control simulation utilizing V&P optical properties for dust produces the common map of precipitation patterns around the world. We see the largest precipitation events occurring in the deep tropics, particularly in the Indo-Pacific basin where there are values reaching 40 mm day$^{-1}$. There are also peaks in the mid-latitude storm tracks. Conversely, the smallest precipitation occurs in the subtropical deserts of the world where precipitation values hardly reach above 0 mm day$^{-1}$.

Changing the optical regime of dust to become more scattering leads to heterogeneous changes across the globe. We observe a significant decrease of nearly $-2$ mm day$^{-1}$ in precipitation across the area of the West African monsoon in the most scattering regime. This is in line with our simulations of adding scattering dust to the climate and is due to anomalous circulations developing which hinder convective motions. Similarly, we see a decrease of about $-1.5$ mm day$^{-1}$ in precipitation across much of the regions affected by the South Asian monsoon system including the Indian subcontinent and southeast Asia in the 0.1x0.9hem (Fixed LW) case.

Conversely, we observe large increases across much of the tropical and subtropical North Atlantic peaking around 1.5 mm day$^{-1}$. This is in part due to the anomalous circulation which sets up over the West African monsoon region as it is counterbalanced by rising motion over the North Atlantic. This difference stretches through central America and into both North and South America in nearly every perturbation optical regime with a peak difference of over 4 mm day$^{-1}$ in the annual average over Central America.

Interestingly, it appears that across the Pacific there is a shift of the main convective regions towards the Equator as we see substantial negative changes on the poleward flank of the Intertropical Convergence Zone (ITCZ) noted in the V&P control simulation and balancing increases on the equatorward flank. This same equatorward shift appears to
happen in the main storm tracks across the West Pacific and the North Atlantic, indicating a possible shift in the hemispheric jet circulation.

### 4.3.5 Tropical Cyclone Differences

Spurred by the results of Chapter 3, we calculate the impact of changing the optical regime of dust on the globe’s tropical cyclone climatology (Fig. 4.4). Using the results of our TC tracking algorithm, our control simulation using the V&P dust optical regime appears similar to the standard CM2.5-FLOR simulations of TCs (Vecchi et al. 2014). We see the largest density of TCs occurs in the West Pacific basin in agreement with the observational record, with over 30 days of TCs per year in some parts. This is much larger than in the North Atlantic where we simulate an annual average of just about 3 days of TCs per year, a low bias smaller than the observed record. One large bias is the CM2.5-FLOR simulates too many TCs in the Central North Pacific which are quite uncommon in the observational record.

Turning to the changes in the other optical regimes, we focus first on the North Atlantic basin. In every case we see an increase in the TC density across the North Atlantic, with the largest difference off the Cape Verde Islands in the Main Development Region. The largest change occurs in the 0.9hem case with an increase of well over 2 storm days per year, which equates to nearly 80% of the model’s climatology. Interestingly, we find a similar result to Chapter 3 in that as we continue to increase the amount of scattering beyond the 0.9hem regime, the magnitude of the positive difference begins to decrease. This is likely because the forcing from decreasing the absorption properties of dust begins to be outweighed by the forcing from increasing the scattering properties of dust by the 0.1x0.9hem regime, leading to a reversal in the trend.

These positive differences appear to translate across Central America into the Eastern Pacific basin, but only along the coastline. Here, we have changes peaking over 4 more storm days per year in the 0.1x0.9hem (Fixed LW) case. However, there is a very sharp longitudinal
Figure 4.4: The tropical cyclone lifetime-track density differences (days with storm present per year) due to changes in dust optical properties. The shading in all but the top row denotes the percent change while the contours (shading in top row) represent the absolute difference with a contour interval of 0.5 day year\(^{-1}\). The values are indicative of changes over a 10x10 degree grid centered on each grid cell and only those values with absolute magnitude larger than 0.25 day year\(^{-1}\) are shaded. The rows, from top to bottom, show the annual average in the V&P regime and the differences from it for the 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem [Fixed LW] optical regimes, respectively, and are ordered by increased scattering of dust.
gradient around 250E where to the east we have the aforementioned strong positive difference, but to the west we have strong negative differences nearly equal in absolute magnitude stretching throughout the Central Pacific basin and into the higher latitudes of the West Pacific basin. In the 0.1x0.9hem (Fixed LW) regime, negative changes reach over −4 storm days per year. It should be noted though, that while these differences are large in absolute magnitude, they represent less than 50% of the simulated climatology. Furthermore, in contrast to the changes in the North Atlantic, those in the East Pacific appear to grow in absolute magnitude monotonically with increasing scattering suggesting a decoupling from the absorption-scattering balance.

