

E. Nowotnick¹, P. Colarco², A. da Silva³, D. Hlavka⁴, M. McGill⁴

¹Atmospheric and Oceanic Sciences, University of Maryland, College Park, MD, USA

²Code 613.3, NASA Goddard Space Flight Center, Greenbelt, MD, USA

³Code 610.1, NASA Goddard Space Flight Center, Greenbelt, MD, USA

⁴Code 613.1, NASA Goddard Space Flight Center, Greenbelt, MD, USA

Corresponding Author – E. Nowottnick – epnowott@atmos.umd.edu

24 **Abstract**

25

26 Saharan dust was observed over the Caribbean basin during the summer 2007 NASA
27 Tropical Composition, Cloud, and Climate Coupling (TC⁴) field experiment. Airborne
28 Cloud Physics Lidar (CPL) and satellite observations from MODIS suggest a barrier to
29 dust transport across Central America into the eastern Pacific. We use the NASA GEOS-
30 5 atmospheric transport model with online aerosol tracers to perform simulations of the
31 TC⁴ time period in order to understand the nature of this barrier. Our simulations are
32 driven by the Modern Era Retrospective-Analysis for Research and Applications
33 (MERRA) meteorological analyses. We evaluate our baseline simulated dust
34 distributions using MODIS and CALIOP satellite and ground-based AERONET sun
35 photometer observations. GEOS-5 reproduces the observed location, magnitude, and
36 timing of major dust events, but our baseline simulation does not develop as strong a
37 barrier to dust transport across Central America as observations suggest. Analysis of the
38 dust transport dynamics and loss processes suggest that while both mechanisms play a
39 role in defining the dust transport barrier, loss processes by wet removal of dust are about
40 twice as important as transport. Sensitivity analyses with our model showed that the dust
41 barrier would not exist without convective scavenging over the Caribbean. The best
42 agreement between our model and the observations was obtained when dust wet removal
43 was parameterized to be more aggressive, treating the dust as we do hydrophilic aerosols.

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46

47 **1. Introduction**

48

49 During boreal summer, Saharan dust is transported to the Caribbean and northern South
50 America by the prevailing tropical easterly winds [Karyampudi et al., 1999; Carlson and
51 Prospero, 1972]. Mineral dust aerosols influence Earth's radiation budget directly
52 through the scattering and absorption of light [Zhu et al., 2007; Haywood et al., 2003;
53 Sokolik and Toon, 1996] and indirectly by serving as cloud condensation nuclei (CCN)
54 [Rosenfeld et al., 2001] and ice nuclei [DeMott et al., 2003] and so affecting the
55 properties of clouds. Dust aerosols are thought to modulate tropical cyclogenesis over the
56 tropical North Atlantic by modifying wind fields during development and reducing sea
57 surface temperatures through the absorption of short wave radiation [Lau and Kim, 2007;
58 Dunion and Velden, 2004]. Additionally, insoluble iron in dust aerosols can be converted
59 into a soluble form via photochemistry and cloud processing [Hand et al., 2004; Kieber et
60 al., 2003; Desbouefs et al., 2001; Zhu et al., 1997], which when deposited at the Earth's
61 surface can serve as a nutrient source for aquatic and terrestrial ecosystems [Mahowald et
62 al., 2005; Jickells et al., 2005; Falkowski et al., 2003].

63

64 An apparent barrier to dust transport from the Caribbean into the eastern Pacific was
65 suggested by aircraft observations made during July-August, 2007, NASA Tropical
66 Composition Cloud and Climate Coupling (TC⁴) field campaign [Toon et al., 2010]. We
67 have identified this Central American dust barrier as a persistent feature in satellite
68 imagery during the boreal summer. While other studies have focused on the broader
69 transport and deposition of dust in the Caribbean [Kaufman et al., 2005; Mahowald et al.,

1999, Tegen and Fung, 1995; Duce et al., 1991], we are not aware of any studies identifying this barrier or its causes. We find this barrier is also present in chemical transport model simulations of Saharan dust transport, but that the ability of the model to reproduce the observations is sensitive to the treatment of dust loss processes.

In this paper we explore the controls on establishing and maintaining the observed Central American dust transport barrier, in particular exploring the relative roles of atmospheric dynamics and dust removal processes. We describe our aerosol transport model in Section 2. We present the Central American dust barrier and evaluate our simulated dust distributions using satellite observations from the Moderate Resolution Imaging Spectroradiometer (MODIS) and the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP), airborne observations from the Cloud Physics Lidar (CPL), and ground-based observations from the Aerosol Robotic Network (AERONET) (Section 3). We then explore the cause of the Central American dust barrier by analyzing the dynamical and loss transport pathways of the dust in this region (Section 4). We additionally explore the sensitivity of our analyses to uncertainties in our parameterization of dust loss through wet processes (Section 5). We discuss our conclusions in Section 6.

2. Model Description

Our aerosol transport model is based on the Goddard Earth Observing System (GEOS-5) model, the latest version of the NASA Global Modeling and Assimilation Office

93 (GMAO) earth system model. GEOS-5 contains components for atmospheric circulation
94 and composition (including atmospheric data assimilation), ocean circulation and
95 biogeochemistry, and land surface processes. Components and individual
96 parameterizations within components are coupled under the Earth System Modeling
97 Framework (ESMF) [Hill et al. 2004]. The GEOS-5 earth system model serves as a state-
98 of-the-art modeling tool for studying climate variability and change, and provides
99 research quality reanalyses for use by NASA instrument teams and the scientific
100 community. In addition to traditional meteorological parameters (winds, temperatures,
101 etc.) [Rienecker et al. 2008], GEOS-5 includes modules representing the atmospheric
102 composition, notably aerosols [Colarco et al. 2010] and tropospheric/stratospheric
103 chemical constituents [Pawson et al. 2008], and includes the impact of these constituents
104 on radiative processes within the atmosphere.

105

106 GEOS-5 has the capability to run at various horizontal spatial resolutions, from $4^\circ \times 5^\circ$
107 latitude by longitude for long climate integrations to $\sim 3 \times 3 \text{ km}^2$ using advanced
108 dynamical cores. The version we use here is run at $0.5^\circ \times 0.625^\circ$ latitude by longitude, to
109 match the spatial resolution of the meteorological analyses used to drive our simulations.
110 The model has 72 vertical layers distributed in a hybrid coordinate system that is terrain
111 following near the surface and transforms to pressure coordinates near 180 hPa, with a
112 model top at about 85 km.

113

114 GEOS-5 can be run as a climate model or in a data assimilation stream. Here, we exploit
115 the GEOS-5 capability to “replay” from a prior data assimilation run. This functions as a

116 data assimilation run in that the model makes a forecast to the analysis time (typically
117 every six hours), however, rather than performing the data assimilation step at that point,
118 the model dynamical state (winds, pressure, temperature, and specific humidity) is simply
119 replaced by fields from the assimilation data set. In our case we use fields from the
120 Modern Era Retrospective Analysis for Research and Applications (MERRA) [Rienecker
121 et al., 2011] analysis, available every six hours at a spatial resolution of $0.5^\circ \times 0.625^\circ$
122 latitude by longitude.

123

124 The aerosol module in GEOS-5 is based on the Goddard Chemistry, Aerosol, Radiation,
125 and Transport (GOCART) model [Chin et al. 2002], as previously integrated into an
126 earlier version of the GEOS model framework [Colarco et al., 2010]. GOCART provides
127 a treatment of five tropospheric aerosol species (dust, sea salt, black carbon, organic
128 carbon, and sulfate), including their sources, sinks, and chemistry. Our treatment of dust
129 follows from GOCART and the description given in Nowottnick et al. [2010]. The dust
130 size distribution is partitioned into five non-interacting size bins spaced between 0.1 and
131 10 μm radius. Dust mobilization follows from Ginoux et al. [2001] with sources
132 preferentially located in large-scale topographic depressions (see also Prospero et al
133 [2002]). Dust losses are through dry and wet removal processes, including turbulent dry
134 deposition, sedimentation, and wet removal by large-scale and convective cloud systems.
135 Further details of our treatment of dust, including dust optics, are provided in Nowottnick
136 et al. [2010] and Colarco et al. [2010].

137

138 **3. Evidence for the Central American dust barrier and model evaluation**

139

140 To evaluate Saharan dust transport to the Caribbean and understand the Central American
141 dust barrier we performed a baseline GEOS-5 replay simulation using the MERRA
142 analyses. We simulate all aerosol types with radiative feedback to represent the effect of
143 aerosol absorption and scattering (direct effect) on the atmosphere. After 75 days of
144 model spin-up, we conduct our simulation from June 15, 2010 through August 31, 2010.

145

146 **3.1 Data Sources**

147

148 In this section we introduce the observational data sources that show evidence of the
149 Central American dust barrier and which we use to evaluate dust transport in GEOS-5
150 during TC⁴.

151

152 **3.1.1 MODIS**

153

154 The Moderate Resolution Imaging Spectroradiometer (MODIS) was launched on
155 December 12, 1999 aboard the Terra spacecraft. A second MODIS instrument was
156 launched on the Aqua satellite as a part of the NASA A-Train on May 4, 2002. The
157 MODIS instruments provide multispectral observations of the Earth system using 36
158 channels at 10:30 AM (Terra) and 1:30 PM (Aqua) local time. MODIS aerosol retrievals
159 are made at a spatial resolution of at $10 \times 10 \text{ km}^2$ using separate retrieval algorithms for
160 ocean and land. Over oceans, the MODIS algorithm uses retrieved radiances from six
161 channels (550, 660, 870, 1240, 1630, and 2130 nm) to provide aerosol information at

162 seven wavelengths, using the six retrieved channels and an additional fitted wavelength at
 163 470 nm [Remer et al., 2005]. Over land, an empirical relationship between radiance
 164 retrievals at two visible channels (470 and 660 nm) and one near-IR channel (2130 nm) is
 165 used to determine the surface reflectivity to provide aerosols properties at 470, 550, and
 166 660 nm [Remer et al., 2005]. For our analysis, we use MODIS aerosol optical thickness
 167 (AOT) observations at 550 nm from collection 5.1. MODIS provides semi-quantitative
 168 quality assurance (QA) flags, where QA ranges in integer from QA=0 (low confidence in
 169 aerosol retrieval) to QA=3 (high confidence in retrieval). Over land we aggregate only
 170 highest quality (QA=3) retrievals, whereas over ocean we aggregate all retrievals but
 171 weight them by their respective QA flag value, similar to the MODIS canonical Level 3
 172 gridded product [Levy et al., 2009].

173

174 **3.1.2 AERONET**

175

176 The Aerosol Robotic Network (AERONET) of ground-based sunphotometers provide
 177 measurements of direct solar beam extinction every 15 minutes at 340, 380, 440, 500,
 178 670, 870, and 1020 nm to provide AOT measurements at 440, 670, 870, and 1020 nm
 179 with an accuracy of +/-0.015 [Holben et al., 2001]. AERONET utilizes principle plane
 180 and almuncantar scans to invert aerosol properties and to determine size information
 181 [Dubovik and King, 2000]. To determine the AERONET AOT at 550 nm for comparison
 182 to our model, we first determine the 470-870 nm Angstrom parameter α , defined:

$$183 \quad \tau_1 = \tau_2 \left(\frac{\lambda_1}{\lambda_2} \right)^{-\alpha} \quad (1)$$

184 where τ_1 and τ_2 are AERONET AOT values at $\lambda_1 = 470$ nm and $\lambda_2 = 870$ nm,
185 respectively. Once the Angstrom parameter is determined, we use Equation 1 to
186 determine τ at $\lambda = 550$ nm. For evaluation of our model, we use AERONET version 2,
187 Level 2 cloud-screened and quality-assured daily averaged AOT values [Smirnov et al.,
188 2000] at AERONET sites that are near and downwind of the source region (Figure 1).

