Properties and influences of the Saharan air layer on tropical cyclogenesis

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Submitted 17 December 2004

Abstract

This paper extends previous work on the mechanical effects of mineral dust particles on tropical cyclone activity. The goals of this analysis are to explain empirical observations of dust, easterly wave disturbances with regards to diagnostic investigations of dust outbreaks, environmental parameters conducive to tropical storm development, and the implied role of the Saharan air layer as a modulator of storm intensity. These various studies lay the groundwork and provide an enormous impetus for the decoupling of the three-way partnership of African dust, the Saharan air layer, and African wave disturbances. A suitable framework is developed indicating that each behavior in this triad is not mutually exclusive to another, but may, on the other hand, represent an interconnected part in a more general, larger, phenomenon. Joint interactions are advanced through middle-altitude platform research aircraft and numerical simulations from the North Atlantic and Caribbean. The nature and behavioral aspects of each component is developed in tandem with mathematical expressions as a companion to the observations.

1 Introduction

Terrestrial dust is ubiquitous, is transported to all corners of the Earth (Reid et. al. 2003), and greatly affects its local and global environment. An understanding of the nature and interactions of one behavior often provides information of various aspects of another. Sand is the most common feature of the desert environment covering as much as 20 percent of an arid surface, the rest comprising hard-pack gravel, boulders, mountains, and various soil substrates. Very arid to hyper-arid landscapes are primarily wrought by weathering, eolian, and fluvial processes (Singapore 2004). Desert "sandiness" is attributed to rock fragmentation as a result of cyclic expansion and contraction induced by large temperature fluctuations. Winds transport enormous plumes of dust and impel vast amounts of sand along the ground, all particles behaving as abrasive instruments that carve, facet, and polish rocks (Singapore 2004). Annually, over one billion metric tons of dust particles is elevated and transported from the Sahara to various points around the globe (Reid et. al. 2003). Of the one billion tons transported, 200 million metric tons is deposited over the North Atlantic Ocean and Caribbean, along the path of a hurricane. Also there is 40 to 60 percent less tropical cyclone

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activity in the North Atlantic ocean than elsewhere in the world (Karyampudi and Pierce 2002). The largest contributor of dust on the planet, the Sahara may be the linked to this veracity. Three decades of study suggests that a dust plume and a hurricane's demands for self-perpetuation oppose if not vehemently antagonize one another. With their domains so close in proximity and their initial conditions so dissimilar, an encounter has not only been predicted to occur, it has also been predicted to be sensational. Theoretical studies suggest intense destructive interference or at least partial strength attenuation as a result of dust befalling into a hurricane's eyewall. This idea underpins the general framework of this discussion and is divided as follows: in part 2) a treatment of dust and its various aspects, part 3) the Saharan air layer and its two-way partnership with African easterly waves, part 4) a brief account of tropical storm development, part 5) effects on tropical cyclone activity, and in part 6) a summary of limitations followed by concluding remarks.

2 Dust

a. Aerosol properties

To convey even the roughest reasonably comprehensive description of the aerosol we must cover the gamut from the smallest particle undergoing large Brownian motion movements to the largest particle condensing into the size of a cloud droplet. If we admit cloud, fog, rain and snow as part of the aerosol (which they certainly are, unless one is thinking only of the purported "dry aerosol") then the upper-most limit would be adjusted to 10 cm (large hail) (Twomey 1977, hereafer T77). The upper limit is tenuous and strictly arbitrary. Nevertheless, this rather broad criterion covers nine orders of magnitude and is thus quite unreasonable to the prevailing spatial range of Saharan dust. The most common size ranges from 10^{-4} μm to 1 mm and therefore we concentrate on this scale for the remainder of the study.

Neither mass nor radius describes a particle to any degree of completeness. Suspended particles in the earth's atmosphere vary greatly both in their chemical composition and the size distribution of individual constituents (Jennings 1992). This variation in their physical and chemical properties makes the task of numerical prediction particularly challenging. Even within four orders of magnitude, the physics and chemistry of aerosols is complicated and often counterintuitive. We restrict our discussion to the most egregious and non-contentious aspects within the above mentioned range in order to retain not only a definite focal-point to the analyses, but to also avoid the time-intensive task of conveying what may be too controversial or erroneous even. The radiative and optical effects of aerosols are beyond the scope of this article but we point the reader to two exemplary texts by Twomey[1] and Jennings[2].

It is clear that some geographic areas produce aerosols more than others. Dust storms are common features in arid and semi-arid areas because the surface is mostly covered by mobile, loose sediments (Warner 2004, hereafter W04). Dust storms demand a dry, granular surface substrate, as well as winds to initiate suspension.

[1] Twomey, S. (1977). *Atmospheric Aerosols*. Elsevier.

[2] Jennings, S.G. (1993). *Aerosol Effects on Climate*. University of Arizona Press.

West African dust is primarily composed of quarts ($Si0_2$ 49.34 percent), aluminum oxide (Al_2O_3 10.34 percent), sodium oxide (Na_2O 4.4 percent), and variable amounts of titanium oxide (TiO_3), magnesium oxide (MgO), calcium oxide (CaO), and diphosphorous hexaoxide (P_2O_6), to name a few.

