A Case Study of Mobilization and Transport of Saharan Dust

DOUGLAS L. WESTPHAL*
Department of Meteorology, Pennsylvania State University, University Park, Pennsylvania

OWEN B. TOON
Space Sciences Division, NASA/Ames Research Center, Moffett Field, California

TOBY N. CARLSON
Department of Meteorology, Pennsylvania State University, University Park, Pennsylvania
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ABSTRACT

Numerical models of the atmosphere and aerosols are used to investigate mobilization and transport of Saharan dust over West Africa and the tropical Atlantic Ocean for 23–28 August 1974. We have found that mobilization during this period was related to the passage of a shallow easterly wave and was not initiated by dry convective mixing of a midlevel easterly jet, as has been previously suggested, since high static stability beneath the midlevel easterly jet inhibited significant boundary layer development and transport of momentum in the jet down to the surface. Instead, mobilization was done by dry convective mixing of low-level jets associated with the easterly wave. Another easterly wave present in the domain during the period did not contribute significantly to dust mobilization while over Africa yet became a strong tropical storm over the Atlantic Ocean in early September. The periodicity of the outbreak was reinforced by scavenging of dust by precipitation associated with the easterly waves.

The model simulations show that the aerosol at any one point can be a complicated mixture of particles lifted at different times and different places. Bimodal size distributions developed when dust was mobilized within a dust plume that was generated on a previous day. An elevated layer of dust developed over the ocean as the northeast trade winds advected clean air underneath the dust-laden air as it moved westward. The size and spatial distributions of aerosol in the marine layer depended upon the undercutting process, the amount of background mineral aerosol present, and transport across the marine layer inversion by sedimentation and turbulent mixing.

1. Introduction

The general characteristics of Saharan dust outbreaks over the Atlantic Ocean have been determined from satellite imagery, field studies, and diagnostic studies (Prospero and Carlson 1981). A conceptual model of an outbreak has emerged in which strong insolation and dry convection over the desert mix momentum from the midlevel easterly jet (MLEJ) down to the surface where it mobilizes dust. This usually occurs in an anticyclonically rotating eddy of air, called the Saharan air layer, or SAL, which travels westward behind an easterly wave. The MLEJ is a common feature behind the wave axis and along the boundary between the hot desert air to the north over the Sahara and the cooler air to the south over the Sahel and Sudan. Measure-ments of the daily averaged mineral aerosol concentration at Sal Island (Fig. 1) and other locations in the tropical Atlantic Ocean and Caribbean Sea display a midday periodicity that has been associated with the more or less regular passage of easterly waves and dust-laden eddies.

While the general features of dust outbreaks over the ocean are known, many interesting questions remain regarding the dynamics of mobilization and of transport and the periodic nature of outbreaks. For example, it has been assumed that dust mobilization, or deflation, is greatest in the afternoon when the boundary layer is deepest and most efficient at mixing momentum from aloft down to the surface where it is transferred to the soil. However, satellite imagery reveals strong deflation events just after sunrise as well. Additionally the MLEJ is considered to be the elevated source of momentum for deflation during dry convection, but analyses of some Saharan dust outbreaks by Estoque et al. (1986) show that dust is often mobilized well to the north of the preferred location of the MLEJ. These points indicate that additional dynamical process may be involved.

* Present affiliation: NASA/Ames Research Center, Moffett Field, California.

Corresponding author address: Dr. Douglas L. Westphal, NASA/Ames Research Center, MS 245-3, Moffett Field, CA 94035.
We are interested in these and other aspects of Saharan dust outbreaks. If enough observations were available, these topics could be addressed by conducting case studies of individual outbreaks, as Bertrand et al. (1974) and Morales (1979, 1981) have done for the Harmattan and Sudanese dust storms. However, little is known about the atmospheric circulation in the preferred region for mobilization during summer since the region covers an area roughly 10° latitude by 15° longitude with no regularly reporting surface or upper air stations. Using meteorological observations alone, it is difficult to determine the situation leading to an outbreak and to follow the plume across the Sahara.

However, recent work by Karyampudi and Carlson (1988; hereafter KC) affords an opportunity to investigate the dynamics of dust lifting. Using a dynamical model, they have carried out a five-day numerical simulation of the tropical atmosphere from the Red Sea to the Caribbean Sea. Their dynamical model produced an array of meteorological variables in the data-sparse regions. The dates of the simulation, 23–28 August 1974, were chosen based on the availability of a meteorological dataset developed by the Geophysical Fluid Dynamics Laboratory (GFDL) and the fact that easterly waves were particularly intense during late August and early September of that year. Although not as strong as those observed earlier in the summer, the dust outbreak that occurred during 23–28 August was observed at the surface stations near the coast, detected in infrared and visible satellite imagery, and detected in aerosol measurements taken at Sal Island (Fig. 1).

Our goal is to explain the dynamics of dust mobilization and transport during this period. We have adapted a predictive numerical aerosol model to desert dust storms and have coupled it to the dynamical model used in KC. Previous numerical investigations of desert dust storms have been made with one-dimensional (1D) or two-dimensional (2D) mixed layer models (Berkośky 1982; Lee 1983), 2D steady state models (Schütz 1980), and 2D time-dependent grid-point models (Westphal et al. 1987). The mixed layer models lack the vertical resolution necessary for modeling the layered structure of dust outbreaks and incorrectly assume that all dust, even the largest particles, are well mixed in the layer at all times. Schütz’s model had finer vertical resolution but did not consider mobilization or transport of dust over the desert. The characteristics of these models were discussed in Westphal et al. (1987, hereafter WTC). In that paper we used 2D versions of the 3D dynamical and aerosol models to gain valuable insights into dust mobilization, vertical mixing, and horizontal transport, but the inherent assumptions made in reducing the models to two dimensions rendered them incapable of modeling easterly waves, the multiday periodicity of outbreaks, under-cutting, and other 3D features of outbreaks. We now use the 3D models to investigate these other aspects of Saharan dust outbreaks.

In section 2 we present the available observations of the outbreak that took place during 23–28 August and compare the outbreak with the conceptual model of outbreaks. We describe the meteorological and aerosol models in section 3 and present the dynamical and aerosol model results separately in sections 4 and 5, respectively. A summary and discussion are presented in the final section.

Throughout this paper, we will use Coordinated Universal Time (UTC) which is equivalent to Greenwich Mean Time (GMT). Because of the proximity to the Greenwich meridian, UTC is close to local time for most of the model domain. As in WTC we will use the following terms to describe aerosol size.
ranges: “submicron” for particles with radius $r < 1$ μm, “large” for $1 \mu m < r < 3 \mu m$, “giant” for $3 \mu m < r < 30 \mu m$, and “ultragiant” for $r > 30 \mu m$.

2. Observations

The observed (GFDL analyses) and simulated synoptic-scale circulations have been described extensively in KC and are discussed briefly here. Because of the paucity of data, little is known about the circulation in the dust source area. If the model simulations agree with the available observations in other regions then we can, with some confidence, use the simulations to study the dynamics of deflation in data-sparse regions.

a. The meteorological analysis

The observed synoptic situations at 700 mb for 1200 UTC 23 and 28 August 1974 (Fig. 2) exhibit the general features of the conceptual model. Two wave axes or troughs, denoted by T-1 and T-2, are seen on 23 August at 18°W and 8°E, respectively. By 28 August, T-1 and T-2 have moved to 48° and 22°W, respectively, and a third wave T-3 is located at 3°W. A MLEJ is found behind the wave axes. On a climatological basis, precipitation and cloud amounts associated with easterly waves are greatest at and ahead of the trough axis at a latitude of 10°N and smallest at and ahead of the ridge (Burpee and Reed 1982; Simpson et al. 1968). The areas of precipitation (shown in Fig. 2a, b) follow this general pattern with the largest areas of precipitation associated with the trough axes and confined mostly to the area south of 17°N. During the days following 28 August, the first easterly wave became Hurricane Carmen, the second dissipated in the mid-Atlantic Ocean, and the third became Tropical Storm Elaine. Thus, this period is particularly interesting in terms of the dynamics and energetics of tropical disturbances. An investigation of the growth mechanisms and energetics of the easterly waves is described in KC.

Two upper air stations that regularly reported during GATE were Sal, Cape Verde Islands (16.7°N, 23.0°W) and Dakar, Senegal (14.7°N, 17.5°W). Time-height cross sections of wind, relative humidity, and temperature deviation from the time mean were constructed from the radiosonde data for these two stations and are presented in Figs. 3 and 4. Also shown are the surface weather reports and measured aerosol concentrations (Prospero et al. 1976). Sal Island is north of the monsoonal trough and the path of the surface cyclones so the low-level winds are generally from the north or east and no precipitation is reported. The time section shows the SAL arriving with the east-northeast wind between 500 and 900 mb early on 26 August, two days after the passage of the trough axis. The SAL has a dry and relatively warm base and a moist and relatively cool top.

The surface cyclones associated with the first and second easterly waves cross the African coast close to the latitude of Dakar so the low-level winds at Dakar are more variable than at Sal Island and even take on westerly and southerly components at times (Fig. 4). The middle and lower levels of the troposphere at Dakar are moist throughout the period, and showers or thunderstorms are reported on four of the six days. The SAL does not reach Dakar, except possibly between 700 and 500 mb on 26 August.

b. Analysis of 23–28 August dust plumes

We now analyze the dust plumes during 23–28 August by combining the 1200 UTC surface station reports of weather type and horizontal visibility with dust plume outlines observed in the satellite imagery and the aerosol measurements taken at Dakar and Sal Island. The haze and dust weather types that were reported during the period of study are defined in Table 1. We recognize that horizontal visibility is a subjective measure of atmospheric aerosol loading in the surface layer. Surface stations may be biased toward frequent
reports of dust storms and low horizontal visibility because the stations are usually located at open, wind-swept airports and the area around the towns has been overgrazed and otherwise disturbed by human activity, making it more susceptible to deflation.

