Impacts of Dust–Radiation versus Dust–Cloud Interactions on the Development of a Modeled Mesoscale Convective System over North Africa

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ABSTRACT

This study evaluates the impact of dust–radiation–cloud interactions on the development of a mesoscale convective system (MCS) by comparing numerical experiments run with and without dust–radiation and/or dust–cloud interactions. An MCS that developed over North Africa on 4–6 July 2010 is used as a case study. The *CloudSat* and *Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO)* satellites passed over the center of the MCS after it reached maturity, providing valuable profiles of aerosol backscatter and cloud information for model verification. The model best reproduces the MCS's observed cloud structure and morphology when both dust–radiation and dust–cloud interactions are included. Our results indicate that the dust–radiation effect has a far greater influence on the MCS's development than the dust–cloud effect. Results show that the dust–radiative effect, both with and without the dust–cloud interaction, briefly delays the MCS's formation but ultimately produces a stronger storm with a more extensive anvil cloud. This is caused by dust–radiation-induced changes to the MCS's environment. The impact of the dust–cloud effect on the MCS, on the other hand, is greatly affected by the presence of the dust–radiation interaction. The dust–cloud effect alone slows initial cloud development but enhances heterogeneous ice nucleation and extends cloud lifetime. When the dust–radiation interaction is added, increased transport of dust into the upper portions of the storm—due to a dust–radiation-driven increase in convective intensity—allows dust–cloud processes to more significantly enhance heterogeneous freezing activity earlier in the storm's development, increasing updraft strength, hydrometeor growth (particularly for ice particles), and rainfall.

1. Introduction

Mineral dust (hereafter, just dust) is one of the most important natural aerosol types present in the atmosphere. The number of studies related to dust has steadily increased since the late twentieth century, due to the availability of new observational datasets and sophisticated numerical models (Ginoux et al. 2001; Kaufman et al. 2005; Gryspeerdt et al. 2014; Twohy 2015). These studies have advanced our understanding of dust's impact

on weather and climate. Similar to other aerosols, the two major ways dust can alter ambient meteorological conditions, formation and development of cloud, and large-scale circulations are by interacting with 1) radiation (i.e., the dust–radiation interaction, dust-direct effect, or dust–radiative effect) and 2) clouds (i.e., the dust–cloud interaction, dust-indirect effect, or dust–microphysical effect) (Shi et al. 2014, Fan et al. 2016).

Generally, a layer of suspended dust heats the atmosphere within the layer and cools the surface and atmosphere below by absorbing and scattering incoming solar radiation. This pattern of shortwave (SW) heating and cooling is opposed by longwave (LW) emission within the dust layer (cooling) and the trapping of LW radiation below the layer (warming). The net radiative

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forcing by dust can be positive or negative (Tegen et al. 1996; Miller and Tegen 1998; Jacobson 2001; Balkanski et al. 2007; Yoshioka et al. 2007), the sign and magnitude of which are largely controlled by dust's optical properties, size and vertical distribution, and the albedo of the underlying surface (Highwood and Ryder 2014). Although many studies agree that the dust-radiative forcing can alter the atmospheric thermal structure, there is some disagreement regarding how these changes affect convective activity and precipitation. For example, as discussed in Wong and Dessler (2005), the net dust-radiative effect within the Saharan air layer warms and dries the atmosphere below 700 hPa, increasing low-level atmospheric stability and suppressing convection. On the other hand, Ma et al. (2012) suggest that dust-induced heating of the lower troposphere reduces local stability but enhances the meridional temperature gradient, intensifying African easterly waves (AEWs) and convective activity in late summer. Other studies indicate that the dust-radiative effect on storm strength and precipitation is highly dependent on whether dust is in a convectively active (e.g., the Atlantic Ocean) or suppressed regime (e.g., the Sahel) (Miller et al. 2004; Yoshioka et al. 2007).

In addition to directly altering the energy budget, dust is capable of indirectly redistributing the energy in the atmosphere via dust-cloud effects. Dust particles are widely understood to be effective ice nuclei (IN), owing to their nucleation efficiency, and can thus enhance heterogeneous ice nucleation processes in ice and mixed-phase clouds (DeMott et al. 2003; Twohy et al. 2009; Koehler et al. 2010; Atkinson et al. 2013). The additional latent heat release caused by a dust-induced increase in ice can potentially invigorate cloud development, prolong the life of ice clouds, and increase cold precipitation (Li et al. 2017; Paukert et al. 2017). The effectiveness of dust as cloud condensation nuclei (CCN), on the other hand, depends on its physical and chemical properties. Dust particles, which are largely insoluble, are not expected to contribute significantly to CCN concentrations compared to other soluble aerosols. However, dust may mix with other particulate matter (e.g., sea salt, anthropogenic pollutants, and secondary organic aerosol particles) or water (i.e., water adsorption), enhancing hygroscopicity and thus becoming effective CCN aerosols (Kumar et al. 2009). Suspended dust particles, activated as CCN, can increase the number concentration of cloud droplets. This delays warm rain processes and increases the opportunity for these droplets to be advected high enough to freeze, releasing additional latent heat in the upper troposphere (Lohmann et al. 2007; Karydis et al. 2011). Dust-CCN effects cannot only increase cloud albedo (Albrecht 1989) but also strengthen updrafts and potentially extend the

lifespan of cloud systems, all of which can influence the radiation budget and precipitation (Khain et al. 2005; Koren et al. 2005; Jenkins and Pratt 2008; Khain and Lynn 2009; Lebo et al. 2012). Over source regions, dust particles can be large enough to serve as giant CCN (GCCN), leading to the rapid, efficient formation of raindrop embryos through an enhanced collision-coalescence process (Feingold et al. 1999; Wurzler et al. 2000).

Of the world's major dust-source regions, North Africa (primarily the Sahara Desert) is the largest contributor to global dust emissions (Engelstaedter et al. 2006). Thus, it is not surprising that Saharan dust has received considerable attention over the past two to three decades (Prospero et al. 2002; Kaufman et al. 2002; Huneus et al. 2011), with studies examining a variety of topics over North Africa and the Atlantic Ocean, including the effects of Saharan dust on the African easterly jet (AEJ; Chen et al. 2010; Reale et al. 2011; Grogan et al. 2016, 2017; Bercos-Hickey et al. 2017), AEWs (Jones et al. 2003; Nathan et al. 2017; Bercos-Hickey et al. 2017), and tropical cyclones (Dunion and Velden 2004; Evan et al. 2006; Braun 2010; Storer et al. 2014; Herbener et al. 2014; Reale et al. 2014; Chen et al. 2015; Nowottnick et al. 2018). This study focuses on understanding how dust-radiative and dust-microphysical effects influence the development and evolution of an observed mesoscale convective system (MCS)—an organized, long-lived cluster of convective cells. MCSs produce a large portion of the total annual rainfall over central and West Africa (Mathon et al. 2002). They often initiate just south of the mobile interface, which is known as the intertropical discontinuity (ITD; Lélé and Lamb 2010) separating moist southwesterly monsoonal and dry northerly Harmattan flows during summer. Here, convective cells that take advantage of monsoonal moisture and a favorable shear environment can organize into MCSs. At times, the evaporatively cooled outflow (i.e., rainfall drag and downdraft of cold air mass) from an MCS cannot only move over arid desert surfaces and produce dust storms (also known as haboobs) (Roberts and Knippertz 2012), but also preferentially trigger convective cells at the edge the cold pools (Roberts and Knippertz 2014; Trzeciak et al. 2017). Because the storm and the dust plume are in close proximity, the likelihood that the two interact is high.

Several studies have explored the effects of either dust-radiation or dust-cloud interaction on North African or tropical Atlantic MCSs. Martínez and Chaboureau (2018) used numerical simulations of a dust outbreak in North Africa to examine the effect of dust-radiation interaction on the distribution of MCS-caused precipitation. They found that the dust-radiative effect reduced the number

of simulated MCSs and decreased total rainfall by increasing low-level stability, but increased the energy available to those MCSs that did develop. Studies exploring the dust-microphysical effect on MCSs have found that it tends to reduce total rainfall and anvil cloud extent. Min et al. (2009) suggested that higher dust concentrations increase CCN and IN concentrations, which may suppress precipitation due to increased moisture competition. Li and Min (2010) and Gibbons et al. (2018) further pointed out that the dust-microphysical impact on MCS precipitation strongly depends on whether the precipitation is convective or stratiform in nature. In the convective cores, dust invigorates convective updrafts and enhances cloud ice growth but reduces the average size, leading to stronger evaporation and less precipitation. MCS-related stratiform and anvil cloud precipitation increases, however, due to the sedimentation of more numerous, larger snow particles produced by and eventually detraining from the convective cores. The production and rapid settling of larger snow and ice particles also reduces the extent of an MCS's anvil cloud. Seigel et al. (2013) suggested that by serving as CCN and GCCN, which accelerate collision-coalescence processes, dust may increase warm rain production and weaken convection. But, on the other hand, the cloud radiative feedback can invigorate convection and increase precipitation by destabilizing an MCS's anvil region through the processes of enhancing radiative cooling, increasing supersaturation of the anvil, facilitating ice nucleation by dust and releasing more latent heat. While these studies tackle the influence of only dust-radiation or dust-cloud interaction on the MCS development, would their conclusions still be held if the other dust-physics interaction was included?