There are two major meteorological factors important for dust elevation: vertical turbulent mixing in the surface layer and transfer of wind momentum to the ground surface (Nickovic 1996). Warner (2004) argued that air flowing over the granular surface causes a lower pressure on top. As the pressure under the grain is not affected, a net upward force results. The lower pressure above (due to the rushing air) causes the grains to be lifted into the adjacent airflow. If the cohesive forces due to molecular, electrostatic, and chemical bonding are overcome, or if the geometry or packing structure is conducive to removal, the particles will quiver and then lift into the air stream. Particles with greater diameter and density possess a higher wind threshold required for entrainment (Chepil 1945). However, particle sizes below 0.06 mm also possess a high threshold because of the greater electrostatic and molecular cohesive forces, retain moisture more effectively, and benefit from filling in the space among the larger particles (W04). The settling velocity is low and turbulence can mix particles upward throughout the boundary layer where they can be suspended for days. Residence time for such suspended dust can be long by convective-layer turbulence, even in the absence of significant large-scale winds. For larger particles, suspension occurs by various modes (not discussed here) highlighted by a horizontal momentum transferal from the original grain through exposure to the higher-velocity air stream. The grains descend back to the grain bed and impact other grains, ejecting others, and so on. Heathcote (1983) estimates that 90 percent of sand is transported in the lowest 50 cm of the atmosphere, Wiggs (1992) measured up to 35 percent of sand transport taking place in the lowest 2.5 cm, and Butterfield (1991) states that 80 percent of all transport occurs within 2 cm of the surface.

As explained by Gillette et. al. (1993)(hereafter G93), in almost every case, this mixing is connected with a heat flux so that strong heating of the ground is required. In the case of a heat flux from the ground to the atmosphere into which dust may mix, boundary layer height is determined by the strength of heat flux, the history of the heat flux, and the thermal structure of the atmosphere above the boundary layer. Let the height be denoted by z and the parameter L defined by Lumley and Panofsky (1964) as

$$L = -\frac{U^3 c_\rho \rho T}{KgH}, \tag{1}$$

where H is the heat flux (J m^{-2} s^{-1}), U the friction velocity equal to $\sqrt{\tau}/\rho$, τ the surface stress (N m^{-2}), ρ the density of air (kg m^{-3}), g the acceleration of gravity, c_ρ the specific capacity of air at constant pressure, T the temperature (K), and K the von Karmann constant (0.4). For the sake of illustrating meteorological conditions of dust production G93 specified a constant temperature and

$$\frac{z}{L} = A\,\frac{-zH}{U^3}, \tag{2}$$

for some constant A. Large values of the ratio occur for very large heat transfers from the surface, for large boundary layer heights, and for small values of friction velocity. When strong heating of the ground and near-surface air is added to a source of upper-level strong winds, a situation develops conducive to large-scale dust production and transport (G93). In this case, z/L begins with large values due to heating, high momentum is impelled to the surface creating intense large-scale wind erosion, mixing dust to great heights of the atmosphere, and large-scale transport potential (G93).

Dust storms often have a strong seasonal variation in their occurrence (Littmann 1991; Yu et. al. 1993) and functionally depend on impinging weather regimes, substrate moisture, vegetation cover, and prevalence of frozen ground (W04). Windy weather statistically elevates more particles than non-windy weather. Sustained wind more adequately keeps material already suspended in the atmosphere suspended longer than a discrete, periodic one. Though not entirely understood, wet soil surface tension associated with pore moisture binds the grains together making removal by mechanical means more difficult and energy intensive (W04). This explains why sand must be to a large degree moist to be useful in building a "sand-castle." A "wet" substrate has a higher entrainment threshold than a dry substrate. Vegetation cover reduces the mean surface speed by its roughness index increasing the turbulent mixing layer in the process. The corresponding desiccating effect wind has on a surface is staved off from the obstructive nature of protruding flora. Frozen ground will not be impacted by wind as long as its contents are effectively sealed-off from any ambient influences.

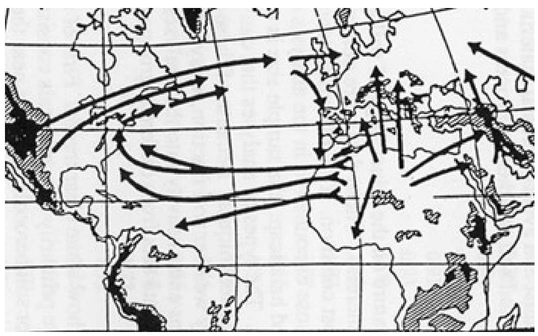

Figure 1: Probable trajectories of soil aerosol transport based on quartz isotope ratios. (Jennings 1993.)

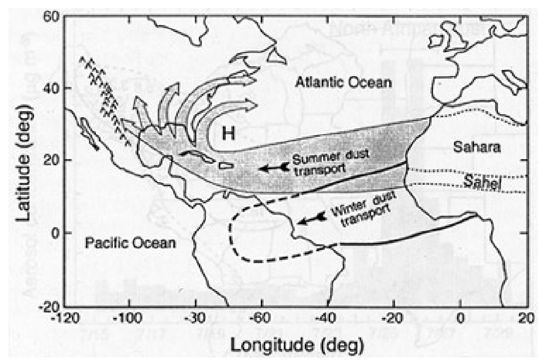

Figure 2: Dust transport over the North Atlantic from North Africa in the northern hemisphere summer and winter. (Warner 2004.)