Infrared satellite imagery taken at half-hourly intervals are used to determine plume location in the remote regions of the Sahara. In the infrared imagery, the dust takes on a gray shade similar to stratocumulus water clouds but can be differentiated from water clouds since the latter have sharp lateral boundaries whereas dust has diffuse boundaries. The dust can be obscured in the infrared imagery when water clouds form over the Sahara. On the other hand, mixed-layer cumulus clouds sometimes form in dust plumes in late afternoon and can be used to track the dust during the night hours when the dust plumes, by themselves, are indistinguishable from the cool desert surface. The dust plumes cannot be tracked in the moist air south of the intertropical front (ITF) because the dust signature is obscured by the strong water vapor emission. In the visible satellite imagery, dust appears as a diffuse gray haze over the ocean that can be followed day to day. However, widespread low-level stratocumulus clouds inhibit the detection of dust over the ocean in visible imagery during much of this five-day period.

Prospero et al. (1976) measured the 24-h average mineral aerosol concentration at Sal Island and the 48-h average concentration at Dakar during the 5-day period. These nine measurements are shown in Figs. 3 and 4 and represent all of the available direct aerosol measurements for the period. The aerosol concentration at Sal Island is low for the first 3 days but then increases to 14.6 \( \mu g \) m\(^{-3}\) on 26 August, 40.5 \( \mu g \) m\(^{-3}\) on 27 August, and 62.1 \( \mu g \) m\(^{-3}\) on 28 August. The aerosol concentration at Dakar is 7.5 \( \mu g \) m\(^{-3}\) on 23 August and increases to 13.3 \( \mu g \) m\(^{-3}\) on 26 August. Prospero et al. (1976) noted that the aerosol concentration at Dakar may have been affected by the local “clean” ocean breeze. Additionally, scavenging may have reduced the local aerosol concentration since precipitation was reported at Dakar on 4 days.

The surface reports, satellite imagery, and surface aerosol measurements are combined in Figs. 5a–f. Also shown are the locations of the MLEJ (when present) and the trough axes. Areas of precipitation as determined from infrared and visible satellite imagery and surface station reports of precipitation are shown. We also show the boundary of the outbreak as simulated by the model (see section 5 for discussion).

At 1200 UTC 23 August (Fig. 5a) there are only a few surface observations of haze or suspended dust
On 24 August, haze and suspended dust are reported in central Algeria and along the Spanish Saharan coast (Fig. 5b). In addition to a general haze seen over the Sahara, a heavier cyclonically curved dust plume is apparent in the infrared imagery just north of the T-2 axis at 20°–25°N, 0°–5°E. Dust is probably present over the ocean north of 20°N because the horizontal visibility off the Spanish Sahara coast and in the Canary Islands (28.6°N, 17.7°W) is less than 10 km.

The cyclonic dust plume from 24 August moves westward during the night and by 1200 UTC 25 August is centered at 22°N, 44°W. The heaviest dust cloud seen in the infrared imagery for the 5-day period appears north of T-2 and the 24 August plume at 0630 UTC (Fig. 5c). The new plume is over 1000 km long and has a distinct cyclonic curvature. Haze and suspended dust are reported in western Mauritania, the Spanish Sahara, and the Canary Islands, and the horizontal visibility is lower in these areas than on 24 August. Dust could not be seen over the ocean in the satellite imagery because of stratocumulus, but several reports of horizontal visibility of 10 km along 23°N and in the Canary Islands suggest the presence of dust there. An isolated dust storm is reported at Zinguinchor, Senegal (12.5°N, 16.3°W).

![Diagram with weather symbols and observations]

**Table 1. Symbols used in Figs. 5a–f.**

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Meaning</th>
</tr>
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<tbody>
<tr>
<td>From satellite imagery</td>
<td></td>
</tr>
<tr>
<td>⍵ ⍵ ⍵ ⍵ ⍵ ⍵ ⍵</td>
<td>Haze with distinct boundary</td>
</tr>
<tr>
<td>⍵ ⍵ ⍵ ⍵</td>
<td>Haze with indefinite boundary</td>
</tr>
<tr>
<td>⍵ ⍵ ⍵ ⍵ ⍵ ⍵ ⍵</td>
<td>Heavy dust</td>
</tr>
<tr>
<td>⍵ ⍵ ⍵</td>
<td>Area of precipitation</td>
</tr>
<tr>
<td>From dynamical analysis</td>
<td></td>
</tr>
<tr>
<td>← MLEJ — — — — — — — — — — — —</td>
<td>Location of MLEJ axis</td>
</tr>
<tr>
<td>T1——</td>
<td>Easterly wave axis</td>
</tr>
<tr>
<td>From surface station reports</td>
<td></td>
</tr>
<tr>
<td>20</td>
<td>Visibility (km)</td>
</tr>
<tr>
<td>⊙</td>
<td>Aerosol concentration (µg m⁻³)</td>
</tr>
<tr>
<td>∞</td>
<td>Haze</td>
</tr>
<tr>
<td>S</td>
<td>Widespread dust in suspension, not raised at time of observation</td>
</tr>
<tr>
<td>$</td>
<td>Dust raised at time of observation</td>
</tr>
<tr>
<td>←</td>
<td>Slight or moderate dust storm; no change during past hour</td>
</tr>
<tr>
<td></td>
<td>Slight or moderate dust storm; has increased during past hour</td>
</tr>
<tr>
<td>From simulation</td>
<td></td>
</tr>
<tr>
<td>⍵ ⍵ ⍵ ⍵ ⍵ ⍵ ⍵</td>
<td>Boundary of outbreak (optical depth = 0.01)</td>
</tr>
</tbody>
</table>

though most of the desert appears to be covered with light dust in the infrared imagery. The single observation of suspended dust is near T-2. Even though dust was not seen in the visible imagery, we have shown a haze area over the ocean in the analysis because of the suppressed cloudiness in the region, typical of the SAL, and because of an eyewitness account by one of the authors (TNC) while aboard an aircraft in the vicinity that day. A MLEJ is associated with T-1 at the coast but there were no reports of dust mobilization in the area.
Fig. 5. Analyzed location of dust plumes for 23–28 August 1974: (a) 23 August, (b) 24 August, (c) 25 August, (d) 26 August, (e) 27 August, and (f) 28 August. Analyses were made using satellite imagery, surface observations, and aerosol measurements. The simulated plume outlines are shown also. Symbols are defined in Table 1.
The heaviest dust lifting episode reported in the surface observations takes place on 26 August in the Spanish Sahara and western Mauritania (Fig. 5d). Several stations reported dust storms and horizontal visibility of less than 5 km. In the visible imagery, desert surface features are obscured and heavy dust with striations paralleling the trade winds can be seen over the ocean between Nouadhibou, Mauritania (21°N,
17°W) and 17°N in the area downwind of the land stations that reported moderate dust storms. An area of lighter haze is also visible between the coast and T-1. The time section of temperature and moisture for Sal Island (Fig. 3) shows the SAL arriving on this day and the dust concentration at Sal Island increasing to 14.6 μg m⁻³. There are no observations to support the presence of significant dust over the ocean north of 20°N but the area is relatively cloud free in the visible imagery, which suggests the presence of a SAL with low dust concentrations. The infrared images show that further inland, haze covers much of the Sahara and a new plume becomes visible near 21°N, 3°E at 1130 UTC. Suspended dust is reported in central Algeria along the northern boundary of the haze seen in the infrared imagery.

On 27 August, haze and suspended dust are reported in the Spanish Sahara and off its coast while heavy dust can be seen off the coast in the visible imagery (Fig. 5e). There are no reports of haze or suspended dust in western Mauritania but low horizontal visibility is reported again. The visible imagery also shows an anticyclonically curved tongue of dusty air with suppressed convection (i.e., the SAL) ahead of T-2. Beneath the SAL at Sal Island the aerosol concentration has risen to 40 μg m⁻³. A single suspended dust observation is reported in eastern Mali near T-3. The outline of the dust plume over the central Sahara cannot be discerned in the infrared imagery.

The heaviest aerosol concentration at Sal Island for the 5-day period is 62.1 μg m⁻³ and occurs on 28 August (Fig. 5f). Sal Island is within the northern boundary of an anticyclonic tongue of dust that is clearly visible in the visible imagery over the ocean between T-1 and T-2. Low horizontal visibility, haze, and suspended dust are reported in the Spanish Sahara and Mauritania but dust cannot be seen in the infrared imagery because of water clouds. The suspended dust reported at surface stations in central Algeria is not apparent in the infrared imagery, but the northeast boundary of a dust plume appears sometime before 0800 UTC in Southern Algeria and well north of the MLEJ associated with T-3. Elsewhere in central Africa, the dust boundary is difficult to define.

In summary, the surface reports, satellite imagery, and aerosol measurements suggest that some dust was in suspension over the ocean on 23 August and that a separate outbreak occurred over the Sahara on 24–26 August and reached Sal Island on 26–28 August more than a day after the passage of the SAL front (Fig. 3a). The north-northeast to northeast winds in the marine layer between Sal Island and the coast during the 5-day period (Fig. 3a) may lead to the undercutting effect discussed by Jaenicke and Schütz (1978) and Prospero and Carlson (1981), wherein cool, moist, dust-free air undercut and replaces the SAL in the lowest 100 mb as it nears the coast and delays the arrival of dust at Sal Island. Dust mobilization appears to be correlated with the position of T-2 on 23–26 August and with the position of T-3 on 28 August. However, the deflation events are not closely, if at all, related to the MLEJ. In most cases the individual dust plumes appear over the desert in the early morning hours, not late afternoon as would be expected if dry convective mixing of momentum from the MLEJ was responsible for mobilization. Generally, the horizontal visibility over the desert decreases over the 5 days, with the exception that the lowest individual visibilities for the entire period were reported on 26 August. There are only two dust storm observations and one haze observation south of 17°N during the 5-day period, and the horizontal visibility in this area is generally greater than 20 km.

3. Model descriptions

We use two numerical models in our investigation of this Saharan dust outbreak. The atmosphere is modeled using the Pennsylvania State University/NCAR limited area tropical model (KC), a version of the regional model originally developed for midlatitude weather systems by Anthes and Warner (1978). Karyampudi and Carlson (1988) improved the cumulus convection and surface moisture flux parameterizations, and added a new lateral boundary condition and a simple atmospheric radiative heating parameterization. Benjamin and Carlson (1986) improved the surface radiation budget calculations. The horizontal grid spacing is 220 km. The spatial domain of the dynamical model is shown in Fig. 6. The model is divided vertically into 13 layers with a 2 mb (~20 m) thick surface layer, 50 mb spacing below 600 mb, and 100 mb spacing above 600 mb.

