When Does the Saharan Air Layer Impede the Intensification of Tropical Cyclones?

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ABSTRACT: We investigate the circumstances under which the Saharan air layer (SAL) has a negative impact on the intensification of tropical cyclones (TCs) over the North Atlantic Ocean. Using hurricane tracking, aerosol optical depth (AOD) data, and meteorological analyses, we analyze the interaction of the SAL with 52 named TCs that formed over the east and central Atlantic south of the Cape Verde islands between 2004 and 2017. Following the categorization of negative SAL influences on TC intensification by Dunion and Velden, only 21% of the investigated storms can be classified (28% of all storms that encountered the SAL), and 21% of the storms continue to intensify despite the presence of the SAL. We show that among TCs that encounter the SAL, there is evidence supporting a weak negative correlation between the magnitude of TC intensification and the ambient AOD. However, above-average Saharan dust abundance in the vicinity of TCs is not a good independent indicator for storm nonintensification. To better understand the specific processes involved, a composite study is carried out, contrasting storms that intensify in the presence of the SAL against those that do not. We find that sheared air masses on the north side and drier air from the northeast of the storm early on during its lifetime, in addition to higher AOD, are associated with TC nonintensification in proximity to the SAL.

KEYWORDS: Tropics; Wind shear; Dust or dust storms; Tropical cyclones; Humidity

1. Introduction

Tropical cyclones (TCs; in this study referring to both tropical storms and hurricanes) are one of the most dangerous natural hazards when making landfall and their impact may be expected to increase in the future (e.g., Emanuel 2005; Pielke et al. 2008). To make accurate forecasts of their tracks and intensity, it is crucial to understand their cyclogenesis and the environmental conditions under which they intensify. For a TC to develop and intensify, a range of prerequisites such as sufficiently high sea surface temperature (SST), low vertical wind shear, high midtropospheric humidity, and a pre-existing disturbance need to be present (e.g., Palmén 1948; Riehl 1948; Gray 1968). In addition, over the Atlantic, Saharan dust may influence storm intensification.

Mineral dust aerosols are frequently lifted from the Saharan desert and are transported westward over the subtropical to tropical Atlantic Ocean in synoptic outbreak events during spring to early fall (e.g., Prospero et al. 2002; Laken et al. 2013).

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These outbreaks are associated with convective disturbances over West Africa that move westward in connection with African easterly waves at a frequency of 3–4 days (Goudie and Middleton 2001). The Saharan air layer (SAL) influences the atmosphere above the Atlantic in many ways. Through scattering and absorption of the dust aerosols, the lower atmosphere within the dust layer is heated and cooled beneath (e.g., Diaz et al. 1976; Carlson and Benjamin 1980; Dunion 2011; Davidi et al. 2012), thus affecting the regional radiative budget and modifying atmospheric stability. In the SAL at roughly 800–550 hPa, the air is characterized by nearly constant potential temperature and water vapor mixing ratio (Carlson and Prospero 1972; Karyampudi and Carlson 1988). Through the modification of the radiative budget the dust also affects SSTs over the tropical Atlantic: increased (decreased) Saharan dust is associated with cooling (warming) of the Atlantic surface temperature during the early hurricane season from July to September (Lau and Kim 2007). All these processes can contribute to distinctive differences in moisture, temperature, and wind profiles between SAL-influenced and SAL-free air, as was found by Dunion and Marron (2008) through examination of over 750 rawinsondes from the 2002 hurricane season.

In addition, dust may also impact cloud microphysics, as it is a significant source of cloud condensation nuclei (CCN) and ice nucleating particles (e.g., Levin et al. 1996; DeMott et al. 2003; Twohy 2015). In the case of TCs, most studies suggest that additional CCN generally weaken storm intensity (Cotton et al. 2007; Khain et al. 2008; Khain and Lynn 2011; Rosenfeld et al. 2011, 2012; Zhang et al. 2007, 2009). For example, when simulating the effects of the increase in CCN concentrations on Hurricane Katrina (2005) during landfall, Khain et al. (2008) and Khain and Lynn (2011) showed that enhanced CCN concentrations led to a reduction of maximum wind speeds by 10–15 m s^{−1}, as well as to a reduction of the area of strong winds.

FIG. 1. Schematic of the hampering influences of the SAL on TC intensification in the Atlantic main development region, with dusty air in orange and dust-free air in blue: 1) microphysical processes, 2) intrusion of dry air, and 3) vertical wind shear.

However, the microphysical processes at play are far from clear cut because of the complexity of various aerosol indirect effects. Jenkins et al. (2008), for instance, noted the presence of SAL outbreaks prior to two cases of tropical cyclogenesis in 2006 and through analysis of satellite and aircraft data, postulated possible rainband invigoration by the SAL. Herbener et al. (2014) also showed that aerosol introduced to the periphery of an idealized TC can lead to decreases in the storm extent but increases in its intensity through aerosol–cloud dynamics interactions.

Due to the strong temperature gradient at the southern and southwestern edge of the SAL and following the thermal wind balance, the geostrophic wind maximizes in this region to form the midlevel African easterly jet [AEJ; Burpee 1972; the exact mechanism for the maintenance of the AEJ is more intricate and is discussed in, e.g., Cook (1999) and Thorncroft and Blackburn (1999)]. The AEJ is associated with large vertical and horizontal wind shear and an induced meridional ageostrophic circulation that results in enhanced upward motion of air south of the jet and subsidence northward within the SAL (e.g., Carlson and Prospero 1972; Karyampudi and Carlson 1988; Braun 2010). South of the AEJ, this circulation supports deep convection, which interacts with the large background cyclonic vorticity in this region (Karyampudi and Carlson 1988; Braun 2010), supporting the development of tropical disturbances.

While the above studies mostly point to a negative impact of dust on TCs through microphysical processes (Fig. 1, label 1), it is not clear whether radiative and associated dynamical effects of dust and the SAL reduce or strengthen TC intensity. In 2004, Dunion and Velden (Dunion and Velden 2004, hereafter DV2004) published an article on various negative influences of the SAL on TCs. Using geostationary operational satellite data, DV2004 identified SAL outbreaks and inferred that they reduced the intensification of the investigated TCs by three mechanisms: First, the intrusion of dry SAL air into a TC suppresses convection by reducing the convective available potential energy (CAPE) and by promoting downdrafts (Fig. 1,

label 2). Second, the SAL is associated with vertical wind shear that decouples the lower circulation from the upper-level circulation of the storm (Fig. 1, label 3). Third, the temperature inversion at the base of the SAL inhibits deep convection and acts to stabilize the environment (Evan et al. 2006; Dunion 2011). Whereas DV2004 examined only few storms, other authors assumed a broader applicability of the results: Evan et al. (2006) showed an inverse correlation between TC days and dust cover, Lau and Kim (2007) found a negative correlation between SAL activity and SST when comparing the 2005 hurricane season with 2004, and similarly Sun et al. (2008) found cooling over the main development region and drying over the western North Atlantic when comparing the active 2005 hurricane season with the dustier 2007 season. When examined on a long-term basinwide seasonal scale starting in the 1950s, Wu (2007) also noted significant inverse correlations between the Atlantic hurricane peak intensity and SAL activity, although the processes responsible for this relationship remained unclear. On another hand, Shu and Wu (2009) stated that the SAL may enhance TC genesis, but it inhibits further intensification once the storm is developed: In their composite study of 274 cases from 37 named TCs, they found a positive influence on TC growth when the SAL is present in the northwestern quadrant of the storm and a negative influence when the SAL is to the south of the storm. Braun (2010) re-evaluated the role of the SAL. Investigating NASA satellite datasets, NCEP global analyses, and composite analyses of the early stages of the storms, he found no statistically significant differences in the characteristics of the SAL for strengthening and weakening storms in the first days after genesis, although he did not rule out the possible role of the SAL in storm evolution when considered simultaneously with other environmental factors. In contrast, in a high-resolution modeling study, Reed et al. (2019) found that dust had a suppressing influence on their simulated TCs: they compared the frequency, duration, and intensity of TCs in the presence of African dust versus a low-dust experiment. In their model, the frequency of North Atlantic TCs increased by 27%, they lived longer by 13% and were slightly stronger by 3% in a low-dust environment relative to a high-dust environment.