189

190 **3.1.3 CALIOP**

191

192 The Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) was launched onboard
193 CALIPSO on April 28, 2006 as part of the NASA A-Train. CALIOP is a two-channel
194 (532 and 1064 nm) spaceborne lidar that provides profiles of cloud and aerosol properties
195 along the satellite subpoint [Vaughan, 2005]. CALIOP has a temporal resolution of
196 20.16 Hz and vertical resolution that varies from 30 m in the troposphere up to 60 m at
197 higher altitudes. Because CALIOP is an active instrument, it provides both a daytime
198 (1:30 pm local time) and nighttime (1:30 am local time) measurement. CALIOP sends
199 out polarized light at 532 nm and is equipped with sensors that measure the parallel and
200 perpendicular components of the backscattered signal. The standard CALIOP retrieval
201 provides measurements of total attenuated backscatter at each channel [Vaughan, 2005].
202 However, polarization information and spectral variation of the backscatter can be used
203 to infer the presence of aerosols and their type [Vaughan, 2005] In the CALIOP
204 algorithm, backscatter from aerosols is differentiated from clouds by defining a lidar
205 color ratio ($\beta_{1064\text{nm}} / \beta_{532\text{nm}}$). At visible wavelengths, aerosols exhibit spectral variation
206 while clouds do not, therefore a lidar color ratio that is approximately one is used to

207 identify clouds [Vaughan, 2005]. Once aerosols are differentiated from clouds,
208 polarization properties can be used to infer aerosol type. Non-spherical aerosols such as
209 dust are depolarizing and contribute to signal return in both the perpendicular and parallel
210 planes. Spherical aerosols are not strongly polarizing and scatter predominantly in the
211 parallel plane. Therefore, a depolarization ratio ($\beta_{\text{perpendicular}} / \beta_{\text{parallel}}$) can be defined to
212 identify the presence of non-spherical aerosols. For our analysis, we use CALIOP
213 version 3.01 data, which offers an improved technique for the daytime 532 nm total
214 attenuated backscatter calibration relative to previous versions.

215

216 **3.1.4 CPL**

217

218 The Cloud Physics Lidar (CPL) is a multi-pulse lidar that has provided observations
219 during several NASA field campaigns [McGill et al., 2004; McGill et al., 2000]. During
220 TC⁴, CPL flew on the NASA ER-2 aircraft, providing profiles of total attenuated
221 backscatter on 16 different days. CPL measures backscatter at 3 wavelengths (355, 532,
222 and 1064 nm) with a frequency of 5 kHz and depolarization ratio at 1064 nm [McGill et
223 al., 2002]. Processed CPL data is available with a temporal resolution of 1 s and has a
224 spatial resolution of 30 m in the vertical and 200 m in the horizontal [McGill et al., 2002].

225

226 **3.2 Evidence of the Central American dust barrier**

227

228 To illustrate the Central American dust barrier, we show the climatology of July MODIS-
229 Aqua (2003-2010) and MODIS-Terra (2000-2010) land and ocean AOT averaged over

230 latitudes of peak Caribbean dust AOT (10° N - 20° N, see Figure 1) in Figure 2. From
231 this, we see that the Central American dust barrier is a persistent feature, marked by a
232 sharp drop in the AOT west of 80° W. Specifically, during July 2007, the MODIS-Terra
233 AOT drops from 0.375 at 80° W down to 0.2 at 90° W.

234

235 Also shown in Figure 2 is the July 2007 AOT from the GEOS-5 model averaged over the
236 same region. For this comparison we sample our modeled aerosol distributions at the
237 times and locations of the MODIS observations, which has been shown to reduce biases
238 between the MODIS and model AOT because of clouds [Colarco et al., 2010]. Over the
239 Caribbean (west of 60° W), the model AOT is comparable to MODIS-Terra. Near the
240 Central American coastline, the model shows evidence of a barrier to dust transport,
241 although not as drastic, decreasing from 0.4 at 80° W to 0.3 at 90° W (Figure 2). This
242 suggests that either our removal processes or atmospheric dynamics that drive transport
243 might not be correct over this region of the Caribbean and will be further explored in
244 Section 5. Despite this, the model shows evidence for a barrier to dust transport that
245 corresponds with the Central American coastline.

246

247 **3.3 Model Evaluation**

248

249 Here we evaluate the location, timing, and magnitude of dust events simulated in our
250 model with AOT observations from MODIS-Aqua and AERONET and vertical profile
251 observations from CPL and CALIOP.

252

253 Figure 3 shows July 2007 monthly means of total AOT from MODIS-Aqua and our
254 simulation (sampled at MODIS-Aqua observations points as described above). Off the
255 west coast of North Africa, the model has the peak AOT in the same location as the
256 sensor, but at a greater magnitude. Moving west across the tropical North Atlantic, the
257 model matches the observed dust plume location and width, and the magnitude of AOT
258 becomes more comparable with observations. Owing to improvements in the model
259 physics and the MERRA analyses, GEOS-5 does better transporting dust from the
260 Saharan source region to the Caribbean relative to previous versions of the model
261 [Colarco et al., 2010; Nowottnick et al., 2010]. However, the model extends its dust
262 plume somewhat into the eastern Pacific ($90^{\circ} - 95^{\circ}$ W), while MODIS-Aqua AOT values
263 are constrained to the Caribbean. This feature is also seen in Figure 2, where the model
264 representation of the Central American dust barrier is not as pronounced as observed by
265 MODIS-Terra.

266

267 To evaluate the timing of simulated dust events in the model we compare our
268 For each AERONET site, we compare the observations to simulated total AOT values
269 from the model grid box that contains the location of the site and calculate mean AOT
270 and square of the Pearson correlation (r^2) values on days when AERONET data is
271 available (Figure 4). During TC⁴, we have data from three AERONET sites near the
272 source region (Ras El Ain, La Laguna, and Capo Verde) and one site downwind (Cape
273 San Juan).

274

275 At the Ras El Ain site, the model has excellent agreement with the magnitude and the
276 timing of observed dust events, marked by comparable mean AOT values and a high
277 correlation coefficient ($R^2 = 0.71$). On the island of Tenerife, the model is well correlated
278 with the elevated La Laguna site ($R^2 = 0.62$), but is somewhat larger in magnitude.
279 AERONET AOT values exhibit more variability than at Ras El Ain, as passing dust
280 events cause a peak in AOT for 2-3 days and then return to almost zero. The model
281 reproduces the daily variability when compared to AERONET, but simulate a greater
282 AOT when dust events are observed. At the Capo Verde site, located downwind of a
283 major Saharan dust source, the model simulates a slightly higher mean AOT value and is
284 moderately correlated ($R^2 = 0.46$) with AERONET. The lower correlation might be the
285 result of fewer coincident data points between AERONET and model, likely requiring
286 more observations for a more meaningful evaluation of the model at this location.
287 Downwind of the Saharan source region at the Cape San Juan site, the model is well
288 correlated ($R^2 = 0.56$) with AERONET, matches the timing of transported dust events,
289 and has a mean AOT value that is nearly identical to the mean AERONET value.
290
291 Overall, GEOS-5 accurately simulates the timing and magnitude of dust events near the
292 Saharan source region and in the Caribbean during the TC⁴ field campaign as compared
293 to the AERONET observations. This contrasts with our comparison to MODIS-Aqua,
294 where the model generally simulated a higher AOT, particularly just downwind of the
295 Saharan source region. The MODIS and AERONET datasets are complementary and
296 have their respective advantages. While MODIS provides a great deal of spatial
297 coverage, there are uncertainties in the retrieved AOT due to uncertain aerosol optical

298 properties, surface characterization, and cloud contamination. On the other hand,
299 AERONET provides a direct measurement of AOT and has a much higher temporal
300 coverage (multiple observations per day). When comparisons to AERONET are
301 combined with those to MODIS-Aqua, we find that the model captures the shape,
302 magnitude, and timing of dust plumes during the TC⁴ timeframe.
303
304 During TC⁴, a Saharan dust plume was observed over the Caribbean on 19 July with the
305 CPL flying on the NASA ER-2 aircraft. Using CALIOP, we tracked this dust event from
306 the Saharan source region (14 July) to the Caribbean (19 July) to evaluate our simulated
307 vertical dust distributions during transport (Figure 5). For an accurate comparison, we
308 sampled GEOS-5 along the CALIPSO track at the model synoptic time nearest to the
309 daytime CALIOP measurement. Shown in Figure 6 are GEOS-5 comparisons to CALIOP
310 532 nm total attenuated backscatter and feature mask from 14 July to 19 July. On 14 July,
311 CALIOP observes a thick, elevated dust plume located from 2-5.5 km that extends from
312 10° - 26° N. The model captures the latitude extent of the dust plume observed by
313 CALIOP, but is lower in altitude ranging from 1-5.5 km. A limitation of CALIOP is that
314 its signal becomes attenuated towards the surface when it encounters thick aerosol
315 plumes. On this day, the CALIOP signal might be partially attenuated at low altitudes, so
316 the CALIOP data may suggest the lowest edge of the dust plume is at a higher altitude
317 than it actually was. In the CALIOP layer identification product, low-level marine clouds
318 are observed north of 15° N below 1 km. While we only show extinction from aerosols,
319 the influence of these clouds can be seen in the aerosol total extinction where the aerosols
320 in this region have swelled in the marine boundary layer and are marked by high

321 extinction values. Moving farther from the Saharan source region, the edge of a dust
322 event is observed on 15 July. CALIOP observes an elevated, thick layer of dust that
323 extends from 2-5 km between 11° - 24° N, which is well represented in the model.
324 Further downwind on 17 July, the model matches the observed horizontal extent and
325 altitude of the observed dust plume. The simulated dust plume extends down to the
326 surface into a region where CALIOP identifies a thin layer of maritime clouds, making it
327 difficult to determine whether the lower extent of the simulated plume is correct. On 19
328 July, the model captures the narrow north-south width and low-altitude dust plume
329 observed below 3 km by CALIOP, although clearly the observations are impacted by the
330 presence of mid- and low-level clouds. In general, we see for this case that GEOS-5 is
331 capturing similar dust plume features to the CALIOP observations during this time
332 period.

333
334 Figure 7 shows the spatial distribution of AOT at 550 nm retrieved from MODIS-Aqua
335 and GEOS-5 at 18Z, with the ER-2 flight track overlaid on 7/19. The flight originated
336 from Costa Rica, heading southwest over the Pacific Ocean to 90° W, then turned around
337 and headed northeast back towards Costa Rica. The aircraft continued past Central
338 America over the Caribbean Sea to 75° W and then headed southwest back to Costa Rica.
339 During the flight, CPL provided an approximately east-west transect of total attenuated
340 backscatter that extends from the Pacific Ocean into the Caribbean. Comparing the
341 model to MODIS-Aqua on this day, the model matches the observed AOT location and
342 magnitude over the Caribbean. Over the Pacific Ocean MODIS-Aqua is partially
343 obscured by clouds, but the model shows a majority of the model AOT confined to the

344 Caribbean and over Central America. This phenomenon is more clearly seen in the CPL
345 profile of the 532 nm total attenuated backscatter and column AOT when compared to
346 GEOS-5 profiles of extinction and AOT at 550 nm that have been sampled along the ER-
347 2 track at the nearest model synoptic time on 7/19 (Figure 7). Although the CPL signal is
348 frequently attenuated by clouds over the Pacific and only occasionally over the
349 Caribbean, both CPL and GEOS-5 provide an illustration of the Central American dust
350 barrier along the eastern coastline of Costa Rica (9° N, 84° W, marked by a mountain).
351 To avoid cloud contributions to the AOT, we compare the column AOT from 5 km to the
352 surface for CPL and GEOS-5 (Figure 7). CPL observes high AOT values over the
353 Caribbean, and a sharp decrease in AOT that corresponds with the Central American
354 coastline. A similar feature is seen in the simulated AOT but at a lower magnitude and as
355 in Figure 2, the representation of the Central American dust barrier is not as well defined,
356 indicating that our transported dust loading might be too low on this day.