The NASA Ames aerosol model (Toon et al. 1988) uses a subset of the dynamical fields produced by the dynamical model (Fig. 6) to predict the 3D distribution of aerosols segregated into 30 bins ranging from 0.1 to 80 μm in radius. The aerosol model includes physical processes such as 3D advection and diffusion, sedimentation, dry deposition, wet removal, and dust mobilization. Currently, aerosol radiative heating rates are not calculated for input to the dynamical model so the two models operate in a “one-way interaction” mode. Westphal et al. (1987) described a 2D version of the combined model as applied to Saharan dust outbreaks. Coagulation is not included in this study since the 2D simulations in WTC showed that coagulation does not significantly alter the size distribution for the low number concentrations and short time scales considered here. Of particular importance to our study are the models’ treatment of the mobilization flux and wet removal. These aspects of the models are discussed below in more detail.

a. Dust mobilization parameterization

The value of the surface wind speed above which deflation is first observed is called the threshold wind speed \( V_t \). Helgren and Prospero (1987) have estimated
at eight stations in the Sahara from the reported wind speed during 41 raised dust events during 1 July to 15 August 1974. Most of the heavy outbreaks that were detected at Sal Island during GATE took place during this period (Fig. 1). The stations sampled by Helgren and Prospero were in the western and northern Saharan desert and may or may not represent conditions in central Sahara. Helgren and Prospero found a great deal of variability in $V_i$ from station to station with values of $V_i$ ranging from 5 to 12.5 m s$^{-1}$. The average $V_i$ for all stations was 8.2 m s$^{-1}$ and the lowest $V_i$ was 5 m s$^{-1}$. The average wind speed during all events was 10.4 m s$^{-1}$. The variability in $V_i$ from station to station was attributed to differences in soil type and terrain at the different stations.

We carried out the same analysis for the same eight stations for 23 August to 23 September 1974. The statistical significance of the analysis is questionable because there were only 12 raised dust events at the stations during this period. Hence, we report the average wind speed during all 12 events instead of the lowest wind speed at which raised dust was observed for each station. The average wind speed for all events was 4.6 m s$^{-1}$. The wind speed during a raised dust event in the second period never exceeded 5.6 m s$^{-1}$, whereas dust was not raised in the first period for winds less than 5 m s$^{-1}$. The data suggest that the wind speeds and $V_i$ decreased in late summer. Lighter winds could be due to movement or weakening of the ITF. Apparently, the decrease in wind speed dominates the possible decrease in $V_i$ since the Sal Island aerosol concentrations (Fig. 1) are substantially lower after 22 August.

We calculated the average wind speed at all surface stations in the Sahara that reported conditions of no dust, haze and suspended dust, and raised dust during 23–28 August and found the values to be 2.8, 3.2, and 4 m s$^{-1}$, respectively. This shows that despite the lower wind speeds observed during our period of study, deflation is still correlated with higher wind speeds. The lower threshold wind speeds during the second period could be due to drier, more weathered soils in late summer. There are no data to substantiate this possibility but we note that precipitation was not reported at or near these eight stations during the 5-day period. Seasonal, annual, and spatial trends in $V_i$ might become apparent if many years of surface data for all of West Africa were analyzed in the same manner used by Helgren and Prospero (1987).

Although the dynamical model predicts the surface wind speed, we prefer to use the threshold friction velocity $u_*$ instead of $V_i$ to determine the onset of deflation because $u_*$ accounts for the influence of atmospheric stability on the surface stress. Making the usual surface layer assumptions, $u_*$ (cm s$^{-1}$) is defined as

$$u_* = V_i \kappa / \left[ \ln(z_i/z_0) - \psi_m(z_i/L) \right]$$  \hspace{1cm} (1)

where $V_i$ is the wind speed at the midpoint $z_i$ of the surface layer, $\kappa$ is the von Karman constant, $z_0$ the surface roughness, $\psi_m$ the stability parameter for momentum, and $L$ the Monin-Obukhov length (e.g., Panofsky and Dutton 1984). The choice of $z_0$ is discussed in KC and is generally 0.01 m for the desert. For neutral conditions, $z_i/L = 0$ and $\psi_m$ is zero. The atmosphere over the desert is usually unstable in daytime. Under these conditions $z/L < 0$ and $\psi_m > 0$ so more momentum is transferred to the ground. The values of $V_i$ from Helgren and Prospero (1987) have been converted to $u_*$ for neutral ($z/L = 0$) and unstable ($z/L = -2$) conditions and are shown in Table...
TABLE 2. Estimates of \( u_a \) for the values of \( V_a \) suggested by Fernandez–Partagas et al. (1986), assuming \( z_a = 0.01 \) m (desert surface).

<table>
<thead>
<tr>
<th>( V_a ) (m s(^{-1}))</th>
<th>Neutral ((z/L = 0))</th>
<th>Unstable ((z/L = -2))</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>29</td>
<td>35</td>
</tr>
<tr>
<td>8</td>
<td>46</td>
<td>55</td>
</tr>
<tr>
<td>11</td>
<td>64</td>
<td>77</td>
</tr>
</tbody>
</table>

2. The corresponding values for unstable conditions range from 25 to 77 cm s\(^{-1}\). Based on the analysis by Helgren and Prospero (1987), measurements by Gillette et al. (1980, 1982), and our own analysis, we have chosen a value of 60 cm s\(^{-1}\) for \( u_a \), for all but one of the simulations. Even though variations in \( u_a \) within the source region are likely, we use 60 cm s\(^{-1}\) everywhere since the available data are insufficient to define a spatially dependent \( u_a \).

The source flux \( F_a \) (g cm\(^{-2}\) s\(^{-1}\)) of particles between 0.1 and 80 \( \mu \)m in radius from erodible desert surfaces was estimated in WTC to be

\[
F_a = 2.9 \times 10^{-14} u_a^4
\]

and is shown in Fig. 7. We distribute \( F_a \) among the model's 30 size bins according to a power law \( dF_a/ d\log r \propto r^{1.5} \). With this distribution, only 4.4% of the lifted dust has radius less than 10 \( \mu \)m. This power law size distribution was chosen in WTC after studying several size distributions measured by Schütz and Jaenicke (1974) during dust storms and has the benefit that any changes to the initial distribution can be easily detected in size distribution plots since the power law represents a straight line when displayed in the usual log–log convention.

Not all of the surface of the Saharan desert is susceptible to deflation or is a good source of transportable dust. The dust source region for the model was determined by combining the observations of dust storms for 23–28 August with the outlines of young dust storms for the entire summer of 1974 (Estoque et al. 1986). Data from more recent years is available but may not apply to 1974 since the record of dust storm occurrence at stations in Africa and at Barbados (Middleton 1985; Prospero and Nees 1986) indicate year-to-year variation in frequency and intensity which may be due to changes in the location of the source region brought about by rainfall variations. The resulting source region is shown in (Fig. 6) and consists of the Spanish Sahara, Mauritania, Northern Mali, and central and Southern Algeria. This area includes the Ahaggar Massif (24°N, 6°E) the flanks of which are considered a primary source of dust (Martin 1975; Estoque et al. 1986). The southern boundary coincides with the location of the ITF in summer. Deflation is less likely south of the ITF because precipitation and vegetation are more common. The surface observations discussed in section 2b support this assumption. Furthermore, precipitation was particularly widespread south of 17°N on 21–22 and 24 August 1974. Niger and Chad have landforms similar to that of Mali and are dust sources in winter (Kalu 1979) but evidently not during GATE because dust was not observed there in the infrared satellite imagery or reported at the surface stations during the period of study. Morocco is excluded since dust was not observed or reported there either. Intercomparison between the surface observations and the plume outlines is desirable but not possible because the two data sources are mutually exclusive: there are no inhabitants where the atmosphere is dry and clear enough to view dust via satellite, and the high water vapor mixing ratios and clouds preclude satellite analysis in inhabited regions. Areas that witnessed an isolated dust event but were bordered by nondust reports [such as at Zinguinchor on 25 August (Fig. 5c)] were not included in the source area.

A variety of surface types are likely to be found within individual grid boxes that lie in the source area. Based on a land type survey by Clements et al. (1957), we estimate that, on average, only 13% of a source area grid box has a deflatable surface. The factor of 0.13 has been included in (2) but was not used in the 2D study (WTC) so \( F_a \) may have been overestimated by a factor of 7.7 in that study. This omission was partly compensated by the low values of \( u_a \) that were predicted by the 2D model.

![Fig. 7. The surface aerosol flux as a function of \( u_a \). The solid line represents the function used in the numerical model, (2). The dashed line is the expected value of the surface flux when the assumed multivariate probability distributions of \( u_a \) and \( u_c \) are integrated over the range of possible values, i.e., (5). Also shown is the ratio of the expected value to the modeled value (use the scale at right). For \( u_a > 40 \) cm s\(^{-1}\), the modeled value is within a factor of 2 of the expected value.](image-url)
We now comment on the applicability of our model to simulations of Saharan dust storms. In particular, we discuss the problem of resolving the small-scale features of dust mobilization with a numerical model that has 220-km grid spacing. The general problem of relating the surface flux of a quantity to the large-scale model variables is not a new one. Indeed, much effort has been spent in deriving suitable parameterizations of the momentum, heat, and moisture fluxes for use in numerical models. Mahrt (1987) has recently highlighted some of the problems associated with subgrid-scale flux parameterizations. Because of the nonlinear response of surface fluxes to surface variables, the flux determined from the average or grid-scale surface variables is not equal to the average flux determined from the local surface variables.