Finally, looking at the Indo-Pacific basin, we observe what appears to be a shift towards the Southern Hemisphere in terms of TC activity. In the West Pacific there is a significant latitudinal gradient in TC activity with increases over 3 storm days per year towards the Equator and decreases over −2 storm days per year poleward. In the Indian Ocean, we see very large increases of well over 5 storm days per year with general positive differences throughout the rest of the ocean. While these changes are the largest across the globe, they are small in percent difference considering the large climatology of the region. This hemispheric shift may be due to the fact that the majority of aeolian dust is in the Northern Hemisphere potentially making the entire hemisphere less conducive for TCs.

### 4.4 Atmospheric Circulations

To better understand the non-local responses mentioned in the prior sections, we analyze in this section changes in atmospheric circulation. As dust is directly emitted into the atmosphere, we will begin there by analyzing the hemispheric wind patterns and jet structure (Fig. 4.5). The largest changes in 10m winds occur in the North Atlantic. As we increase the scattering properties of the dust we see a decrease of North Atlantic anticyclonic circulation by −1 m s$^{-1}$. There is a slight increase on its southern flank, just north of the Equator.
Figure 4.5: The global and zonally averaged wind differences due to changes in dust optical regimes. The values are annually averaged and only those values that pass a Student’s t-test to 95% are shaded/plot ted. The first, third, and fifth columns show the response in the 10m, 500mb, and 250mb wind fields (m s$^{-1}$), respectively, where the shading represents the magnitude of the wind difference and the arrows represent the direction of the wind difference. The second, fourth, and sixth columns show the zonal average of these responses, respectively, with the black curve representing the global zonal average and the red curve representing the zonal average only over the Atlantic Ocean. The rows, from top to bottom, show the annual average of the V&P regime and the differences from it for the 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem [Fixed LW] optical regimes, respectively, and are ordered by increased scattering of dust.

It’s not until scattering increases to the 0.9hem case that a similar response appears in other basins of the world. In the North Pacific there is an increase of equatorial easterlies, except for near the Central American coastline where we see anomalous westerlies. This has significant implications for El Niño which depends on anomalous westerly wind bursts along the Equatorial Pacific. Meanwhile, across the Southern Ocean there appears to be a general weakening of the winds.

While more heterogeneous, there are non-negligible changes in 10m winds over North Africa which have the potential to induce a feedback in dust emission. As the amount of scattering by dust is increased to the 0.1x0.9hem case, we note a slight decrease in surface
winds across much of the Northern Sahara desert, but a significant increase of over 1 m s\(^{-1}\) in surface winds along a band centered around 15N. This happens to roughly coincide with the Bodélé Depression in southwestern Chad, the source of the majority of dust emissions from North Africa (Ginoux et al. 2012b). Because dust emissions are strongly tied to the surface winds (Gillette et al. 1980), and that we also have a coincident decrease in precipitation leading to surface drying, it would appear that more scattering dust leads to a positive feedback in dust emissions.

Ascending to the middle troposphere, we focus on the 500mb level as it is at this level that we have the African Easterly Jet, a dominant mode of advection for Saharan dust across the North Atlantic. We note that the circulation changes are much more global in extent. Across much of the Northern Hemisphere we see a decrease in the wind speed of the poleward flank of the mid-latitude jet and a compensating increase on the equatorward side of up to 2 m s\(^{-1}\). This difference is largest in the peak of the West Pacific jet, just off the east Asian coastline. These shifts may be the cause of our observed precipitation changes in the main storm track regions of the Northern Hemisphere. Comparing the zonal average, we see a much sharper negative difference than the broader positive difference to the south, possibly indicating a slowing of the jet. Across North Africa we see what appears to be a weakening of the African Easterly Jet, but a modest increase of the equatorial easterlies just to the south of this feature. Further up in the atmosphere, at the 250mb level which is the primary level of the mid-latitude jet streams, we see a similar southward shift and partial weakening of the Northern Hemisphere jet.