357

358 **4. Controls on Saharan Dust During Transport**

359

360 To understand the cause of the Central American dust barrier, we must understand the
361 roles of the controls on dust distributions during transport. Once emitted from the source
362 region, dust is further lifted into the atmosphere through dry convection and turbulent
363 eddies to form an elevated layer, often penetrating into the so-called Saharan Air Layer
364 (SAL) of hot, dry air [Karyampudi, 1999]. During summer months, a surface north-south
365 temperature gradient forms between the hot Sahara and the relatively cooler Sahel [Cook
366 et al., 1999]. Through thermal wind balance, this leads to the summertime African

367 Easterly Jet (AEJ). During AEJ formation, the SAL converges on the north side of the
368 AEJ axis and is then transported along the north side of the AEJ, delivering dust to the
369 Caribbean. During the journey from the Sahara to the Caribbean, dust distributions are
370 controlled by both dynamical and loss processes. Atmospheric dynamics controls the
371 direction and magnitude of the transported dust mass flux or flow, while loss processes
372 control the overall dust burden. Therefore, we suspect that the Central American dust
373 barrier is caused by increases in wet removal, a change in transport direction resulting
374 from a shift in the prevailing atmospheric dynamics, or some combination of both.

375

376 Ideally, we would have airborne measurements while tracking several dust plumes to
377 help understand cause of the Central American dust barrier. Unfortunately,
378 measurements of this sort are extremely limited. However, from our comparisons to
379 observations of mean dust plume position, event timing, and vertical distributions near
380 and downwind of the Saharan source region, GEOS-5 provides a reasonable
381 representation of dust distributions during the TC⁴ timeframe, while simulating the
382 aforementioned processes that are not easily measured. The accuracy of our simulated
383 wet removal processes are directly linked to our ability to accurately simulate the timing
384 and intensity of precipitation events. Figure 8 shows the July 2007 mean precipitation
385 (mm day⁻¹) from the Global Precipitation Climatology Project (GPCP) [Huffman et al.,
386 2009; Adler et al., 2003] and GEOS-5. GPCP provides monthly mean precipitation data
387 at 1° x 1° resolution using rain gauges, microwave satellite observations from the Special
388 Sensor Microwave Imager (SSM/I), and infrared satellites observations from many global
389 geostationary satellites [Adler et al., 2003]. The precipitation patterns in GEOS-5 are

390 generally consistent with GPCP, matching peak values located over Central and South
391 America. However, GEOS-5 produces a broad area of convective precipitation over the
392 Caribbean that is not seen in the GPCP data. Over the Caribbean, the average GEOS-5
393 precipitation rate is 5 mm day^{-1} while the average GPCP precipitation rate is 1.5 mm day^{-1} .
394 This presents an interesting feature of the model. Figure 2 suggests that our removal
395 rates are not aggressive enough in removing dust, particularly in the region of the Central
396 American dust barrier. However, on average, our precipitation rate is greater by a factor
397 of three (Figure 8). This quandary suggests that the relationship between precipitation
398 and wet removal is not strong enough in our model. We could, alternatively, simply
399 rescale our dust emissions lower, which would remove most of the bias seen in Figure 2,
400 but crucially this would not produce the abrupt dust barrier evident in the data at
401 approximately 90° W .

402

403 In addition to possible errors in our representation of loss processes, our simulated dust
404 distributions are sensitive to atmospheric dynamics. By using a replay simulation, we are
405 providing the model with assimilated winds, so that it will be forced with actual
406 dynamics at each synoptic time. Our estimation of dust transport is therefore sensitive to
407 our ability to reproduce the actual dynamical state and will be limited by errors in the
408 representations of advection, planetary boundary layer mixing, and convective mixing.

409 In addition to sensitivities to the internal dynamical processes, simulated dust
410 distributions will also be sensitive to the accuracy of observations used in the analysis.
411 Despite these potential sources of error, our July 2007 simulated dust distributions are
412 comparable to MODIS-Aqua, AERONET, and CALIOP, as shown in Section 3.

413 Therefore, we use our dust distributions from GEOS-5 to understand the relative roles of
 414 the processes that control the Central American dust barrier.

415

416 **4.1 Dust Mass Budget**

417

418 We begin our investigation of the controls on the Central American dust barrier by
 419 employing the vertically integrated mass divergence form of the continuity equation for
 420 mean values from July 2007:

$$421 \quad \frac{\partial q}{\partial t} = (P - L) + \nabla \cdot \vec{Q} \quad (2)$$

422 where q is the column dust loading defined:

$$423 \quad q = \sum_{z=0}^{z=top} \gamma \cdot \rho_{air} \cdot dz \quad (3)$$

424 and \vec{Q} is the vertically integrated dust mass flux:

$$425 \quad \vec{Q} = \sum_{z=0}^{z=top} \gamma \cdot \rho_{air} \cdot u \cdot dz \cdot \hat{i} + \sum_{z=0}^{z=top} \gamma \cdot \rho_{air} \cdot v \cdot dz \cdot \hat{j} \quad (4)$$

426 Here, γ is the dust mass mixing ratio (kg kg^{-1}), ρ_{air} is the atmospheric air density (kg m^{-3}),
 427 u and v are the east-west and north-south components of the wind field (m s^{-1}), and dz is
 428 the thickness (m) of each model layer in the vertical column.

429

430 After integrating in the vertical, Equation 2 has three terms: the storage term ($\frac{\partial q}{\partial t}$), the

431 production-loss ($P - L$) term, and the divergence, or transport, term ($\nabla \cdot \vec{Q}$). The storage

term represents the net local change in the dust column loading, the $P - L$ term is defined as the sum of the column emission fluxes minus fluxes due to dry and wet removal, and the transport term represents any dust column convergence and divergence resulting from transport. All terms in Equation 2 are in flux form and have the units ($\text{kg m}^{-2} \text{s}^{-1}$). Equation 2 can be interpreted as any accumulation of dust mass within an atmospheric column results from the sum of the net production minus loss and dust import/export via transport. Figure 9 shows the July 2007 monthly mean storage, $P - L$, and transport terms. We analyze each term separately to understand their respective influence on our simulated dust distributions over the Caribbean. Our analysis of Equation 2 uses monthly mean components that have been computed from instantaneous model output at every 3 hours; thus, the fields examined include both the mean flow and the contribution from transient eddies.

4.1.1 Storage Term

At each grid cell, the storage term represents the mean local change in the column dust loading q (kg m^{-2}) (Equation 3). During July 2007, the largest variations in the dust column loading occur away from regions of semi-persistent dust flow (Figure 9). This can be seen north of 20°N off the west coast of North Africa during July 2007, where removal rates are small (Figure 9). Eventually, this dust will be removed from the atmosphere via loss processes or transport. Over the Caribbean, the storage term is significantly less than the $P - L$ and transport terms, indicating that the other terms are in near-balance over this region. Over longer time periods, we expect the storage term to

455 approach zero, as deviations in the mean dust flow will become less significant and
456 averaged out. In this case, the $P - L$ term will balance the transport term.

457

458 **4.1.2 $P - L$ Term**

459

460 The mean $P - L$ term for July 2007 shows positive values over the global source region
461 and negative values downwind, corresponding to regions where emissions and losses
462 prevail, respectively (Figure 9). Once dust is emitted from the source region, the total
463 atmospheric burden is controlled by losses through dry and wet removal processes. In
464 the Atlantic, losses peak immediately downstream of the source region, although a broad
465 area of high dust losses carries into the Caribbean.

466

467 Figure 10 shows the relative contributions of our modeled dust loss processes as a
468 function of distance from the source region. By mass, gravitational sedimentation is the
469 dominant removal process near the Saharan source region, as the largest, most massive
470 dust particles fall quickly from the atmosphere. Sedimentation becomes less efficient
471 further downwind as the largest particles are removed. Wet removal becomes the
472 dominant loss process, first via large-scale precipitation immediately west of the source
473 region and then through convective precipitation in the western Caribbean and near
474 Central America. This region where convective removal dominates coincides with the
475 location of the Central American dust barrier.

476

477 **4.1.3 Transport Term**

478

479 The transport term represents any column accumulation or loss of dust from the divergent

480 component of the transported dust mass flux. To analyze the contribution of transport to

481 Equation 2, we begin with our vertically integrated dust mass flux \vec{Q} ($\text{kg m}^{-1} \text{s}^{-1}$)

482 (Equation 4). Because dust is typically located at low altitudes, \vec{Q} will be weighted

483 toward the mass concentration and the near-surface wind direction and magnitude.

484

485 Consider the Helmholtz decomposition [Brown, 1991]:

486
$$\vec{Q} = \vec{Q}_{rot} + \vec{Q}_{div} \quad (5)$$

487 where \vec{Q}_{rot} and \vec{Q}_{div} are the rotational and divergent components of the vertically

488 integrated mass flux vector \vec{Q} , with $\nabla \cdot \vec{Q}_{rot} = 0$ and $\nabla \times \vec{Q}_{div} = 0$ by definition. The

489 corresponding mass flux streamfunction ψ and velocity potential χ can be obtained by

490 solving Poisson's equations [Brown, 1991]:

491
$$\nabla^2 \Psi = \nabla \cdot \vec{Q}_{div} \quad (6)$$

492
$$\nabla^2 \chi = \hat{k} \cdot \nabla \times \vec{Q}_{rot} \quad (7)$$

493 from which we obtain the divergent and rotational components of \vec{Q} :

494
$$\vec{Q}_{rot} = -\frac{\partial \Psi}{\partial y} \hat{i} + \frac{\partial \Psi}{\partial x} \hat{j} \quad (8)$$

495
$$\vec{Q}_{div} = \frac{\partial \chi}{\partial x} \hat{i} + \frac{\partial \chi}{\partial y} \hat{j} \quad (9)$$

496 The rotational component depicts the recirculation of dust in the atmosphere, while the
 497 divergent component of the vertically integrated mass flux is associated with the P – L
 498 process ($\nabla \cdot \vec{Q}_{div} = \nabla \cdot \vec{Q}$) (Equation 2). Shown in Figure 11 are the July 2007 mean
 499 streamfunction and velocity potential contours with the rotational and divergent dust flow
 500 vectors overlaid. We recall that the rotational component of the dust flow is proportional
 501 to the curl of the streamfunction; therefore, rotational flow will be strongest where
 502 streamlines are closest. By definition, the rotational flow will be cyclonic surrounding
 503 relative minima of the streamfunction, and anti-cyclonic surrounding the relative
 504 maxima. We see strong rotational dust flow leaving the Sahara as part of the SAL and
 505 riding on the northern side ($15^\circ - 25^\circ$ N) of the AEJ across the Atlantic Ocean. In this
 506 region, the rotational component of the dust flow is strong for two reasons: 1) dust
 507 concentrations are high within the SAL and 2) strong, non-divergent easterlies within the
 508 AEJ persist. The effect is a narrow band ($15^\circ - 25^\circ$ N) of strong rotational flow that
 509 transports dust from the Sahara to the Caribbean. Upon reaching the Caribbean, the
 510 rotational flow weakens because: 1) dust loss processes have reduced the overall dust
 511 load during transport and 2) easterly wind speeds are reduced. Additionally, the flow
 512 direction shifts from primarily westward to north-westward over the Caribbean as it is
 513 now influenced by the Azores subtropical high-pressure system that exists over the
 514 Atlantic Ocean. The rotational dust flow eventually turns eastward and returns dust back
 515 to the Saharan source region. Thus, when following a constant streamline, the rotational
 516 component of Saharan dust flow is an anti-cyclonic recirculation, where dust leaves the
 517 source region as part of the AEJ and returns with the westerlies as part of the Azores