For our application, two important variables display subgrid-scale or subtime-step variations. These are $u_*$, which controls the magnitude of the surface flux, and $u_{*r}$, which controls when dust is lifted. The value of $u_*$ predicted by the model represents the mean value for the entire grid box during the time step. Statistically, the local values of the friction velocity, $u_{*r}$, are distributed about $u_*$ according to some probability density function so that $u_{*r}$ may exceed $u_*$ somewhere in the grid box or during the time step even when $u_* \ll u_{*r}$. This might represent a case of localized dust lifting and could explain the persistent haze and isolated observations of dust storms in the Sahara even when light winds are reported. Often a Rayleigh probability distribution function is assumed to represent the distribution of wind speed and $u_{*r}$ over time and space (Christofferson and Gillette 1987; Pavia and O'Brien 1987):

$$g(u_{*r}; u_*) = \frac{2u_{*r}}{u_*^2} \exp\left[-\left(\frac{u_{*r}}{u_*}\right)^2\right].$$  

(3)

For $u_*$ $>$ $u_{*r}$, the contribution to the surface flux by positive deviations from the mean will outweigh that from negative deviations of the same magnitude because of the power law nature of (2). Hence, it is possible that the flux based on the average $u_*$, i.e., $F_0$, will underestimate the average of the subgrid-scale aerosol flux, $f_0$. If we substitute (3) into (2) and integrate analytically from zero to infinity we find that $f_0[g(u_{*r}; u_*)]$ is exactly twice $F_0$. This integration assumes that $u_{*r} = 0$ and therefore represents the maximum value of $f_0$, for a given $u_*$ and any choice of $u_{*r}$.

Because different surface types are expected within a grid box, it is likely that $u_*$ has some distribution of values within the grid box, but not within a time step. Table 1 of WTC gives some indication of the relative frequency of deflatable surface types and of the range of $u_{*r}$ for those types but is not suitable for determining the distribution of $u_{*r}$. In order to estimate the effects of gustiness and surface variability on deflation we will assume that $u_{*r}$ has uniform probability $h(u_{*r})$ between 25 and 150 cm s$^{-1}$ over 13% of the grid box and 0 for the remainder of the grid box:

$$h(u_{*r}) = \begin{cases} 0.13/(150 - 25), & 25 < u_{*r} < 150 \text{ cm s}^{-1} \\ 0, & \text{otherwise}. \end{cases}$$  

(4)

The expected value of the average of the subgrid-scale surface flux $f_0$ allowing for variations in $u_*$ and $u_{*r}$ is determined by integrating the multivariate probability distribution over the range of values for $u_*$ and $u_{*r}$, i.e.,

$$f_0 = \int_{u_{*}}^{\infty} \int_{u_{*r}}^{\infty} F_0(u_{*r}) g(u_{*r}; u_*) h(u_{*r}) d(u_{*r}) d(u_*) .$$  

(5)

We have calculated $f_0$ for a range of values of $u_*$ and have plotted $F_0$, $f_0$, and the ratio of $f_0$ to $F_0$ vs $u_*$ in Fig. 7. The graph illustrates that the flux $f_0$ is nonzero, even when $u_*$ is less than the lowest value of $u_{*r}$ allowed, or 25 cm s$^{-1}$. This is due to the nonzero probability of all values of $u_{*r}$ when the Rayleigh distribution is used. Nevertheless, the ratio of the fluxes is less than 0.2 for $u_* < 20$ cm s$^{-1}$ because much of the Rayleigh probability distribution is below 25 cm s$^{-1}$. Thus, $F_0$ is not a good approximation of $f_0$ when $u_{*r}$ is small.

We have already shown that for $u_* \gg u_{*r}$, the ratio of $f_0$ to $F_0$ is bounded by 2.0. In Fig. 7 we see that the model flux $F_0$ is within a factor of 2 of $f_0$ when 40 $< u_*$ $< 100$ cm s$^{-1}$. This error is acceptable considering the uncertainties in the source region and the source function. We conclude that for large $u_*$, the dust storm becomes grid scale, and the model and (2) are reasonable approximations of Saharan dust storms.

b. Wet removal parameterization

Prodi and Fea (1979) have presented evidence that Saharan dust is removed from the atmosphere by accumulation in cloud droplets (rainout) or by collision with raindrops (washout). They found significant amounts of Saharan dust 0.1-8 $\mu$m in radius in rainwater samples collected in Italy during the passage of a Saharan dust outbreak. Thus we have included a wet removal parameterization in our model. Unfortunately, theoretical models of rainout and washout do not completely explain the magnitude or size dependence of the observed scavenging rates but an estimate of the scavenging rate can be made. Since the time scale of the accretion process is considered to be longer than that of scavenging by cloud droplets, the details of the latter process were ignored by Pruppacher and Klett (1978) in deriving the following expression for an upper limit on the scavenging rate $\Lambda_a$ (h$^{-1}$) in and below the cloud that is independent of particle size and droplet size:

$$\Lambda_a = 1.5E R^{0.79},$$  

(6)
where $\dot{E}$ is the collection efficiency ($=0.83$) and $R$ the rainfall rate (millimeters per hour). For large $R$, accretion is more rapid than fallout and Eq. (6) will overestimate $\Lambda$. Under these conditions, Pruppacher and Klett (1978) suggest

$$\Lambda_f = 10.8 R^{0.16} / H$$

where $H$ is the thickness of the cloud or model layer in kilometers. If $H = 1$ km then $\Lambda_a = \Lambda_f = 18.7$ h⁻¹ when $R = 30.5$ mm h⁻¹.

Washout and evaporation of droplets below the cloud are not considered in the model. Fortunately, collisions are unlikely since the ratio of the dry dust radius to raindrop radius is small and evaporation below cloud generally results in smaller droplets, rather than the complete evaporation of a few droplets (Pruppacher and Klett 1978).

Not all aerosol that enters a cloud is removed by rainout or washout. Cumulus clouds have been found to be efficient at "venting" aerosols from the lower troposphere to the upper troposphere. The aerosol is transported vertically in cloud droplets or the interstitial air, and is then detrained at the sides and the top of the cloud. The particles inside the detrained droplets are released when the droplets evaporate outside the cloud. The convective precipitation parameterization described below does not allow venting of dust by individual cumulus clouds since most of the condensed water is immediately removed from the model. The convective precipitation parameterization does allow for vertical redistribution of some of the water that converges in the column (term b in Anthe 1977; see Appendix) but we do not vertically redistribute the aerosol contained in that water.

The rainfall rates used in the scavenging parameterization are calculated by the dynamical model using two different parameterizations. The choice is based on the atmospheric stability. During ascent of saturated air in a conditionally stable column, the "nonconvective" precipitation parameterization is used. Under these conditions, water in excess of saturation is removed from the dynamical model grid element as nonconvective rainfall $R_n$.

When the atmosphere is conditionally unstable, and sufficient low-level moisture convergence exists, the "convective" precipitation parameterization is used in the dynamical model. Unlike nonconvective clouds, cumulus clouds usually cover only a small fractional area of the atmosphere (or model grid box) and often extend through the depth of the troposphere. The Anthe–Kuo (Anthe 1977) cumulus parameterization scheme, with modifications by KC, is used to describe the impact of subgrid-scale cumulus elements on the resolved scale. We assume that the convective rainfall rate $R_c$ is proportional to the condensational heating rate diagnosed by the cumulus parameterization. The precipitation is assumed to fall over a fraction, $a_p = 0.2$, of the grid element at a rate equal to $R_c/a_p$. The value of $a_p$ was based on GATE observations (Burpee and Reed 1982; see Appendix).

The total scavenging rate due to nonconvective and convective precipitation in a grid element is the sum of the scavenging due to $R_n$ over $(1 - a_p)$ of the grid and the scavenging due to $R_n + R_c$ over $a_p$ of the grid,

$$\Lambda = (1 - a_p) \Lambda(R_n) + a_p \Lambda(R_n + R_c/a_p).$$

Here $R_n$ and $R_c$ represent the vertical sum of the non-convective and convective rainfall rates, respectively, for a precipitating model layer and all precipitating layers above. Below the precipitation region, the scavenging rate is $0$. Values of $\Lambda$ for several values of $R_n$ and $R_c$ are shown in Table 3.

4. Dynamical simulations

Karyampudi and Carlson (1988) has already shown that the dynamical model adequately simulates the observed synoptic-scale circulation for the 5-day period. Here we first discuss the large-scale aspects briefly and then focus on the dynamics behind occurrence of high $\mu_\ast$.

The strength of the first and second waves and the shallowness of the second wave are accurately simulated by the model. On a day-to-day basis, the simulated wave might be ahead or behind the observed wave location by up to 6° longitude but some of the difference in locations may be due to errors in the analyzed wave location. Comparing Fig. 8 with Fig. 2b shows that at the end of the simulation, the location of the three wave axes is in agreement with the analyses and the model simulates the large areas of precipitation associated with each wave. Smaller areas of precipitation, such as those at Dakar on 28 August (Fig. 5f), are either not simulated or are displaced by several degrees.

The essential features of the SAL and marine layer at Sal Island, such as the wind fields and the timing of the passage of the trough axis and the arrival of the SAL, are captured by the model (Fig. 3). The SAL was present in the Dakar radiosonde data only on 26 August, but the simulation shows the characteristic features of the SAL at Dakar from 26 August until the end of the simulation (Fig. 4). The simulated winds between 900 and 750 mb have a larger northerly wind component than the observed winds which leads to less moisture and rain at Dakar than was observed.

<table>
<thead>
<tr>
<th>$R_n$</th>
<th>$\Lambda(R_n)$</th>
<th>$R_c$</th>
<th>$a_p \Lambda(R_c/a_p)$</th>
<th>$\Lambda(R_n + R_c)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1.2</td>
<td>1</td>
<td>0.9</td>
<td>2.0</td>
</tr>
<tr>
<td>10</td>
<td>7.7</td>
<td>10</td>
<td>4.0</td>
<td>10.3</td>
</tr>
<tr>
<td>30.5</td>
<td>18.7</td>
<td>30.5</td>
<td>4.8</td>
<td>19.9</td>
</tr>
</tbody>
</table>

Table 3. Scavenging rates $\Lambda$ (h⁻¹) for values of nonconvective $R_n$ and convective $R_c$ rainfall rates (mm h⁻¹) for a 1 km thick water cloud.
During the 5 days of the simulation, $u_*$ exceeds $u_{*e}$ on a number of occasions and at several locations within the source area. To summarize the behavior of $u_*$ in the simulation, we have averaged $u_*$ over the dust source area (Fig. 6) and plotted the average against time in Fig. 9. The diurnal fluctuation of the average $u_*$ is readily apparent in Fig. 9 with high values beginning at 0800 UTC and ending at 1900 UTC each day. This is due to erosion of the nocturnal inversion each morning and the entrainment of momentum by the growing boundary layer. At first it appears that $u_*$ never exceeds $u_{*e}$ (60 cm s$^{-1}$) but recall that the value of $u_*$ in Fig. 9 is the average over the entire source area. On each day, $u_*$ reaches its maximum value in the morning before the boundary layer attains its greatest depth. This conflicts with the usual assumption that the flux of momentum from aloft to the surface is greatest in the afternoon when the boundary layer is deepest. Of the six diurnal cycles shown, $u_*$ is greatest on 25 and 26 August. It is shown below that the episodes of high $u_*$ on these days are directly related to the second easterly wave. In contrast, the 2D model of WTC, which did not resolve easterly waves, predicted only a monotonic increase in maximum daily values of $u_*$ as the PBL grew deeper each day.