A very limited number of studies have examined both the individual and combined influence of dust-radiation and dust-cloud physics on MCSs. Shi et al. (2014) used the NASA Unified Weather Research and Forecasting (WRF) Model to study the effects of dust-cloud-radiation physics on the development of an MCS in August 2006. Their results indicate that the dust-direct radiative effect had a far greater impact on the storm than the dust-indirect effect. Specifically, they found that the dust-direct effect delayed the onset of MCS precipitation while the dust-indirect effect altered the growth of cloud hydrometeors in low to midlevels. These conclusions remained true regardless of whether dust-radiation and dust-cloud physics were simulated together or separately. Additionally, the inclusion of at least one dust-physics process caused a slight increase in domain-averaged precipitation.

Similar to Shi et al. (2014), this study also investigates the influence of dust-cloud-radiation interactions on

the development of an MCS. However, this study differs from Shi et al. (2014) in three significant aspects. First, Shi et al. (2014) used a one-moment bulk microphysics scheme with CCN/IN diagnosed from the WRF Model coupled with Chemistry (WRF-Chem)/Goddard Chemistry Aerosol Radiation and Transport (GOCART)-predicted aerosol mass concentrations, while this study explicitly represents dust-CCN-GCCN-IN processes in a two-moment bulk microphysics scheme. Studies have shown that two-moment microphysics schemes can more accurately simulate storm structures and precipitation development than their one moment counterparts (e.g., Milbrandt and Yau 2005; Morrison et al. 2009; Bryan and Morrison 2012). Second, our simulations were initialized with a predefined dust field consistent with the initial meteorology (see section 2c for additional details), while the numerical experiments of Shi et al. (2014) were initialized with no dust. This helps maintain the accuracy of the dust distribution in the vicinity of the MCS before and after development and allows dust in our experiments to interact with and affect the MCS's synoptic-scale environment from the start of the simulations. Since Moderate Resolution Imaging Spectroradiometer (MODIS) aerosol optical depth (AOD) data for the case study in Shi et al. (2014) shows a moderate amount (>1) of dust in the vicinity of the MCS prior to development, the absence of dust at the initial time of their simulations might cause the dust effects to be underestimated. Last, we study and verify the structure of the modeled MCS against satellite observations after 74 h of simulation time, which is longer than the total simulation time in Shi et al. (2014) (2 days). Extending the forecast time is critical for studying dust-induced changes to the dynamics of MCSs and their environments at lower latitudes, since dynamical adjustments (e.g., changes in vertical wind shear) due to changes in horizontal temperature gradients take longer to fully materialize at low latitudes. Since it is generally accepted that both the storm's environment and in-cloud dynamics and thermodynamics play key roles in its development and morphology (Tao et al. 2007; Fan et al. 2009; Lee 2011; Lebo and Morrison 2014), a complete assessment of the impact of dust on an MCS should include the dynamical adjustments to dust-altered meteorological fields (i.e., temperature). In this study, the MCS of interest developed near a moderate dust plume that initially appeared about 2 days prior to the formation of the MCS, suggesting that the dust plume could have altered the MCS's prestorm environment. Thus, simulations that are several days long are necessary to understand the influence of both dust-radiation and dust-cloud interactions on the MCS's development.

FIG. 1. False color red–green–blue (RGB) images from Meteosat High Rate SEVIRI Level 1.5 data at (a) 1200 UTC 2 Jul, (b) 0200 UTC 4 Jul, (c) 0200 UTC 5 Jul, and (d) 0200 UTC 6 Jul 2010. Data were obtained from the EUMETSAT website (<https://www.eumetsat.int/website/home/index.html>). This imagery is produced by setting the RGB color triplet for each pixel using the observed 12.0–10.8- μm brightness temperature (BT) difference (red), 10.8–8.7- μm BT difference (green), and 10.8- μm BT (blue), resulting in a “false-color” image. Dust plumes appear pink, thick high-level clouds are red, black areas indicate thin cirrus or contrails, thick midlevel clouds appear brown, and thin midlevel clouds are colored green. More detailed color interpretations can be found on the EUMETSAT website. Tracks of *CloudSat* (yellow line) and *CALIPSO* (red line) passing through the MCS appear in (d). The white oval in each image encapsulates the dust plume, while the white boxes in (b)–(d) surround either the (c),(d) MCS or its (b) precursor convection.

The article is organized as follows: a brief review of the studied case, model details, and numerical experiments is given in section 2. In section 3, model results are presented and verified. Section 4 discusses the impacts of dust–radiation and dust–cloud interactions, both separately and together, on the MCS, and conclusions are presented in section 5.

2. Numerical experimental design

a. Case description

We selected an MCS that developed over North Africa during 4–6 July 2010 primarily for its proximity to a moderate dust plume (Fig. 1), but also since the NASA *CloudSat* and *Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations* (*CALIPSO*) satellites passed over the MCS after it reached maturity (Fig. 1d), providing useful vertical profiles of backscatter and cloud information for model verification. Deep convection first

appeared in the southeastern corner of the dust plume at 0200 UTC 4 July 2010 (Fig. 1b) and became better organized in the late evening. Over the next 2 days, the MCS experienced two distinct convective cycles. In the first cycle, convection initiated at about 1400 UTC 4 July, reached maturity around 0200 UTC 5 July (Fig. 1c), and decayed in the early morning of 5 July. Redevelopment of the storm (i.e., the second convective cycle) began on the afternoon of 5 July, followed by the merging of multiple convective clusters into an MCS as it propagated westward along with the dust plume. The redeveloped MCS matured in the early morning hours of 6 July (Fig. 1d) and became more compact as it moved toward the Atlantic Ocean and dissipated on 7 July (not shown).

While the convection that ultimately spawned the MCS first appeared at 0200 UTC 4 July, the birth of the dust plume can be traced back to 1200 UTC 2 July (Fig. 1a). Studies have shown that surface winds strong enough to lift dust can be produced by several

mechanisms, including 1) convective downdrafts and near-surface outflow from moist convection (i.e., haboobs) (e.g., Roberts and Knippertz 2012), 2) the downward mixing of momentum in an AEW (e.g., Jones et al. 2003; Knippertz and Todd 2010), 3) the transfer of momentum from nocturnal low-level jets (LLJ) to the surface (e.g., Chen et al. 2008; Knippertz and Todd 2012), and 4) warm and dry downslope winds on the lee side of mountains (e.g., Smith 1985). In this case, the dust plume was generated via mechanism 3, where momentum associated with a nocturnal LLJ was mixed to the surface in the morning as the boundary layer deepened. Additional details regarding the birth of the plume can be found in section 3. Later, downdrafts produced by convective clusters within the mature MCS enhanced the dust loading and increased the spatial extent of the dust plume as the MCS propagated westward with time (Figs. 1b–d).

b. Model

This study uses the WRF Model V3.7.1 coupled with an online dust module (Chen et al. 2010, 2015). In this model, the evolution of the dust field is described by a dust continuity equation that accounts for the dust-flux divergence, mixing due to subgrid turbulence and cumulus convection, sedimentation, dry and wet deposition, and surface emission. Simulated dust can interact with radiation and cloud microphysics. In this study, dust emission occurs only where the surface type is “barren,” the volumetric soil moisture is less than 0.2, and the 10-m wind speed (u_{10}) exceeds a critical wind speed (u_{10c}) of 6 m s^{-1} . The calculation of the surface vertical dust flux (F_d) follows Tegen and Fung (1994), where $F_d = C(u_{10} - u_{10c})u_{10}^2$. The dimensional constant C is set to $0.4 \mu\text{g m}^2 \text{ s}^{-5}$ in this study. In general, small dust particles remain in the air longer than large ones and thus have a higher chance of influencing an MCS. Thus, our emission scheme partitions the dust mass flux into 5 size bins with median dust radii of 0.25, 0.5, 1, 2, and $4 \mu\text{m}$ using the scale-invariant physical theory of Kok (2011), which states that the particle size distribution of emitted dust is independent of meteorological conditions or soil characteristics. Note that the dust emission scheme used in this study is relatively simple compared to those used in other studies that predict the saltation mass flux and thus dust emission using soil moisture and mineralogy, surface texture, roughness length, and other soil and/or surface properties (e.g., Marticorena and Bergametti 1995; Shao 2001). While using a more physically realistic dust emission scheme is ideal, we use this simpler scheme because it has produced reasonable results for us in the past and improving dust emission is not the focus of this study. We use aerosol data assimilation to reduce the impact of inaccuracies in the simulated dust

distribution that may be due, in part, to our choice of emission scheme.