The Saharan desert is bereft of soil moisture, is never frozen, and is absent of any vegetation that could impede wind motion. As a result, the critical energies and velocities required to elevate dust in the Sahara are exceptionally lower than in other environments.

3 Saharan air layer

a. Evolution

For the past three decades, the significant role Saharan aerosols play in modulating the atmospheric environment over the tropical Atlantic has become increasingly recognized and better understood (Rothman et. al. 2003). However, its link to tropical cyclone (TC) activity has never been fully explained (Dunion and Velden 2004, hereafter DV04). The SAL occurs primarily from late spring through early fall over extensive portions of the North

Atlantic Ocean between the Sahara desert, the Caribbean, and the United States (Carlson and Prospero 1972, hereafter CP72). The development of the SAL can be described as follows (Karyampudi and Carlson 1988, hereafter KC88): during summer months there is a flow of maritime air from the Mediterranean across North Africa. The air undergoes dramatic transformation in the lower troposphere as a result of strong sensible heating over the arid land mass. After several days of heating, the air is characterized by both a deep isentropic mixed-layer (potential temperatures between 42-46°C) and a slowly varying mixing ratio (2-6 g kg^{-1}). As the air traverses across the arid African continent, the intense heating produces vigorous dry convection throughout the layer of air in convective contact with the ground (CP72). Because of the vastness of the desert terrain and the intensity of the solar heating, a column of air entering the Sahara remains in convective contact with the ground for a period of at least a few days (acquiring the above mentioned uniform potential temperature in the mixing layer) (Carlson and Ludlam 1965). As it approaches the coast the base rises rapidly with distance to perhaps 900 m off Africa to perhaps 2 km off the Caribbean (Diaz 1976). The top of the SAL extends to and tapers-off at approximately 5.5 km. It is still not certain whether the rise of the SAL base occurs through erosion of the layer by cumulus convection and differential, radiative, cooling (KC88), or as a result of some other mechanism or combination of mechanisms.

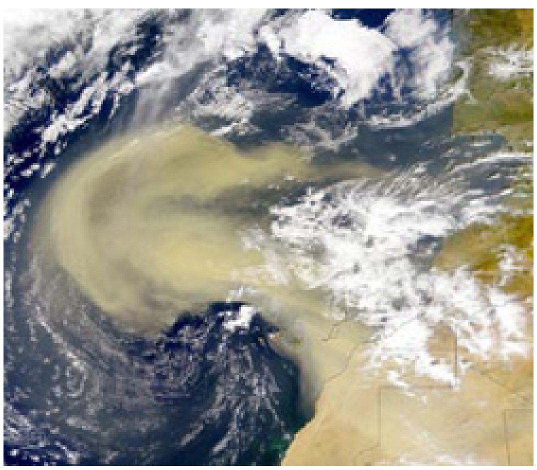

Figure 3: Extricated dust plume over west Africa. (NASA 2004.)

Once over the ocean, the SAL moves westward with an elongated plume subject to the vagaries of the Atlantic Ocean. The SAL appears to be internally self consistent by retaining its continental characteristics (treated below) as it traverses a path of 7000 km in approximately five to six days (7.7 m s^{-1}) to the West Indies and the United States. It is not uncommon for the SAL to be as large as the contiguous US. Thus the Sahara is the principal staging area of dust mobilization.

The SAL is hot (5-6° warmer than non-SAL air, DV04), dry (20-30 percent drier than ambient air, DV04), is environmentally stable (reinforcing a temperature inversion), touts a wind regime comparably faster than the local Trades (by 7-10 m s^{-1}, DV04), possesses a small mixing ratio, and retains a large optical depth. Low soil moisture is directly responsible for the production of a deep mixing layer (KC88). The SAL base is so dry that vertical ascent is not accompanied with cloud formation. The warmth of the air above the base is maintained by absorption of short-wave radiation by the suspended mineral dust (DV04). The daytime heating can transcend the overall long-wave cooling in the layer, further its base (Carlson and Benjamin 1980). Dust measurements indicate that dust concentration can vary by a factor of twenty-seven between SAL and non-SAL regimes and can be greatest at three kilometers than at the surface.

The SAL loses little internal energy to its environment between West Africa and the Caribbean. The SAL cools at approximately 0.7°C per day, or just over 4°C during its trans-oceanic passage to the Caribbean. And because it cools, it loses internal energy, and, as result, dust cascades out from the base onto the ocean surface. Consequently, the SAL top will sink 50-100 mb (1-2 mm s^{-1}), or 450-900 m between the African coast and Barbados. This is significant as some constituents can be up to several orders of magnitude more dense than dry air.