It is impractical to study all of the individual mobilization events simulated by the 3D model so we will study only the major deflation events of 25 and 26 August (Figs. 5c, d). These two events are due to different dynamical forcing mechanisms and are discussed separately below.

**a. Dynamics of deflation in the central Sahara on 25 August**

The simulated fields of $u_*$ for 0600, 0900, and 1200 UTC 25 August are shown in Figs. 10a, 11a, and 12a, respectively. At 0600 UTC, the surface wind speed and $u_*$ are low except south of the ITF where the monsoon is strong throughout the night. The small local maxima at 20°N, 3°E has developed since sunset the previous day. By 0900 UTC, the sun has risen and $u_*$ has increased at all locations over Africa. The area of high $u_*$ at 20°N, 3°E has expanded greatly in size and values as high as 90 cm s$^{-1}$ are reached. This episode of strong surface stress is responsible for the high value of average

![Fig. 8. Simulated 700 mb wind field (thin solid lines, contour interval of 5 m s$^{-1}$) and precipitation rates (heavy dashed lines, contour interval of 1 mm h$^{-1}$) for 1200 UTC 28 August. Compass wind vectors indicate direction and magnitude of wind. In this and all other figures, an upward-pointing arrow represents a southerly wind. Use vector at upper right corner for scale. The easterly wave axes are shown with bold dash-dot lines. Compare with the analyses in Figs. 2b and 5f.](image)

![Fig. 9. The time dependence of simulated $u_*$ averaged over the source area shown in Fig. 6.](image)
\( u_0 \) in Fig. 9 at 0900 UTC 25 August. The episode is short-lived, however, and by 1200 UTC is reduced considerably in size and magnitude with the maximum \( u_0 \) being just over 60 cm s\(^{-1}\). The same scenario is repeated on a smaller scale before local noon in other locations on other days during the simulation so the dynamical forcing mechanism warrants closer inspection.

The zonal cross section of potential temperature \( \theta \) and wind speed and direction along 21°N at 0600 UTC 25 August (Fig. 10b) reveals two shallow, low-level jets: an easterly jet in the central Sahara between 0° and 25°E with wind speeds in excess of 15 m s\(^{-1}\) that develops after sunset the previous day, and a northerly jet along the coast between 12° and 8°W with weaker winds. The coastal jet is stronger on 26 August so discussion of the dynamics of coastal deflation is deferred until subsection b. Here we concentrate on the eastern or “central Saharan jet.”

The closely spaced horizontal isentropes near the surface in Fig. 10b indicate that a nocturnal inversion is present at 0600 UTC. The small local maximum of
in Fig. 10a is due to mechanical turbulence beneath the central Saharan jet. As the sun rises and heats the surface, the nocturnal inversion is eliminated and buoyancy-generated turbulence mixes the lowest kilometer of the atmosphere (Fig. 11b). The surface wind speed beneath the central Saharan jet at 0900 UTC exceeds 10 m s⁻¹ and \( u_\eta \) exceeds 90 cm s⁻¹ (Fig. 11a). The momentum of the jet is not limitless, however, and by 1200 UTC the surface wind speeds are only 5 m s⁻¹ and \( u_\eta \) is just over 60 cm s⁻¹ (Fig. 12). Note in Fig. 10b that the midlevel winds over the desert are generally weak so there is little momentum available from aloft for deflation at the surface.

A detailed analysis of the simulation revealed that the central Saharan jet of 25 August forms as the zonal wind accelerates in response to a localized 4-mb fall in surface pressure at 18°N, 1°E during the first 24 h of the simulation. In a simulation by KC in which sensible heating of the atmosphere was not allowed, the surface pressure falls over the entire desert during the first 24 h were greatly reduced but were still largest (1 mb) at 18°N, 1°E. This indicates that the surface pressure falls over the desert are, in general, due to sensible heating of the atmosphere and that the second easterly wave adds an additional perturbation to the surface pressure at 18°N, 1°E. In their simulation the maximum wind
speed in the low-level jet was only 11 m s\(^{-1}\), and \(u_\ast\) never exceeded \(u_{\ast 1}\) even after sunrise.

\section*{b. Dynamics of deflation along the coast on 26 August}

We now analyze the significant deflation event of 26 August that occurred over the Spanish Sahara and western Mauritania (Fig. 5d). The two low-level jets present at 0600 UTC 25 August (Fig. 10b) recur on 26 August (Fig. 13a). The eastern jet has reformed during the night farther westward at the same longitude as the T-2 axis, which now is at 7\(^\circ\)W, but the eastern jet is weaker than on 25 August and quickly loses momentum to the surface and boundary layer once the sun rises.

The coastal jet is stronger on 26 than 25 August because the approaching easterly wave strengthens the existing land–ocean pressure gradient. In contrast to the central Saharan jet, the coastal jet remains strong throughout the day because the dynamical forcing (i.e., the pressure gradient) remains strong and perhaps is even strengthened during the daytime by heating over the desert. This is supported by the dynamical simulation in KC in which sensible heating was not allowed. In that simulation, the coastal jet was 6 m s\(^{-1}\) weaker at 1200 UTC 26 August than in the case with sensible heating.
The model simulation (Fig. 13a) shows that before sunrise on 26 August, the surface wind speeds along the coast are low, but as the sun rises, sensible heating destabilizes the trade wind flow over land and momentum is mixed in a shallow boundary layer. The friction velocity at 1200 UTC exceeds 60 cm s$^{-1}$ over an area 2200 km by 400 km and exceeds 80 cm s$^{-1}$ over smaller areas until sunset (Fig. 14).

The coastal low-level jet persists after sunset but $u_*$ falls below $u_{0f}$ as the formation of the nocturnal inversion terminates the vertical coupling to the jet. The same deflation process does not repeat on 27 or 28
August because T-2 moves westward over the coast, weakening the pressure gradient at the coast. On 27 and 28 August, lighter winds prevail over the Spanish Sahara and western Mauritania and the strongest winds are found off the coast, in agreement with analyses by Sadler and Oda (1978).

c. The role of the MLEJ in deflation

In the conceptual model, strong surface winds and deflation are attributed to dry convective mixing of the momentum in the MLEJ, but in the observations and our simulation of 25 and 26 August 1974 it appears that the MLEJ does not play a role in deflation. What then is the relationship between deflation and the MLEJ? The jet forms along the boundary between the hot SAL and the cooler air south of the ITF. The existence of the jet implies that there is strong vertical shear of the thermal wind and that the atmosphere directly below the jet is more stable. Under these conditions, the boundary layer does not grow as deep and vertical mixing of momentum beneath the jet is reduced. As an example, we show in Fig. 15 the meridional cross section of potential temperature and wind through the MLEJ (5°W) at 1200 UTC 28 August. The atmospheric stability beneath the MLEJ suppresses dry convective mixing of momentum so the surface winds are weak and deflation unlikely. The cool monsoonal air is particularly far north at this longitude because this is the area of southerly winds on the east side of the vortex of T-3 (Fig. 8). Several hundred kilometers to the west, north, and east of this tongue of cool air, the boundary layer extends to 600 mb and mixing is more efficient, but the vertical shear of thermal wind, and therefore the MLEJ, are weaker in those locations so less momentum is available. Hence, the average $u_8$ in the source area on 28 August is less than on 25 or 26 August (Fig. 9) and, as we shall see, the third easterly wave mobilizes less dust than the second wave (Table 4).

Deflation seems to be related to the MLEJ at other times during the summer. Satellite imagery reveal instances when mobilization begins in the afternoon and...
Table 4. Dust budget for model simulations, in M: M, mobilized during period; R, remaining in suspension at end of period; and S, scavenged during period.

<table>
<thead>
<tr>
<th>Model time period</th>
<th>Central Saharan* plume</th>
<th>Coastal plume</th>
<th>Combined simulation</th>
<th>Low scavenging</th>
<th>( \nu_{st} = 40 \text{ cm s}^{-1} )</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>M</td>
<td>R</td>
<td>S</td>
<td>M</td>
<td>R</td>
</tr>
<tr>
<td>1200–2400 UTC 23 August</td>
<td>0.39</td>
<td>0.48</td>
<td>0.00</td>
<td>0.32</td>
<td>0.01</td>
</tr>
<tr>
<td>0000–2400 UTC 24 August</td>
<td>0.23</td>
<td>0.04</td>
<td>0.12</td>
<td>0.08</td>
<td>0.00</td>
</tr>
<tr>
<td>0000–1200 UTC 28 August</td>
<td>0.20</td>
<td>0.02</td>
<td>0.10</td>
<td>0.39</td>
<td>0.07</td>
</tr>
</tbody>
</table>

* Mobilization not allowed in source area west of 2°W.

when daytime transport of dust is very rapid (>20 m s\(^{-1}\)). These events are most easily explained by dry convective mixing of the MLEJ. However, our simulations and the satellite imagery do not show evidence of either of these phenomena during the period of study.