In this study, we use an extended dataset of Atlantic TCs (tropical storms and hurricanes) together with satellite data of aerosol optical depth (AOD) and reanalysis data of the European Centre for Medium Range Weather Forecast (ECMWF) for a systematic study of the SAL and Atlantic TCs of the period 2004–17. In the next section, we give an overview of the data used. Section 3 describes the methods applied, and in section 4, the results are presented: while section 4a focuses on evaluating AOD and storm intensity, section 4b presents results from a composite study, focusing on dynamical processes related to the SAL that may impede the intensification of TCs (Fig. 1), and last the geographical locations of the storm tracks are contrasted in section 4c. Section 5 contains the conclusions.

2. Data

In this study, we investigate tropical storms and hurricanes ($>17.5 \text{ m s}^{-1}$) that form over the eastern and central Atlantic

FIG. 2. July–September climatological mean SEVIRI AOD over the Atlantic for the period 2004–17. The yellow-outlined rectangle denotes the TC genesis region considered in this study, spanning 5°–20°N and 15°–45°W.

south of the Cape Verde islands and are first recorded in the HURDAT database within a box of 5°–20°N and 15°–45°W (Fig. 2), in the period from 2004 through 2017. This is the part of the Atlantic where TCs emerge from African easterly waves and where the SAL may potentially and significantly interact with the developing TCs for several days (Fig. 2). Filtering all available storms for the above region, we find cases occurring during the months from July to early October, when African easterly waves are most active (e.g., Grist 2002). Altogether, this selection yields a total of 56 TCs (at tropical storm and hurricane strength), of which 52 are considered (Table 1). Karl (2004), Colin (2010), Isaac (2012), and Bertha (2014) are excluded because of incomplete aerosol data, particularly during initial storm intensification, which prevents categorization of the TC.

To track the TCs, we use data from the National Hurricane Center's (NHC) revised Atlantic hurricane database (HURDAT2), provided by the National Oceanic and Atmospheric Administration (NOAA). From HURDAT2, we use 1-min-averaged maximum sustained wind speeds (kt) and geographical coordinates of the storm center at 6-hourly intervals at 0000, 0600, 1200, and 1800 UTC.

In addition, to assess the effects of other parameters on the intensification of the TCs, we evaluate relative humidity at 700 hPa and tropospheric horizontal wind data at various pressure levels from the ECMWF operational archive. Data were retrieved at 900 (925 prior to 2007), 850, 700, 300, 250, and 200 hPa and 0000, 0600, 1200, and 1800 UTC at a high resolution of $0.25^\circ \times 0.25^\circ$. Daily mean high-resolution SST data are obtained from NOAA's daily Optimum Interpolation Sea Surface Temperature (daily OISST) dataset at a spatial resolution of $0.25^\circ \times 0.25^\circ$ (data available from <https://www.esrl.noaa.gov/psd/data/gridded/data.noaa.oisst.v2.highres.html>).

For analyzing the daily dust abundance above the Atlantic Ocean, we use a retrieval product from the Spinning Enhanced Visible and Infrared Imager (SEVIRI) radiometer that flies onboard the Meteosat Second Generation (MSG-1, -2, and -3)

geostationary satellites. Meteosat is centered above 0° longitude and the equator and provides an image of the Atlantic Ocean approximately every 15 min. The measurements are retrieved in near real time by the ICARE Cloud–Aerosol–Water–Radiation Interactions Thematic Center, operated by the University of Lille (<http://www.icare.univ-lille1.fr/>) to derive AOD at 550 nm from the 635- and 810-nm channels (Thieuleux et al. 2005). We use the level-3 daily aerosol product, which is generated from all individual data of the day (i.e., from 0400 to 1945 UTC). The spatial resolution is 3 km at the crossing of the equator and the Greenwich meridian. Comparison of the ICARE SEVIRI AOD with sun photometer measurements from the Aerosol Robotic Network (AERONET) shows a slight bias in the SEVIRI data with large AODs being slightly larger (Bréon et al. 2011, their Fig. 2). However, considering the high spatial resolution and availability of the SEVIRI data over the whole investigated time period 2004–17 and the fact that we use these data in a qualitative rather than in a quantitative manner, small potential biases will not significantly affect the results.

As a dynamic measure of storm size, the radius of the outermost closed isobar (ROCI) is obtained from the NHC tropical cyclone extended best track dataset for each storm at 6-hourly intervals (Demuth et al. 2006). While TCs can vary greatly in size, there is variety in its measure and different definitions exist in the literature focusing on various aspects of the TC structure (e.g., Merrill 1984; Kimball and Mulekar 2004; Chavas and Emanuel 2010; Knaff et al. 2014; Schenkel et al. 2017; Feldmann et al. 2019). In this study, ROCI is chosen to represent the extent of the storm circulation, which provides a measure of the distance at which environmental factors of interest (e.g., AOD and wind shear) may directly interact with the TC. For every day of the storm, ROCI is determined as the average distance in nautical miles from the center of the storm to the outermost closed isobar. This is converted to degrees latitude by dividing the nautical miles value by 60 and

TABLE 1. Tropical storms (TS) and hurricanes (HU) considered in this study, grouped by categories described in Table 2, below. The table provides storm name, year of storm occurrence, category on the Saffir–Simpson scale, the maximum Saffir–Simpson category reached within the first five storm days, the initial longitude and latitude according to the HURDAT dataset, and the mean $2 \times \text{ROCI}$ averaging radius over the first five days in degrees of latitude. An asterisk indicates the AD1 storms that are *not* included in the AD1b category.

Storm name	Year	Storm category	Max storm category in the first five days	Initial lon/lat	5-day mean averaging radius ($^{\circ}$ lat)
DV1					
HU Igor	2010	HU4	HU4	−22.7/14.0	6.3
HU Leslie	2012	HU1	TS	−27.4/12.9	7.9
DV2					
TS Josephine	2008	TS	TS	−21.9/12.3	6.4
TS Fiona	2010	TS	TS	−41.8/14.0	5.2
TS Florence	2012	TS	TS	−23.1/12.2	4.9
TS Joyce	2012	TS	TS	−31.7/10.7	6.6
DV3					
HU Ivan	2004	HU5	HU4	−27.6/9.7	6.1
HU Lisa	2004	HU1	TS	−32.4/13.3	4.5
HU Karen	2007	HU1	HU1	−35.9/10.0	6.0
HU Danielle	2010	HU4	HU2	−31.1/10.7	7.4
HU Jose	2017	HU4	HU4	−33.5/9.3	4.9
AD1					
HU Danielle*	2004	HU2	HU2	−21.8/12.3	5.1
HU Florence*	2006	HU1	TS	−39.4/14.1	7.1
HU Dean	2007	HU5	HU5	−28.9/12.2	6.1
HU Fred	2009	HU3	HU3	−24.0/11.8	7.0
HU Bill	2009	HU4	HU4	−32.0/11.8	7.5
HU Earl*	2010	HU4	TS	−19.3/12.4	7.0
HU Julia	2010	HU4	HU4	−20.5/12.9	6.7
HU Katia	2011	HU4	HU1	−19.0/9.5	6.1
HU Nadine	2012	HU1	HU1	−38.0/15.5	7.8
HU Fred*	2015	HU1	HU1	−17.5/11.6	5.6
HU Gaston	2016	HU3	HU1	−19.4/11.0	6.4
AD2					
HU Frances	2004	HU4	HU4	−35.2/11.1	5.8
TS Debby	2006	TS	TS	−21.7/11.6	6.1
HU Helene	2006	HU3	HU1	−22.0/11.9	7.4
TS Melissa	2007	TS	TS	−25.8/14.0	5.6
HU Bertha	2008	HU3	HU3	−22.9/12.7	6.2
HU Ike	2008	HU4	HU4	−37.0/17.2	7.0
TS Ana	2009	TS	TS	−24.0/14.3	5.2
TS Gaston	2010	TS	TS	−32.2/12.2	6.0
HU Maria	2011	HU1	TS	−35.9/11.5	6.0
HU Ophelia	2011	HU4	TS	−37.0/11.6	5.5
HU Philippe	2011	HU1	TS	−17.5/8.3	6.9
TS Erin	2013	TS	TS	−23.3/13.9	5.3
HU Humberto	2013	HU1	HU1	−17.6/13.0	7.7
HU Danny	2015	HU3	HU3	−29.3/9.6	5.3
TS Grace	2015	TS	TS	−23.1/12.0	4.1
TS Fiona	2016	TS	TS	−32.2/12.0	4.5
HU Lee	2017	HU3	TS	−23.1/10.4	4.6
AD3					
HU Emily	2005	HU5	HU4	−42.4/10.7	5.5
HU Irene	2005	HU2	TS	−33.5/12.9	5.4
TS Ingrid	2007	TS	TS	−43.6/13.0	5.4
TS Nana	2008	TS	TS	−36.6/15.5	6.0
HU Lisa	2010	HU1	HU1	−32.0/15.2	7.0
TS Oscar	2012	TS	TS	−38.0/12.4	6.4
TS Chantal	2013	TS	TS	−41.4/9.3	5.4
TS Dorian	2013	TS	TS	−20.0/11.3	4.6