Aerosol studies conducted in Barbados over a four year period show that the amount of airborne dust reaching the island is highly variable, especially during the summer months when dust concentration is greatest. The average dust loading variance can be greater by one or two orders of magnitude in the summer than in the winter (Prospero 1968; Prospero et al. 1970). The onset of dust transport is often rather abrupt and corresponds to the inception of airflow from the western-Sahara. From August to October the average dust density content at Barbados is fairly high with exceptionally dusty periods becoming less prevalent as October tapers-off into November (CP72).

b. African Easterly Waves

African dust is transported along high amplitude, low frequency African easterly waves (AEWs) (Fig. 4). These wave disturbances tend to be the most intense in the vicinity of the coast (KC88). Local effects owing to surface conditions create a unique situation. Strong sensible heating generating a strong positive temperature gradient in the lower troposphere is responsible for the existence of a strong easterly jet core near 650 millibars (Holton 1992).

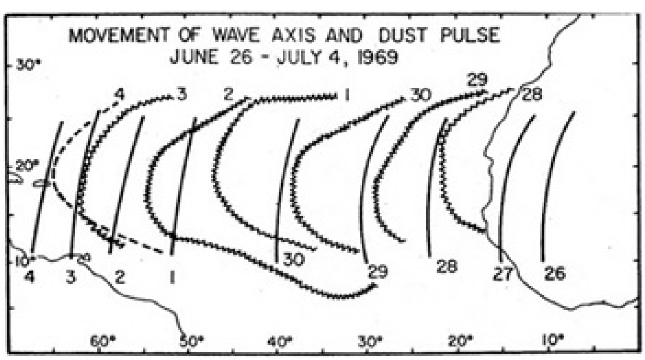

Figure 4: Dust movement 26 June-4 July, 1969. Serrated edges signify dust pulses (SAL), and solid lines wave axes (AEWs). (Carlson and Prospero 1972.)

Synoptical scale disturbances are observed to form and propagate westward in the cyclone shear zone to the south jet core. Carlson and Prospero (1972) showed that these dust pulses appear to be generated near the African coast by an interruption in the westward transport by the passage of traveling AEWs from equatorial Africa.

Numerical models run by Karyampudi and Carlson (1988) illustrate that surface heating over the Sahara is essential for the generation of the SAL and the development of the AEWs. When the frontal structure in the SAL was suppressed, a weaker midlevel easterly jet, weaker AEWs, reduced energy transformations, and a weaker to non-existent frontal structure resulted. The results confirmed that a weaker initial structure of the SAL appeared to produce noticeably weaker leading wave disturbances. To condense the findings the following important conclusions were obtained from the study:

- *Sensible heating over the Sahara desert is responsible for the SAL and the midlevel easterly jet.*
- *Widespread heavy dust within the SAL impedes convection on the equatorial side of the mid-level easterly jet from compensatory sinking forced by the rising motion within the SAL.*
- *The SAL is central for the maintenance and possible growth of some AEWs due to a higher combination of circulations and greater latent heat release in cumulus convection along the lateral boundary of the SAL.*
- *The presence of the SAL enhances the meridional temperature gradient and therefore the strength of the mid-level easterly jet.*

It is not suggested, however, that AEWs in reality produce the dust pulses by literally uplifting dust. On the other hand, it is likely that the alternate slackening and strengthening of the pressure gradient over Africa caused the by the cyclic passage of the AEWs contributes to variations in low-level wind speeds which, consequently, may produce fluctuations in dustiness (CP72). A weak correlation has been shown statistically between the frequency of dust haziness over West Africa and the passage of a travelling AEW. Carlson and Prospero (1972) claim the weakness of the correlation may be due to the surface observations failing to show any systematic variations in the daily occurrence of dust, haze, or visibility along the coast.

4 Hurricanes

Hurricanes, cyclones, and typhoons, depending on where one is in the world, exact billions of dollars in damage and claim the lives of thousands each year as nothing can prevent these roiling tempests from devastating vast swaths of land and inundating extensive coastal acreage. Populated areas have never quite learned to cope with the fearful force, they merely have acclimated to being knocked into periodic silence. Some have rid themselves of the responsibility of dealing with hurricanes by simply absconding from their homes and heading towards the country's interior, but, for many, and especially for those living in poor, developing nations, moving elsewhere is economically unrealistic.

Understanding the mechanisms involved in cyclone generation will serve as a springboard towards its interactions with the Saharan air layer in the next section. In developing a sense of understanding, it may also be possible one day to nudge a hurricane onto a more benign

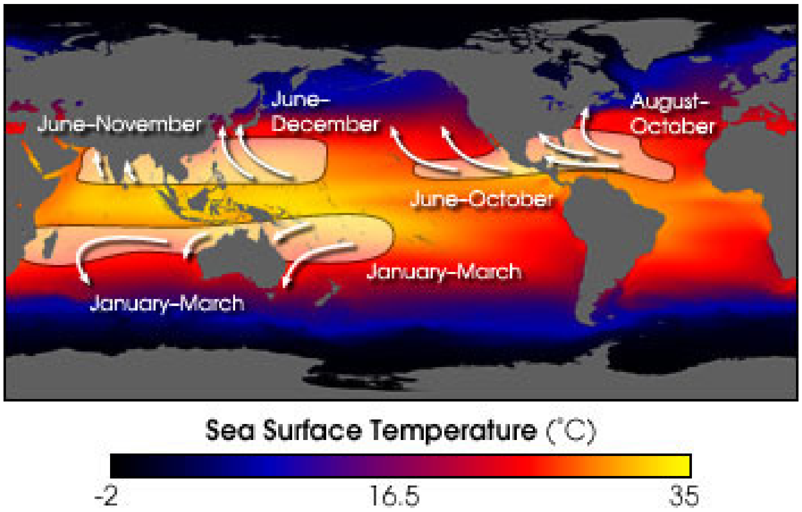

Figure 5: Global occurrence of cyclone activity. (NASA 2003.)

path or otherwise diffuse it[3].