The objective of this study is to investigate the impact of dust–radiation–cloud interactions on the development of the MCS described previously. The Goddard Space Flight Center (GSFC) SW and LW radiation schemes (Chou and Suarez 1999; Chou et al. 2001) are chosen since dust-radiative effects have been implemented into these schemes (Chen et al. 2010). Absorption and scattering by dust in each size bin are computed using size and wavelength-dependent dust optical properties, including the single scattering albedo, asymmetry factor, and extinction coefficient, all of which are generated using the Optical Properties of Aerosols and Clouds software (Hess et al. 1998). A two-moment microphysics scheme developed by Cheng et al. (2007, 2010) is modified to include dust–cloud interactions. In this scheme, number concentrations of cloud droplet, ice, and rain are directly predicted (i.e., two moment), while those of snow and graupel are diagnosed (i.e., single moment). Dust can act as CCN, GCCN, and IN. Dust’s CCN activity is controlled by the hygroscopicity parameter κ , which is set to 0.05 based on Koehler et al. (2009).

Additional CCN and IN in this scheme are provided by prescribed background aerosols. Background CCN-aerosols are treated as ammonium sulfate with an average background size distribution equal to that described in Whitby (1978), while the background IN-aerosol number concentration is 400 per liter following the suggestion of Chen and Lamb (1994). Heterogeneous ice nucleation processes for dust aerosol, such as immersion freezing and deposition, are individually and directly simulated for dust aerosols, while those for background IN are parameterized with the same empirical bulk formulas used in the original version of this two-moment scheme (e.g., Fletcher 1962; Huffman 1973; Cooper 1986; Meyers et al. 1992; DeMott et al. 1998) due to a lack of detailed information about the size and chemical composition of the background IN aerosols.

c. Experimental design

Four numerical experiments, YRYM, YRNM, NRYM, and NRNM, are conducted in this study, differing only in whether or not dust–radiation and/or dust–cloud interactions are actively simulated (Table 1). The letters R and M indicate the dust-radiative and dust-microphysical effect, respectively. The letters Y and N indicate that the dust effect identified by the letter immediately afterward (i.e., R or M) is activate and inactive, respectively. For example, YRNM is an experiment that includes the dust-radiative (YR) effect, but excludes the dust-microphysical (NM) effect. Each WRF dust simulation uses three two-way nested domains with horizontal grid

TABLE 1. The design of the four numerical experiments. The letters R and M indicate the dust–radiation and dust–microphysics (i.e., dust–cloud) effect, respectively. The letters Y and N indicate the activation and deactivation of a dust–physics process (i.e., R and M), respectively.

Interaction	Experiment			
	YRYM	YRNM	NRYM	NRNM
Dust–radiation	ON	ON	OFF	OFF
Dust–cloud	ON	OFF	ON	OFF

spacings of 27 (d01), 9 (d02), and 3 (d03) km, respectively, and 41 vertical levels. The d01 covers North Africa and the east Atlantic, while d03 covers the portion of southern West Africa where the selected MCS developed (Fig. 2a). Analyzing and comparing these experiments will help us understand the impact and relative importance of the dust–radiation and dust–cloud interactions on the development, intensity, rainfall, and large-scale environment of the MCS over North Africa.

Figure 2b illustrates the design of the numerical experiments. In each experiment, the WRF dust model is integrated from 0000 UTC 3 July to 0000 UTC 7 July. To account for the presence of a significant dust plume at the initial time, forming almost 36 h prior to the MCS’s development, a separate, two-domain WRF dust simulation, forced by four-dimension data assimilation (FDDA), is conducted from 0000 UTC 1 July to 0000 UTC 3 July to produce a dust distribution that is consistent with the reanalysis meteorology at the initial time. To further improve the initial dust field and ensure that it remains reasonable during and prior to the MCS’s development, AOD level-2 data from the MODIS on board the *Terra* and *Aqua* satellites are assimilated at 1200 UTC 3 July using the three-dimensional variational data assimilation (3DVAR) method of the community Gridpoint Statistical Interpolation (GSI) analysis system. Only the observations available within a 6-h time window centered at 1200 UTC are assimilated.

FIG. 2. (a) The WRF domain configuration and (b) schematic outline of the numerical experiment. A more detailed interpretation can be found in section 2c.

FIG. 3. The 10 m wind (m s^{-1} , arrows), surface dust flux ($\mu\text{g m}^{-2} \text{s}^{-1}$, color shading), and 900-hPa geopotential height (m, blue contours) over a portion of d01 during the FDDA-driven portion of the experiment at (a) 0600 and (b) 0800 UTC 2 Jul 2010. Bodies of water, including the Atlantic Ocean and Mediterranean Sea, are masked with light gray shading. The area enclosed by the black box in (a),(b) is used for computing the area-mean values shown in (c). (c) Time series of area-mean 10-m wind speed (m s^{-1} , green), dust flux ($\mu\text{g m}^{-2} \text{s}^{-1}$, red), 900-hPa wind speed (m s^{-1} , blue), and planetary boundary layer height above ground level (m, black) on 2 Jul 2010.

The physics schemes chosen for the simulations include the Goddard radiation schemes (Chou and Suarez 1999; Chou et al. 2001), a two-moment microphysics scheme (Cheng et al. 2007, 2010), the Kain–Fritsch cumulus parameterization scheme (Kain 2004), and the Medium-Range Forecast Model (MRF) boundary layer scheme (Hong and Pan 1996). The cumulus parameterization scheme is only used in d01 and d02, while the remaining schemes are used in all three domains. The meteorological initial and boundary conditions are taken from ERA-Interim reanalysis data (Dee et al. 2011).

3. Model results

a. Dust plume emission mechanism

Figure 3 shows the modeled surface dust flux, 10-m wind, and 900-hPa geopotential height over the high dust emission region, and a time series of area-averaged quantities between early morning and around noon during the FDDA period on 2 July 2010. The surface dust flux, which is a function of the 10-m wind speed, begins to increase at 0600 UTC 2 July (Fig. 3a), peaks at

around 0800 UTC (Fig. 3b), and decreases thereafter. Our analysis shows that during the morning growth of the boundary layer, dust is lifted from the surface due to the downward mixing of high momentum associated with a nocturnal LLJ. Previous studies have described this emission mechanism by tracing the evolution of the planetary boundary layer (PBL) and the vertical wind profile before and during a dust emission event (e.g., Chen et al. 2008, Knippertz and Todd 2012). Here, we add discussion of the dust emission cycle together with the development of the PBL depth and the change in surface wind.

Before 0700 local time (i.e., 0600 UTC), the LLJ wind maximum is located at approximately 900 hPa (about 500 m above the ground), separated from the weak winds within the shallow nighttime inversion layer near the surface. These winds remain strong until dawn (Fig. 3c). With the onset of solar heating after sunrise, a near-surface superadiabatic layer develops and dry convection begins, resulting in a deepening of the PBL and an increase in surface winds and dust emission. Both surface winds and dust emission increase as higher momentum above is mixed downward (green and red

FIG. 4. (a) ERA-Interim and (b) WRF-simulated (YRYM) 900-hPa temperature (K, color shading), wind barbs (kt) ($1 \text{ kt} \approx 0.5144 \text{ m s}^{-1}$), and sea level pressure (hPa, red contours). (c) ERA-Interim and (d) YRYM 650-hPa wind speed (m s^{-1} , color shading), wind barbs (kt), and geopotential height (km, yellow contours). All figures represent conditions at 1200 UTC 5 Jul 2010.

lines in Fig. 3c), peaking at around 0800 UTC when the PBL top reaches the aforementioned wind maximum near 900 hPa (Fig. 3c). As the PBL top pushes above the level of the LLJ core, PBL wind speeds decrease, as do the surface dust emission fluxes.

b. Model verification

Before assessing the impact of dust–radiation–cloud interactions on the MCS’s development, we evaluate the model’s performance using the YRYM experiment, which includes both dust–radiation and dust–cloud interactions, to ensure that the simulated atmospheric conditions are realistic. The model’s synoptic-scale meteorological features are evaluated by comparing model results to ERA-Interim reanalysis data, while the simulated dust fields are evaluated against MODIS AOD observations. The model representation of the MCS is verified with the global infrared radiation (IR) brightness temperature (BT) dataset created by National Centers for Environmental Prediction (NCEP) Climate Prediction Center (CPC). This dataset is generated by merging IR channel data (between 10.2 and $12.5 \mu\text{m}$) from several geostationary satellites and remapping it to a 4-km horizontal resolution grid. The resulting dataset has a temporal resolution of 30 min (Janowiak et al. 2001).