TABLE 1. (Continued)

Storm name	Year	Storm category	Max storm category in the first five days	Initial lon/lat	5-day mean averaging radius (° lat)
HU Edouard	2014	HU3	HU2	−33.2/13.7	6.1
TS Ida	2015	TS	TS	−24.2/9.4	5.6
TS Karl	2016	TS	TS	−19.0/13.1	5.6
TS Lisa	2016	TS	TS	−26.7/13.2	6.6
HU Irma	2017	HU5	HU3	−26.9/16.1	6.4

subsequently doubled (also denoted as $2 \times \text{ROCI}$) for all applications within this study; $2 \times \text{ROCI}$ has been chosen to dynamically account for insufficient AOD data close to the storm center due to the storm’s cloud canopy, which generally extends to between ROCI and $2 \times \text{ROCI}$ distance. For all subsequent references and applications in the current study, ROCI values listed as undefined in the extended best track database are replaced by the database’s first-5-days average across the 56 storms examined (3.1°). In addition, to avoid including regions too far away to be relevant for the TC, ROCI is capped at 1 standard deviation above the mean (4.05°). After applying the above conditions and for the storm days examined in the current study, the 5-day storm average $2 \times \text{ROCI}$ radius used ranges from 4.5° to 7.9° latitude (Table 1), with an overall average value of 6.0° .

3. Methods

To test whether the presence of Saharan dust influences the intensification of TCs, we categorize storms in a style similar to that presented in Fig. 7 of DV2004. For every TC in our study and all storm days, we average AOD over the four quarters of a circle around the storm with $2 \times \text{ROCI}$ radius (Fig. 3). A circle mean is calculated over the quarter averages if the data cover at least 5% of the total circle area. If there is less than 5% data coverage, the storm average AOD for that day is removed

from the analysis. As can be seen in Fig. 3, the criterion is necessary as the SEVIRI AOD data are only available over cloud-free regions excluding the direct TC environment. The method is chosen to balance the need to minimize the inclusion of irrelevant data captured through an increased averaging radius and to minimize false representation by averaging over too little data. Additionally, it weighs each quarter of the circle where data are available equally during averaging to avoid significant local biases. In particular, the latter can become a problem if a simple average over the whole circle is used. Note, however, that in our approach, quadrants without any available data will still be omitted from the analysis if there is otherwise sufficient coverage ($>5\%$) from other quadrants. A sensitivity study investigating the dependence of our results on the choice of averaging radius and threshold is included in the online supplemental material.

Furthermore, we separate “low dust” and “dust laden” averages by a simple threshold concept, similar to DV2004. While DV2004 distinguish between “no dust/not dry” and “dusty/dry” air in the low-to-midlevels using GOES split window imagery, we identify the signature of the SAL by a threshold median AOD value of 0.33 from the probability density distribution of AOD in the boxed region in Fig. 2 during the time periods of consideration in the current study (Fig. 4). An AOD above 0.33 thus reflects that the storm encounters an environment that is more dusty than typical in this region. The time

FIG. 3. Schematic showing the averaging method of AOD around a TC for the example of Hurricane Igor (2010) on 8 Sep 2010. The circle indicates a region with radius $2 \times \text{ROCI}$ (6° in this instance) around the storm center. For more details, see the text.

FIG. 4. AOD over the TC genesis region (5° – 20° N, 15° – 45° W) shown in Fig. 2; probability density function of daily mean SEVIRI AOD data for the months of July–September of the years 2004–17.

series of storm intensity and storm-relative AOD are then examined for each storm, with a focus on the first five days of storm development. The TCs are then categorized accordingly as storms that interact with AOD in an expected manner according to DV2004 (the DV1, DV2, and DV3 categories), those that intensify to hurricane strength despite persistent presence of above-median AOD (AD1), those developing with no clear correlation to ambient AOD (AD2), and those that go through their lifetime in the absence of elevated AOD (AD3; Table 2). These categories therefore summarize the AOD-based aspect of our analysis for each storm.

In addition to investigating AOD, we evaluate three common criteria potentially important for impeding the intensification of TCs, which are lower-to-midtropospheric relative humidity (RH) at 700 hPa, deep-tropospheric wind shear, and SST. The first and last properties are obtained directly from the respective datasets as described in the data section (section 2). Deep-tropospheric vertical wind shear, on the other hand, is calculated following the approach of Fitzpatrick (1997): The shear is determined by taking the zonal and meridional wind differences between upper- and lower-tropospheric layers and computing the vector difference. The upper- and lower-tropospheric-layer winds are calculated as the arithmetic mean of the 300-, 250-, and 200-hPa and 900-, 850-, and 700-hPa levels, respectively. Note that, prior to 2007, ECMWF data are not available at 900 hPa, but only at 925 hPa. Hence, for this period, the lower-tropospheric wind means were calculated using the latter pressure level data. While aligning with the commonly used definition of deep vertical wind shear between 200 and 850 hPa (in studies of wind shear over the Atlantic relevant for TCs: e.g., Frank and Ritchie 1999, 2001; Aiyyer and Thorncroft 2006; Nolan and McGauley 2012), note that the SAL midlevel easterly jet has typical altitudes of around 600–700 hPa (Carlson and Prospero 1972). As such, the vertical wind shear calculated and shown herein will be a lower estimate of the maximum shear in the air column in the presence of the AEJ.

In section 4b, a composite study of the storms is presented. For the purpose of compositing, a coordinate transformation

TABLE 2. Storm categories according to Dunion and Velden (2004), and additional categories defined in this study.

Label	Description
<i>Storm categories, DV2004</i>	
DV1	First days adjacent to the SAL with minimal intensification, then moving into SAL-free environment and intensifying to hurricane
DV2	Storm consistently adjacent to the SAL, with no or only little intensification
DV3	Storm first in SAL-free environment; when moving into proximity of the SAL, weakening
<i>Additional storm categories</i>	
AD1	Storm intensifies and reaches hurricane strength in proximity to the SAL
AD1b	The subset of AD1 storms that reaches hurricane strength and does not permanently leave SAL-influenced regions within the first five days
AD2	Storms with no clear correlation between intensification and the SAL
AD3	Storms with no exposure to the SAL

to a storm-relative coordinate system has been carried out for each TC during the first five storm days recorded in the HURDAT database. For the sake of clarity, we restrict the number of graphs and the analysis to 1200 UTC data only. In this procedure, a TC center relative coordinate system with a horizontal resolution of $0.25^{\circ} \times 0.25^{\circ}$ is introduced: the storm-relative coordinate center is defined by the HURDAT geographical coordinates of the TCs. The ECMWF and OISST data fields, which have the same horizontal resolution as the new coordinates, are then shifted by the vector difference between the original coordinates and the storm center. In addition to the coordinate shift, the AOD data, which have a much higher horizontal resolution (cf. section 2), are averaged into the above mesh. As AOD data are sparse in cloudy regions close to the TCs, regions where observations are available from less than three storms are excluded from the composite. Storm category average fields are then computed for RH and deep wind shear at 1200 UTC. For SST and AOD, the daily mean values of the first five storm days in the HURDAT database are used.