518 High. A similar—but weaker—cyclonic feature is seen south of 15° N, transporting dust
519 to South America.

520

521 The divergent component of the flow is proportional to the gradient of the velocity
522 potential. Therefore, regions of divergence correspond to relative minima of the velocity
523 potential correspond, while regions of convergence correspond to relative maxima. In
524 Figure 11, we see a dipole in the divergent flow field between the Saharan source region
525 and the Caribbean. Over the source region strong divergent flow persists, as a divergent
526 component to the dust flow is required for dust to leave the source region. During
527 transport, the divergent flow is significantly reduced and there is a broad, region of
528 convergence over the Caribbean where loss processes prevail. The significant reduction
529 in the divergent flow can be the result of a weakening of the wind field or a reduction in
530 the dust burden caused by the various loss processes during transport. As previously

531 mentioned, the divergence of the divergent flow ($\nabla \cdot \vec{Q}_{div}$) is the transport term in
532 Equation 2. In Figure 9, as expected, the July 2007 transport term is positive (divergent)
533 over the source regions, as dust is transported outward from the sources. Downwind of
534 the Saharan source region, the transport term is negative (convergent), which corresponds
535 with the convergent flow field in Figure 11. One striking feature of the divergence field
536 is that it aligns with the $P - L$ term in regions where production and loss occur. Because
537 these regions have a semi-persistent flow of dust for this month and the storage term is
538 small, there is a near-balance between the transport and $P - L$ terms. Thus, over these
539 regions, regions of dust emission ($P - L > 0$) correspond with divergent outflow (positive
540 transport term) and regions of dust loss ($P - L < 0$) correspond with convergent inflow

541 (negative transport term). We expect that convergent flow increases dust loss rates in
542 two ways. First, the convergent flow will accumulate dust within the atmospheric
543 column. This accumulation will increase the potential for removal in regions where the
544 storage term is small. Second, we find vertical motion over convergent regions (not
545 shown), which is associated with convection. This second process is more relevant for
546 wet removal as we expect greater wet deposition and scavenging rates in the presence of
547 precipitation and clouds.

548

549 Despite the link between $P - L$ and divergent flow, it is clear that rotational flow has a
550 greater magnitude and is in a different direction (predominantly westward) than the
551 divergent flow (predominantly eastward). However, this alone does not lend much
552 insight into any influences that transport might have on the Central American dust
553 barrier. In addition to the effects of loss processes, the dust barrier could be influenced by
554 a slight change to the flow field over the Caribbean or a combination of the rotational and
555 divergent components. To better understand this, we further break the rotational and
556 divergent components into their east-west and north-south components. Figure 12 shows
557 the east-west and north-south total, rotational, and divergent flow components. Over the
558 Caribbean, the rotational component of the east-west flow is strongly westward while the
559 divergent component is weakly eastward. Despite cancellation between the two
560 components near the coast of Costa Rica, the net east-west flow is westward and acts to
561 transport dust across Central America. The north-south flow for the rotational
562 component shifts from southward to northward near 12.5° N over the Caribbean, while
563 the divergent flow shifts from northward to southward flow at 17.5° N . However, the net

564 north-south flow is northward over the entire Caribbean. Thus, there is a northward
565 turning of the dust flow as it enters the Caribbean, which when combined with the net
566 westward flow causes a northwestern migration of the overall dust flow and serves as a
567 possible explanation of the Central American dust barrier.

568

569 **4.2 Loss Processes vs. Transport**

570

571 We investigate the dust mass budget in the latitude band of peak dust AOT ($10^{\circ} - 20^{\circ}$ N)
572 to understand the relative roles of dust loss processes and transport in the Central
573 American dust barrier. Figure 13 shows the mass of dust removed from loss processes,
574 from transport out of the northern (20° N) and southern (10° N) sides of the latitude band,
575 and the change in the east-west mass flux (flux in minus flux out) as a function of
576 longitude. To obtain the amount of dust lost via removal, we integrate the $P - L$ rates
577 spatially and temporally and sum over the latitude band at each longitude (black curves in
578 Figure 13). To quantify the net north-south dust mass flux out of the band, we subtract
579 the net spatially and temporally integrated north-south dust flux at 20° N from that at 10°
580 N at each longitude (Figure 13). To obtain the change in the east-west mass flux, we first
581 integrate the net east-west component of the dust flow spatially and temporally at each
582 grid box. The change in the east-west mass flux is then determined by differencing the
583 east-west flow in the westward direction and then summing along all latitudes (Figure
584 13). Negative mass values in Figure 13 correspond with net loss via removal processes or
585 transport out of the latitude band, or a reduction in the westward mass flux. It should be
586 noted that the sum of the net north-south mass flux and the change in the westward mass

587 flux is the divergence term in Equation 2. This sum is approximately equal to the mass of
588 dust removed by loss processes, with any residual related to the storage term.

589

590 Over the Caribbean, removal from loss processes and northward transport were shown to
591 serve as possible causes of the Central American dust barrier. In Figure 13, the
592 longitudes of the Central American dust barrier (80° - 90° W) correspond with increases
593 in dust mass loss and northward transport. To quantify their relative contributions, we
594 integrate the production-loss and north-south transport curves in Figure 13 over the
595 region of the Central American dust barrier. From this, we estimate that loss processes
596 remove 1.67 Tg of dust while the north-south dust flow transports 1.46 Tg of dust out of
597 the Central American dust barrier region during July 2007 (Table 1).

598

599 Based on these estimations, it is clear that both loss processes and atmospheric dynamics
600 have a contribution to the Central American dust barrier. Of the two processes, dust loss
601 from removal processes has a slightly greater contribution (53%) to the Central American
602 dust barrier than northward transport (47%).

603

604 **5. Discussion**

605

606 We have shown that loss processes have a greater contribution towards the Central
607 American dust barrier than northward transport for July 2007. From Figure 10, it is clear
608 that wet removal by large scale and convective scavenging dominate the loss processes
609 downwind of the Saharan source region between 10° – 20° N and serve as the major

610 pathways for dust removal over the Caribbean. However, as discussed in Section 3.2.1,
611 we suspect that our wet removal rates are not aggressive enough over the Caribbean and
612 serves as the cause of our weaker representation of the Central American dust barrier in
613 Figure 2. To explore the controls of wet removal on our transported dust distributions,
614 we perform additional simulations of July 2007 where we modify our parameterization of
615 wet removal processes relative to our baseline simulation setup.

616

617 Table 1 presents a budget analysis for our baseline simulation, as well as the sensitivity
618 analyses we will discuss here. Included are the dust mass removal by loss processes,
619 north-south transport, and their contribution to the Central American dust barrier. Also
620 shown are the $10^{\circ} - 20^{\circ}$ N net east-west mass transported across the planes at 80° W and
621 90° W and their difference. This difference, when combined with the north-south
622 transport is the mass divergence and should approximately balance the mass removed by
623 loss processes, with any residual attributable to the storage term in Equation 2. Table 1
624 lists a dust mass barrier efficiency of the Central American dust barrier defined as the
625 difference between the $10^{\circ} - 20^{\circ}$ N net east-west transported dust mass at 80° W (flow
626 in) from that at 90° W (flow out) divided by the transported dust mass at 80° W (flow in).
627 Additionally, after sampling consistently with MODIS-Terra, Table 1 lists a total AOT
628 barrier efficiency and a coarse mode (dust plus sea salt) AOT efficiency that can be
629 compared to the MODIS coarse mode AOT after averaging from $10^{\circ} - 20^{\circ}$ N.

630

631 Our baseline simulation has a dust mass barrier efficiency of 0.36, meaning that the
632 Central American dust barrier removes 36% of the dust mass between 80° W to 90° W

633 (Table 1). Our baseline simulation has a total AOT barrier efficiency of 0.21 and a coarse
634 AOT efficiency of 0.17. Comparisons to MODIS-Terra show that our removal rates are
635 not aggressive enough, as MODIS-Terra has a total AOT barrier efficiency of 0.37 and
636 coarse AOT barrier efficiency of 0.30 (Table 1).

637

638 As our model does not include a detailed representation of aerosol-cloud-precipitation
639 interactions, we parameterize aerosol wet removal in terms of the model grid box
640 convective updraft mass flux (for convective scavenging) and precipitation rate (for large
641 scale wet removal). An efficiency factor is assigned to each aerosol species that
642 represents its susceptibility to wet removal (i.e., its hygroscopicity) [Colarco et al. 2010].
643 For dust we have assumed its wet removal efficiency is approximately half as efficiency
644 as for hydrophilic carbonaceous and sulfate aerosols. In our first sensitivity test we
645 double the dust convective scavenging efficiency so that it is equivalent to that for
646 hydrophilic aerosols. In Figure 13, we see that doubling the convective scavenging rate
647 increases the mass of dust lost to removal while reducing the north-south and east-west
648 dust flow. If we integrate along our longitudes of the Central American dust barrier,
649 doubling the convective scavenging rate increases the loss contribution to 61% (1.90 Tg)
650 and reduces the contribution by northward transport to 39% (1.24 Tg), increasing the
651 mass barrier efficiency to 0.48 (Table 1). Figure 14 shows the MODIS-Terra sampled
652 AOT from our baseline and sensitivity tests, the ratio of the MODIS-Terra and simulated
653 AOT, and the slope of the AOT ($\Delta\tau/\Delta x$). After doubling the convective scavenging rate,
654 we see a reduction in the high AOT bias in the model and improvement in the slope of
655 AOT as a function of longitude (Figure 14). This corresponds with a significant

656 improvement in the representation of the Central American dust barrier as the simulated
657 AOT reduces from 0.34 at 80° W to 0.25 at 90° W (Figure 14). This corresponds to
658 greater AOT barrier efficiencies of the total (0.25) and coarse (0.21) representations of
659 the Central American dust barrier.

660

661 We performed a second sensitivity test where in addition to doubling the dust convective
662 scavenging rate, we increased the large-scale scavenging rate so that dust wet removal is
663 treated the same as for hydrophilic aerosols. While this further increases the mass of dust
664 lost to removal and reduces the north-south and east-west flow, we find that our
665 simulated dust distributions are more sensitive to modifications to convective scavenging
666 than large-scale scavenging in this region. However, the combined effect of increasing
667 the large-scale and convective scavenging rates consistent with other aerosol types
668 corresponds with an increased contribution from loss processes (66%, 1.97 Tg), a
669 reduced contribution (34%, 1.02 Tg) from northward transport, and an increase in the
670 barrier mass efficiency (0.52) of the Central American dust barrier (Table 1). Treating
671 the wet removal of dust the same as other aerosols yields further improvement in the
672 representation of the AOT magnitude and slope when compared to MODIS-Terra (Figure
673 14). Over the region of the Central American dust barrier, the simulated AOT reduces
674 from 0.31 at 80° W to 0.23 at 90° W (Figure 14), corresponding with an improved total
675 AOT barrier efficiency of 0.28 and a coarse AOT barrier efficiency of 0.22 (Table 1).

