4C.4 THE CLIMATOLOGICAL EFFECT OF SAHARAN DUST ON GLOBAL TROPICAL CYCLONES IN A FULLY COUPLED GCM

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1. INTRODUCTION

North Africa, and in particular the Sahara desert, is the largest contributor to global, aerosolized mineral dust in the world (Ginoux et al., 2012). Dust emissions from North Africa have a significant interannual scale of variability with a nearly 5x increase from the 1960s to the 1990s (Ridley et al., 2014). In addition, Saharan-borne dust is commonly not of a single homogeneous composition, but is instead a combination of a multitude of regional mineral deposits (Caquineau et al., 2002).

The choice of mineralogy dataset has been show to considerably affect the interactions with radiation (Sokolik and Toon, 1999). The scattering and absorptive properties of dust lead to interactions with both the shortwave and longwave spectrums which in combination lead to a redistribution of radiative heating in the atmospheric column and in turn influence the climate (Miller et al., 2014). Such climatolgically important changes could further modulate the impact of dust on tropical cyclone frequency.

2. METHODOLOGY

We use the Geophysical Fluid Dynamics Laboratory (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. Using the model's six-hourly output, we track simulated tropical cyclones (TCs) with the method described by Zhao et al. (2009). In calculating the density of TCs, we define the TC density as the number of days with a tropical cyclone present in a box 10 by 10 degrees centered on each 1 degree grid box.

We used 6 different optical regimes to model the radiative forcing of dust, the properties of which are detailed in Figure 1. The first regime (V&P or Volz & Patterson) is derived from a combination of the observations of Volz (1973) and Patterson et al. (1977). The remaining optical regimes are calculated using Mie theory with the refractive indicies given by Balkanski et al. (2007) (2.7hem, 1.5hem, and 0.9hem). In addition we create two other artificial optical regimes. The first we multiply the imaginary part of the refractive index by 0.1 for the 0.9hem case (0.1x0.9hem). The second we use the same as 0.1x0.9hem, but set the LW refractive index to a constant value (0.1x0.9hem [Fixed LW]).

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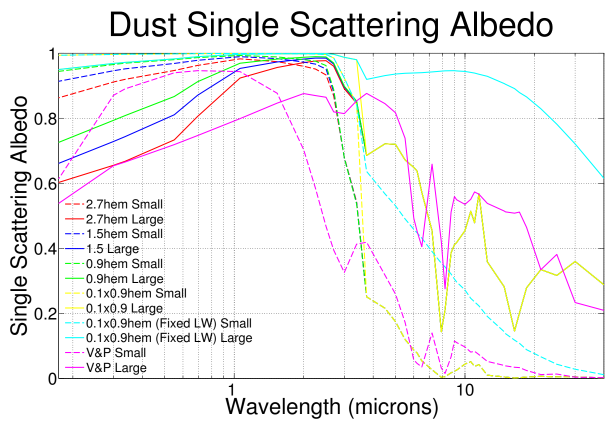

Figure 1: The single scattering albedo as a function of wavelength of all regimes of dust used in this experiment. The values are averaged over the fine (small) dust bin sizes (0.1 µm, 0.2 µm, 0.4 µm, and 0.8 µm; dashed lines) and the coarse (large dust bin sizes (1 µm, 2 µm, 4 µm, and 8 µm; solid lines).

We use the same prescribed climatological annual cycles of monthly, global mineral dust aerosol burden as in Figure 2 of Strong et al. (2015). The model is initialized in a spun-up state for each of the 6 optical regimes using the base dust climatology and run for 300 years. Using the climate state at years 100 and 200 as initialization we create 2 perturbation simulations for each optical regime where we keep the optical properties the same, but change the annual dust climatology to the reduced state. Each of these perturbations is allowed to run for 100 years in parallel to their respective control simulations. We calculate the impact of dust by aligning and then subtracting the perturbation simulations from their respective control simulations.

3. MEAN CLIMATE STATE ANOMALIES

Saharan-born dust causes significant anomalies in the net clear-sky radiative flux at the Top of the Atmosphere (ToA) in the regions closest to the main dust plume. There is a consistent negative anomaly across the tropical North Atlantic Ocean due to an increased column-averaged albedo over the relatively dark ocean surface. There is a positive anomaly over much of North Africa in the most absorbing case (V&P) which becomes negative as the dust scattering increases relative to its absorption. This is due to dust decreasing or increasing the net column albedo over the relatively reflective Sahara, respectively.

Conversely, dust almost unanimously decreases the amount of radiative flux at the surface across all optical regimes, particularly under the region of the main Saharan dust plume. This is due to the increased atmospheric attenuation of radiation.