Figure 4 displays the synoptic-scale meteorology from reanalysis data and YRYM shortly before the second

convective development (i.e., 1200 UTC 5 July). We select this time only for evaluation since the synoptic-scale differences between reanalysis and YRYM are very similar immediately prior to the first and second convective cycles (comparison not shown), but those prior to the second cycle are slightly larger in magnitude and coverage. Since most dust particles reside beneath the midtroposphere, model verification focuses on synoptic-scale features in meteorological fields within the same portion of the atmosphere, including sea level pressure (SLP) and 900- and 650-hPa temperature, winds, and geopotential height. Reanalysis and YRYM data share many synoptic-scale features and characteristics, including the southwesterly monsoonal flow, the position and intensity of the Saharan heat low (SHL; centered at about 21°N , 2°W), and the position and intensity of the 650-hPa AEJ. However, some discrepancies between these two datasets also exist. Compared to reanalysis, the 900-hPa temperature in YRYM is about 1–2 K colder north of the AEJ, 1–2 K warmer from about 5°N to just south of the AEJ, and about 1 K colder south of 5°N . The 650-hPa winds from YRYM are about $1\text{--}3 \text{ m s}^{-1}$ stronger south of the AEJ and $1\text{--}3 \text{ m s}^{-1}$ weaker within the jet at the West African coast. In addition, the SLP gradient and associated near-surface winds northeast of the SHL are stronger in YRYM due to a stronger simulated high pressure system northeast of the SHL in d01, particularly on 5 July. This ultimately

FIG. 5. (left) Observed *Terra* and *Aqua* MODIS AOD and (right) model-predicted AOD at 550 nm at (a),(b) 1200 UTC 3 Jul 2010, (c),(d) 1200 UTC 4 Jul 2010, and (e),(f) 1200 UTC 5 July 2010. Gray-shaded areas in (b),(d),(f) indicate where cloud top temperatures are less than or equal to 280 K. The red box in (f) shows the location of the modeled MCS.

causes simulated dust emission on 5 July to be overestimated.

Figure 5 shows the MODIS-observed and modeled (YRYM) AOD fields at 1200 UTC each day between 3 and 5 July 2010. The time of 1200 UTC is chosen since most MODIS AOD data over North Africa from the *Terra* and *Aqua* satellites are available around this time (approximately a 3-h time window centered at 1200 UTC). Recall that AOD data assimilation is performed at 1200 UTC 3 July 2010. Figures 5a and 5b compare observed and modeled AODs immediately after data assimilation. During the subsequent free integration, the model reproduces the observed AOD pattern north of

20°N over North Africa but overestimates the maximum AOD throughout much of the simulation period. At 1200 UTC 4 July 2010, shortly before the first convective development, the modeled and observed AOD magnitudes near the MCS are comparable (Fig. 5c versus Fig. 5d). At 1200 UTC 5 July 2010, before the convective redevelopment (second convective development), the model reasonably reproduces the observed AOD distribution near the MCS, but overestimates AOD magnitudes, particularly over the northern part of the dust plume (Fig. 5e versus Fig. 5f). As discussed previously, this is caused by an overstrengthening of the high pressure system northeast of the SHL in the model.

FIG. 6. IR brightness temperature from (left) NCEP CPC (K) and (right) YRYM modeled cloud top temperature (K) at (a),(b) 0200 UTC 5 Jul 2010 and (c),(d) 0200 UTC 6 Jul 2010.

Finally, a comparison of observed IR BT and modeled cloud top temperature (CTT) (Fig. 6) shows that the observed position and morphology of the MCS are well predicted by the model. During the first convective development period, simulated cloud development is too vigorous and results in the presence of more cloud systems than observed. Additionally, the modeled MCS of interest features colder CTTs and moves southwestward at a slightly quicker pace compared to observations (Fig. 6a versus Fig. 6b). Nevertheless, the model reasonably captures the observed evolution of the storm. During the redevelopment period, the model continues to reasonably reproduce the observed strength and morphology of the mature MCS, as demonstrated by comparisons against observations at the beginning of 6 July, but the modeled storm still features colder cloud tops (Fig. 6c versus Fig. 6d). The discrepancies between modeled and observed MCS CTTs could be caused by an overly intense modeled storm, but may also be due to the assumptions used to derive CTT from observations, particularly over areas where cirrus clouds are present.

The accuracy of the modeled MCS's position and morphology during the convective redevelopment period made it possible to compare the properties of the modeled storm in its mature stage to observed vertical profiles from the NASA *CloudSat* and *CALIPSO* satellites, which are available for the center of the MCS at about 0200 UTC 6 July. To make a direct comparison possible, the Goddard Satellite Data Simulator Unit (G-SDSU) is used to transform model data into

equivalent satellite-observed quantities from *CloudSat* and *CALIPSO* (e.g., reflectivity and backscatter) along the satellites' tracks (Fig. 7a) using the same scanning geometry (Matsui et al. 2013, 2014). While the observed pattern of dust-caused backscatter is well-captured in the YRYM simulation, a lack of modeled dust in the upper portion of the observed dust layer causes modeled dust backscatter to be underestimated there (Fig. 7b versus Fig. 7c). The underestimation of backscatter might be because of the ignorance of smaller dust particles in our dust parameterization. As for the MCS, the model accurately simulates the height of the brightband, the position of convective cores within the MCS, and the morphology of the storm (Fig. 8a versus Fig. 8b and Fig. 8f versus Fig. 8g). However, the model underestimates reflectivity above the bright band, with over-predicted ice content (snow in particular) in the convective cores but underpredicted cloud ice in the anvil clouds. These biases of modeled reflectivity and ice water content could be due to many reasons, such as errors in the assumed size distribution and shapes of ice phase particles used by the G-SDSU to simulate *CloudSat* reflectivity and different partitioning of cloud water content used in the *CloudSat* and the model. The exact reasons why remain under investigation.

Overall, the modeled (YRYM simulation) large-scale meteorological patterns, dust AOD and backscatter distribution, and MCS characteristics are realistically simulated compared to reanalysis data and satellite observations. This increases our confidence in the model's

FIG. 7. (a) Simulated AOD (unitless, shading) and the 260-K cloud-top temperature isoline (K, red contours) from YRYM at 0200 UTC 6 Jul 2010. Vertical cross sections of 532 nm backscatter coefficients from (b) CALIOP L2 data and (c) G-SDSU processed YRYM results along the instrument track [black line in (a)] from A to B are shown. Brown shading in (b) represents missing data. Black contours in (b),(c) bound areas where valid CALIOP data is present.

ability to reasonably simulate dust–physics processes for this case study, which allows us to assess their impact on the MCS’s development, as described in the next section.

4. Impacts of dust physical processes on MCS development

a. Improvement of model MCS structure by simulating dust–physics processes

Before we investigate the influence of different dust–physics processes on the MCS’s development and the mechanisms behind them, we first confirm that the WRF-Dust model produces a better MCS structure when dust–physics processes are included by comparing modeled clouds from four numerical experiments to observations. Figure 8 shows that all four experiments are able to produce the MCS of interest but with different structures, intensities, and cloud properties. With respect to *CloudSat* observations (Figs. 8a,f), YRYM has a slightly better representation of the convective cores, asymmetric anvil clouds, bright band, and anvil coverage than YRNM, and YRNM in turn performs better than NRYM and NRYM for the same characteristics. However, all of them overestimate the ice water content (snow in particular), especially in the convective cores, while underestimate the content and extent of ice crystals in the anvil clouds. Overall, the modeled cloud structure is better predicted in the experiments that include the dust–radiation interaction (i.e., YRYM and YRNM). YRYM, however, best captures the weak-echo anvil clouds above 10 km located between (12.7°N, 5.9°W) and (10.1°N, 6.4°W) (Fig. 8a versus Fig. 8b). This weak-echo zone is composed primarily of small ice

particles that have been transported vertically in the convective cores before detraining and spreading horizontally on the downshear side of the MCS, leading to an anvil that is asymmetric about the convective cores, and can only be accurately reproduced in a simulation that reasonably reproduces the dynamics, thermodynamics, and microphysics of the MCS. The improved representation of the weak echo and asymmetric anvil cloud in YRYM, which includes both sets of dust–physics processes, implies that the impacts of dust on the MCS are properly captured and can be further dissected by analyzing and comparing the four experiments.