676 Although still not as efficient as indicated by MODIS-Terra, this result suggests that the
677 dust wet removal rates in GEOS-5 are too slow and treating the wet removal of dust in a
678 fashion similar to other (more ostensibly hygroscopic) aerosol types yields better

679 comparisons to observations in regions where wet removal is dominant. Because the
680 representation of the dust barrier improves with increases to the wet removal rates, the
681 contribution from loss processes to the Central American dust barrier is likely greater
682 (66%) than originally estimated from our baseline simulation (53%).
683

684 We performed two additional sensitivity tests aimed at understanding if the Central
685 American dust barrier exists when the effects of convective and large-scale scavenging
686 are not simulated. In the first sensitivity test, we did not simulate wet removal from the
687 large-scale scavenging of dust, leaving only convective scavenging as a source of wet
688 removal. As shown in Figure 13, large-scale scavenging over the Caribbean has a small
689 effect on the dust load, as the northward and westward flows are slightly increased and
690 losses are reduced when the effects of large-scale scavenging are not simulated. When
691 we integrate over the longitudes of the Central American dust barrier, we see a shift in
692 the relative significance of northward transport and loss. Northward flow transports 1.85
693 Tg of dust out of the region (59% of the total removal) while dust losses remove 0.95 Tg
694 of dust (41% of the total removal), corresponding with a barrier mass efficiency of 0.33
695 (Table 1). When the AOT is sampled consistent with MODIS-Terra, we see a small
696 increase in the magnitude of the AOT and slope from the coast of North Africa (20° W)
697 to Central America (80° W), but there is still evidence of a Central American dust barrier
698 (Figure 14). When the effects of large scale scavenging are not simulated, the total AOT
699 efficiency and coarse AOT efficiency decrease to 0.19 and 0.16, respectively (Table 1).
700 This result is consistent with the simulations already discussed and suggests that large-
701 scale convective scavenging has a small effect to the Central American dust barrier.

702

703 In a final sensitivity test, we performed a simulation where the effects of all wet removal
704 (convective scavenging and large-scale scavenging) were not simulated. In Figure 13, we
705 see a large increase in the northward and westward dust flows and a significant reduction
706 in the dust loss. Over the Central American dust barrier region, northward transport
707 accounts for 78% (3.35 Tg) of dust removal from the atmospheric column, while loss
708 processes account for 22% (1.28 Tg), corresponding with a mass barrier efficiency of
709 0.25 (Table 1). When compared to MODIS-Terra, we see a nearly constant increase in
710 the AOT from the coast of North Africa (20° W) to the beginning of the Caribbean (60°
711 W) (Figure 14). However, over the Caribbean where convective scavenging has the
712 largest contribution to the overall removal (Figure 10), the model AOT relative to
713 MODIS-Terra increases non-linearly (Figure 14) and reduces the total and dust AOT
714 barrier efficiency to 0.17 and 0.13, respectively (Table 1). Finally, when all wet removal
715 processes are not included, there is no evidence of the Central American dust barrier
716 (Figure 14). Therefore, we determine the Central American dust barrier could not exist
717 without convective scavenging. In practice, however, the Central American dust barrier
718 is the result of two processes working in tandem: 1) Loss processes significantly reducing
719 the dust loading during transport and 2) Atmospheric dynamics redirecting the reduced
720 dust flow northward near the Central American coastline.

721

722 **6. Conclusions**

723

724 We used the GEOS-5 model to simulate the distribution of aerosols during the period of
725 the NASA TC⁴ field campaign (July – August, 2007). In this simulation, we have shown
726 that GEOS-5 simulates dust distributions that are spatially and temporally comparable to
727 MODIS, CALIOP, and AERONET data. Downwind of Africa, GEOS-5 has a similar
728 plume shape to the MODIS observations, but in our baseline simulation overestimates the
729 AOT. GEOS-5 has a better agreement with AERONET AOT values and is well
730 correlated with the AOT time series from sites within and nearby the Saharan source
731 region. GEOS-5 accurately reproduced the latitudinal, longitudinal, and vertical extent of
732 a Saharan dust event during its transport from North Africa to the Caribbean when
733 compared to CALIOP. Over the Caribbean, GEOS-5 AOT magnitude is comparable to
734 MODIS and well correlated with the Caribbean AERONET site, but provided a weak
735 representation of the Central American dust barrier. This feature suggested that our loss
736 processes be explored and possibly adjusted in future implementations of the model.

737

738 In a series of sensitivity analyses with our model we explored the relationship between
739 wet removal parameterization and transport in defining the Central American dust
740 transport barrier. The best agreement between our model and the observations was
741 obtained when dust wet removal was treated as we treat the removal of hydrophilic
742 aerosol species. Conversely, we showed that in the absence of dust wet removal there is
743 essentially no dust transport barrier set up in our model. The implication of appealing to
744 an increase in dust wet removal efficiency is that perhaps processing of dust during
745 transport results in a more hydrophilic aerosol. Such an aerosol would likely be more
746 bioavailable to oceanic organisms once it is eventually deposited.

747

748 Our analysis shows that both wet removal and transport play a role in creating a semi-
749 permeable barrier to dust transport across Central America into the Pacific. Of the two
750 processes, for our best case simulation we find wet removal has a factor of two greater
751 contribution toward defining the barrier than northward transport. Moreover, of the
752 removal processes, the Central American dust barrier is more sensitive to removal by
753 convective scavenging and is not evident when convective scavenging is not simulated.

754

755 Our results should be taken with a few caveats. First, our component analysis is valid for
756 July 2007. While we have shown that the Central American dust barrier is a persistent
757 feature in July (Figure 2), we expect that the barrier will be somewhat sensitive to the
758 variability of inter-annual meteorological conditions over the Central American region.
759 Pfister [2010] found that La Nina conditions in 2007 caused an increase in westward flow
760 and a significant reduction in Caribbean cold clouds and corresponding increase in
761 Pacific cold clouds during the TC⁴ field campaign. This suggests that under less
762 anomalous conditions, transported dust would be more confined the Caribbean and the
763 Central American dust barrier would have a greater presence. The presence of the
764 Central American dust barrier also has implications for equatorial aquatic ecosystems
765 located to the west of the Central American coastline. In this region, high phytoplankton
766 growth inferred from chlorophyll concentration observations during July [Falkowski et
767 al., 1998], suggest that the Central American dust barrier serves as a natural inhibitor of
768 carbon sequestration in the Pacific. Additionally, transported dust distributions will be
769 sensitive to variability in Saharan dust emissions, AEJ strength, and Inter-Tropical

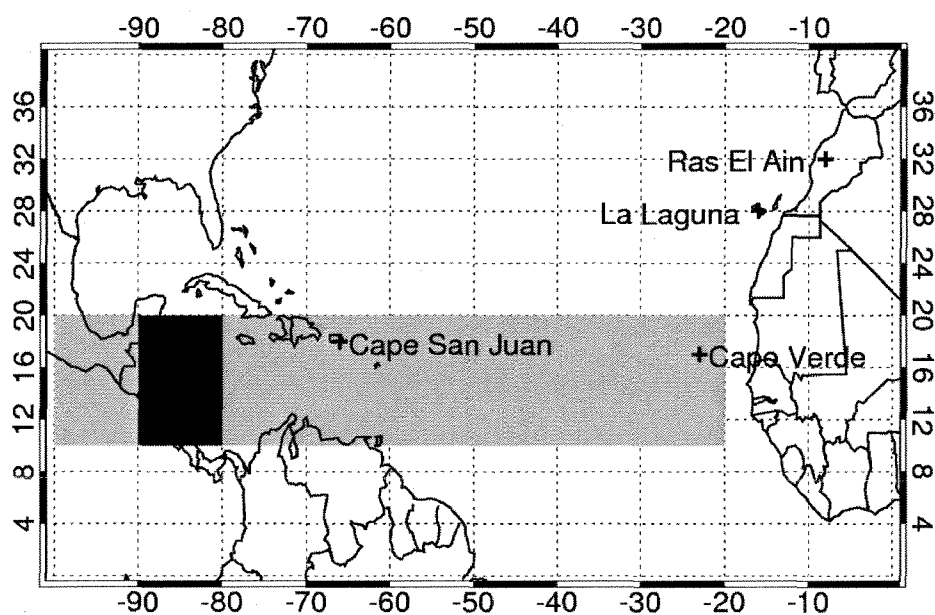
770 Convergence Zone (ITCZ) position. Prospero and Lamb [2003] showed that dust
771 transported from the Sahara to the Caribbean is linked to Sahel precipitation from the
772 previous year. Another caveat is that we expect the Central American dust barrier to
773 exist only in summer months. The AEJ forms during northern hemisphere summer and
774 corresponds with peak dust transport from the Sahara to Caribbean. Offline analysis of
775 the MODIS-Terra 2000 – 2010 monthly climatology suggests that transported dust
776 loadings are too low to see evidence of a Central American dust barrier during non-
777 summer months. One final caveat is that the strength of our results lies in the ability of
778 our model to accurately simulate dust loss processes. In particular, our analysis relies
779 heavily on the ability of the model to provide a realistic representation of convection,
780 which subsequently influences wet removal over the Caribbean. Because wet removal
781 rates are not typically measured in the field, it is difficult to determine whether our
782 parameterization of wet removal is accurate and therefore we are limited to relying on
783 proxies, such as column AOT. As previously discussed, our baseline simulation provided
784 a weak representation of the Central American dust barrier when compared to MODIS-
785 Terra, suggesting that our wet removal rates were too relaxed in the model (Figure 2).
786 However, when compared to the GPCP observations, the July 2007 mean GEOS-5
787 precipitation was slightly greater over most of the Caribbean (Figure 8). These results
788 suggest that the connection between wet removal and precipitation should be
789 strengthened in GEOS-5, in particular that our simulation which best captured this dust
790 barrier was the one that treated dust the same as hygroscopic aerosol species with respect
791 to wet removal processes, suggesting that the best representation of dust in our model is
792 one which allows that dust has mixed or been processed so as to be more hydrophilic.

793 Figure 1.

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798 Figure 1. AERONET site locations and dust barrier-averaging regions (shaded).

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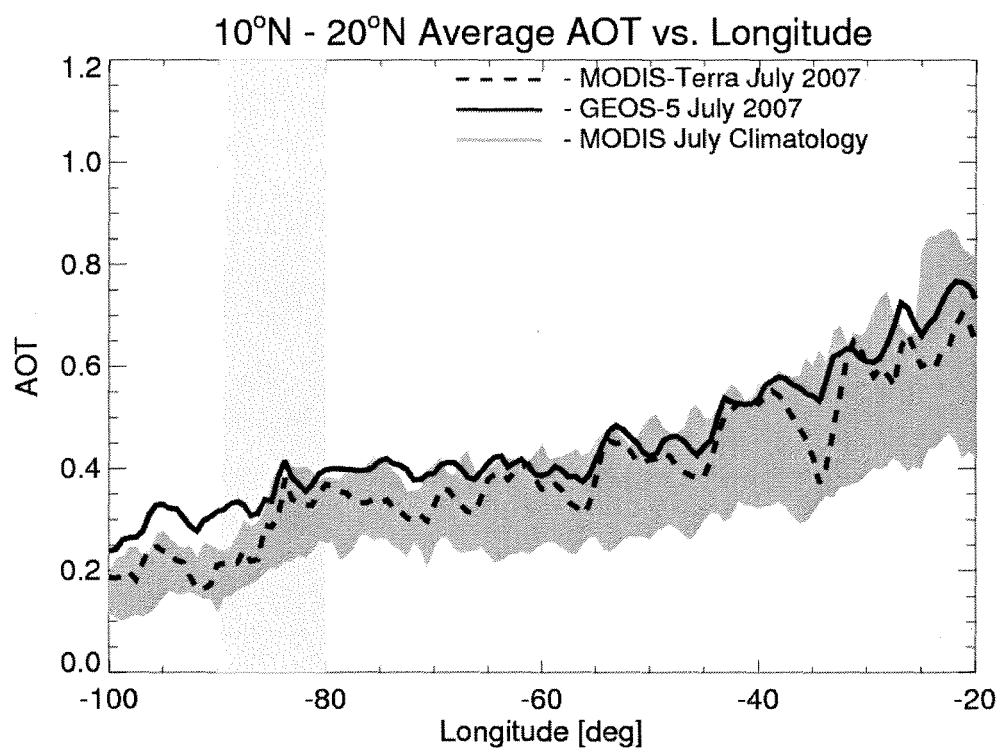
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806 Figure 2.