The surface air temperature anomalies are more

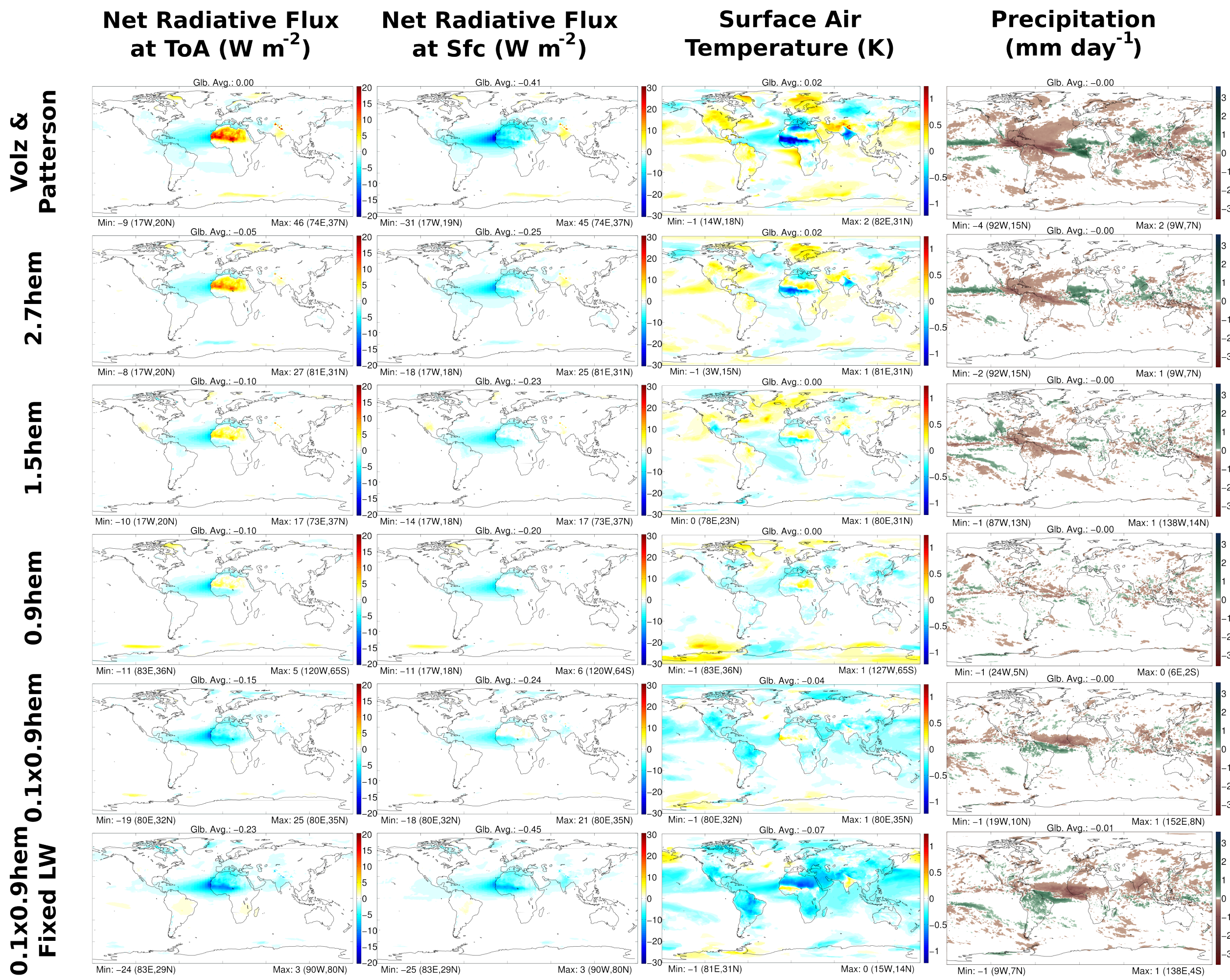

Figure 2: The simulated net clear-sky radiative flux anomalies at the top of the atmosphere (positive downwards; W m^{-2}), net clear-sky radiative flux anomalies at the surface (positive downwards; W m^{-2}), 2 meter surface air temperature anomalies (K), and precipitation anomalies (mm day^{-1}) due to an increase in Saharan dust, averaged across the summer and autumn seasons (JJASON) for various optical regimes of dust.

global with a net warming effect in the most absorbing cases and a net cooling effect in the most scattering cases. There is generally a strong cooling across the tropical North Atlantic Ocean. A dual dipole structure develops across North Africa with the Mediterranean and West Africa being out of phase with the central Sahara and the Gulf of Guinea.

In the most absorbing case there is an increase in precipitation over West Africa which becomes negative as we switch to more scattering regimes as shown by Strong et al. (2015). There is a consistent decrease in precipitation across the tropical North Atlantic Ocean, but there is a dipolar effect along the Equatorial Atlantic and Amazonian Basin. There are also significant changes through the tropical Northern Pacific Ocean and over the Indian subcontinent.