To attempt to understand these apparent shifts in the mid-latitude jet system, we examine the zonal average of the zonal wind against temperature (Fig. 4.6). As we increase the amount of scattering, we note the distinct pattern of an equatorward shift of the jet with zonal average increases of over 1 m s\(^{-1}\) to the south and decreases of over \(-1.5\) m s\(^{-1}\) to the north of the V&P control simulation jet, centered around 300mb and decreasing in altitude as it shifts equatorward. There are also significant decreases in the Southern Hemisphere stratospheric
circulation. Meanwhile we see a significant cooling throughout the troposphere, focused in the upper tropical tropopause. There is also moderate cooling in the Northern Hemisphere polar stratosphere and Southern Hemisphere mid-latitude stratosphere alongside a general stratospheric warming across the tropics. Unfortunately, these changes cannot explain the zonal jet differences using a simple thermal wind relationship as the principal cooling occurs in the tropics meaning there should be an increase on the flank of the temperature differences.

Instead, we look to changes in isobars and geopotential heights for an explanation of the jet shift (Fig. 4.7). In both the 500mb and 250mb geopotential height fields, we observe a general decrease of the entire geopotential height surface, but most focused on the regions concurrent with the strongest northern hemisphere jet shifts over the North Atlantic and West Pacific. In these regions we see height drops of nearly $-40$ m and $-65$ m, respectively. These height falls are in agreement with the zonally averaged cooling throughout the tropopause that we saw (Fig. 4.6) and can potentially explain the observed shifts in the jet structure. Geopotential height falls would generally lead to a lowering of the jet structure towards the surface. This coupled with stronger height falls in the tropics would lead to an equatorward shift of the jet system as the main baroclinic regions shift equatorward. As these differences are most pronounced in the Northern Hemisphere where the majority of aeolian dust is located, this is where we see the strongest shift in the jets.

While on the topic of isobars, we can examine the impact of changing dust optical properties on the annually averaged sea level pressure field. What we see is a general increase in sea level pressure across the polar regions of the world and a general decrease across much of the mid-latitude and tropical oceans. These effects are most pronounced in the North Atlantic where there is a sharp latitudinal gradient in the mid-latitudes between a decrease of $-2$ hPa over the subtropical anticyclone and an increase of $2$ hPa over the high-latitude cyclone. This anomalous pattern correlates well with a decrease in the North Atlantic Oscillation/Arctic Oscillation (NAO) mode of variability. Studies have shown that the NAO is correlated with dust emissions from North Africa (Ginoux et al. 2004), so this may indicate
Figure 4.6: The globally, zonally averaged wind and temperature differences due to changes in dust optical regimes. The values are annually averaged and only those values that pass a Student’s t-test to 95% are shaded/plotted. The shading represents the magnitude of the temperature difference (K) and the contours represent the magnitude of the zonal wind difference (m s\(^{-1}\), intervals of 0.25 m s\(^{-1}\)). The rows, from top to bottom, show the annual average of the V&P regime and the differences from it for the 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem [Fixed LW] optical regimes, respectively, and are ordered by increased scattering of dust.
Figure 4.7: The global and zonally averaged pressure and geopotential height differences due to changes in dust optical regimes. The values are annually averaged and only those values that pass a Student’s t-test to 95% are shaded/plotted. The first, third, and fifth columns show the response in the sea level pressure (hPa), 500mb and 250mb geopotential height fields (m), respectively, where the shading represents the magnitude of the difference. The second, fourth, and sixth columns show the zonal average of these responses, respectively, with the black curve representing the global zonal average and the red curve representing the zonal average only over the Atlantic Ocean. The rows, from top to bottom, show the annual average of the V&P regime and the differences from it for the 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem [Fixed LW] optical regimes, respectively, and are ordered by increased scattering of dust.
a possible feedback of the system whereby more scattering dust leads to less dust emissions. However, this is at odds with our finding that an increase in scattering leads to an increase in winds across the southern Sahara which should lead to an increase in dust emissions. Thus, it appears as though the NAO is not correlated with dust emissions in this model simulation.