4. Results

a. Relationship between TC intensity and AOD

DV2004 discussed three types of storms (Table 2a), which may be negatively affected by Saharan air at different stages of development: Storms that experience minimal intensification while initially or during the first days in the proximity of the SAL and then emerge from the dust-laden air and become hurricanes are classified as DV1 (DV2004; TCs Cindy 1999, Floyd 1999, Erin 2001, and Felix 2001). Category DV2 comprises storms that stay in the vicinity of dusty air and do not or only slightly intensify (DV2004; TCs Debby 2000 and Chantal 2001). This category includes TCs that remain below hurricane

strength during the whole record and those developing into category-1 hurricanes later in their lifetimes. DV2004 find another case, where the storm first intensifies to hurricane strength in SAL-free air masses and then weakens when encountering dust-laden air (DV2004; TC Joyce 2000), which we denote as DV3. These cases are identified in our study when a clear correspondence of storm weakening and the mean AOD changing from below to above the median AOD of 0.33 is observed in its lifetime. These three categories encompass the cases where the SAL appears to have an impeding effect on storm intensification. Note that, although the DV2004 definitions for these categories are retained, the method by which the criteria are analyzed differs in our study. DV2004 used GOES split window imagery to detect a combination of dry and dusty air within two degrees of the TC center, while this study uses AOD within $2 \times \text{ROCI}$ to detect the SAL (i.e., we include AOD values at greater distances from the storm center).

For the period 2004–17, the above classification describes only 28% of the storms that encounter the SAL in our sample [calculated as $(\text{DV1} + \text{DV2} + \text{DV3})/(\text{DV1} + \text{DV2} + \text{DV3} + \text{AD1} + \text{AD2})$; storm categorization is shown in Table 1, and categories are summarized in Table 2]. This is only 21% of all the TCs we examined $[(\text{DV1} + \text{DV2} + \text{DV3})/\text{all cases}]$, which further include cases without exposure to the SAL (AD3). For this reason, we introduce three additional storm categories to classify the remaining storms in the sample (Table 2): Category AD1 describes cases in which the storm intensifies in the presence of the SAL, AD2 cases are those that encounter the SAL but develop with no clear correlation to the SAL, and AD3 TCs those that develop without the presence of the SAL. All storms examined in the current study are subsequently categorized individually according to the time series of AOD surrounding the storm, and a summary is provided in Table 1. Note, however, that the relationship between the proximity to high-AOD air and strength of TC (non-)intensification is not always clearly defined. Some subjective interpretations are applied during the categorization process, although best efforts are made to ensure that they are most closely in line with the chosen category definition.

While we do not claim this additional categorization to be complete, we can clearly show that TC development is more complex than indicated in DV2004, a concern also raised by Braun (2010). We find only 11 of 52 cases (21%) that fit the DV1 (2), DV2 (4), and DV3 (5) criteria (Table 1). In these cases, it appears that the SAL plays a role in either impeding or delaying the intensification of Atlantic TCs. However, we also find 11 AD1 cases (21% of the whole sample) where the presence of dust around the storm does not prevent the TC from intensifying or even undergoing rapid intensification in some cases (summarized in Fig. 5a). In the case of AD1 storms, they all intensify at first within the proximity of the SAL, as indicated by high values of AOD. Additionally, Fig. 5b shows 17 AD2 cases (33% of the whole sample) where the SAL is only marginally involved, and no consistent relationship can be noted between the SAL and TC intensification. This is often exhibited as sporadic exposures to above-median AOD air for short durations during intensification, prior to intensification, or in storms that do not intensify without any clear impact

(positive or negative) of the SAL. In some cases, the TCs strengthen slightly in a dusty environment but do not reach hurricane strength before moving to a low-AOD surrounding (Fig. 5b). There are also cases hardly affected by Saharan dust that show a nonintensification similar to those surrounded by the SAL (a subset of AD3 storms). A high average AOD around the storm thus appears to be neither a sufficient nor a necessary condition for the nonintensification of TCs in the Atlantic.

To investigate possible correlations between the presence and strength of the SAL and storm intensity, the 52 storms are plotted contrasting the maximum 24-h intensity change in the first 5 days in the HURDAT record against the mean AOD over those days until the maximum intensity in the first 5 days is reached (Fig. 6a). Spearman's ranked correlation analysis yielded no statistically significant correlation between these properties (p value of 0.95). Focusing on only the storms that encounter the SAL (i.e., excluding AD3 cases), we find a weak negative correlation (r_s of -0.26 ; p value of 0.11). The increase in statistical significance is mainly driven by the exclusion of a large number of low-AOD nonintensifying AD3 cases that are determined by other environmental factors. Despite this, a p value of 0.11 is insufficient to reject the null hypothesis of the correlation being purely due to chance. Furthermore, the R^2 statistic for the reduced sample excluding AD2 cases indicates that only 5% of the variance in maximum intensification can be explained by the mean AOD. When the AOD is averaged across a larger radius (as summarized in the online supplemental material), a more statistically significant correlation is found. For instance, a Spearman's correlation of -0.42 for the reduced sample is found to be significant at the 99% level when AOD excluding AD3 cases is averaged over 3 times the ROCI (Fig. 6b). The explained variance also increases to 15%. This increase in statistical significance with increasing averaging radius may be indicative of noise introduced by averaging across insufficient data in the standard setup. However, given the large averaging radius in the $3 \times \text{ROCI}$ case, the identified correlation may also be capturing the impact of other large-scale environmental characteristics that occur concurrently with the overall high AOD in the region (possibly also in association with the SAL), but does not necessarily reflect close proximity of high-AOD/SAL air. Additional composite studies examining other environmental factors are therefore discussed in the next section.

b. Composite study

In this section, we aim to elucidate the difference between storms that do not intensify in the presence of high AOD (DV2 cases) and those that continue to develop despite the high AOD (AD1 cases). For this purpose, we create storm-centered composites of RH, vertical wind shear, AOD, and SST. To contrast DV2 cases against storms that clearly intensify in the presence of the SAL within the time period examined, the composites are produced using a subset of AD1 cases called AD1b (Table 2b; Fig. 5b) that excludes TCs that move permanently into SAL-free regions within the first two days (Hurricanes Danielle 2004 and Fred 2015) or those that do not reach hurricane strength before the sixth day (Hurricanes Florence 2006 and Earl 2010).

FIG. 5. Time series of HURDAT intensity for the (a) AD1, (b) AD2, and (c) AD3 categories. Colors indicate the average AOD surrounding the storm, where yellow, orange, and red refer to AODs higher than the median, 68th, and 85th percentile, respectively, within the TC genesis region considered (see Fig. 4). Blue lines indicate periods during which the storms are surrounded by relatively low dust burdens. Dotted black lines indicate insufficient AOD information. The thin dotted horizontal lines indicate the 33 m s^{-1} wind threshold that needs to be exceeded for category-1 storms according to the Saffir–Simpson hurricane scale.

1) AOD

In the first set of composites, RH is shown in combination with AOD for the first five days (Figs. 7 and 8). We find that, relative to AD1b cases, DV2 storms on average appear to be subject to a higher-AOD environment for a longer period of time. Over time in the first five days, the overall storm relative AOD decreases in the AD1b composite while the DV2 composite storm remains surrounded by a relatively high AOD environment even until day 5. However, no consistent pattern in statistical significance can be found in the difference (Fig. S2 in the online supplemental material). Thus, from a pure AOD standpoint, there may only be weak evidence to postulate that a very high dust environment prevents the intensification of tropical storms into a hurricane.

2) RH

RH immediately in the storm center region does not differ strongly between DV2 (Fig. 7) and AD1b (Fig. 8) cases in the first two days, but a clear and statistically significant (Fig. S3 in

the online supplemental material) drying of the storm area by roughly 10% RH can be noted in DV2 cases by day 3 while AD1b cases retain a high RH around the storm. Upon inspection of individual storm cases (not shown), we attribute this drying of the storm center in the DV2 composite to both the weakening of individual storms and asymmetric displacements of the moisture peaks from a concentric circle around the TC center, which can result as the TC weakens and loses its axisymmetric shape likely due to the presence of vertical wind shear. When averaged across composite members, this latter aspect results in lower RH in the storm center in the composite. This nonsymmetric flow of humid air around the storm is also indicative of DV2 TCs becoming less organized with time than AD1b storms, which continue to intensify.