b. Impact of dust–radiation versus dust–cloud interactions on MCS development

To evaluate the impact of dust–radiation and dust–cloud interactions on the MCS’s development, we compare the results of the four experiments within a box that encapsulates and follows the MCS with time. Since the morphology and position of the MCS differ between experiments, the box used for each experiment can also differ slightly at any given time. For comparisons focusing strictly on the MCS, only the grid points within the MCS are considered. The analysis that follows involves area-summed cloud properties and area-averaged dynamical parameters. Figure 9 shows a time–height cross section of MCS-area-summed profiles of total hydrometeors (both liquid and solid phases) and MCS-area-averaged convective upward mass fluxes. The convective updraft mass flux is defined as the product of the vertical velocity (w) and the air density (ρ) for all grid points within the box at which $w > 1 \text{ m s}^{-1}$ (Lebo and Morrison 2014). Convection first develops after about 38 h into the free-run forecast period (i.e., 1400 UTC 4 July). Local maxima

FIG. 8. Vertical cross sections of (left) radar reflectivity (dBZ) and (right) ice water content (mg m^{-3}). (a),(f) *CloudSat* data, and (b)–(e),(g)–(j) G-SDSU processed model data along the instrument track (black line in Fig. 7a) at 0200 UTC 6 Jul 2010. The radar reflectivity in (a) is from the *CloudSat* 2B-GEOPROF product (Marchand et al. 2008), and the modeled radar reflectivity is estimated by G-SDSU. The retrieved estimates of ice water content in (f) is from the *CloudSat* Radar-Only Cloud Water Content Product (2B-CWC-RO) (Austin et al. 2009), and the modeled ice water content includes cloud ice and snow.

of total hydrometeors appear at about 6-km height during the next two evenings, evidence of a strong diurnal pattern in the life cycle of the MCS. The storm redeveloped in the afternoon of 5 July. The end of the time series (i.e., 0200 UTC 6 July) marks the time after which the modeled storm merges with and becomes indistinguishable from a coastal convective system.

Before delving further into the results, it should be noted that differences among the MCSs in each experiment during the first diurnal cycle are mainly due to different dust-physical processes (i.e., dust-radiation and/or dust-cloud interactions), while differences present during the second diurnal cycle (i.e., convective redevelopment) are not only due to different dust-physical processes, but also to the effects of each experiment's unique MCS on its environment during the first diurnal cycle (i.e., first convective development). For example, differences in the cold pools produced during the first convective development can alter each storm's environment differently prior to the second cycle. In short, the "initial conditions" for each experiment at the start

of the second diurnal cycle are different, making it more difficult to attribute differences appearing in the second diurnal cycle to the use of different dust-physical processes versus differences generated in the first diurnal cycle that the model has carried forward and likely amplified in time. Thus, we will primarily use the results of the first diurnal cycle to determine the impact of dust-radiation and/or dust-cloud processes on MCS's development and rainfall amounts. Nevertheless, the results of the second diurnal cycle are still important and will be analyzed for two reasons: 1) the dust-radiative effect on vertical shear requires a longer integration time to be fully realized in simulations over low-latitude areas and 2) *CloudSat* samples the observed MCS during the second diurnal cycle, offering a great opportunity to evaluate model results near and within the MCS as discussed in section 3.

While the hydrometeor distributions in all four experiments are generally similar during the first convective cycle (Figs. 9 and 10), including two peaks in precipitation (the rain field in Figs. 10e–h), the experiments with the

FIG. 9. Model-predicted, area-summed total cloud hydrometeors (kg m^{-3} , color shading) and area-averaged convective updraft mass flux ($\text{kg m}^{-2} \text{s}^{-1}$, black contours) between 1200 UTC 4 Jul and 0200 UTC 6 Jul 2010 for experiments (a) YRYM, (b) YRNM, (c) NRYM, and (d) NRNM. Area summation occurs only over cloudy grid points within a box surrounding the MCS at each time. Cloud hydrometeors include cloud water, cloud ice, rain, snow, and graupel. A grid point is considered “cloudy” if the total hydrometeor mixing ratio at that point is greater than or equal to $10^{-6} \text{ kg kg}^{-1}$. The solid and dashed red lines represent the 0° and -40°C isotherms, respectively.

same dust–radiation setting (YRYM and YRNM, NRYM and NRNM) are particularly similar, mirroring each other with regards to MCS formation time and changes in intensity and cloud depth (Figs. 9 and 10). Comparing experiments that differ only by the presence of the dust–radiation physics reveals that the dust–radiative effect delays the formation of the storm by about an hour but increases the intensity of its initial convection (Fig. 9a versus Fig. 9c and Fig. 9b versus Fig. 9d). This convective invigoration brings more cloud water aloft, facilitating ice nucleation and growth (Fig. 10a versus Fig. 10c and Fig. 10b versus Fig. 10d). As a result, a subsequent increase in mass of snow and graupel, which predominantly grow by autoconversion (i.e., ice crystals to snow), accretion (i.e., riming), and collision–coalescence process, is present (Fig. 10e versus Fig. 10g and Fig. 10f versus Fig. 10h), leading to more precipitation at the first rainfall peak (between 1800 UTC 4 July and 0000 UTC 5 July; the rain field in Figs. 10e–h). On the other hand,

the dust–microphysical effect is shown to extend the lifetime of hydrometeors in the MCS (Fig. 9a versus Fig. 9b and Fig. 9c versus Fig. 9d). As cloud ice and snow persisted longer in the MCS (Fig. 10a versus Fig. 10b and Fig. 10c versus Fig. 10d for blue contours; Fig. 10e versus Fig. 10f and Fig. 10g versus Fig. 10h for green shading), the probability of collection of small hydrometeors by large ones can increase and result in more precipitation. Thus the MCS in the simulations with the inclusion of dust–cloud interaction, such as YRYM and NRYM, produces more rainfall than that without the dust–cloud interaction during the second rainfall peak (between 0300 and 0900 UTC 5 July; the rain field in Figs. 10e–h) of the first diurnal cycle. In all, YRYM and YRNM produce the strongest and second strongest MCSs, respectively, in terms of convective updraft mass flux and cloud top temperature (not shown) during the first diurnal cycle, while NRYM and NRNM produce noticeably weaker storms. In addition, as with the total

FIG. 10. Model-predicted densities (kg m^{-3}) for cloud water (gray shading), cloud ice (blue contours, only the 0.5, 1.0, 1.5, 2.0 isolines are shown), rainwater (red shading), snow (green shading), and graupel (black contours, only the 1, 2, 3, 4 isolines are shown) summed over cloudy grid points within a box surrounding the MCS at all times between 1200 UTC 4 Jul and 0200 UTC 6 Jul 2010 for experiments (a),(e) YRYM, (b),(f) YRNM, (c),(g) NRYM, and (d),(h) NRNM. A grid point is considered “cloudy” if the total hydrometeor mixing ratio at that point is greater or equal to $10^{-6} \text{ kg kg}^{-1}$.

hydrometeors (Fig. 9), maximum instantaneous area-summed values of individual hydrometeor types, including rain, ice, snow, and graupel, are larger in simulations that include the dust–radiation interaction, with the largest maximum values occurring in YRYM for all five types (i.e., including cloud; Fig. 10). This supports the notion that, at least in this case, the impact of dust–radiation interactions on the MCS’s development overpowers that of dust–cloud interactions when dust concentrations are high. However, with the inclusion of dust–cloud interaction YRYM and NRYM produce the highest and the second highest rainfall, respectively, during the first diurnal cycle. Compared to NRNM, the YRYM, NRYM, and YRNM simulations produce 39%, 18%, and 14% more of rainfall, respectively. This result lies in stark contrast to the results of many other studies, which have shown that the dust–cloud interaction decreases rainfall totals (e.g., Min et al. 2009; Seigel et al. 2013). Further investigation will be required to understand exactly why this is. Shi et al. (2014), on the other hand, found, as we did, that the inclusion of at least one of the two dust–physics processes increased the rainfall produced by their simulated

MCS. However, the dust-induced percent increase in rainfall in their study is far smaller than ours, ranging from 0.09% to 3% for different experiments. This could be due to the use of a two-moment microphysics scheme in this study as opposed to a one-moment microphysics scheme in theirs, the selection of completely different real-case MCSs, or differences in numerical experiment design (e.g., differences in dust-initial condition and model integration time).