Tropical cyclones (TC) require sea surface temperatures of 26°C and the influence of earth's rotation to initiate spinning. They occur where the conditions are *humid*, *unstable*, and *calm*. A wet environment continually provides the necessary fuel from the ocean surface ensuring growth and development. An environmentally unstable lapse rate promotes buoyant, warm, parcels of air to convect upward unabatedly. A developing storm situated in a zone of negligible wind-shear (whereby wind speed and direction does not change appreciably with height) is important in maintaining the delicate balance of both uplift and structure.

To see why hurricane strength may be attenuated by entrained dust, one must understand their nature and origins. Hoffman (2004) (hereafter H04) explains how a hurricane is engendered: they grow as clusters of thunderstorms over the tropical oceans. Low-latitude seas continuously provide heat and moisture to the atmosphere producing warm, humid air above the surface. As the air rises, the water vapor in it condenses to form clouds and precipitation. Condensation releases latent-heat – the solar it heat it took to evaporate the water at the ocean surface – in a still reinforcing feedback process (see Fig. 6). The heat released above the tropical seas creates a surface low, where additional most air from the peripheries converges. The continuous movement in the ever enlarging thunderstorm shifts huge amounts of heat, air, and water skyward. This upward transfer further enhances the convergence of the surrounding air toward the center, which starts to circulate under the influence of the earth's rotation. It continues apace (Fig. 7). As the storm intensifies, a calm low pressure hub typically forms. By now it is encircled by a ring of clouds and high winds called the eye wall. It is now a hurricane (Fig. 8) . At the same time, the rising air, now heated and having lost much of its moisture, can rise no further as the stratosphere acts as a lid, consequently downdrafting occurs in the eye-wall.

[3]See the article by Hoffman, R.N. (2004). "Controlling Hurricanes," *Scientific American*, **294**(4), 69-75.

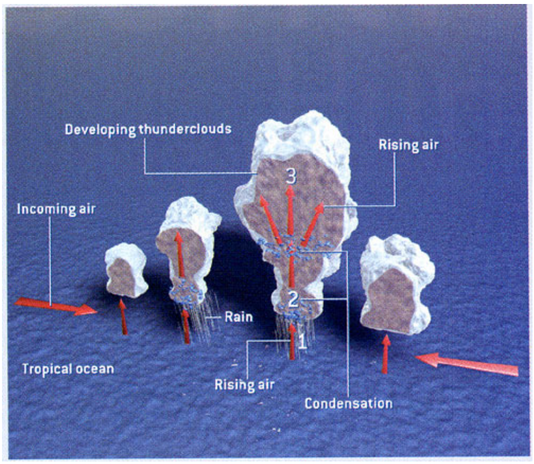

Figure 6: Progression model of a cyclone during first major initiation phase. (*Scientific American* 2004.)

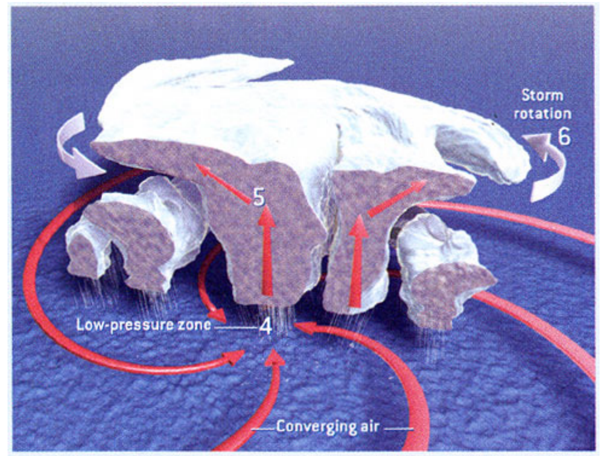

Figure 7: Progression model of a cyclone during second major initiation phase. (*Scientific American* 2004.)