### 4.5 Oceanic Circulations

Shifting attention to the ocean, we similarly examine how changing the optical properties of dust affects oceanic circulations and in particular the Atlantic Meridional Overturning Circulation (AMOC), a principal driver of inter-hemispheric oceanic heat transport (Fig. 4.8). Plotted in density space, we see that an increase in scattering of dust by any amount leads to appreciable differences in the AMOC. In all cases there is a slow down of over 1 Sv in the densest water in the North Atlantic and that as we increase the scattering, this difference modestly grows in magnitude and expands to encompass more intermediate density waters. There is also a modest increase in the overturning circulation around the tropics in the near surface waters. This change was noted in Chapter 2 while describing anomalous near-surface currents around the Equatorial Atlantic which act to draw heat into the Northern Hemisphere. Here we observe that this northward transport seems to only extend to about 15N before transitioning to a decreased circulation.

Recent research suggests that the NAO has an appreciable effect on the AMOC through its impact on surface heat fluxes across the North Atlantic Ocean (Delworth and Zeng 2016; Delworth et al. 2016). A positive phase of the NAO corresponds to increased winds across the subpolar gyre which extract heat and drive deep water formation leading to an enhanced AMOC. In our simulations we observe mean state changes that resemble a negative phase of the NAO, with decreased sea level pressure over the subtropical gyre and increased sea level pressure over the subpolar gyre and poles (Fig. 4.7, first column). Simultaneously, we note
Figure 4.8: The Atlantic Meridional Overturning Circulation (AMOC) differences due to changes in dust optical regimes, plotted in density versus latitude space. The values are annually averaged and only those values that pass a Student’s t-test to 95% are shaded/plotted. The shading and contours (1 Sv interval) represent the magnitude of the AMOC difference (Sv). The rows, from top to bottom, show the annual average of the V&P regime and the differences from it for the 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem [Fixed LW] optical regimes, respectively, and are ordered by increased scattering of dust.
a decrease in the AMOC which overall corresponds to the expectations from the processes described by Delworth and Zeng (2016).

This decrease in the strength of AMOC as well as the aforementioned shifts in atmospheric circulations would imply shifts in the meridional heat transport of the Earth. Examining the meridional heat transport components in the atmosphere and ocean show that this is indeed the case (Fig. 4.9). In the net annual average V&P control simulation we see the atmosphere transporting over 5 PW poleward from the Equator and the ocean transporting a smaller 2 PW into the Northern Hemisphere and 1 PW into the Southern Hemisphere. As we increase the scattering properties of dust we see an increased northward transport of energy by the atmosphere from nearly 60S to 20N and a southward transport of heat from 20N to over 70N. This leads to a convergence of energy into the region around 20N which is around the peak of the dust plume advecting away from the Sahara desert. Similarly, the ocean appears to transport energy northward from around 15S to 15N, as we saw in Figure 4.8, and southward from about 15N to 45N. This similarly leads to a convergence of heat into the region around 15N. So while it appears that the AMOC shifts, it is the changes in the surface currents that drive much of the global meridional heat transport.