It is interesting to see that the results from the convective redevelopment period are slightly different from those of the first convective development. As in the first development cycle, YRYM is still slower to initiate convection but ultimately produces a storm with the largest convective mass fluxes (Fig. 9a), the greatest amount of hydrometeors (Fig. 9a), and the coldest cloud-top temperature (figure not shown). The results of the other three experiments, on the other hand, show fewer differences. The storms in both NRYM and NRNM develop slightly stronger than those during the first development cycle, in part due to more moisture and a larger MCAPE during the redevelopment cycle. For YRNM, the early development of large hydrometeors between 1400 to

1600 UTC 5 July (Fig. 10) induces downdrafts at midlevels (figure not shown) that suppress the development of updrafts, leading to a more similar intensity to those in NRYM and NRNM and a weaker intensity than that in YRYM. Specifically, the absence of the dust-microphysical effect during the early stages of the redevelopment period in YRNM causes water vapor from evaporating midlevel supercooled droplets to effectively deposit onto ice particles left over from the first convective development (i.e., the Wegener–Bergeron–Findeisen mechanism; Wegener 1911; Bergeron 1935; Findeisen 1938). These ice particles quickly grow in size and produce large hydrometeors (snow, graupel and rain) through accretion (i.e., riming) and collision–coalescence processes, enhancing downdrafts and causing simulation results to depart from those of YRYM.

In the following subsections, we discuss the impact of dust–radiation and dust–cloud interactions on the MCS development in more detail. Since most previous studies have examined the direct and indirect effects of dust on storm development separately, a logical question to ask is: “Would the impact of individual dust–physics process on the MCS development change when the other dust–physics process is simulated together?” For example, would the impact of dust–radiation interaction on the MCS development differ with and without the dust–microphysics interaction? In the following subsections, our discussion will address the impact of dust–radiation (section 4d) and dust–cloud (section 4e) interactions on the MCS development separately. However, in each subsection we further investigate whether the impacts of individual dust–physics interaction on the MCS development would change when the other dust–physics interaction is also included. Since the dust–radiative effect can directly modify an MCS’s environment, potentially affecting storm development, we first analyze the impact of dust–radiation interaction on the storm’s environment (section 4c), which will be used to help interpret the radiative effect on the MCS development.

c. Impact of dust–radiation interactions on MCS environment

In this section, the analysis only considers non-cloudy grid points within the boxed area encompassing each experiment’s MCS. Figures 11a and 11b are time series of area-averaged environmental surface radiation fluxes. It is clear that the dust–SW interaction, in which suspended dust absorbs and scatters incoming solar radiation, reducing the net downward SW radiation at the ground (Fig. 11a), accounts for much of the total dust–radiative effect on the surface energy budget during the day (Fig. 11a versus Fig. 11b). The impact of the dust–LW interaction, on the other hand, is

evident both day and night as dust warms the lower atmosphere (figure not shown), which increases the surface downward LW radiation flux (Fig. 11b). Figures 11c and 11d show maximum convective available potential energy (MCAPE) and maximum convective inhibition (MCIN) respectively, which are computed at each grid point by adiabatically lifting an air parcel starting at an altitude below 3000 m where the equivalent potential temperature (EPT) is maximized (Colman 1990). During the first convective development, the dust–radiative effect consistently increases the maximum of MCAPE by about $150\text{--}400\text{ J kg}^{-1}$ throughout the simulation (Fig. 11c), but only increases MCIN by about 20 J kg^{-1} during the day when MCIN magnitudes are small. At night, when MCIN values are large, the dust–radiative effect decreases MCIN by about $50\text{--}100\text{ J kg}^{-1}$. Changes in both MCIN and MCAPE can be attributed to changes in environmental temperature and low-level temperature and moisture where an air parcel is lifted. Since changes in MCIN have a strong diurnal cycle, while changes in MCAPE are fairly consistent throughout the day, it must be that the dust-induced changes in MCIN are not only due to changes in low-level temperatures (where the maximum EPT is located), since these changes would affect both MCIN and MCAPE, but also to temperature changes at other levels. The reasons why changes in MCIN and MCAPE behave differently are discussed further below.

Over the desert, the level of free convection (LFC), which is the starting point of the vertical integral used to compute MCAPE, is relatively high and often above the dust layer. Thus, temperature changes due to the dust–radiative effect mainly occur below the LFC. This has far greater consequences for MCIN than MCAPE. During the daytime, dust heats both the lower troposphere, where the dust–LW heating exceeds the dust–SW cooling, and midtroposphere (e.g., 700 hPa), where the dust–SW heating outweighs the dust–LW cooling (Chen et al. 2010). Heating the lower troposphere decreases MCIN while heating the midtroposphere increases it. The net result is a slight increase in MCIN (e.g., 1200–2200 UTC 4 July in Fig. 11d). At night, the dust–LW effect continues to heat the lower troposphere but now, in the absence of the dust–SW effect, also cools the midtroposphere, both of which decrease MCIN. As a result, dust-induced changes in MCIN feature a strong diurnal cycle. For MCAPE, since dust has only a minor impact on temperatures above the LFC, any changes in MCAPE are mainly controlled by changes in low-level temperatures, which are in turn primarily governed by the dust–LW interaction, ultimately heating the lower troposphere both day and night. Thus MCAPE increases are seen at all times of the day. Note that the

FIG. 11. Convective energy and surface radiation fluxes for all four experiments, averaged over noncloudy horizontal grid points within a box surrounding the MCS at each model output time between 1200 UTC 4 Jul and 0200 UTC 6 Jul 2010. Average (a) net downward surface shortwave radiation flux (W m^{-2} , downward is positive), (b) downward surface longwave radiation flux (W m^{-2} , downward is positive), (c) maximum convective available potential energy (MCAPE, J kg^{-1}), and (d) maximum convective inhibition (MCIN, J kg^{-1}). A horizontal grid point is considered “cloudy” when at least one point in the associated vertical column is cloudy (i.e., total hydrometeor mixing ratio $\geq 10^{-6} \text{ kg kg}^{-1}$). The two thick, dashed horizontal lines in (a),(b) mark 350 and 400 W m^{-2} , respectively. The y-axis tick-mark interval for both (a),(b) is 25 W m^{-2} , while (c),(d) both use a y-axis tick-mark interval of 100 J kg^{-1} .

dust–radiation-induced temperature changes described here also affect the static stability beneath the midtroposphere. Specifically, static stability increases during the daytime and decreases at night (figure not shown). Changes to environmental MCAPE, MCIN, and static stability can all impact storm development.

It is notable that the influence of dust on the net downward SW and downward LW radiation fluxes at the surface is similar during both diurnal cycles. Dust–physics processes still increase environmental MCAPE during the second cycle, but the differences in MCAPE between all four experiments at night and in the early morning are reduced (figure not shown). Differences in MCIN between the experiments are also reduced, particularly at night (figure not shown), which is partially due to the lingering effects of low-level cold pools produced during the first convective development.

By altering the thermal structure of the atmosphere, dust–radiation interactions can also affect dynamical fields (i.e., vertical wind shear) through the thermal wind relation if the interaction lasts long enough (i.e., about 2–3 days for low latitudes; [Chen et al. 2015](#)). [Figure 12](#) shows the vertical wind shear between 600 and 900 hPa in the late afternoon for both convective cycles. At 1800 UTC 4 July (42 h into the free run), during the first convective development, the dust–radiative effect mainly enhances vertical wind shear in and near the AEJ ([Fig. 12a](#) versus [Fig. 12b](#)). At 1800 UTC 5 July (66 h into the free run), during the second convective development ([Fig. 12c](#) versus [Fig. 12d](#)), continuous dynamical adjustments to dust–radiative changes in the temperature field over the previous 24 h cause changes in vertical shear to increase and expand. Dust–radiative heating of the low to midtroposphere within the high

FIG. 12. (a),(c) 600–900-hPa vertical wind shear (m s^{-1} , arrows and brown shading) and thickness (km, blue contours) for NRYM at 1800 UTC 4 Jul and 5 Jul 2010, respectively. (b),(d) 600–900-hPa vertical wind shear (m s^{-1} , arrows and brown shading) and AOD (unitless, red contours starting from 0.5 with a 0.5 interval) for YRYM data, and 600–900-hPa thickness differences (m, green contours) between YRYM and NRYM at 1800 UTC 4 Jul and 5 Jul 2010, respectively. Thicker red contours represent larger AOD values. Areas shaded in black indicate cloudy regions where the cloud top temperature is less than or equal to 280 K in (a),(c) NRYM and (b),(d) YRYM.

dust concentration region enhances vertical wind shear (Fig. 12d) and produces a clockwise horizontal circulation anomaly surrounding the dust plume at 600 hPa (figure not shown), similar to that in Chen et al. (2015). These changes in vertical wind shear cause the MCS in YRYM to have a larger and more elongated and asymmetric anvil/stratiform region than that of NRYM—in better agreement with *CloudSat* reflectivity profiles of the observed MCS (cf. Figs. 8a,b,d).