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808



809

810 Figure 2. MODIS-Terra/Aqua July climatological (2002-2010) AOT (shading), MODIS-
811 Terra July 2007 AOT (dashed) and GEOS-5 sampled (solid) July 2007 AOT averaged
812 from 10°-20° N.

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819 Figure 3.

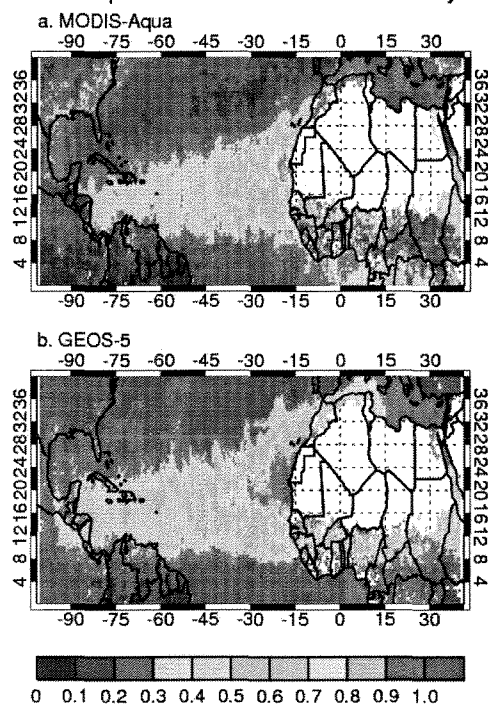
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MODIS-Aqua and GEOS-5 AOT for July 2007



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825

826 Figure 3. MODIS-Aqua (a) and GEOS-5 sampled (b) July 2007 AOT

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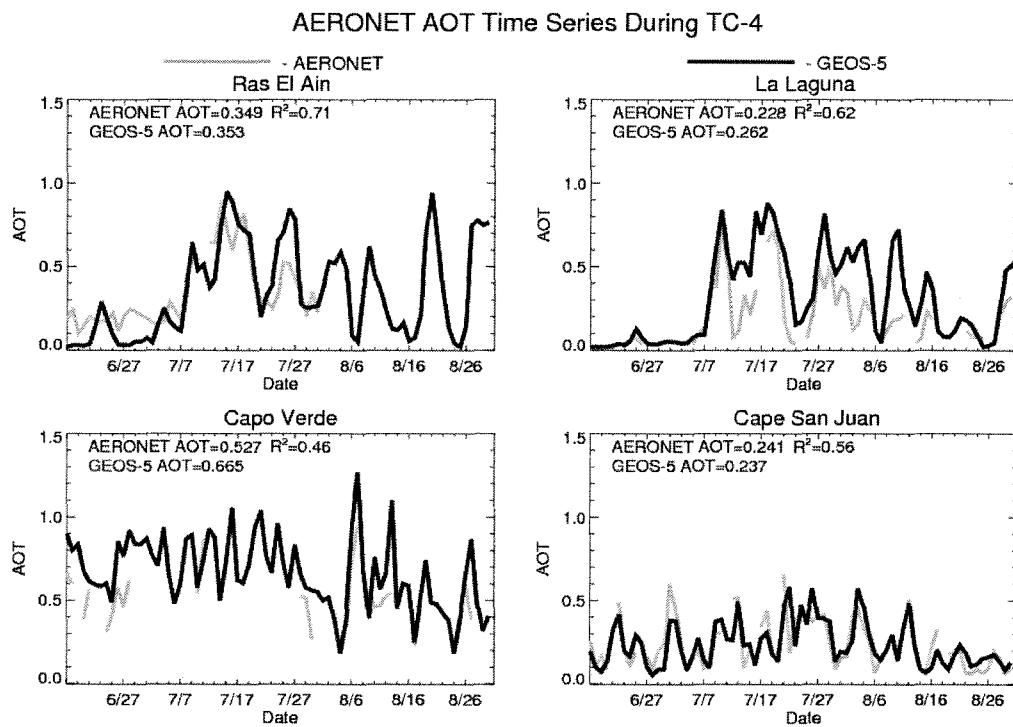
832 Figure 4.

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838 Figure 4. AERONET and GEOS-5 AOT time series.

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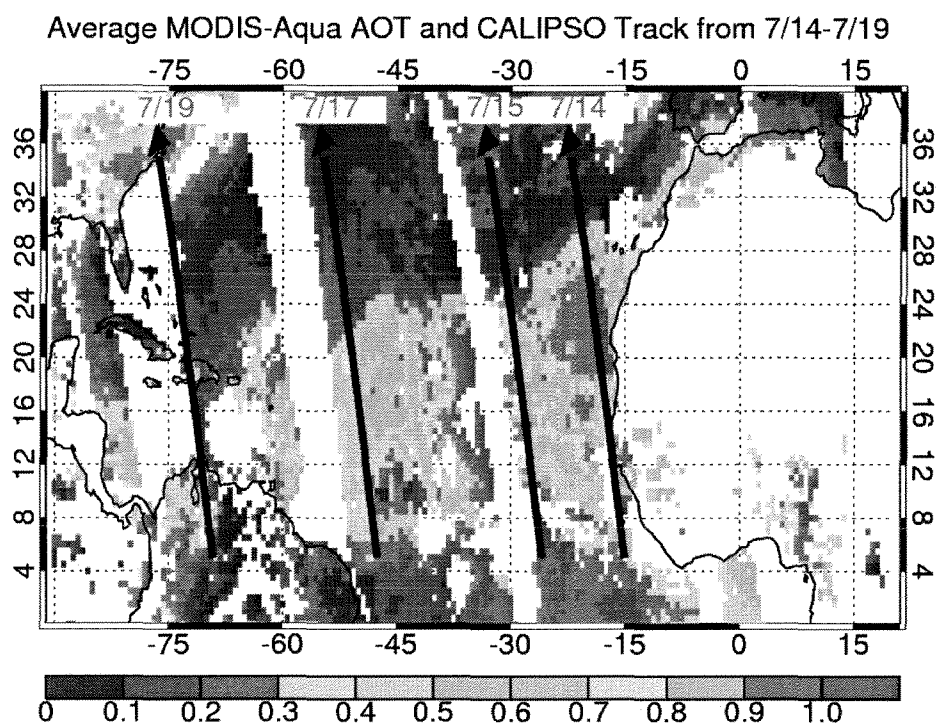
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845 Figure 5.

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850 Figure 5. Average MODIS-Aqua AOT and CALIPSO track from 14-19 July.

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Figure 6.

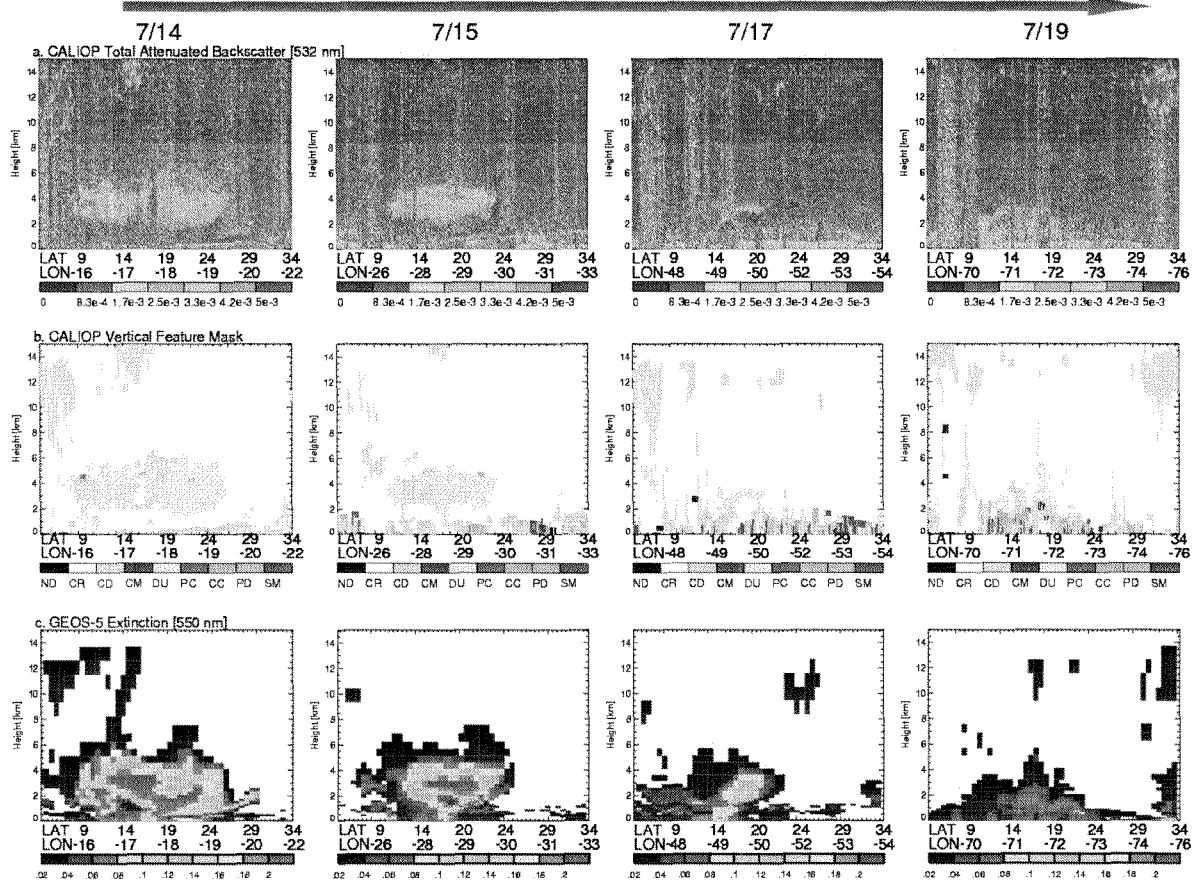
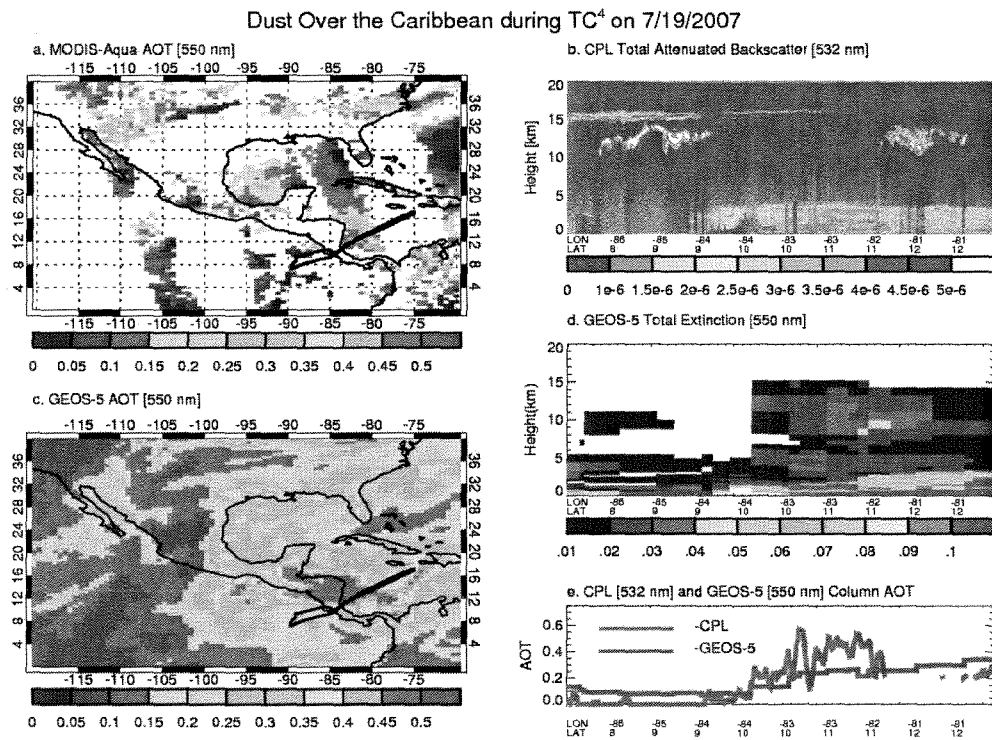


Figure 6. CALIOP total attenuated backscatter [$\text{km}^{-1} \text{sr}^{-1}$] (a), CALIOP vertical feature mask (b), and GEOS-5 extinction [km^{-1}] (c) for a dust event tracked from Africa (14 July) to the Caribbean (19 July).