Farther away from the storm center, around or beyond 5° relative latitude and longitude, we note that the high AOD to the north of the DV2 storm composite is accompanied by drier air masses closer to the storm on the first day when compared to the AD1b composite (Figs. 7a and 8a). This difference is largely not

FIG. 6. Maximum 24-h intensity change ($\text{m s}^{-1} \text{day}^{-1}$) during the first five storm days shown against the mean AOD until the maximum intensity within the first five storm days is reached. Symbols denote the category as listed in Table 2, the linear best fit excluding AD3 cases is drawn in solid gray, and the dotted vertical line indicates the regional median AOD (0.33; see Fig. 4). Spearman's ranked correlation coefficient r_s and p value are noted in the upper-right corners of (a) and (b) for the whole sample of storms and excluding AD3 cases. The average AOD is calculated within a radius of (a) $2 \times \text{ROCI}$ and (b) $3 \times \text{ROCI}$. Note that the storm categorization differs slightly between the two averaging choices, as discussed in more detail in the online supplemental material.

statistically significant, however (supplemental Fig. S3a). Instead, we note statistically significant lower-RH air in the DV2 cases relative to the AD1b cases wrapping from the east-northeast of the storm on the first two days to the north and northwest of the storm on days 3 and 4 (Figs. 7 and 8 and supplemental Fig. S3). Despite the individuality of the different cases and a noisy composite due to the limited sample size, we find drying of air masses around the entire storm in the DV2 storm composite (Figs. 7d,e). In contrast, AD1b storms tend to retain a steadier supply of humid air to the southeast during the later days of storm development (Figs. 8c–e). From such a composite approach, however, it is not

possible to discern whether this lack of connection to a supply of humid air is a consequence or a cause of storm weakening.

Given that dry air masses are present at some distance from the storm center, we postulate that one of the mechanisms leading to storm nonintensification in the DV2 cases may be due to the entrainment of dry air with the descending air mass into the storm peripheral boundary layer, reducing the energy generated by the Carnot cycle that fuels the storm (Riemer et al. 2010). Alternatively, horizontal transport of dry air toward the storm center above the boundary layer (Montgomery and Smith 2017; Houze 2014) may also be responsible for suppressing the storm intensification.

While the SAL can be a cause for the observed drying, it is important to note that dry air advection can occur both in association with Saharan air masses and without (Zhang and Pennington 2004; Huang et al. 2010). In the latter cases dry air is usually caused by subtropical dry air outbreaks from north of the SAL and/or as a result of subsidence (e.g., Braun et al. 2013; Fritz and Wang 2013). Similarly, a high dust load can also be present without the coexistence of dry air. This calls for a more thorough investigation on a case-by-case basis while additionally considering the vertical dimension, but this is beyond the scope of this paper.

3) SST

SST composites (Figs. 9 and 10) exhibit an expected gradient of warmer temperatures to the south-southwest of the storm and cooler to the north-northeast. Within a 5° radius from the storm center, both DV2 and AD1b categories yield average daily SSTs higher than 26.5°C in the first five storm days, indicating that the SST is generally not an inhibiting factor to storm intensification in these cases (McTaggart-Cowan et al. 2015). Aside from slightly warmer waters south of the storm center on the first two days and slightly colder waters directly to the west-northwest and northeast of the storm center on day 4 in the DV2 compared to the AD1b composite, we do not find any significant differences between the DV2 and AD1b composites. As such, SST is not expected to influence the analysis of the SAL impact on TC intensification.

4) VERTICAL WIND SHEAR

When examining the composite of vertical wind shear surrounding the storm in the first five days, we find statistically significantly higher vertical wind shear on the northern side closer to the storm center in the DV2 cases in the first day, whereas the AD1b storm composite shows higher vertical wind shear on the southern side (Figs. 9a and 10a; see also Fig. S5a in the online supplemental material). Day 3 exhibits a transitional stage where lower wind shears are observed more than 5° away from the storm center while higher shear persists directly adjacent to the storm in the DV2 composite (Figs. 9c and 10c and supplemental Fig. S5c). By days 4 and 5, the pattern is reversed, with AD1b storms experiencing higher shear to the north while shear is higher to the south of the storm for DV2 cases (Figs. 9d and 10d; see also supplemental Fig. S5d). While slightly stronger lower-level easterly winds can be noted in the DV2 composite on the first two days (not shown), the difference between the categories mainly originates from differences in their upper-level

FIG. 7. Storm-centered composite of 700-hPa RH at 1200 UTC and daily mean AOD for the first five days of DV2 storms. Colored contour lines in the foreground indicate the RH (%), with dry regions denoted by dark blue ($\leq 50\%$ RH) and black ($\leq 40\%$ RH) hatchings. High-RH regions with values equal to or above 70% and 80% are shaded light and bright blue, respectively. AOD is indicated in colored shading according to the upper color bar for regions where at least three storms contain data for the composite average. For reference, circles are drawn around the composite center using the composite-average $2 \times \text{ROCI}$ for each day. The included table lists the daily composite-average storm properties.

wind fields, with stronger southerly outflow feeding into a southwesterly jet streak to the north/northeast of DV2 storms that is initially absent in the AD1b composite (Figs. 11a,d and Fig. S6a in the online supplemental material). Conversely, stronger upper-level winds are found to the south of storm center in the AD1b composite, especially on the second day (Figs. 11b,e and supplemental Fig. S6b). As drier air masses and lower SSTs are found to the north of the storm (as discussed above), the

presence of vertical wind shear in combination with this environment can have a more detrimental impact on TC intensification. This is in agreement with findings by Tao and Zhang (2014), in which they noted a weakened impact of shear in a higher-SST and moister environment relative to a lower-SST and drier environment. Vertical wind shear to the north of the storm in the AD1b composite approaches the storm center only after three days (Fig. 10). This allows the TC time to develop into a

FIG. 8. As in Fig. 7, but for AD1b storms.

stronger storm before eventually weakening due to the presence of wind shear.

c. Storm paths

Last, the geographic locations of DV2 (Fig. 12a) and AD1b (Fig. 12b) storms are compared. Focusing on the first five storm days analyzed in the composites, we note a large spread of tracks in the AD1b cases resulting from the range of synoptic situations faced by the individual storms. The DV2 tracks place well within the geographical distribution of the AD1b storms (Fig. 12b). Although the composite daily average storm center latitude and longitude are slightly more northward and westward for the DV2 category (Figs. 7 and 8), the differences

are not statistically significant. As the definition of the DV2 category stipulates nonintensifying storms, we find three cases of relatively short lived TCs in this group. Strikingly, high-AOD environments can be noted around Tropical Storm Fiona's (2010) far-reaching but nonintensifying path, with dusty air found over the western Atlantic Ocean close to the Caribbean Sea. This is indicative of strong dust events having taken place during Fiona's journey across the Atlantic.

5. Conclusions

In this study, we evaluated 52 named Atlantic Ocean tropical storms and hurricanes that originated from south of the

FIG. 9. Storm-centered composite of deep-layer vertical wind shear between approximately 850 and 250 hPa at 1200 UTC (m s^{-1} ; see section 3 for the exact definition) and daily mean SST ($^{\circ}\text{C}$) for the first five days, for DV2 storms. Contours indicate the vertical wind shear in increments of 5 m s^{-1} , and SST is color shaded. Composite-average storm properties for each day are provided in the table.

Cape Verde islands over the period 2004–17 to investigate whether and, if so, under what conditions the SAL delays or impedes their intensification.