4.6 Chapter Summary

Mineral dust arises from mineralogically distinct regional sources which means the resultant aerosol can have substantially different optical properties. Often, the optical properties of dust aerosols are taken to be representative of the Sahara desert, as this is the single largest source of global aeolian dust. We have shown that given a range of realistic optical properties of Saharan dust, the choice of these optical properties drastically affects the mean climate of the globe. Variations in the scattering properties of dust can lead to direct radiative changes on the order of or larger than simulations where strictly Saharan-born dust atmospheric burden is increased.
Figure 4.9: The global meridional heat transport differences due to changes in dust optical regimes. The values are annually averaged and only those values that pass a Student’s t-test to 95% are plotted. Each line represents the magnitude of the meridional heat transport (PW). The top row shows the annual average of the V&P regime. The middle row shows the atmospheric transport differences from the V&P regime for the 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem [Fixed LW] optical regimes. The bottom row shows the oceanic transport differences from the V&P regime for the 2.7hem, 1.5hem, 0.9hem, 0.1x0.9hem, and 0.1x0.9hem [Fixed LW] optical regimes.
These radiative flux differences lead to global-scale changes in several other parameters as well. As dust’s optical regime is shifted to be more scattering, there is a net global cooling with regional variations in the main monsoon regions of the world. These are associated with large changes in precipitation both in the monsoon regions and the larger tropics as a whole. On average, as dust becomes more scattering we simulate a decrease in monsoonal precipitation and a shift of tropical precipitation towards the Equator. These differences also appear in more meteorological features such as tropical cyclone density where, outside a significant increase in TC activity in the North Atlantic and far East Pacific, there is a general shift of TC activity towards the Southern Hemisphere.

The choice of dust optical properties also influences the atmospheric circulation patterns throughout the troposphere, but particularly within the Northern Hemisphere. We observed an equatorward shift and weakening of the Northern Hemisphere mid-latitude jet and a slight decrease in the strength of the Southern Hemisphere mid-latitude jet as the amount of scattering by dust increased. This translated to a weaker subtropical high in the North Atlantic and an anomalously negative phase of the NAO. The principal reason for this is the significant cooling of the troposphere, particularly in the tropics which lowers the geopotential height surfaces of the atmosphere and moves the main jet axes equatorward.

We note similar changes in ocean circulations and in particular a weakening of deep water formation and a slowing of the AMOC. However, there is an increase in shallow, tropical cross-equatorial circulation which affects the meridional heat transport of the ocean. As we increase the scattering of dust, this leads to a significant cooling particularly in the Northern Hemisphere which must be balanced by increased meridional heat transport into the main dust belts by both the atmosphere and the ocean.
Chapter 5

Conclusions

The incorporation of the impacts of aerosols into models is important for accurate simulations of regional and global climate. Mineral dust is one of the largest contributors to global atmospheric aerosol burden, albeit constrained tightly to regional sources. These regional sources can drastically vary in mineralogy leading to significantly different optical properties of the eventual aerosol. Saharan dust in particular accounts for one of the largest regional aerosol forcings in the world. Due to it’s climatological advection westward, it can induce substantial direct and semi-direct interactions with the climate across both land and ocean. The principle goal of this dissertation was to understand how perturbations of Saharan dust affect climate, given different optical regimes of dust aerosols. This can be broken down into two distinct questions: What are the climatological impacts of changes to the amount of dust in the atmosphere?, and what are the impacts of changes to the type of dust being emitted to the atmosphere?
5.1 Saharan Dust and the Climates of West Africa and the North Atlantic

We started to answer these questions in Chapter 2 by using the fully coupled GFDL CM2.1 to explore the impact of realistic dust perturbations to the climate system of North Africa and the tropical North Atlantic across a range of viable optical properties. The sign of the ToA radiative imbalance is critical in determining the atmospheric response to dust (Miller et al. 2014). In line with the opposite anomalies in ToA radiation, we also noted opposing responses in cloud fields and precipitation patterns between the two optical regimes. An increase in absorbing dust lead to a significant increase in cloud cover and precipitation across much of North Africa, but particularly in the West African monsoon (WAM) region. Conversely, an increase in SCT-dust lead to a decrease in precipitation and cloud cover, again focused over the WAM region. The changes in cloud cover affect the all-sky radiative budget which leads to feedbacks on the vertical energy fluxes and a resultant dipole structure over North Africa. These changes modulate the column radiative budget and cause a general broadening of the radiative response at the surface. Simultaneously, the changes in precipitation and soil moisture content concur with significant changes in surface turbulent fluxes, particularly the latent heat flux, and surface temperature over North Africa. In general, an increase of dust in both optical regimes lead to a general cooling across the Sahara desert, but opposite responses over the WAM region with a cooling in the ABS-dust regime and a warming in the SCT-dust regime. However, over the ocean the turbulent fluxes increase the heterogeneity of the surface energy balance.