863 Figure 7.

864

865



866

867 Figure 7. MODIS Aqua AOT (a), CPL total attenuated backscatter [$\text{km}^{-1} \text{sr}^{-1}$] (b),
868 GEOS-5 AOT (c), GEOS-5 extinction [km^{-1}] (d), CPL AOT below 5 km (red) and
869 GEOS-5 AOT below 5 km (blue) (e) on 19 July 2007.

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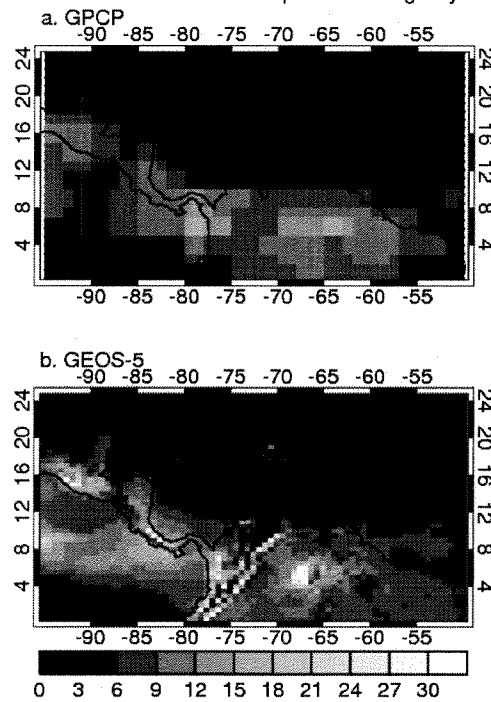
876 Figure 8.

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879

GPCP and GEOS-5 Total Precipitation during July 2007



880

881 Figure 8. July 2007 monthly mean GPCP (a) and GEOS-5 (b) total precipitation [mm dy

882 ¹].

883

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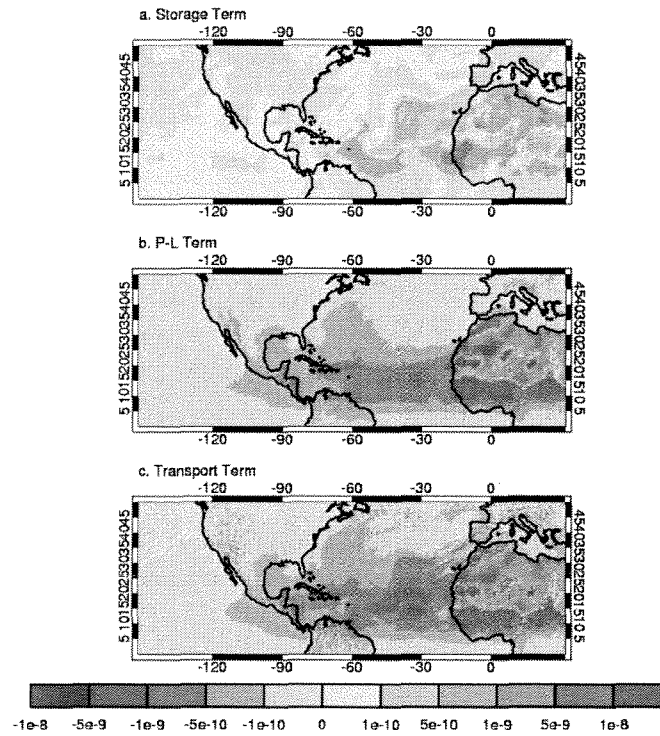
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889 Figure 9.

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894 Figure 9. July 2007 storage(a) production minus loss (P – L) (b), and divergence terms

895 (c) in kg m⁻² s⁻¹.

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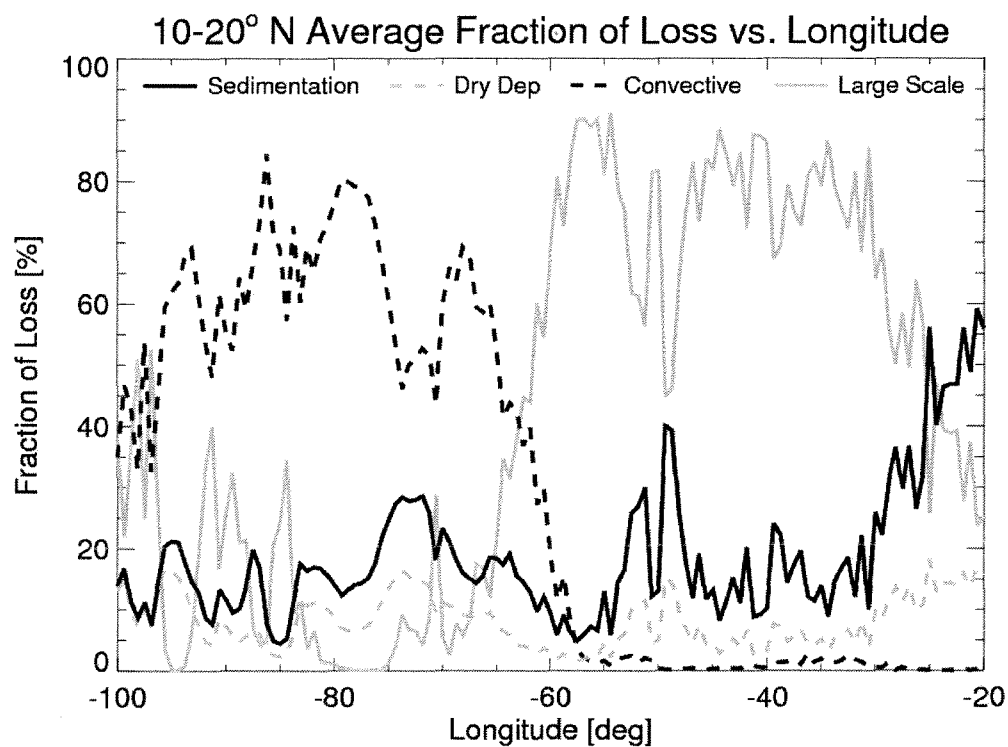
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902 Figure 10.

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907 Figure 10. 10°-20° N July 2007 average significance of sedimentation, dry deposition,
908 large scale scavenging, and convective scavenging.

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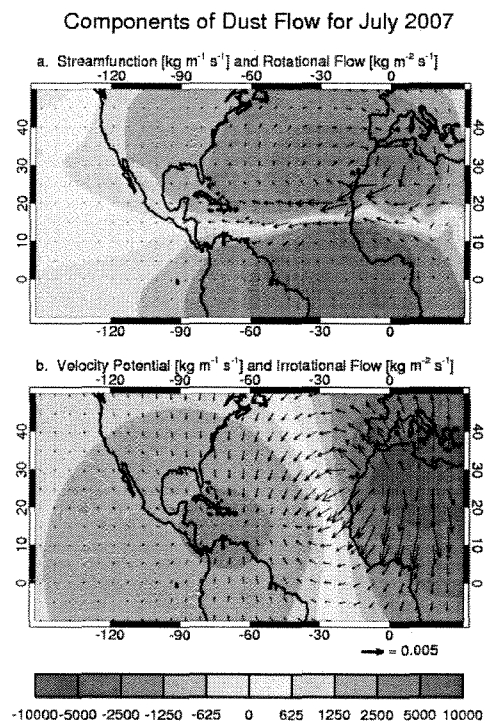
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915 Figure 11.

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919 Figure 11. July 2007 mean streamfunction (a) and velocity potential (b). Rotational (top)

920 and irrotational (bottom) flows are indicated by vectors.

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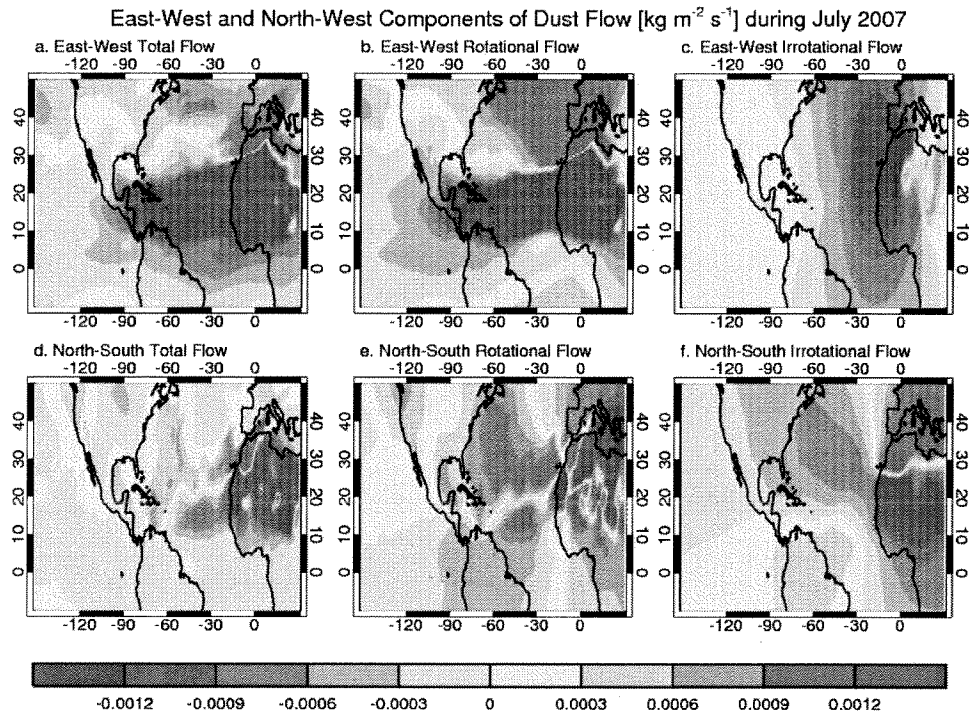
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928 Figure 12.

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933 Figure 12. East-west (top) and north-south components (bottom) of the total (left),

934 rotational (center), and divergent (right) flow.