5 Effects on Hurricanes

In three main ways, the SAL may affect tropical cyclone activity:

- *Imposes low humidity while promoting convectively driven downdrafts*
- *Stabilizes the atmosphere by inducing a temperature inversion*
- *Enhances strong vertical wind shear by its mid-level easterly jet*

The dry SAL air can act to suppress convection by enhancing evaporatively driven downdrafts (Emmanuel 1989; Powell 1990). Dunion and Velden (2004) summarized the following: the low-to-midlevel inflow of TC's advect the SAL's low humidity into the TC circulation. This dry air is also associated with the reduced vales of convective available potential energy, a measure of the stability of the atmosphere. The base of the SAL is often 5-10°C warmer than the Jordan (1958) mean tropical sounding (Diaz et. al. 1976). As mentioned above, the warmth of the SAL is primarily due to the intensity of the solar heating and the vastness of the arid desert terrain. The absorption of solar radiation by the suspended minerals can exceed the overall longwave cooling in the layer, thereby warming the SAL and reinforcing the temperature inversion at the base (DV04; Carlson and Benjamin 1980). The southern or southwestern edge of the SAL usually coincides with an easterly wind maximum near 700 millibars. The wind speed at the maximum is often 10-17 m s^{-1} and is thought to be as high as 25 m s^{-1}, generally 7-10 m s^{-1} faster than the typical Trade Winds (HV04). Satellite tracking suggests the SAL's mid-level wind maximum suppresses TC formation. These embedded strong winds significantly increase the local vertical wind shear by increasing the low-to-mid-level easterly flow. Several hurricanes (examined by DV04) (see Fig. 10, Fig. 11) embedded in the SAL had low-level circulations that transcended their mid-and upper-level deep convection, mainly due to the SAL's mid-level easterly jet. As depicted in Fig. 10, Hurricane Joyce was relegated to a moderate storm only three days after being upgraded to a hurricane from being overrun by the SAL. In a case study performed by a joint venture

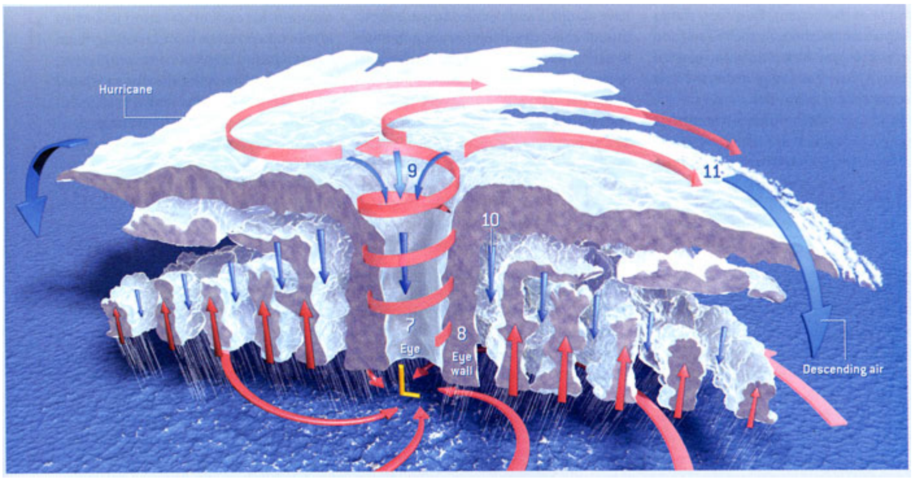

Figure 8: Fully developed final phase model of a cyclone (full rotation). (*Scientific American* 2004.)

between NOAA and the US Navy (as reported by DV04), in September 2000, Joyce formed from an AEW positioned several hundred kilometers ahead of a large SAL outbreak. Favorable environmental conditions allowed this AEW to develop from a weak tropical depression late on 25 September 2000 to a 41 m s^{-1} hurricane early on 28 September. Overtaking Joyce on 27 September, the storm immediately began to follow a weakening trend as the plume impinged its characteristic suppressive qualities on it. During the next 48 hours, Joyce deteriorated greatly and within 96 hours became a disorganized tropical depression. The SAL likely imposed low humidity and strong vertical wind shear on the main circulation while stabilizing the environmental lapse rate by a temperature inversion.

Figure 10 is a graphic representation of a time series for several Atlantic TCs in 1999 through 2001. Red shading indicates the TC was under the suppressing influence of the SAL. Green shading indicates periods when the SAL was not impacting the TC. Immediately visible is the unmistakable smothering effect of the SAL on each hurricane as each one struggles to intensify or maintain structure. This significant feature demonstrates how antagonistic the SAL is on TCs.

6 Limitations

a. Likely objections

Some aspects of the studies remain unresolved or ambiguous. Summarized are some notables: During the TC project run conducted and reported by DV04, trajectory calculations disabled models attempting to quantify spatial and temporal concentrations. Computer forecasts became inaccurate because knowledge of the SAL was not well represented and hence not convincingly well-matched to observations. The SAL was not detectable by visible and infrared bands. However, suspended dust reduced brightness and so could be reasonably followed in certain channels. Because of the SAL's limited vertical extent, its dry air be

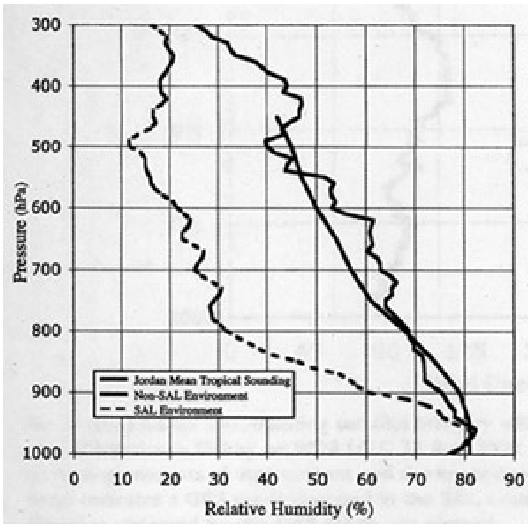

Figure 9: Composite GPS sonde profiles from sondes launched in the environments of Hurricane Danielle and Georges of 1998 and Hurricane Debby and Joyce of 2000. (Dunion and Velden 2004.)

difficult to quantify using moisture channels on the GOES[4] and MSG[5] satellites. The most effective means in studying its low humidity was by first identifying it with tracking imagery and then directing aircraft to make in situ measurements of its thermodynamics structure.