When we investigate the West African monsoon, we see that an increase in ABS-dust causes a significant increase in precipitation whereas an increase SCT-dust leads to a significant decrease in precipitation. These are associated with an anomalous Hadley-like and Walker-like circulation over the tropical North Atlantic. In the absorbing dust case, we see an anomalous ascent over the WAM region balanced by subsidence over the central tropical
North Atlantic. Conversely, in the scattering dust regime, we show anomalous subsidence over the WAM region and marginal increases in ascent over the tropical North Atlantic. These changes are in accordance with the convergence or divergence of energy in the atmospheric column over the Sahara and Sahel, as per the theory of Chou and Neelin (2003). However, this theory cannot explain the results in the WAM region. One solution could be the oceanic transport of energy to offset the aerosol induced heating or cooling, so we turned to the North Atlantic Ocean.

Over the tropical North Atlantic Ocean, the results of the ABS-dust simulation were nearly opposite those of the SCT-dust simulation. Under an increase of ABS-dust, we found a significant decrease in the UOHC of the central tropical North Atlantic. Meanwhile, an increase in SCT-dust lead to a significant increase of UOHC in this same area. These anomalies could be explained by changes in the surface wind stress curl imposing either a positive or negative Ekman pumping anomaly, respectively. This was verified by mixed layer depth anomalies across the tropical North Atlantic. It was found that the ITCZ significantly shifts and the resultant anomalies were in accordance with the wind stress anomalies. In addition, we found non-negligible surface current anomalies which aid in the transport of energy across the Equator.

### 5.2 Saharan Dust and Tropical Cyclones

Expanding upon the previous chapter’s results, we used the higher-resolution, yet still fully coupled GFDL CM2.5-FLOR model forced by the same realistic variations in atmospheric burden of Saharan dust alongside a wider range of aerosol optical regimes. Our results in the area of North Africa and the tropical North Atlantic are fundamentally the same as the previous chapter. The largest anomalies are found in the most absorbing and scattering regimes with more muted responses for the more middle-of-the-road optical regimes. We extended our analysis to encompass the rest of the globe and found significant anomalies in
other regions such as the equatorial Pacific and the Indian ocean regions and surrounding continents. The more absorbing dust leads to a net positive radiative flux anomaly at the ToA and surface over the Himalayas whereas the more scattering dust causes a net negative radiative flux anomaly at the ToA and surface. These anomalies are due to changes in snowpack across the Himalayas affecting the surface albedo. These radiative anomalies are associated with surface air temperature anomalies over the entire Indian subcontinent. There are also substantial anomalies in temperature and precipitation across the greater tropics as a whole. In the Equatorial Pacific we see a general warming and an increase in ITCZ precipitation associated with the ITCZ in the most absorbing dust regimes and an opposite cooling and decrease in ITCZ precipitation in the most scattering cases.

Using a TC tracking scheme, we examined how Saharan-born dust influences the climatology of TCs around the globe. The most consistent result is a significant decrease in TC activity across the North Atlantic as Saharan-born dust load increases. This change is largest in magnitude for the most absorbing and scattering regimes and is focused heavily in the North Atlantic main development region. These changes account for over 75% of the model’s natural TC climatology. For perspective, we compared the simulated changes in TC activity to observations over the period between the late 1960s and the mid-1980s, the period we idealized with our variations in atmospheric burden of Saharan dust. We found that our simulated TC changes could explain nearly all of the observed trend in our most extreme optical regimes and around 10% of the observed change in our more moderate optical regimes. Interestingly, there were also TC density anomalies in nearly every other tropical basin. The largest absolute anomalies occurred in the West Pacific and Indian Oceans with values exceeding those seen in the North Atlantic. However, these changes were small in percent difference as gauged by the model’s natural climatology and unsurprisingly explained little of the observed changes in these basins.