DV2004 categorized TCs by relating storm intensity and proximity to the SAL’s dry and/or dusty air for sample storms where the SAL had a negative influence on these TCs. We adopted their categorization to an AOD perspective and extended it to describe the whole sample of storms formed over the eastern and central Atlantic in the 2004–17 period based on the time series of storm intensity and AOD surrounding the

storm. The interactions of a TC with the SAL during storm development were found to be more complex than those explored by DV2004. In particular, only 28% of the TCs that encounter high-AOD environments comply with their original categorization, showing a negative influence of high-AOD environment on storm intensification (DV1, DV2, and DV3 cases; 21% of the whole sample of TCs). The same number of TCs show an opposite relation, in which the storms intensified despite their proximity to a high-AOD environment (AD1 cases; 21% of all storms). The remaining 44% of the storms

FIG. 10. As in Fig. 9, but for AD1b storms.

that encounter the SAL are only sporadically exposed to dusty air masses or do not show any relationship between the presence of the SAL and storm intensification (AD2 cases; 33% of all storms). Notably, TCs that encounter the SAL account for 75% of all the TCs examined. We find 13 of 52 TCs originating from the region of the Atlantic influenced by African easterly waves between 2004 and 2017 to not encounter any high-AOD environment in their lifetime (AD3 cases). Note also that the limited number of storms represented by the DV2004 categorization could have contributed to the difference between their findings and those by Braun (2010), who examined the entire population of storms developed in association with the SAL.

The role of ambient AOD on storm intensification was examined through two perspectives in this study. First, a correlation analysis was performed between the average AOD around the storm until the maximum intensity within the first five storm days was reached and the storm's maximum 24-h intensity change within the first five days. No correlation was found between AOD and the storm intensity properties when analyzed across all cases. However, when only TCs that encountered the SAL are considered (all except AD3 cases), we find a weak negative correlation that increases in statistical significance with increasing averaging radius. This indicates that if a TC encounters the SAL during intensification, there is an increased likelihood for it to intensify more strongly if the

FIG. 11. (a)–(c) DV2 and (d)–(f) AD1b storm-centered composite of the 200-hPa wind field on days (top) 1, (middle) 2, and (bottom) 3 at 1200 UTC.

ambient AOD, even at a distance, is lower. However, given the significant overlap in the range of average AODs for intensifying versus nonintensifying TCs, an above average AOD alone is not a good predictor for storm nonintensification. Second, storm-centered composites were created to elucidate the differences between TCs that intensify to hurricane strength in the proximity of high-AOD air within the first five days (AD1b category storms) and those that do not (DV2 category storms). While slightly higher composite AOD is found surrounding the storm for a longer period of time in the DV2 nonintensifying cases, this is largely not statistically

significant and clearly distinguishable differences cannot be found when comparing individual storms. Overall, we find some evidence of weaker TC intensification with higher ambient AOD among cases that encounter the SAL, but a high AOD alone is not a suitable proxy for identifying storm nonintensification. Note that our AOD-based analyses are strongly limited by the scarce availability of data close to the TC due to satellite signal contamination by clouds. However, given the spatial extent of the SAL, its qualitative presence or absence in the proximity of the TC can generally be captured despite the limited observations, as evidenced by the relative insensitivity

FIG. 12. Geographical paths of (a) DV2 and (b) AD1b storms over the Atlantic Ocean. The line style indicates the TC strength, where dotted lines indicate a low pressure system or a tropical depression, dashed lines indicate tropical storms, and solid lines denote hurricanes or extratropical storms. Colors indicate the mean AOD around the storm, with black denoting insufficient AOD data. Black dots along the storm paths indicate the position of the TC in the first five 1200 UTC records in the HURDAT dataset. In addition, DV2 tracks are overlaid as solid gray lines in (b).

of TC categorization to the choice of averaging radius as discussed in the online supplemental material. The exact value of the AOD, however, may not be well represented (also notable from the sensitivity of storm-average AOD to the choice of averaging radius), and we are unable to conclude if there is any direct interaction of the high-AOD air with the TC near the storm center. This is particularly relevant when comparing to DV2004, where the SAL influence is examined within a much closer 2° distance.

To investigate the role of other environmental factors accompanying the high AOD in the SAL, the composite study also examined differences in the moisture, vertical wind shear, and SST distributions between intensifying and nonintensifying storms in the proximity to the SAL. While AD1b storms,

intensifying despite the high AOD, retain a connection to moist air masses to their south/southeast over time, the DV2 composite shows initially drier air to the northeast and overall drying of the entire storm region over time as the TC weakens. Higher wind shear can also be noted with statistical significance closer to the storm on the northern side over colder waters and in drier air masses in the first two days of TC development in the DV2 composite. This results mainly from differences in the upper-level wind field between the categories instead of differences in the lower-level winds associated with the SAL. While a strong outflow into a jet streak to the north of the storm is found in the upper troposphere in the DV2 case, this feature is not present in the AD1b composite until the third day. With regard to the SST, we find that all TCs are exposed to similarly favorable

conditions. Additionally, an examination of the geographical location of the DV2 and AD1b storms shows the former cases lie within the wider spread of tracks of the AD1b storms. Thus, a bias in the storm tracks is not a factor distinguishing the two groups.

Overall, we find that the early and close presence of sheared air to the north of the storm and dry air from the northeast may cause a TC to not intensify in proximity to the SAL. As these factors may also act together independently of the SAL, further research is necessary to isolate the exact role of the SAL on TC intensification. This is in line with conclusions by Braun (2010), who noted that other factors (such as large-scale subsidence) may be behind the unfavorable environmental conditions (such as dryness) otherwise associated with the SAL. Additional factors not examined in the current study include, in particular, changes in the microphysical structure of the TCs induced by dust particles and their influence on cloud-dynamical processes, the vertical dependence of dry air advection, and directional wind shear. Furthermore, we approached the analysis in this study by selecting SAL-influenced cases based on the ambient average AOD, which has limited data availability close to the storm. It is also important to note that a high-AOD environment does not necessarily imply the presence of the SAL and vice versa. Clearly better aerosol observations under cloudy conditions are needed to disentangle this problem and to reach more robust results of the influence of the SAL on TC intensification.