5. Dust plume simulations

We now use the aerosol model to simulate the deflation events of 23–28 August. We investigate, individually, the central Saharan dust plume of 25 August and the coastal deflation event of 26 August in two separate simulations that begin at 0000 UTC on 25 August and 0000 UTC on 26 August. Both simulations end at 1200 UTC 28 August. In order to better isolate the dust plumes, deflation is allowed only during the first 24 h in each of these simulations. Furthermore, deflation is not allowed west of 2°W (i.e., along the coast) in the central Saharan simulation. In a third simulation, beginning at 1200 UTC 23 August, deflation is allowed everywhere in the source region on all days. At the end of the section we discuss several additional simulations that are used to investigate various physical processes.

a. The central Saharan plume of 25 August

The deflation event in the central Sahara on this day is intense, with \( \nu_{st} \) in excess of 90 cm s\(^{-1}\), and short-lived, lasting from 0700 to 1200 UTC. The size distribution by mass during mobilization (Fig. 16) is similar to the 2D results presented in WTC with the original power law distribution below 1 \( \mu \)m and a flattening above 20 \( \mu \)m due to sedimentation. The deviation from the power law between 1 and 10 \( \mu \)m is due to preferential removal of these sizes by dry deposition and is greater than in the 2D simulations because \( \nu_{st} \) and therefore the deposition velocity are higher in the 3D simulations. The size distributions simulated by the 2D model were compared extensively with the available measurements in WTC. In Fig. 16 we show the dust storm size distribution measured at Camp Derj, Libya by Schütz and Jaenicke (1974) upon which our source size distribution is based. Note that the simulated distribution resembles the measured distribution at radii below 30 \( \mu \)m. The difference between measured and simulated size distributions above 30 \( \mu \)m has been at-
tributed to some subgrid-scale deflation mechanism not included in the model whereby ultragiant particles are preferentially lifted and suspended in the atmosphere (Bagnold 1941; Gillette and Walker 1977).

Early in the day when the deflation event begins, the boundary layer is shallow and the aerosol is confined to a thin layer. But by the end of the diurnal heating cycle, the boundary layer and dust extend to 600 mb (Fig. 17). Of the 39 Mt of dust lifted during the 5 h that $u_h$ exceeds $u_{cr}$, only 0.5 Mt remain at 0000 UTC 26 August (Table 4). The ultragiant particles in suspension during mobilization fall out quickly once mobilization ceases leaving the size distribution steep-sided above 10 $\mu$m (Fig. 16). The meridional cross section of aerosol concentration at 1200 UTC 26 August along 1°E (Fig. 18) shows a deep layer of dust north of 21°N, while south of 20°N, dust is found in the elevated layer that is formed when the SAL overrides the monsoon.

The simulated winds in the middle and lower troposphere over the desert on 25 August (Fig. 11b) are 7–8 m s$^{-1}$, on average, or roughly 6° longitude per day. At this speed, the dust mobilized on 25 August at the Greenwich meridian takes 3 days to reach the coast, in agreement with the slow movement of dust plumes seen in the satellite imagery for the period. We mentioned earlier that dust plumes on other days of GATE have been observed to travel from Greenwich meridian to the coast in less than 1 day, or at an average speed of over 22 m s$^{-1}$. The synoptic situation for those cases must be quite different from this period.

The distribution of optical depth at 1200 UTC 28 August (Fig. 19) shows how the dust plume has stretched meridionally. The northern extension of the plume is dominated by the circulation about the Northern Sahara anticyclone while the southward transport of dust is due to cyclonic circulation ahead of the developing third wave which follows the plume across the Sahara (Fig. 8). The optical depth at the center of the plume is only 0.05 and must be considered quite small when compared to other dust outbreaks during GATE. Some of those had optical depths greater than unity (Carlson 1979). The dust front reaches the coast at levels aloft but not at the surface at Sal Island and Dakar. Thus, this slow-moving plume with origins far from the coast does not contribute significantly to the aerosol measured at Sal Island by Prospero et al. (1976).
The location of the area of high $u_*$ in the central Sahara on 25 August (Fig. 11a) leading to the dust plume described above is 6 deg longitude east of the observed dust plume of 25 August shown in Fig. 5c. However, the simulated position of the T-2 wave axis was also 6 deg longitude east of the analyzed wave so the simulated wave and area of high $u_*$ had the same relative spatial relationship as did the observed wave and dust plume. Moreover, the observed plume appeared just after sunrise as did the simulated area of high $u_*$. We suggest that the mobilization mechanism described in section 4a is responsible for the deflation event of 25 August. We and D. Helgren (personal communication 1986) have often seen plumes in satellite imagery originating before noon so deflation by low-level jets may not be unique to 25 August.

b. The coastal dust plume of 26 August

As we noted previously, the 26 August coastal deflation event (Fig. 14) has peak values of $u_*$ lower than those of the 25 August central Saharan episode, but the coastal episode covers a larger area and lasts longer, and 242 Mt of dust are lifted. The zonal cross section of mass concentration at 1200 UTC 26 August along 21°N (Fig. 20) shows that the boundary layer west of 8°W is shallow and the mobilized dust is concentrated in the lowest 100 mb. The maximum concentration exceeds 20 000 $\mu$g m$^{-3}$. A weak central Saharan deflation event, similar in nature to that of 25 August, occurs east of 7°W where the PBL is deeper. This event makes only a minor contribution to the 242 Mt of dust lifted that day.

The distribution of optical depth at the final model time (Fig. 21) shows a plume of dust nearly 1000 km wide extending past Sal Island to 28°W with a shape similar to that shown in our analysis for 1200 UTC 28 August (Fig. 5f). Since coastal deflation took place within the northeast trade winds, the dust leaves the continent not as an elevated layer of air as in the conceptual model, but instead as a layer of air in contact with the ocean. The vertical distribution and size distribution of dust in the plume over the ocean will be discussed in a later section.

c. The combined dust plume

We now study the simulated dust plume made up of all of the individual sources on all days and compare the simulation with our analyses. Hereafter this simulation will be referred to as the "combined simulation." Table 4 shows how much dust is lifted on each
day and how much is still in suspension at the end of each day. (The value of 118 Mt for 25 August does not agree with the value of 39 Mt given in the earlier discussion of the 25 August dust plume because deflation was restricted to the area east of 2°W in the earlier simulation.) Only 1.9 Mt remain in suspension at 1200 UTC 28 August, the final model time, since 95.6% of the mobilized dust has radii greater than 10 μm and is subject to rapid depletion by sedimentation. This value is low but within the range of values estimated for other outbreaks during GATE (Carlson 1979). The aerosol measurements at Sal Island (Fig. 1) and the satellite imagery support the idea that this storm was weak, relative to others during GATE.

In Table 4 we see a distinct peak in deflation on 25 and 26 August that is preceded and followed by weaker deflation. Thus, temporal variations in deflation may be responsible for the periodic outbreaks of dust from Africa. Other possible factors, such as transport and scavenging, will be discussed later in this section. The spatial distribution of the 413 Mt of dust lifted during the 5-day period is shown in Fig. 22. The major source areas are Mali, Mauritania, the Spanish Sahara, and the area southwest of the Ahaggar Massif. In general, dust that is mobilized west of 10°W is lifted throughout the day by the northeast trade winds according to the mobilization scenario for the coastal jet. Dust mobilized east of 10°W is lifted by low-level jets before noon according to the mobilization scenario for the central Saharan jet.

The outlines of the combined simulation plume at 1200 UTC on each day are superimposed on the analyses in Figs. 5b–f. On 24 August, there is dust in suspension over the ocean in the analysis but not in the simulation, probably because this dust was in suspension at the initial time whereas the model was initialized with no dust in suspension. On subsequent days, the analyzed plume near the coast represents a fresh outbreak and there is good agreement between the shape of the simulated plume and the shape of the analyzed plume, especially over the ocean on the last three days of the period. Both the simulated and analyzed plumes exhibit anticyclonic curvature and a general southward shift over the ocean.

Dust could not be seen in the satellite imagery south of 17°N because of the high water vapor content so comparison between the southern edge of the simulated plume and that of the analyzed plume is not possible. Dust that is south of 17°N may not be reported by surface stations because the dust is above the monsoon (e.g., Fig. 18) and does not directly influence the surface weather. The simulated plume over Morocco and western Algeria extends farther north than in the analysis of the satellite imagery.

The optical depth in the combined plume at the final time is shown in Fig. 23. We have not determined the actual optical depth for 28 August by analyzing the satellite imagery, as Carlson (1979) has done for other cases, but based on our experience the simulated optical depth of 1.0 at 1200 UTC 28 August is reasonable.

In WTC we discussed how during mobilization the aerosol size distribution and the soil size distribution at the location of the aerosol measurement may differ because of size selective lifting and removal processes, mixing of aerosols of different age or origin, and the dependence of the degree of agglomeration on the magnitude of \( \nu_\ast \). The simulated aerosol size distribution in Fig. 16 shows how dry deposition (a size-dependent process) modifies the aerosol size distribution near the ground by preferentially removing particles 1 to 10 μm in radius.

An example of the mixing process is presented in Fig. 24 where we show the surface layer size distribution at 21°N, 15°W at 3-h intervals beginning at 0900 UTC 27 August. At 0900 UTC, the aerosol distribution is dominated by a mode at 1.2 μm with a lesser mode at 6.5 μm. At 1200 UTC, dust is being mobilized within the preexisting aged cloud. By 1500 UTC, deflation has ceased, the giant and ultragiant particles have fallen out, and a bimodal size distribution has been generated. This occurs even though the new and aged aerosols have the same power law source size distribution. Thus, aerosol size distributions measured near the source, such as those discussed in WTC, will be complex. The effect of this process on the size distribution of dust is less noticeable far from the source region since the giant mode is eliminated in a short time by sedimentation.

d. Deflation events not simulated by the model

During the period of study, there were several dust plumes that were not simulated by the model, such as
the dust plume that appeared north of the second easterly wave at 22°N, 2°E on 24 August (Fig. 5b). This plume had sharp boundaries in the satellite imagery and therefore was probably lifted that day and was not merely the residue of a plume from the previous day. The degree of similarity between the plumes of 24 and 25 August (both appeared before noon and both had cyclonic curvature) suggests that the mobilization mechanisms may have been similar. The model predicted low values of $u_\ast$ (25 cm s$^{-1}$) in this area on 24 August, however. The actual wind speeds are not known since there were no meteorological reports at the site of the 24 August plume. No explanation could be found for the discrepancy between the model and the satellite imagery.

The observed winds in central Algeria north of 30°N on 24 and 26–28 August are light, yet suspended (but not raised) dust is reported there on these days (Figs. 5b, d–f). It is possible that the surface station reports of suspended dust referred to dust of nonlocal origin. Indeed, the final distribution of dust in the combined plume simulation (Fig. 23) shows that by the final model time, dust has been transported to central Algeria from other areas by the anticyclonic flow around the North African high.