We attempted to understand these differences in TC climatology by examining several common predictive genesis potential indices (GPIs). However, we were unable to find a single
GPI capable of explaining the observed trends across optical regimes and even across basins. These inconclusive results translated to their individual components as well. We found that North Atlantic basin-wide ACE anomalies could be explained by Saharan-dust induced ToA radiative anomalies across the MDR with larger anomalies in ACE corresponding to a larger ToA radiative imbalance. We also explained the results in other basins with several theories based on larger scale circulation patterns. In the East and Central Pacific, we hypothesize that the simulated TC anomalies are due to the established-linkage between East Pacific circulations and the North Atlantic whereby certain El-Niño-like states are stimulated. Conversely, in the West Pacific and Indian Oceans, we theorize that the TC anomalies are due in part to large scale hemispheric shifts of the primary zones of convective activity.

5.3 Mean State Sensitivity to Dust Optical Regime

Finally, in Chapter 4 to complement the previous studies, we compared the same CM2.5-FLOR model simulations using a high level of Saharan-born dust atmospheric burden to explore differences in the mean state under varying optical regimes of dust. We have shown that the choice of dust optical properties substantially impacts the mean climate of the globe. By increasing the scattering properties of global dust within realistic ranges, we were able to simulate direct radiative anomalies on the order of or larger than those which arose by changing Saharan-born dust atmospheric loading by 80%. In addition, as we transition to a more scattering dust optical regime, we end up cooling the globe with regional increases in the areas of the West African and Indian monsoons. In these same areas, we noted large decreases in precipitation which is in line with the results of Chapter 2 for the most scattering dust regime. The largest surface temperature anomaly, however, is located in the northern high latitudes and is associated with changes in sea ice. On the larger scale of the Equatorial Pacific, we show a shift of tropical precipitation towards the Equator as dust
optical properties become more scattering. These anomalies also appear in the climatology of TC density where in the West Pacific and Indian Oceans we see a substantial shift of TC activity towards the Southern Hemisphere. Meanwhile, there is a significant increase in TC activity in the North Atlantic and far East Pacific as dust becomes more scattering.

To understand some of the non-local responses, we explored the influence of changing dust optical properties on the general circulation patterns of the Earth. We find that the choice of dust optical properties can lead to substantial differences in atmospheric circulation patterns throughout the troposphere, particularly aloft. The maxima of these changes occurs in the Northern Hemisphere where the majority of dust sources occur. As dust transitions to more scattering regimes, the Northern Hemisphere mid-latitude jet weakens and shifts southward. There is also a slight decrease in the Southern Hemisphere mid-latitude jet. These differences arise from the significant cooling of the troposphere which lowers the geopotential height surfaces of the atmosphere and moves the main jet axes equatorward. This translates through to the mid-troposphere where we also see a weakening of the African Easterly Jet. At the surface we see an increase of 10m wind speed over the main dust producing regions of North Africa which could lead to potential feedbacks in dust emissions. Meanwhile, as dust becomes more scattering we simultaneously simulate a weaker subtropical high in the North Atlantic and an anomalously negative phase of the NAO.

In addition to changes in atmospheric circulation patterns, we simultaneously simulated changes in ocean circulations. We focus first on the Atlantic Ocean and find that as dust becomes more scattering, we see a decrease in the Atlantic Meridional Overturning Circulation. This is attributed to the negative phase of the NAO which has been shown to drive a weakening of North Atlantic deep water formation Delworth and Zeng (2016). At the same time we note an increase in shallow, tropical cross-equatorial circulation which affects the meridional heat transport of the ocean and corroborates the results of Chapter 2 where we see significant cross equatorial flow in the upper ocean. Finally, these circulation pattern differences should lead to anomalous meridional heat transport, particularly because as we
increase the scattering of dust, we simulate a significant cooling of the Northern Hemisphere. This is in part balanced by increased meridional heat transport into the main dust belts by both the atmosphere and the ocean.