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REFERENCES

- Aiyyer, A. R., and C. Thorncroft, 2006: Climatology of vertical wind shear over the tropical Atlantic. *J. Climate*, **19**, 2969–2983, <https://doi.org/10.1175/JCLI3685.1>.
- Braun, S. A., 2010: Reevaluating the role of the Saharan air layer in Atlantic tropical cyclogenesis and evolution. *Mon. Wea. Rev.*, **138**, 2007–2037, <https://doi.org/10.1175/2009MWR3135.1>.
- , J. A. Sippel, C.-L. Shie, and R. A. Boller, 2013: The evolution and role of the Saharan air layer during Hurricane Helene (2006). *Mon. Wea. Rev.*, **141**, 4269–4295, <https://doi.org/10.1175/MWR-D-13-00045.1>.
- Bréon, F.-M., A. Vermeulen, and J. Descloîtres, 2011: An evaluation of satellite aerosol products against sunphotometer measurements. *Remote Sens. Environ.*, **115**, 3102–3111, <https://doi.org/10.1016/j.rse.2011.06.017>.
- Burpee, R. W., 1972: The origin and structure of easterly waves in the lower troposphere of North Africa. *J. Atmos. Sci.*, **29**, 77–90, [https://doi.org/10.1175/1520-0469\(1972\)029<0077:TOASOE>2.0.CO;2](https://doi.org/10.1175/1520-0469(1972)029<0077:TOASOE>2.0.CO;2).
- Carlson, T. N., and J. M. Prospero, 1972: The large-scale movement of Saharan air outbreaks over the northern equatorial Atlantic. *J. Appl. Meteor.*, **11**, 283–297, [https://doi.org/10.1175/1520-0450\(1972\)011<0283:TLSMOS>2.0.CO;2](https://doi.org/10.1175/1520-0450(1972)011<0283:TLSMOS>2.0.CO;2).
- , and S. G. Benjamin, 1980: Radiative heating rates for Saharan dust. *J. Atmos. Sci.*, **37**, 193–213, [https://doi.org/10.1175/1520-0469\(1980\)037<0193:RHRFSD>2.0.CO;2](https://doi.org/10.1175/1520-0469(1980)037<0193:RHRFSD>2.0.CO;2).
- Chavas, D. R., and K. A. Emanuel, 2010: A QuikSCAT climatology of tropical cyclone size. *Geophys. Res. Lett.*, **37**, L18816, <https://doi.org/10.1029/2010GL044558>.
- Cook, K. H., 1999: Generation of the African easterly jet and its role in determining West African precipitation. *J. Climate*, **12**, 1165–1184, [https://doi.org/10.1175/1520-0442\(1999\)012<1165:GOTAEJ>2.0.CO;2](https://doi.org/10.1175/1520-0442(1999)012<1165:GOTAEJ>2.0.CO;2).
- Cotton, W., H. Zhang, G. M. McFarquhar, and S. M. Saleeby, 2007: Should we consider polluting hurricanes to reduce their intensity? *J. Wea. Modif.*, **39**, 70–73, <https://www.journalofweathermodification.org/index.php/JWM/article/view/204>.
- Davidi, A., A. B. Kostinski, I. Koren, and Y. Lehn, 2012: Observational bounds on atmospheric heating by aerosol absorption: Radiative signature of transatlantic dust. *Geophys. Res. Lett.*, **39**, L04803, <https://doi.org/10.1029/2011GL050358>.
- DeMott, P. J., K. Sassen, M. R. Poellot, D. Baumgardner, D. C. Rogers, S. D. Brooks, A. J. Prenni, and S. M. Kreidenweis, 2003: African dust aerosols as atmospheric ice nuclei. *Geophys. Res. Lett.*, **30**, 1732, <https://doi.org/10.1029/2003GL017410>.
- Demuth, J., M. DeMaria, and J. A. Knaff, 2006: Improvement of Advanced Microwave Sounder Unit tropical cyclone intensity and size estimation algorithms. *J. Appl. Meteor. Climatol.*, **45**, 1573–1581, <https://doi.org/10.1175/JAM2429.1>.
- Diaz, H., T. N. Carlson, and J. M. Prospero, 1976: A study of the structure and dynamics of the Saharan air layer over the equatorial Atlantic during BOMEX. NOAA Tech. Memo. ERL WMPO-32, 61 pp.
- Dunion, J. P., 2011: Rewriting the climatology of the tropical North Atlantic and Caribbean Sea atmosphere. *J. Climate*, **24**, 893–908, <https://doi.org/10.1175/2010JCLI3496.1>.
- , and C. S. Velden, 2004: The impact of the Saharan air layer on Atlantic tropical cyclone activity. *Bull. Amer. Meteor. Soc.*, **85**, 353–366, <https://doi.org/10.1175/BAMS-85-3-353>.
- , and C. S. Marron, 2008: A reexamination of the Jordan mean tropical sounding based on awareness of the Saharan air layer: Results from 2002. *J. Climate*, **21**, 5242–5253, <https://doi.org/10.1175/2008JCLI1868.1>.
- Emanuel, K., 2005: Increasing destructiveness of tropical cyclones over the past 30 years. *Nature*, **436**, 686–688, <https://doi.org/10.1038/nature03906>.
- Evan, A. T., J. P. Dunion, J. A. Foley, A. K. Heidinger, and C. S. Velden, 2006: New evidence for a relationship between Atlantic tropical cyclone activity and African dust outbreaks. *Geophys. Res. Lett.*, **33**, L19813, <https://doi.org/10.1029/2006GL026408>.
- Feldmann, M., K. Emanuel, L. Zhu, and U. Lohmann, 2019: Estimation of Atlantic tropical cyclone rainfall frequency in the United States. *J. Appl. Meteor. Climatol.*, **58**, 1853–1866, <https://doi.org/10.1175/JAMC-D-19-0011.1>.
- Fitzpatrick, A., 1997: Understanding and forecasting tropical cyclone intensity change with the Typhoon Intensity Prediction Scheme (TIPS). *Wea. Forecasting*, **12**, 826–846, [https://doi.org/10.1175/1520-0434\(1997\)012<0826:UAFICI>2.0.CO;2](https://doi.org/10.1175/1520-0434(1997)012<0826:UAFICI>2.0.CO;2).
- Frank, W. M., and E. A. Ritchie, 1999: Effects of environmental flow upon tropical cyclone structure. *Mon. Wea. Rev.*, **127**, 2044–2061, [https://doi.org/10.1175/1520-0493\(1999\)127<2044:EOEFUT>2.0.CO;2](https://doi.org/10.1175/1520-0493(1999)127<2044:EOEFUT>2.0.CO;2).
- , and —, 2001: Effects of vertical wind shear on the intensity and structure of numerically simulated hurricanes. *Mon. Wea. Rev.*, **129**, 2249–2269, [https://doi.org/10.1175/1520-0493\(2001\)129<2249:EOVWSO>2.0.CO;2](https://doi.org/10.1175/1520-0493(2001)129<2249:EOVWSO>2.0.CO;2).
- Fritz, C., and Z. Wang, 2013: A numerical study of the impacts of dry air on tropical cyclone formation: A development case