It is also possible that the dust was raised near the station. During the simulation, $u_\ast$ in the area was as high as 50 cm s$^{-1}$ on 28 August but never exceeded $u_\ast$. However, the threshold friction velocity may be less than 60 cm s$^{-1}$ in this area. From Herghren and Prospero (1987), we estimate that the average threshold wind speed for three stations near 30°N, 0°E was 6.5 m s$^{-1}$ which corresponds approximately to $u_\ast = 45$ cm s$^{-1}$ (Table 2).

The model is sensitive to the choice of $u_\ast$. In a separate simulation with $u_\ast = 40$ cm s$^{-1}$, twice as much dust was mobilized and nearly three times as much dust remained in suspension at the final model time (Table 4). The large area for which $40 < u_\ast < 60$ cm

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**Fig. 23.** Distribution of optical depth at 0.55 μm at end of the combined simulation. Symbols show locations of vertical profiles of mass concentration presented in Fig. 27.

**Fig. 24.** Simulated surface layer size distributions at 21°N, 15°W showing development of a bimodal size distribution. Distributions are from before (0900 UTC), during (1200 UTC), and after (1500 UTC) a coastal mobilization event on 27 August.
s\(^{-1}\) (e.g., Figs. 10a and 14) more than compensates for the power law dependence of \(F_0\) on \(u_0\). Nevertheless, we consider 60 cm s\(^{-1}\) a reasonable choice for \(u_0\) until better data become available and more detailed simulations are carried out. Note that the maximum amount of dust is still lifted on 25 and 26 August.

e. Comparison with Sal Island and Dakar aerosol measurements

In Figs. 3 and 4 we show the observed and simulated aerosol concentrations at Sal Island and Dakar. The measurements at Sal Island show that the concentration increases after 26 August. The simulated concentrations also increase after 1200 UTC 26 August, reaching 115 \(\mu g\) m\(^{-3}\) on 27–28 August, or nearly three times the measured value. The model was not run past 1200 UTC 28 August so it is not known whether the model would have shown higher concentrations on 28–29 August, or lower concentrations as were observed (Fig. 1).

The high simulated Sal Island aerosol concentrations may be due to overestimating the source strength, either because the predicted values of \(u_w\) are too high, the choice of \(u_w\) is too low, or the source equation or source region is incorrect. Land-type surveys of the Spanish Sahara and western Mauritania show more sand dunes than for the Sahara as a whole, so the factor of 0.13 used in (2), and hence \(F_0\), may be too large for the coastal source area. The occurrence of mobilization is supported by the observations of mobilized and suspended dust from the coast to the easternmost station at 13°W (Fig. 5d). The simulated deflation area extends further east to 8°W (Fig. 14) but we cannot verify whether deflation took place between 13° and 8°W since there are no stations there and the satellite imagery was inconclusive. A higher value of \(u_w\) would reduce the total surface flux and therefore the concentration at Sal Island.

Owing to the sharp meridional gradient of aerosol in the plume (Figs. 23 and 30), the simulated aerosol concentration at Sal Island is sensitive to small meridional displacements of the dust plume. A 200 km southward shift in the plume location would have reduced the simulated concentration at Sal Island by a factor of two.

The simulated aerosol concentrations for Dakar are much greater than the measured concentrations (Fig. 4). The difference between the measured and simulated Dakar aerosol concentrations may be due to overestimating the source as discussed in the previous section. The fact that the simulation shows the presence of the SAL at Dakar while the observations do not (Fig. 4), indicates that the difference between the measured and simulated aerosol concentrations at Dakar may be due to advection of hot, dry, dusty air above 950 mb directly from the desert to Dakar followed by vertical mixing. At first glance we would expect less aerosol at Dakar in the simulation because the simulated ocean breeze is stronger than observed. However, this air had passed over the coastal dust source region before turning southward over the ocean and finally southeastward to Dakar (Fig. 14).

The discrepancy between observed and simulated aerosol concentrations at Dakar may also be due to differences in the observed and simulated areas of precipitation. The precipitation at Dakar occurred on a small scale (Figs. 5a, d-f) and the model only showed precipitation south of but not at Dakar on these days. The lack of precipitation at Dakar in the simulation is related to the presence of the hot dry SAL. Hence, the cooling and moistening of the SAL and the scavenging of the dust that are required for the simulation to match the observations did not take place in the model.

f. Wet removal of dust

The data in Table 4 show that the ratio of mass of scavenged dust to dust remaining in suspension is small for the first several days of the combined simulation. This is because the dust and precipitation are not collocated early in the simulation. By the end of the simulation, the ratio increases to over 10% as the third wave develops into a large convective system within a region where dust previously existed and more dust is drawn into the system. In another simulation (not shown), we initialized the model with a cloud of dust just behind the first easterly wave similar in shape to the area of low static stability shown in Fig. 2a. Mobilization was not allowed during this 5-day simulation. Since the typical wind speed at levels where the dust is transported are twice the phase speed of easterly waves, the dust was eventually incorporated into the leading wave circulation and scavenged so that the wave acted as a barrier to dust transport. In the combined simulation, the leading edge of dust is within 300 km of the convective region of the first easterly wave at 1200 UTC 28 August, the final model time (Figs. 8 and 23). Thus, we expect an increasing ratio of scavenged dust to suspended dust on subsequent days as dust overtakes the first easterly wave, or is drawn into the third easterly wave. In this manner, scavenging of dust in both the leading and trailing easterly waves may separate the dust mobilized during this period from dust generated on subsequent days and emphasize the periodicity of dust outbreaks.

As an example of the efficiency of scavenging, we show in Figs. 25a, b the meridional cross sections of aerosol concentration along 5°W at 1200 UTC 27 August (before the third easterly wave entered the area), and at 1200 UTC 28 August (after the wave strengthened and moved into the area). The dust south of 20°N between 900 and 600 mb at 1200 UTC 27 August is scavenged completely during the following 24-h period. The 700 mb wind field (Fig. 8) shows a convergent wind field pattern at 13°N, 5°W associated with the
third easterly wave. All dust that enters the convective region of the third wave is subjected to scavenging rates as high as 3 h\(^{-1}\).

The scavenging parameterization suggested by Pruppacher and Klett (1978) represents an upper limit of the scavenging rate. We tested the sensitivity of simulation to \(\Lambda\) by repeating the simulation with \(\Lambda\) reduced by a factor of 10. In the simulation, scavenging still removed 61\% of the mass of dust that was removed in the original case (Table 4), indicating that the time scale for scavenging is less than that for transport of the dust through the precipitating region. This fact reduces the sensitivity of the model to the scavenging parameterization.

An interesting result is obtained when the simulation is repeated without any scavenging. In Fig. 25c we see that without scavenging, dust is transported as high as 250 mb at the core of the third easterly wave by the resolved vertical motions [up to 15 cm s\(^{-1}\) (KC)] that develop in response to the spatial distribution of latent heating in the wave. (The first wave redistributed the aerosol in a similar manner in the simulation with the initial dust cloud that was described above.) The simulated upward vertical velocities in the third wave were remarkable but, as KC notes, not unreasonable for easterly waves over the continent. These vertical motions are larger than the sedimentation velocities of particles 10 \(\mu\)m in radius and transport significant
quantities of these and smaller particles to the upper troposphere (Fig. 26), thereby greatly increasing the lifetime of the dust. Though all of this dust is removed in the simulation when scavenging is allowed, it is possible that a fraction is vented to 250 mb in interstitial air or released from droplets that evaporate after being detrained. Improved cumulus and scavenging parameterizations are required to study this process.

g. The elevated dust layer over the ocean

In the conceptual model, dust mobilized in the central Sahara leaves the continent as an elevated layer (the SAL) above the northeast trade winds. This scenario has been used to explain the layered structure observed during field programs at Sal Island and Barbados (Carlson and Prospero 1972; Prospero and Carlson 1972; Talbot et al. 1986). In our numerical simulation, however, the dust that was lifted in the central Sahara and advected westward in the SAL, namely the 25 August plume, does not reach the coast in significant amounts. Instead, the dust over the ocean at the final model time is comprised mostly of dust that originated west of 8°W on 25 and 26 August and was transported westward below the SAL. Yet an elevated layer of dust still develops in the model over the ocean. This is shown in Fig. 27 where the vertical profiles of mass concentration for the locations indicated in Fig. 23 are presented. The aerosol is well mixed near the coast (19°W). Undercutting does not take place because the trade winds contain significant amounts of dust. Downwind along the axis of the plume, the vertical gradient increases and an elevated dust layer develops

![Graph showing dust distribution](image)

**Fig. 26.** Simulated dust size distributions at 13°N, 5°W (the center of the third easterly wave; see Figs. 28 and 25c) at various levels in the atmosphere the end of the simulation without scavenging. Significant amounts of dust have been lofted to the middle troposphere by resolved vertical motions with little depletion of the 6 to 10 μm fraction. Contrast these size distributions with those from a nonconvective region shown in Fig. 28.

![Graph showing vertical profiles](image)

**Fig. 27.** Vertical profiles of mass concentration for the combined simulation at the locations indicated by symbols in Fig. 23 at the final model time. Profiles show the development of an elevated layer of aerosol. Size distributions at 15°N, 35°W (triangle symbol) are shown in Fig. 28.
as the dust-free trade winds replace the dusty air in the lowest 100 mb. At the westernmost point in Fig. 27 (21°N, 37°W), the ratio of SAL to surface layer mass concentration is greater than twenty. Thus, the elevated layer of aerosol develops in our model simulation in much the same manner as that in the conceptual model except that the elevated mixed layer evolves over the ocean, instead of over the continent.

In WTC we found that the 2D model underestimated the marine layer concentration of particles with radii less than 8 μm either because a background mineral aerosol was not included in the initial conditions or because the model was deficient in mixing dust from the SAL into the marine layer. We also found that increasing the vertical turbulent diffusion of dust from the SAL into the marine layer increased the concentration of particles less than 8 μm and yielded a size distribution that more closely resembled the measurements. The higher diffusion coefficient, however, had the deleterious effect of reducing the vertical gradient of mass concentration to values less than that observed. This prompted us to suggest that the undercutting process was required to keep the overall marine layer concentrations low in spite of increased diffusion.