Most sensing platforms had difficulty detecting and measuring the strong easterly wind surge associated with the SAL because of its spatial and temporal limitations. GPS sondes/models misrepresented the relative-humidity at times by a factor of two or three. There is no regular four-dimensional dust observing system established.

The intensity change of TC influenced by the SAL may not be well predicted by the operational SHIPS[6] model (used in the study) and may not specifically consider the SAL in its methodology. Furthermore, SHIPS relies on model data that may not accurate represent the SAL's thermodynamics properties. On the other hand, the salient aspect of the study was to illustrate on real terms the effects of entrained dust into a phenomenon formerly deemed inexorable. The shortfalls of the models are tolerable in light of the successful results attained in the case studies and the unambiguous demonstration of the physical interactions involved.

And what evidence is there that transported dust is of African origin? Firstly, in contrast to winter, field studies conducted in Barbados show that the dust content of the atmosphere is grey or black than light red-brown common in summer. Observation showed that major dust outbreak events terminate by early November, ample time before the onset of winter in the northern hemisphere (CP72). Secondly, parcels of dust-laden air over the ocean are generally visible in satellite imagery during the first few days of its trans-oceanic passage to the West Indies. Because the five to six day transit time is relatively short, estimates of position and areal extent are relatively truthful. And thirdly, the mineralogy of the dust sampled in hazy

[4]Geostationary Operational Environmental Satellite
[5]Meteosat Second Generation
[6]Statistical Hurricane Intensity and Prediction Scheme

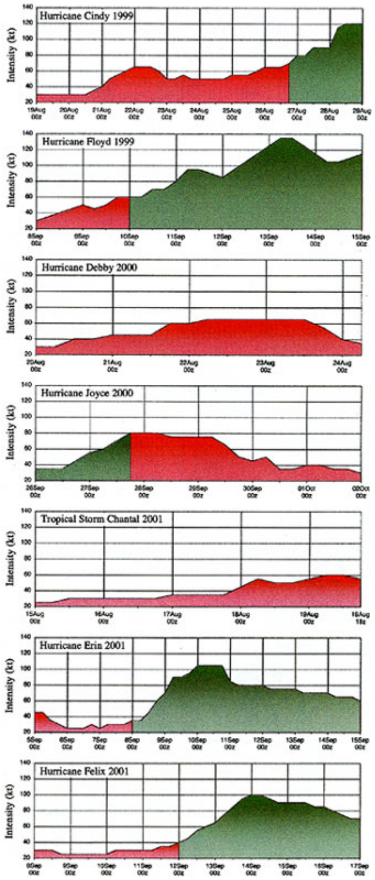

Figure 10: Cross-sectional plot of best-track intensity for several Atlantic TCs in 1999 through 2001. Red shading indicates that the TC was under the suppressing influence of the SAL. Green shading indicates periods when the SAL was not impacting the TC. (Dunion and Velden 2004.)

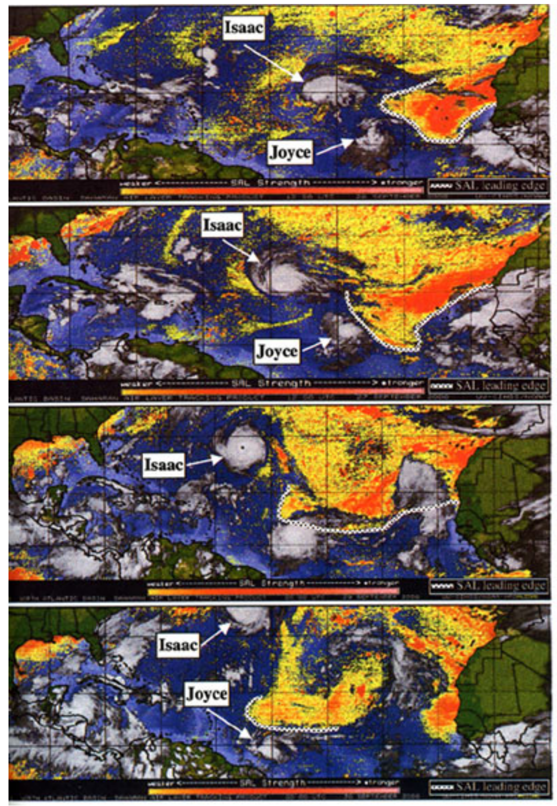

Figure 11: Plan-form view time-series satellite imagery of Hurricane Isaac and Joyce during a four-day period. (Dunion and Velden 2004.)

areas is uniform regardless of the site location, whereas local sampling consists almost entirely of highly calcareous matter, as might be expected due to the large areas of exposed coral in the region (Prospero and Carlson 1971).