- and a nondevelopment case. *J. Atmos. Sci.*, **70**, 91–111, <https://doi.org/10.1175/JAS-D-12-018.1>.
- Goudie, A. S., and N. J. Middleton, 2001: Saharan dust storms: Nature and consequences. *Earth-Sci. Rev.*, **56**, 179–204, [https://doi.org/10.1016/S0012-8252\(01\)00067-8](https://doi.org/10.1016/S0012-8252(01)00067-8).
- Gray, W. M., 1968: Global view of the origin of tropical disturbances and storms. *Mon. Wea. Rev.*, **96**, 669–700, [https://doi.org/10.1175/1520-0493\(1968\)096<0669:GVOTOO>2.0.CO;2](https://doi.org/10.1175/1520-0493(1968)096<0669:GVOTOO>2.0.CO;2).
- Grist, J. P., 2002: Easterly waves over Africa. Part I: The seasonal cycle and contrasts between wet and dry years. *Mon. Wea. Rev.*, **130**, 197–211, [https://doi.org/10.1175/1520-0493\(2002\)130<0197:EWOAPI>2.0.CO;2](https://doi.org/10.1175/1520-0493(2002)130<0197:EWOAPI>2.0.CO;2).
- Herbener, S. R., S. C. van den Heever, G. G. Carrió, S. M. Saleeby, and W. R. Cotton, 2014: Aerosol indirect effects on idealized tropical cyclone dynamics. *J. Atmos. Sci.*, **71**, 2040–2055, <https://doi.org/10.1175/JAS-D-13-0202.1>.
- Houze, R. A., Jr., 2014: *Cloud Dynamics*. 2nd ed. Elsevier/Academic Press, 432 pp.
- Huang, J., C. Zhang, and J. M. Prospero, 2010: African dust outbreaks: A satellite perspective of temporal and spatial variability over the tropical Atlantic Ocean. *J. Geophys. Res.*, **115**, D0502, <https://doi.org/10.1029/2009JD012516>.
- Jenkins, G. S., A. S. Pratt, and A. Heymsfield, 2008: Possible linkages between Saharan dust and tropical cyclone rain band invigoration in the eastern Atlantic during NAMMA-06. *Geophys. Res. Lett.*, **35**, L08815, <https://doi.org/10.1029/2008GL034072>.
- Karyampudi, V. M., and T. N. Carlson, 1988: Analysis and numerical simulations of the Saharan air layer and its effect on easterly wave disturbances. *J. Atmos. Sci.*, **45**, 3102–3136, [https://doi.org/10.1175/1520-0469\(1988\)045<3102:AANSOT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1988)045<3102:AANSOT>2.0.CO;2).
- Khain, A., and B. Lynn, 2011: Simulation of tropical cyclones using a mesoscale model with spectral bin microphysics. *Recent Hurricane Research—Climate, Dynamics, and Societal Impacts*, A. R. Lupo, Ed., InTech, 197–227.
- , N. Cohen, B. Lynn, and A. Pokrovsky, 2008: Possible aerosol effects on lightning activity and structure of hurricanes. *J. Atmos. Sci.*, **65**, 3652–3677, <https://doi.org/10.1175/2008JAS2678.1>.
- Kimball, S. K., and M. S. Mulekar, 2004: A 15-year climatology of North Atlantic tropical cyclones. Part I: Size parameters. *J. Climate*, **17**, 3555–3575, [https://doi.org/10.1175/1520-0442\(2004\)017<3555:AYCONA>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<3555:AYCONA>2.0.CO;2).
- Knaff, J. A., S. P. Longmore, and D. A. Molenar, 2014: An objective satellite-based tropical cyclone size climatology. *J. Climate*, **27**, 455–476, <https://doi.org/10.1175/JCLI-D-13-00096.1>.
- Laken, B. A., H. Parviainen, E. Pallé, and T. Shahbaz, 2013: Saharan mineral dust outbreaks observed over the North Atlantic island of La Palma in summertime between 1984 and 2012. *Quart. J. Roy. Meteor. Soc.*, **140**, 1058–1068, <https://doi.org/10.1002/qj.2170>.
- Lau, W. K. M., and K.-M. Kim, 2007: How nature foiled the 2006 hurricane forecasts. *Eos, Trans. Amer. Geophys. Union*, **88**, 105–107, <https://doi.org/10.1029/2007EO090002>.
- Levin, Z., E. Ganor, and V. Gladstein, 1996: The effects of desert particles coated with sulfate on rain formation in the eastern Mediterranean. *J. Appl. Meteor.*, **35**, 1511–1523, [https://doi.org/10.1175/1520-0450\(1996\)035<1511:TEODPC>2.0.CO;2](https://doi.org/10.1175/1520-0450(1996)035<1511:TEODPC>2.0.CO;2).
- McTaggart-Cowan, R., E. L. Davies, J. G. Fairman, T. J. Galarneau, and D. M. Schultz, 2015: Revisiting the 26.5°C sea surface temperature threshold for tropical cyclone development. *Bull. Amer. Meteor. Soc.*, **96**, 1929–1943, <https://doi.org/10.1175/BAMS-D-13-00254.1>.
- Merrill, R. T., 1984: A comparison of large and small tropical cyclones. *Mon. Wea. Rev.*, **112**, 1408–1418, [https://doi.org/10.1175/1520-0493\(1984\)112<1408:ACOLAS>2.0.CO;2](https://doi.org/10.1175/1520-0493(1984)112<1408:ACOLAS>2.0.CO;2).
- Montgomery, M. T., and R. K. Smith, 2017: Recent developments in the fluid dynamics of tropical cyclones. *Annu. Rev. Fluid Mech.*, **49**, 541–574, <https://doi.org/10.1146/annurev-fluid-010816-060022>.
- Nolan, D. S., and M. G. McGauley, 2012: Tropical cyclogenesis in wind shear: Climatological relationships and physical processes. *Cyclones: Formation, Triggers and Control*, Nova Science Publishers, 1–36.
- Palmén, E., 1948: On the formation and structure of tropical cyclones. *Geophysica*, **3**, 26–38.
- Pielke, R., J. Gratz, C. Landsea, D. Collins, M. Saunders, and R. Musulin, 2008: Normalized hurricane damage in the United States: 1900–2005. *Nat. Hazards Rev.*, **9**, 29–42, [https://doi.org/10.1061/\(ASCE\)1527-6988\(2008\)9:1\(29\)](https://doi.org/10.1061/(ASCE)1527-6988(2008)9:1(29)).
- Prospero, J. M., P. Ginoux, O. Torres, S. E. Nicholson, and T. E. Gill, 2002: Environmental characterization of global sources of atmospheric soil dust identified with the Nimbus 7 Total Ozone Mapping Spectrometer (TOMS) absorbing aerosol product. *Rev. Geophys.*, **40**, 1002, <https://doi.org/10.1029/2000RG000095>.
- Reed, K. A., J. T. Bacmeister, J. J. A. Huff, X. Wu, S. C. Bates, and N. A. Rosenbloom, 2019: Exploring the impact of dust on North Atlantic hurricanes in a high-resolution climate model. *Geophys. Res. Lett.*, **46**, 1105–1112, <https://doi.org/10.1029/2018GL080642>.
- Riehl, H., 1948: On the formation of typhoons. *J. Meteor.*, **5**, 247–265, [https://doi.org/10.1175/1520-0469\(1948\)005<0247:OTFOT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1948)005<0247:OTFOT>2.0.CO;2).
- Riemer, M., M. T. Montgomery, and M. E. Nicholls, 2010: A new paradigm for intensity modification of tropical cyclones: Thermodynamic impact of vertical wind shear on the inflow layer. *Atmos. Chem. Phys.*, **10**, 3163–3188, <https://doi.org/10.5194/acp-10-3163-2010>.
- Rosenfeld, D., M. Clavner, and R. Nirel, 2011: Pollution and dust aerosols modulating tropical cyclones intensities. *Atmos. Res.*, **102**, 66–76, <https://doi.org/10.1016/j.atmosres.2011.06.006>.
- , W. L. Woodley, A. Khain, W. R. Cotton, G. Carrió, I. Ginis, and J. H. Golden, 2012: Aerosol effects on microstructure and intensity of tropical cyclones. *Bull. Amer. Meteor. Soc.*, **93**, 987–1001, <https://doi.org/10.1175/BAMS-D-11-00147.1>.
- Schenkel, B. A., N. Lin, D. Chavas, M. Oppenheimer, and A. Brammer, 2017: Evaluating outer tropical cyclone size in reanalysis datasets using QuikSCAT data. *J. Climate*, **30**, 8745–8762, <https://doi.org/10.1175/JCLI-D-17-0122.1>.
- Shu, S., and L. Wu, 2009: Analysis of the influence of Saharan air layer on tropical cyclone intensity using AIRS/Aqua data. *Geophys. Res. Lett.*, **36**, L09809, <https://doi.org/10.1029/2009GL037634>.
- Sun, D., K. M. Lau, and M. Kafatos, 2008: Contrasting the 2007 and 2005 hurricane seasons: Evidence of possible impacts of Saharan dry air and dust on tropical cyclone activity in the Atlantic basin. *Geophys. Res. Lett.*, **35**, L15405, <https://doi.org/10.1029/2008GL034529>.
- Tao, D., and F. Zhang, 2014: Effect of environmental shear, sea-surface temperature, and ambient moisture on the formation and predictability of tropical cyclones: An ensemble-mean perspective. *J. Adv. Model. Earth Syst.*, **6**, 384–404, <https://doi.org/10.1002/2014MS000314>.
- Thieuleux, F., C. Moulin, F. M. Bréon, F. Maignan, J. Poitou, and D. Tanré, 2005: Remote sensing of aerosols over the oceans using MSG/SEVIRI imagery. *Ann. Geophys.*, **23**, 3561–3568, <https://doi.org/10.5194/angeo-23-3561-2005>.

- Thorncroft, C. D., and M. Blackburn, 1999: Maintenance of the African easterly jet. *Quart. J. Roy. Meteor. Soc.*, **125**, 763–786, <https://doi.org/10.1002/qj.49712555502>.
- Twohy, C. H., 2015: Measurements of Saharan dust in convective clouds over the tropical eastern Atlantic Ocean. *J. Atmos. Sci.*, **72**, 75–81, <https://doi.org/10.1175/JAS-D-14-0133.1>.
- Wu, L., 2007: Impact of Saharan air layer on hurricane peak intensity. *Geophys. Res. Lett.*, **34**, L09802, <https://doi.org/10.1029/2007GL029564>.
- Zhang, C., and J. Pennington, 2004: African dry air outbreaks. *J. Geophys. Res.*, **109**, D20108, <https://doi.org/10.1029/2003JD003978>.
- Zhang, H., G. M. McFarquhar, S. M. Saleeby, and W. R. Cotton, 2007: Impacts of Saharan dust as CCN on the evolution of an idealized tropical cyclone. *Geophys. Res. Lett.*, **34**, L14812, <https://doi.org/10.1029/2007GL029876>.
- , —, W. R. Cotton, and Y. Deng, 2009: Direct and indirect impacts of Saharan dust acting as cloud condensation nuclei on tropical cyclone eyewall development. *Geophys. Res. Lett.*, **36**, L06802, <https://doi.org/10.1029/2009GL037276>.