d. Impact of dust–radiation interactions on MCS development

To explore the overall impact of dust–radiation physics on the MCS’s development, we examine and compare the differences between YRYM and NRYM, and between YRNM and NRYM, respectively. The experiments comprising each pair differ only in whether or not they simulate dust–radiation interactions. Time series of model total hydrometeors and convective updraft mass flux in Fig. 13 are used in this discussion to describe the MCS’s strength and cloud development. Our discussion of the impact of dust–physics processes

on the MCS mainly focuses on the time period of the first convective development. Differences between the experiments in each pair show that the dust–radiative effect delays the initial formation of the MCS by about 1 h (1500 UTC 4 July) independent of the dust–cloud interaction, but produces a thicker cloud later at night (Fig. 9a versus Fig. 9c and Fig. 9b versus Fig. 9d; Figs. 13a,b). This is a consequence of the previously discussed diurnal changes in low-level stability/MCIN and MCAPE caused by the dust–radiation interaction. The stabilization effect and larger MCIN in the lower atmosphere during the day delay the formation of the MCS and associated hydrometeors. However, a larger MCAPE becomes available ($\sim 50\%$ increase) for storm initiation and this is because of the delay of the convection (i.e., more solar heating on the surface) and the dust–radiative effect (Fig. 11c) ultimately invigorating the MCS. At night, dust–LW effects reduce stability and MCIN in the lower atmosphere, further enhancing the storm development (i.e., increasing hydrometeor concentrations and convective updraft mass fluxes; Figs. 13a,b). Consistent with the rain field in Figs. 10e–h,

FIG. 13. Differences of model-predicted, area-summed total cloud hydrometeor density (kg m^{-3} , color shading) and area-averaged convective updraft mass fluxes ($\text{kg m}^{-2} \text{s}^{-1}$, black contours) over a box enclosing the MCS at each model output time between (a) YRYM and NRYM, (b) YRNM and NRNM, (c) YRYM and YRNM, and (d) NRYM and NRNM from 1200 UTC 4 Jul to 0200 UTC 6 Jul 2010. (e) Changes in the dust-radiation (dust-cloud) effect on total cloud hydrometeors and convective updraft mass fluxes due to the addition of dust-cloud (dust-radiation) physics, computed by taking the difference between either Figs. 13a and 13b or Figs. 13c and 13d. Total cloud hydrometeors include cloud water, cloud ice, rain, snow, and graupel.

both YRYM and YRNM produce more rainfall during the first diurnal cycle than their no-dust-radiation counterparts (i.e., greater red areas than blue areas in Figs. 14a,b).

Although the MCS's development is primarily controlled by the dust-radiation interaction, which is often studied in isolation, dust-microphysical effects can potentially modulate the impact of dust-radiation interactions on the MCS's development. The overall pattern in Fig. 13e, which takes the differences between Figs. 13a and 13b, indicates that the addition of dust-cloud physics further strengthens the MCS (i.e., Figs. 13b,e are in a similar phase during the MCS strengthening stage

between 1700 UTC 4 July to 0000 UTC 5 July). In other words, the simulation of dust-cloud and dust-radiation interactions together causes the storm to strengthen more rapidly during its early development (i.e., less negative in Fig. 13a than in Fig. 13b) and ultimately remain stronger after reaching maturity. The reasons for this are explored below.

The intensification of the storm due to dust-radiative effects increases the amount of dust transported above the freezing level, increasing the heterogeneous freezing—particularly immersion freezing by dust (dotted black line in Fig. 15)—of supercooled water (both solid and dotted black lines in Fig. 15) and thus the amount of latent heat

FIG. 14. Precipitation rates (m h^{-1}), computed as the amount of precipitation accumulated per hour of integration time, summed over a box enclosing the MCS at each model output time between 1200 UTC 4 Jul and 0200 UTC 6 Jul 2010 for experiments (a) YRYM and NRYM, (b) YRNM and NRNM, (c) YRYM and YRNM, and (d) NRYM and NRNM. Shaded areas represent the precipitation difference between each pair of experiments, with positive and negative differences shaded red and blue, respectively.

released between 5- and 9-km height, ultimately promoting buoyancy, stronger convective mass fluxes, and greater ice content. This aerosol-driven “convective invigoration” (e.g., Morrison and Grabowski 2013; Altaratz et al. 2014; Fan et al. 2016) explains why updraft differences between YRYM and NRYM are more positive ($\sim 40\%$) than those between YRNM and NRNM in the beginning of the first diurnal cycle (Fig. 13a versus Fig. 13b).

The impact of dust–physics on surface precipitation is consistent with the dust effects on the MCS development described earlier. Figures 14a and 14b show that the dust–radiative forcing delays the onset of precipitation but produces more vigorous storms with higher precipitation rates. Comparing Figs. 14a and 14b reveals that simulating the dust–microphysical effect together with dust–radiation interactions magnifies the increase in precipitation rate due to the dust–radiative forcing alone. In fact, from late afternoon of 4 July to early morning of 5 July, a period in which the storm was

intensifying rapidly, the increase in precipitation due to dust–cloud effects was comparable to the increase due to the dust–radiative effect alone. We suspect that the increase in precipitation efficiency caused by the dust–cloud effect is due to an increase in the immersion freezing process by dust as discussed previously. Note that the decreased precipitation rate between 0200 and 0600 UTC 5 July produced by adding dust–cloud effects to a simulation that previously only accounted for dust–radiative effects is the result of an earlier decay of the storm (i.e., larger negative value in Fig. 14a).

e. Impact of dust–cloud interaction on MCS development

To evaluate the impact of the dust–cloud interaction on the MCS’s development, one can compare the results of NRYM and NRNM (the dust–indirect effect) and YRYM and YRNM (the dust–indirect effect modified by the dust–radiative effect). In addition to earlier discussion that the dust–cloud effect has a smaller impact

FIG. 15. Time and area-summed vertical profiles of ice nucleation rates due to different processes and for all experiments. Solid lines represent nucleation rates due to homogeneous nucleation, deposition nucleation on dust particles, and heterogeneous freezing on background aerosols ($\mu\text{g kg}^{-1} \text{h}^{-1}$). Dotted lines represent nucleation rates due to immersion freezing on dust particles ($\mu\text{g kg}^{-1} \text{h}^{-1}$). Summation is performed over a box enclosing the MCS at all model output times between 1200 UTC 4 Jul and 0200 UTC 6 Jul 2010.

on the MCS's development than the dust-radiative effect (Figs. 9c,d versus Figs. 9a,b), two more interesting findings are presented below.

First, the dust-microphysical effect strengthens the storm at the first 2 h after storm's initiation, weakens the storm development during the next 4 h, and then enhances the storm development again until the end of the first diurnal cycle (Fig. 13d). In theory, the hygroscopic aerosol indirect effect should delay hydrometeor growth/dissipation and onset of precipitation, invigorate convection subsequently due to more latent heat above the freezing level in clouds, and extend cloud lifetime (Twomey 1974, Rosenfeld et al. 2008). While the MCS in NRYM shows some of these features (i.e., convective invigoration and longer cloud lifetime) (Fig. 9c versus Fig. 9d), dust is not expected to be a significant source of CCN near dust-source regions due to its hydrophobic to low-hygroscopic nature. However, dust can serve as

GCCN that facilitate faster droplet growth and thereby enhance the warm-cloud collision-coalescence process and ice particle riming. This dust-GCCN effect causes the short-term intensification at storm's initial state, which consumes vapor and hydrometeors and produces a little bit more rainfall [tiny red region (indicated by arrow) between 1400 and 1600 UTC in Fig. 14d]. The consumption of water vapor and hydrometeors then leads to a less active storm and reduced growth of hydrometeors in the afternoon and early evening of 4 July (Fig. 13d). The dust heterogeneous nucleation processes, in particular the immersion freezing, invigorate the storm development later at night and delay the decaying of the storm. As the result, compared to NRNM, NRYM has slightly more rainfall when the MCS initiates, followed by reduced hydrometeor concentrations and weaker convection during the beginning of the first diurnal cycle, while the opposite features as well as long-lasting cloud particles are found at the later part of the first diurnal cycle (Figs. 13d and 14d).

Second, the impact of dust-cloud processes on the storm is greatly affected by the simulation of dust-radiative effects (Fig. 13c versus Fig. 13d). Figure 13e, which is the difference between Figs. 13c and 13d, shows that the impact of dust-radiation physics on dust-cloud-induced MCS changes is exactly the same as the impact of dust-cloud physics on dust-radiation-induced changes, which are discussed in the previous subsection (i.e., identical to the difference between Figs. 13a and 13b). A comparison of Figs. 13d and 13e reveals that the dust-radiation-induced changes to the dust-cloud effect are almost always greater in magnitude but opposite in sign (i.e., almost out of phase) relative to the dust-cloud effect alone. Specifically, when dust-radiation physics are included, the dust-cloud effect no longer slows the initial intensification of the storm, but instead accelerates it by catalyzing ice formation within the storm, a reality made possible by dust-radiation-enhanced convective motions transporting greater amounts of IN-capable dust into the upper portions of the storm. This is consistent with our previous discussion indicating that the dust-direct effect has a greater impact on the MCS's development than the dust-indirect effect.