Yet we see in Fig. 28 that the surface layer size distribution simulated by the 3D model shows an abundance of particles less than 8 μm in radius and is already similar in shape to the measurements of Jaenicke and Schütz (1978), even though no background aerosol was specified and the nominal value of 10^4 cm^2 s^{-1} was used for the minimum vertical diffusion coefficient. The size distributions in Fig. 28 show a mode around 8 μm developing with increasing pressure due to sedimentation of particles from aloft, as did the 2D model but the presence of a second mode at 4 μm at 835 and 1010 mb implies that there is an additional pathway in the 3D model, not present in the 2D model, by which particles less than 6 μm reach the marine layer. Further examination of the simulations has shown that the particles less than 4 μm were mobilized along the coast on 24 August and advected westward in the marine layer while the particles greater than 4 μm were mobilized along the coast on 25 August and reached the marine layer via sedimentation. The situation is one of particles from 25 August falling into a preexisting or “background” cloud of dust from 24 August. This was one of the mechanisms proposed in WTC by which the simulated size distributions would more closely resemble the observations.

Over the ocean, undercutting is so efficient at reducing the marine layer dust concentration in the 3D simulations that a higher value of K_0 is required to reduce the gradient and match the observations, rather than to diffuse large and submicron particles into the marine layer as in the 2D model. In a simulation (not shown) with K_0 = 10^5 cm^2 s^{-1}, the ratio of the mass concentrations is reduced to less than 5. In summary, vertical turbulent diffusion and sedimentation from the SAL and horizontal advection in the marine layer are all important in determining the marine layer aerosol size distribution and concentration. Dust in the marine layer is no doubt comprised of particles that have traveled from different sources along different pathways. This makes it difficult, if not impossible, to interpret isolated size distribution measurements.

6. Summary and discussion

The results of our model simulations support the hypothesis that the periodicity of outbreaks observed at Salt Island is tied fundamentally to the passage of easterly waves. Originally, the MLEJ was thought to be important to deflation but from our study it appears that the mechanism behind the 25 August central Saharan dust storm was a nocturnal low-level jet related to the surface cyclone. The coastal dust storm of 26 August was due to a low-level jet enhanced by the approaching surface cyclone. In neither case, nor at any other time during the simulation, was deflation related to the MLEJ as proposed in the conceptual model. The static stability of the atmosphere in the vicinity of the MLEJ inhibited vertical exchange of momentum between the jet and the surface during the period of study.
Because the midlevel winds over the central Sahara are weak, dust mobilized in the central Sahara makes only a small contribution to the total mass transported across the Atlantic Ocean while the coastal sources are the major contributors. In spite of our findings for this case, the MLEJ may play an important role in deflation and transport at other times, such as 28–29 July 1974 when newly formed dust plumes were observed in satellite imagery traveling at \( \sim 22 \text{ m s}^{-1} \) during the day.

It is interesting that although the second wave had a shallower and weaker circulation than the third easterly wave, it nevertheless was responsible for most of the deflation that occurred during this period. The zonal cross section of the relative vorticity through the second wave at 0600 UTC 25 August (Fig. 29) shows a well-developed vertical couplet in the second easterly wave at 2°E. This wave must be considered quite shallow since typical easterly waves, such as the first and third waves, extend to 250 mb (Burpee and Reed 1982; KC).

The reason for the weakness of the second wave is not known but a mechanism is proposed here. Because of the shallowness of the wave, its circulation weakens during the day on 25 August as dry convection mixes cyclonic low-level winds with the anticyclonic midlevel winds. Hence, there is less meridional transport of heat and moisture and the rate of generation of eddy kinetic energy is reduced from that which might have occurred had dry convection not been so deep. In contrast, the third wave developed further south along the ITF than did the second wave so energy conversion by baroclinic processes may have played a greater role in development of the third wave. Also, the circulation of the third wave extended well above the boundary layer and was not disturbed during each diurnal cycle as was the second wave. The growth mechanisms of the individual waves should be studied in more detail with a higher resolution model.

By the time the dust has been advected out over the Atlantic Ocean, the outbreak is no longer related to the surface cyclone but instead is related to the midlevel anticyclonic circulation found between two easterly waves, as in the conceptual model. However, the conceptual model has dust everywhere in the SAL whereas the model simulation has the dust confined to the southern half of the SAL. This can be seen in the meridional cross section of potential temperature and mass concentration at 25°W at the final model time shown in Fig. 30. The absence of any dust in the SAL north of 16°N may be due in part to the fact that the model was initialized without any background dust.

From our discussion of the elevated dust layer over the ocean, we see that Saharan dust outbreaks are a complicated mixture of dust originating at different source regions on different days and following different pathways. For the 2D simulations in WTC, the situation was much simpler with dust lifted at a single source area on a single day. We conclude that the Saharan aerosol is more complicated than previously thought, and that measurements should be interpreted carefully, especially when concentrations are near the level of the background mineral aerosol concentration. Continuous measurements by a network of samplers are required to fully understand the marine layer aerosol.
ground-based lidar would be particularly useful because it would yield information on the vertical distribution of dust.

Prospero et al. (1970) and Carder et al. (1986) have detected significant concentrations of Saharan dust with radii greater than 10 μm in the vicinity of Barbados. In WTC we noted that these particles were depleted in our 2D model as near to the coast of Africa as Sal Island, even when the particle density was decreased or the vertical turbulence increased. The SAL and marine layer size distributions simulated by the 3D model also show depletion of giant particles in the elevated dust layer and no ultragiant particles (Fig. 28).

It is possible that giant particles are observed over the ocean when the midlevel flow is more zonal and a strong MLEJ is present. This needs to be investigated further with more case studies and numerical simulations. Jaenicke et al. (1971) suggested that the giant particles are generated in situ when cloud droplets evaporate and release the aerosol. Carder and Prospero (personal communication 1986) note, however, that the giant particles observed at Barbados and in the Pacific (from Asia) are often individual grains and not agglomerated. It is also possible that dust contained in the droplets is released at the sides and top of cumulus clouds when detrainment and evaporation occur. This venting process would greatly increase the lifetime of Saharan dust. These mechanisms cannot be investigated with the current model because the pertinent physics are not included.

Perhaps, large-scale upward motions suspend the giant and ultragiant particles. We have shown that the simulated vertical velocities in the center of the third wave are large enough to prolong the lifetime of 10-μm particles that escape scavenging. Vertical motions at the southern edge of a SAL are small (~1 cm s⁻¹, KC), compared to the fall velocity of a 10-μm particle (3 cm s⁻¹), and are not likely to alter the lifetime of a 10-μm particle.

We have not yet investigated the radiative effects of Saharan dust but can make several conjectures regarding the importance of radiative heating of dust. Assuming optical depths of 1 to 3, KC estimated positive perturbation vertical velocities of only 1 to 2 cm s⁻¹ due to radiative heating in the dusty regions. The values of optical depth in this simulation (Fig. 23) are low compared to the values used by KC and, based on the magnitude alone, we would expect smaller perturbations to the circulation by radiative heating of dust for this case. The vertical velocities are, however, proportional to $\nabla z \cdot \vec{Q}$ where $\vec{Q}$ is the heating rate (KC). Since the dust is concentrated at the southern edge of the SAL in a region more narrow than that assumed by KC (Fig. 30), the vertical velocities associated with these aerosol concentrations may be significant. The perturbation to the circulation caused by radiative heating of the dust can eventually be determined by coupling the current model with a radiation model. Such work is now in progress.

During mobilization, the optical depths over the desert are much greater than unity. Under these conditions, the amount of solar radiation reaching the ground will be considerably reduced and the surface temperatures will be depressed. Brinkman and McGregor (1983) have reported temperature deficits of 6°C in Nigeria during wintertime Saharan dust outbreaks. Under these conditions, less free convective mixing and deflation are expected, thus posing a self-regulating mechanism for dust storms. This process will be investigated in future simulations.

The transport of Asian dust to the central North Pacific has received considerable attention in recent years (e.g., Uematsu et al. 1983). The results of this study do not apply directly to Asian dust storms because the generation and transport mechanisms are different from those of Saharan dust storms. Asian dust storms are usually generated in late winter and spring by the passage of a cold front associated with an extratropical cyclone. The dust is mixed upwards by mechanical turbulence behind the front and then by large-scale vertical motions within the extratropical system. The polar front, and therefore the wind speeds, are much stronger than the ITF and MLEJ. More observations and numerical experiments are necessary before we can explain the extraordinary transport of dust from Asia to the central North Pacific. By design, the models described in this paper have been formulated in a general manner so that they could easily be applied to a study of Asian dust storms.

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APPENDIX

Fractional Coverage of Precipitation

The model predicts the convective rainfall rate $R_c$ for the entire grid box. The fraction of the grid box covered by convective precipitation $a_c$ is important to the scavenging parameterization since $R_c$ must be divided by $a_c$ before being raised to the power 0.79 in (6). If $R_c$ is raised to the power 0.79 before dividing by $a_c$, then the scavenging rate will be overestimated by a factor $a_c^{-21}$ which has values of 2.6, 1.9, and 1.3 for $a_c = 1, 5$, and 25%. Here we explain how $a_c$ was chosen.

In Anthes’ (1977) cumulus convection parameterization, the fractional coverage of the area of the updraft $a_u$ is calculated with

$$a_u = R_c / \int_0^1 \dot{r}_{c} \cdot \frac{\partial p^*}{\partial \sigma} \, d \sigma = \frac{R_c}{R_c}$$

(9)

where $R_c$ is proportional to the horizontal water vapor convergence in the lower troposphere and is calculated based on grid point variables. The denominator is the vertical convergence of cloud water in a cloud updraft and is specified equal to 192 g m$^{-2}$ s$^{-1}$ instead of being explicitly calculated with a cloud model. The value was determined by KC from GATE data and the cumulus cloud model of Fritsch and Chappell (1980). In terms of a precipitation rate, $R'_c$ equals 690 mm h$^{-1}$. With such a large value for $R'_c$, the percent area coverage given by (9) is less than 1% for typical values of $R_c$. It is our opinion that $a_u$ should not be used in (8) since it is the area of the updraft only. During GATE the precipitation area associated with easterly waves was generally greater than 20% of the area covered by convection (Burpee and Reed 1982). Rather than calculating $a_u$ with (9) and then multiplying by some factor to estimate $a_p$, we have chosen to specify $a_p$ outright with a constant value of 0.2 since $a_p$ seems to be better known from GATE observations. For convective rainfall the scavenging rate becomes $W = a_p \Lambda (R_c/a_p)$.

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