Numerical models in tandem with observations show compellingly each component coupled to the other. Strong surface heating over Africa is essential for the generation of the SAL and the development of the AEWs. The suspended dust (further heated from incoming solar radiation) amplifies the SAL and the AEWs. Carlson and Prospero's (1972) analyses exemplify a correlated three-way partnership in natural agreement with observations: *all three represent a general phenomenon initiated by strong sensible heating from the sun over a surface with low surface moisture, that is, a wave disturbance caused by differential heating.* Numerical models indicate that suppressing or shutting down one component inadvertently suppressed or shut down another, or others. The absorption of short-wave radiation by the suspended dust maintained a deep and warm SAL over the ocean, further suppressing convection, further imposing low humidity, and further strengthening the AEW. The AEW indirectly impelled dust skyward and indirectly guided the plume westward to the US and Caribbean.

• Conclusion

This article gave an overview of a series of studies conducted the past 30 years elucidating the properties of suspended Saharan dust over West Africa and in particular the physical properties of dust and its possible influence on tropical cyclone activity in the North Atlantic. Satellite imagery allowed tracking of the SAL across the North Atlantic basin providing insight into the relationship between the SAL and North Atlantic TCs. Satellite imagery also revealed that the size of the SAL can exceed that of the contiguous US and can maintain internal self consistency for thousands of kilometers. And because 40 to 60 percent fewer cyclones occur in the Atlantic Ocean than elsewhere, a telling link may have been provided signaling the coupled relationship between TC generation and the SAL. A more detailed understanding of SAL-TC interactions can be attained by further involving validation of the GOES SAL-tracking imagery using GPS sondes, research aircraft equipped with in situ instrumentation platforms, and supplemented by improved SAL detection using high-resolution satellite imagery and implementing a regular four-dimensional dust observing system.

• **Acknowledgements**. I am deeply indebted to Professor Warner and his patience for allowing me to incessantly delay handing in the article because of time-table clashes, academic commitments, and the proverbial "real-world." I wish to express my gratitude to Srinath Vadlamani for graciously advancing me his copy of a much needed Latex manual, it was an indispensable tool. I especially wish to express my gratitude to a one Viktor Przebinda for so kindly surrendering his domain and providing me with what seemed like limitless time on his computer when mine seemed unable to cope with the task, without him I wonder what the outcome would be like.

• Appendix

Hydrodynamical forces on aerosol particles

Atmospheric aerosol feedback functionally depends on altitude, size, density, composition, distribution, and albedo, to name a few. Aerosols can scatter radiation forward and backward in addition to absorbing it (T77). It is customary to speak of "optical depth" in such problems, where optical depth is the amount of extinction a beam of light experiences traveling between two points. The dynamics of fluid flow around aerosol particles is important in the discussion of all interactions between particles and the atmosphere. In a viscous flow around a sphere (as derived by (T77)), the Navier-Stokes equation:

$$\frac{\partial}{\partial t}\nu + (\nu \cdot \nabla)\,\nu = -\frac{1}{\rho}\,\nabla\rho + \frac{\eta}{\rho}\,\nabla^2\nu \tag{3}$$

governs the flow velocity ν of an incompressible fluid under isentropic (constant entropy) conditions. The vorticity equation

$$\frac{\partial}{\partial t}\omega = \nabla \times (\nu \times \omega) + \frac{\eta}{\rho}\,\nabla^2\omega,$$

which can be attained by taking the curl of the Navier-Stokes equation, setting $\omega = \nabla \times \nu$, shows that only the dynamic viscosity η/ρ is relevant to the flow field. As soon as a body is inserted into the fluid, however, an external length is introduced to occur over a distance of the order of l, if l denotes any appropriate linear dimension of the body. The non-linear term $(\nu \cdot \nabla)\nu$ is of the order of $|v_0^2|$ if v_0 is the speed of the laminar fluid; the viscous term $(\eta/\rho)\nabla^2\nu$ is of the order of $\eta|v_0|/\rho l^2$ (T77). For the viscous term to dominate the condition as further derived by (T77) is evidently:

$$\frac{|v_0^2|}{l}\left(\frac{\eta|v_0|}{\rho l^2}\right)^{-1} << 1,$$

or

$$\frac{\rho l|v_0|}{\eta} << 1,$$

where $\rho l|v_0|/\eta$ is the Reynolds number. Small values of the Reynolds number imply viscous flow, for which one may write:

$$\frac{\partial \nu}{\partial t} = -\frac{1}{\rho}\,\nabla p + \frac{\eta}{\rho}\,\nabla^2\nu, \tag{4}$$

whereas at large values of the Reynolds number the relevant equation is:

$$\left(\frac{\partial}{\partial t} + \nu \cdot \nabla\right)\nu = -\frac{1}{\rho}\,\nabla p. \tag{5}$$

The Reynolds number is not a property of the fluid alone since it necessarily involves the external dimension l. A more thorough treatment is done by (T77).

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