Figures 14c and 14d show the impact of dust-indirect effects, both with and without dust-radiation interactions, on precipitation rates. These figures support the idea that the MCS is affected by dust-cloud processes differently depending on whether or not dust-radiation interactions are also simulated. In the absence of dust-radiative effects, the dust-cloud effect decreases precipitation rates during the first third of the MCS's life cycle and increases it thereafter. However, when the

dust-radiative effect is included, the dust-microphysical effect enhances precipitation rates throughout the storm's life cycle, but causes it to dissipate sooner (i.e., negative precipitation rate anomaly; Fig. 14c).

5. Summary and discussion

This study evaluates the impact of dust–radiation–cloud interactions on the development of a North African MCS that appeared on 4–6 July 2010 using the WRF–Dust model. This storm is selected primarily because 1) it developed near a moderate dust plume, and 2) vertical profiles of *CloudSat* cloud reflectivity and *CALIPSO* aerosol backscatter are available for the center of the MCS after it reached maturity during its second convective development cycle, allowing a comprehensive evaluation of our numerical experiments. To examine the roles of the dust-radiative and dust-microphysical effects on the storm, four numerical experiments are conducted that differ only in whether or not dust–radiation and/or dust–cloud interactions are included. We show that although the dust distribution and MCS are reasonably predicted in all of the experiments, the simulated MCS in each experiment is affected by different dust-physical processes.

The impacts of dust–radiation–cloud interactions on the MCS are summarized schematically in Fig. 16, with the key conclusions of the study described below.

The dust plume and the position, morphology, cloud structure, and microphysical properties of the MCS are best replicated by the experiment that simulates both dust–radiation and dust–cloud interactions (i.e., the YRYM experiment). While both categories of dust-physic processes affect the MCS in important ways, dust-radiative effects have a greater influence on the MCS's development in areas where dust concentrations are high. This is consistent with the conclusion of Shi et al. (2014).

The dust–radiation interaction alters the storm's environment in areas where dust is present, ultimately producing a more intense MCS (Fig. 16a versus Fig. 16c). Specifically, in the daytime, the dust-radiative effect heats the dust layer in the midtroposphere and, to a lesser degree, the atmosphere below. The former is primarily due to the dust–SW interaction, while the latter is caused by the dust–LW interaction. These effects stabilize the lower atmosphere, increase MCIN, and delay the formation of the storm and associated hydrometeors. However, dust–LW heating near the surface increases environmental MCAPE values, which results in a more intense storm by the afternoon. At night, suspended dust emits and traps LW radiation which cools the dust layer in the midtroposphere and heats the atmosphere below. This

increases MCAPE and reduces both MCIN and the stability of the lower atmosphere, promoting storm intensification late at night. These results are consistent with those obtained by Martínez and Chaboureaud (2018), which reveal that the dust-radiative effect inhibits convective initiation but, by increasing environmental CAPE, produces stronger long-lived MCSs with enhanced hydrometeor growth and more precipitation.

Simulations accounting for dust–radiation physics produce a significantly more intense MCS both with and without the dust–cloud interaction (Fig. 16a versus Fig. 16c and Fig. 16b versus Fig. 16d). The addition of the dust–cloud interaction further strengthens the MCS by increasing the activity of heterogeneous nucleation processes—primarily the immersion freezing of supercooled cloud droplets by dust—which produces more hydrometeors, including cloud, rain, ice, snow, and graupel (Fig. 16c versus Fig. 16d). Moreover, the extended duration of the dust–radiation interaction, about 2–3 days in this study, allows changes in vertical wind shear (low-level westerlies and higher-level easterlies) via a thermal wind adjustment to become apparent and begin affecting the simulated storm, increasing the extent of its anvil cloud in better agreement with *CloudSat* observations (Fig. 16d).

Unlike the impacts of the dust–radiation interaction, which are affected little by the presence of the dust–cloud interaction, the impact of the dust–cloud interaction on the MCS is greatly affected by the inclusion of its counterpart. Simulating the dust–cloud interaction alone weakens the development of clouds early in the forecast period (inset in Fig. 16a versus Fig. 16b), but enhances heterogeneous ice nucleation, and extends cloud lifetime later in the period (Fig. 16b). However, when the dust–radiation interaction is also involved, enhanced convection and increased transport of dust into the upper portions of the storm allows dust–cloud–IN physics to promote heterogeneous freezing processes, which accelerate, rather than slow, the initial intensification of the storm by enhancing updrafts, the growth of hydrometeors—particularly ice particles—and precipitation (Fig. 16d). This result also illustrates the complex nonlinearity of dust–radiation–cloud interactions in general. This nonlinear effect on cloud structure and precipitation is not negligible, and should be considered carefully when studying the dust/aerosol radiative or microphysical effects on cloud development separately.

The dust-radiative effect delays the onset of precipitation, but results in higher precipitation rates (Fig. 14b) and increases accumulated rainfall by about 14%. The dust-microphysical effect suppresses cloud development and precipitation early in the forecast period but

FIG. 16. Schematic depiction of the impact of dust–radiation and/or dust–cloud interactions on cloud development in each experiment. The cloud structure and low-level winds in each panel are representative of conditions present at storm maturity. Symbols representing different hydrometeor species are defined in the legend. Cloud ice production due to pure water freezing (blue hexagons) and dust-induced ice nucleation (gray hexagons) are shown. The number and thickness of the curved arrows in each panel represent the relative strength of the convective updrafts in each experiment. The size of the boxed “initiation” cloud in each panel represents the relative intensity of the storm when it first appeared. The orange layer marks the portion of the atmosphere that is influenced by the dust–radiation interaction. See text for additional details.

ultimately increases accumulated precipitation by about 18% relative to NRNM, slightly larger than the increase in precipitation induced by the dust–radiative effect alone, and extends the duration of precipitation due to enhanced heterogeneous freezing (Fig. 14d). When dust–radiation and dust–cloud interactions are simulated together, the dust–radiation-induced storm intensification as well as the dust–cloud-induced enhancement of immersion freezing activity and increase in storm lifetime increase the precipitation rate and total rainfall by as much as 39% during the first convective development cycle, which is larger than the sum of the precipitation increases due to the dust–radiation and dust–cloud effects individually. This emphasizes the importance of the synergy between dust–radiation and dust–cloud interactions for cloud development.

Similar to the CCN– and IN–cloud effects described in previous studies (e.g., Albrecht 1989; Lohmann et al. 2007; DeMott et al. 2003), our study shows that the dust–cloud effect, via the dust–CCN processes, delays the initial development of the MCS and associated hydrometeor growth, but ultimately produces a stronger storm via dust heterogeneous nucleation processes. However, our study shows that the dust–cloud effect increases

rainfall, while most other studies examining the dust–cloud effect show a decrease in rainfall (e.g., Min et al. 2009; Seigel et al. 2013). Shi et al. (2014) also report an increase in rainfall due to dust–radiation and/or dust–cloud effects, but the percent change found in their study is one order of magnitude smaller than ours. This inconsistency between our study and others could be due to the use of different case studies with different environmental conditions (e.g., humidity, dust physicochemical properties and distribution), the latter of which can strongly modulate the effect of dust–cloud processes on precipitation (Khain 2009). Also, to our knowledge, this study is the first to highlight that the conclusion of one dust–physics effect on the MCS development can be altered by the inclusion of the other dust–physics effect, which shows the importance of accounting for and understanding the nonlinear synergistic effect between dust–radiation and dust–cloud physics when assessing the dust impact on the cloud structure, precipitation, and overall development of an MCS. Additional research focusing on this nonlinearity would have great value in understanding the role of dust in triggering new convection and affecting its subsequent organization into an MCS.

Although the present study demonstrates that the dust-direct effect has a greater influence on the MCS's development than the dust-indirect effect, it is important to keep in mind that the MCS in this case study developed in an environment close to the dust-source areas where dust concentrations were relatively high. The findings of this study, including the relative importance of the dust-radiative and dust-microphysical effects on MCS development, may not apply to cases where organized convection is impacted by lower dust concentrations and/or dust that has been transported far from its source. In addition, surface brightness may also play a role in the overall radiative effect of dust. Additional investigation is needed to determine if and how the impact of dust-physics changes under dust and environmental conditions that differ from those in this case study.

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