5.4 Contributions & Implications

This dissertation presents several contributions to the current literature. Firstly, this is the first study to not only use a fully-coupled climate model to study the impact of variations in dust, but also across a broad (but realistic) range of dust optical properties. This allows us to explore the sensitivity of the evolution of the climate response across full atmosphere, ocean, land, and ice components to dust optical regime. The majority of past studies have used either fixed SSTs or a slab ocean to study the impact of dust as well as only a single, generic set of dust optical properties (Table 2.1). Our results are qualitatively similar to these studies in that we find a marked response in the strength of the West African monsoon. Some work on the multi-decadal Sahel drought has found a negative correlation between precipitation anomalies over the Sahel and dust optical depth over the Atlantic (Wang et al. 2012). These anomalies are of the same order of magnitude as those we see in our most scattering dust simulations, suggesting either dust is not an important forcing agent for the drought or the correct optical properties to use in model simulations are far more scattering than those which are currently used. We also discover several coupled atmosphere-ocean responses across the tropical North Atlantic and clues to coupled responses across the Indo-Pacific basins. These have the potential to expand dust’s impact on climate both in spatial influence through teleconnections but also temporally as longer term modes of variability are affected. Understanding how dust impacts these long-term modes of variability could lead to improved predictability of the climate in both model simulations which use an interactive aerosol emission module as well as forecasts based off trends in atmospheric Saharan dust burden.
Secondly, this work builds on previous studies of dust’s impact on tropical cyclones by being the first to use a fully-coupled model to explore the response of global TC climatologies to Saharan dust, and again innovatively with a sensitivity range of dust optical regime. Tropical cyclones rely on several parameters in both the atmosphere and ocean to determine their genesis, growth, and long-term stability. Using a fully-coupled model has allowed us to definitively determine the response of TCs to the direct radiative effects of Saharan dust, namely that an increase in Saharan dust leads to decreased North Atlantic TC activity. While a direct comparison to observations is not possible given the idealized nature of this work, our findings relating a modeled decrease in TC activity across the North Atlantic to a similar trend in the observational record shows that Saharan dust could have a significant contribution to North Atlantic TC climatology. We also discover several changes to TC climatologies across the Indian and Pacific basins. These differences are likely small in comparison to the response from other, more spatially close, regional forcings, but they do show that Saharan dust can have non-local impacts. These results suggest that incorporating long-term predictions of Saharan dust atmospheric burden into not only forecast algorithms of seasonal TC activity but also climate change simulations could improve the predictability of North Atlantic, and possibly global, TC climatology.

Lastly, we have presented a systematic exploration of several dust optical properties in the same coupled modeling framework. This is a surprisingly rare example in the existing literature, and novel in it’s use of a fully-coupled GCM as well as the large number of diverse optical properties used. While only considering the direct radiative and semi-direct effects of dust, we have shown that the choice of optical property can be as important, if not more so, than variations in amount of dust aerosol in a GCM. By increasing the amount of scattering of dust aerosols, we have been able to show significant changes in the local radiative balance of the atmosphere. These differences extend throughout the globe to cause non-negligible changes in the general circulation of the atmosphere and the ocean. This has enormous implications for the heat budget of the planet, despite having its roots in a very local aerosol
forcing. Including multiple types of dust optical properties in a single model simulation could potentially lead to substantial increases in predictability across the North Atlantic and possibly even the globe.

These results expand our understanding of the coupled dust-climate system and pose several interesting implications. Most importantly, we have shown the extreme benefit of sampling multiple types of optically different dust aerosol in model simulations. Across our range of optical regimes, we have found practically opposite responses throughout the globe. Likely, the actual response to dust is somewhere in between our extreme cases so having more than one type of optical regime running concurrently could help find a balance between the opposite climate responses. Such simulations could result in decreased biases of our most accurate GCMs. This could be further enhanced by introducing the aerosol indirect effect, which we have not considered, or the biogeochemical aspects of mineralogically different dust aerosols. We have also seen strong precipitation and wind responses across several of the main dust producing regions of the world. These have the potential to impact the dust emission rate, a variable we did not explore in this thesis, and lead to feedbacks in the system. For instance, as we transitioned to more scattering dust we found a drying and an increase in wind speed across the Bodélé Depression which would suggest an increase in dust emission. Finally, some of the larger scale climate changes could have potentially important changes not assessed in this study, but possibly leading to improved predictability of the climate system. For instance, the hydrologic cycle anomalies we showed in Chapter 2 across North America have the potential to lead to drastically different climate states. Meanwhile, the suggested response in El Niño has significant implications around the globe. In all, despite dust being a regionally focused forcing it can lead to significant global perturbations through variations in atmospheric aerosol burden as well as across various optical regimes.
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