4. TROPICAL CYCLONE DENSITY ANOMALIES

It's apparent that Saharan dust has a significant impact on the tropical cyclone climatology of the North Atlantic Ocean (Figure 3). The strongest responses are in the extreme cases where we have a decrease of about 2 days per year across the central tropical North Atlantic, a nearly 75% decrease. The changes in the more intermediate dust regimes (1.5hem and 0.9hem) are less strong, but are on the whole still negative. These modeled changes explain between none and half of the observed change over the late 1960's and the mid-1980's as idealized by the forcing of Strong et al. (2015).

There are significant changes elsewhere around the globe, particularly in the northwest Pacific and south Indian Oceans. While these changes are small percentage-wise (only about a 25% change from climatology), the absolute changes are on the order of those observed in the North Atlantic. Again, these changes are small for the intermediate regimes.

5. ACCUMULATED CYCLONE ENERGY AND TOP OF ATMOSPHERE RADIATION

Focusing on the changes in the North Atlantic, we try to relate the modeled changes in tropical cyclone track density to common cyclogenesis parameters, but do not find a clear-cut solution. However, when we compare the North Atlantic accumulated cyclone energy (ACE) with ToA radiative anomalies over the Main Development Region (MDR) we obtain a monotonic relationship (Figure 4).

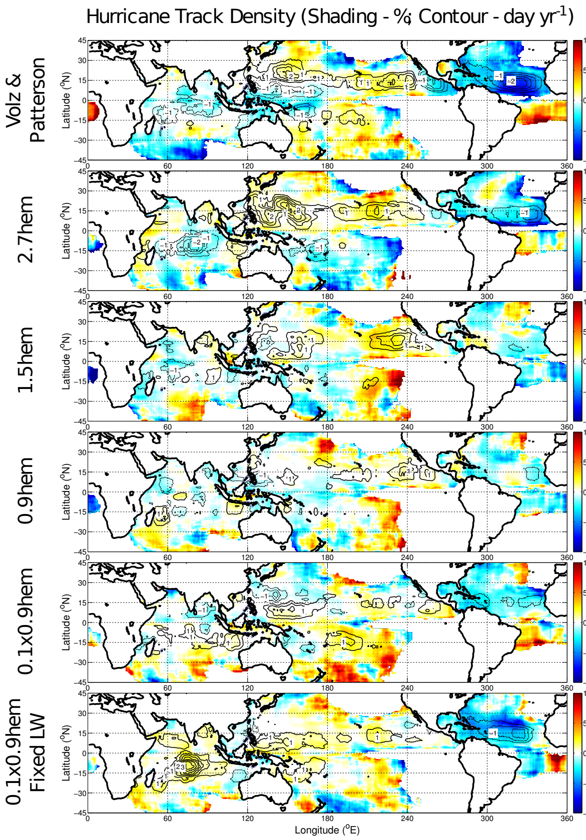

Figure 3: The anomalies in tropical cyclone density due to an increase in Saharan dust for various optical regimes of dust. The density is indicative of changes over a 10x10 degree grid centered on each grid cell. The shading denotes the percent change while the contours represent the absolute anomaly.

For the most scattering and absorbing cases we see the strongest decreases in ACE as well as the largest anomalies in ToA radiation. This relationship is non-linear, but appears to favor longwave anomalies as the single scattering albedo decreases and shortwave anomalies as the single scattering albedo increases.

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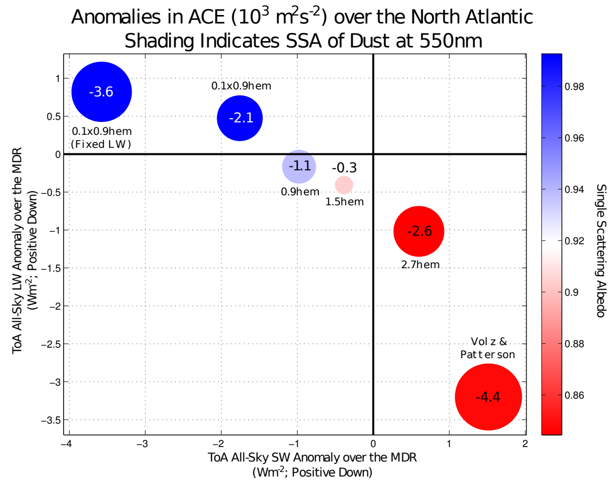

Figure 4: The anomalies in accumulated cyclone energy (10^3 m^2 s^{-2}) in the North Atlantic due to an increase in Saharan dust for various optical regimes of dust plotted against top of atmosphere all-sky shortwave and longwave anomalies (W m^{-2}) averaged across the North Atlantic main development region. The size of the circle correlates with the size of the anomaly in ACE written in the center of each circle. The color of the circle represents the single scattering albedo at 550 nm of 1 μm dust in each optical regime.

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