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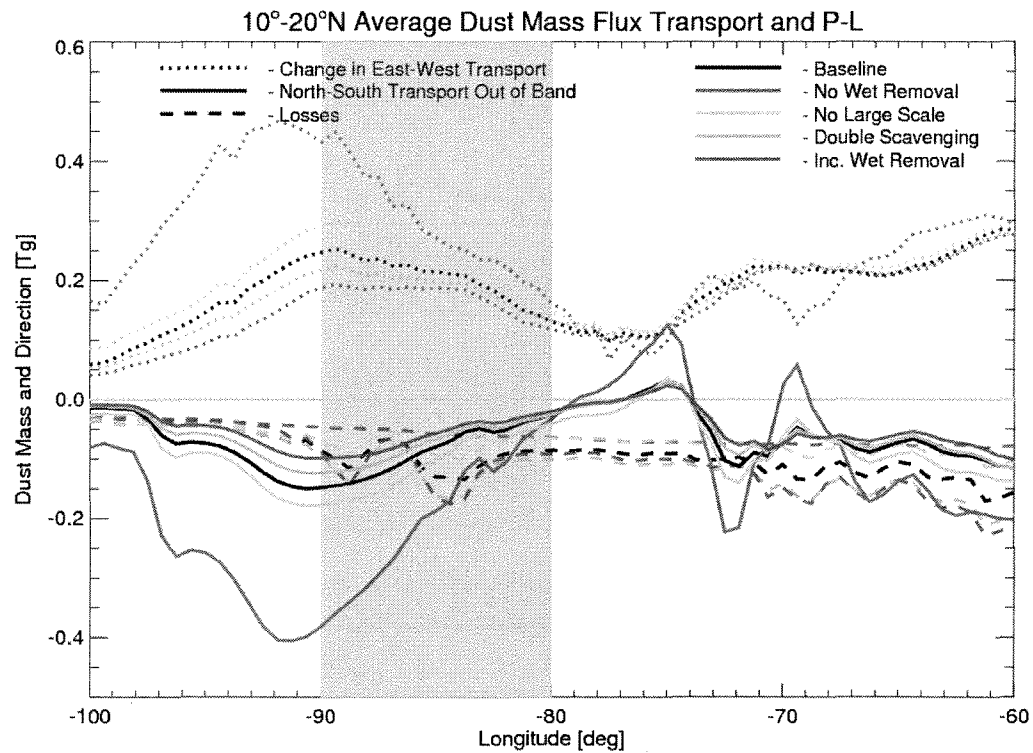
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941 Figure 13.

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945 Figure 13. 10°-20° N July 2007 mass budget for our baseline, no wet removal, no large-

946 scale scavenging, doubled scavenging, wet removal treated as other aerosols sensitivity

947 tests. Shaded region indicates integration region for the Central American dust barrier.

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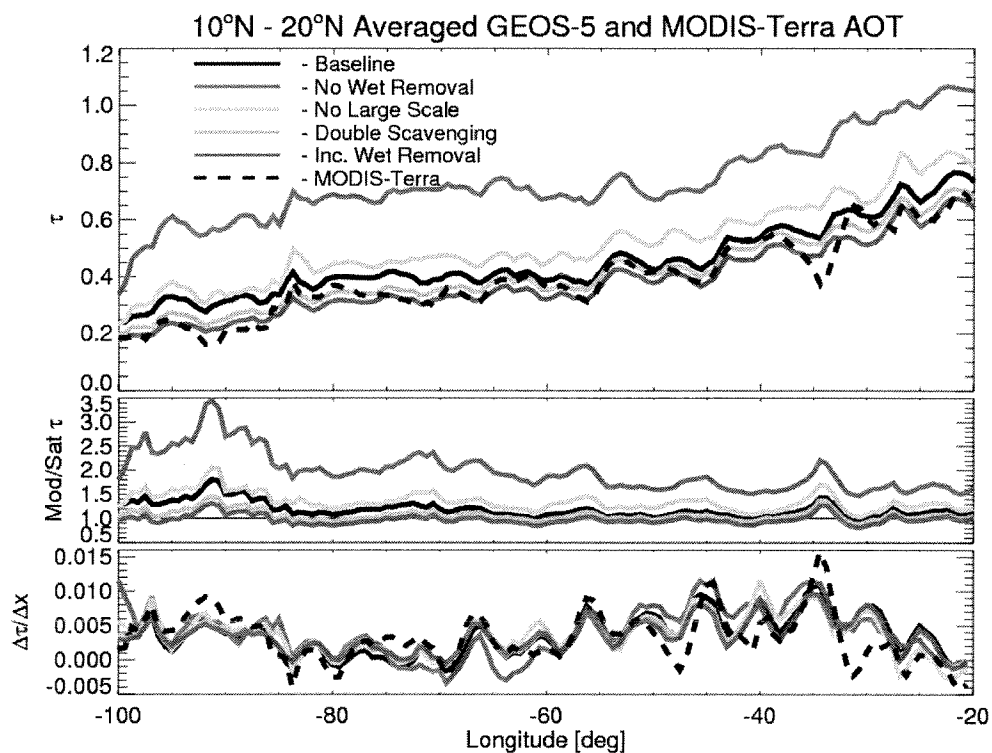
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954 Figure 14.

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959 Figure 14. 10° N - 20° N averaged AOT, model to satellite AOT ratio, and AOT slope

960 for MODIS-Terra and sampled baseline, no wet removal, no large-scale scavenging,

961 doubled scavenging, and wet removal treated as other aerosols sensitivity tests. The thin

962 black line indicates the one-to-one line for ratio plots.

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967 **Table 1**

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Experiment/Satellite	Net Northward Mass Transport (Tg) and Barrier Contribution (%)	Net Mass Loss from Removal (Tg) and Barrier Contribution (%)	80° W, 90° W, and Net Change in Westward Transport (Tg)	Barrier Efficiency		
				Mass	Total AOT	Coarse Mode AOT
1. Baseline	-1.46 47%	-1.67 53%	-7.58 -4.21 3.37	0.36	0.21	0.17
2. Doubled Convective Scavenging	-1.24 39%	-1.90 61%	-6.51 -3.40 3.11	0.48	0.25	0.21
3. Wet Removal Treated As Other Aerosols	-1.02 34%	-1.97 66%	-5.42 -2.60 2.82	0.52	0.28	0.22
4. No Large Scale Scavenging	-1.85 59%	-0.95 41%	-9.76 -6.88 3.87	0.33	0.19	0.16
5. No Wet Removal	-3.35 78%	-1.28 22%	-18.73 -13.96 4.77	0.25	0.17	0.13
6. MODIS-Terra					0.37	0.30

972

973 Table 1. Net northward mass transport and mass loss from removal and relative
 974 contribution, westward mass transport at entrance and exit of barrier region, and mass,
 975 total AOT, and coarse mode barrier efficiencies for all simulations and MODIS-Terra.

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1002 **References**

1003

1004 Adler, R. F., Huffman, G. J., Chang, A., Ferraro, R., Xie, P.-P., Janowiak, J., Rudolf, B.,
 1005 Schneider, U., Curtis, S., Bolvin, D., Gruber, A., Susskind, J., Arkin, P., and
 1006 Nelkin, E.: The Version-2 Global Precipitation Climatology Project (GPCP)
 1007 Monthly Precipitation Analysis (1979–Present). *J. Hydrometeor.*, 4, 1147–1167,
 1008 2003.

1009 Brown, R. A.: *Fluid Mechanics of the Atmosphere*, Academic Press Inc., San Diego, CA,
 1010 1991.

1011 Carlson, T. N., and Prospero, J. M.: The large-scale movement of Saharan air outbreaks
 1012 over the northern equatorial Atlantic, *J. Appl. Meteorol.*, 11, 283-297, 1972.

1013 Chin, M., Ginoux, P., Kinne, S., Torres, O., Holben, B. N., Duncan, B. N., Martin, R. V.,
 1014 Logan, J. A., Higurashi, A., and Nakajima, T: Tropospheric aerosol optical
 1015 thickness from the GOCART model and comparisons with satellite and sun
 1016 photometer measurements, *J. Atmos. Sci.*, 59(3), 461-483, 2002.

1017 Colarco, P. R., Toon, O.B., Reid, J. S., Livingston, J. M., Russell, P. B., Redemann J.,
 1018 Schmid, B., Maring, H. B., Savoie, D., Welton, E. J., Campbell, J. R., Holben, B.
 1019 N., and Levy, R.: Saharan dust transport to the Caribbean during PRIDE: 2.
 1020 Transport, vertical profiles, and deposition in simulations of in situ and remote
 1021 sensing observations, *J. Geophys. Res.*, 108(D19), 8590,
 1022 doi:10.1029/2002JD002659, 2003.

1023 Colarco, P. R., da Silva, A., Chin, M., and Diehl, T.: Online simulations of global
 1024 aerosol distributions in the NASA GEOS-4 model and comparisons to satellite

1025 and ground-based aerosol optical depth, *J. Geophys. Res.*, 115, D14207,
1026 doi:10.1029/2009JD012820, 2010.

1027 Cook, K. H.: Generation of the African Easterly jet and its role in determining
1028 West African precipitation. *J. Climate*, 12, 1165-1184, 1999.

1029 DeMott, P. J., Sassen, K., Baumgardner, D., Rogers, D. C., Brooks, S. D., Prenni, A. J.,
1030 and Kreidenweis, S. M.: African dust aerosols as atmospheric ice nuclei,
1031 *Geophys. Res. Lett.*, 30(14), 1732, doi:10.1029/2003GL017410, 2003.

1032 Desboeufs, K.V., Losno, R., and Colin, J.: Factors influencing aerosol solubility
1033 during cloud processes, *Atmos. Environ.*, 35, 3529-3527, doi:10.1016/S1352-
1034 2310(00)00472-6, 2001.

1035 Dubovik, O., and King, M. D.: A flexible inversion algorithm for retrieval of
1036 aerosol optical properties from sun and sky radiance measurements, *J. Geophys.*
1037 *Res.*, 105(D16), 20673-20696, 2000.

1038 Duce, R. A., and Tindale, N. W: Atmospheric transport of iron and its deposition in the
1039 ocean, *Limnol. Oceanogr.*, 36, 1715–1726, 1991.

1040 Dunion, J. P., and Velden, C. S.: The impact of the Saharan air layer on Atlantic
1041 tropical cyclone activity, *Bull. Am. Meteorol. Soc.*, 85(3), 353 – 365, 2004.

1042 Falkowski, P. G., Barber, R. T., and Smetacek, V.: Biogeochemical controls and
1043 feedbacks on primary production, *Science*, 281, 200-206, 1998.

1044 Ginoux, P., Chin, M., Tegen, I., Prospero, J. M., Holben, B. N., Dubovik, O., and Lin,
1045 S.-J.: Sources and distributions of dust aerosols simulated with the GOCART
1046 model, *J. Geophys. Res.*, 106(D17), 20255-20273, 2001.

1047 Hand, J., Mahowald, N., Chen, Y., Siefert, R., Lou, C., Subramaniam, A., and Fung, I.:

1048 Estimates of soluble iron from observations and a global mineral aerosol model:
 1049 Biogeochemical implications, *J. Geophys. Res.*, 109, D17205,
 1050 doi:10.1029/2004JD004574, 2004.

1051 Haywood, J., P. Francis, S. Osborne, M. Glew, N. Loeb, E. Highwood, D. Tanré, G.
 1052 Myhre, P. Formenti, and E. Hirst (2003),
 1053 Haywood, J., Francis, P., Osborne, S., Glew, M., Loeb, N., Highwood, E., Tanré, D.,
 1054 Myhre, G., Formenti, P., and Hirst, E.: Radiative properties and direct radiative
 1055 effect of Saharan dust measured by the C-130 aircraft during SHADE: 1. Solar
 1056 spectrum, *J. Geophys. Res.*, 108(D18), 8577, doi:10.1029/2002JD002687, 2003.

1057 Hill, C., DeLuca, C., Balaji, V., Suarez, M., da Silva, A., and the ESMF Joint
 1058 Specification Team: The Architecture of the Earth System Modeling Framework,
 1059 *Computing in Sci. and Eng.*, 6, 1-6, 2004.

1060 Holben, B. N., Tanré, D., Smirnov, A., Eck, T. F., Slutsker, I., Abuhassan, N., Newcomb,
 1061 W. W., Schafer, J. S., Chatenet, B., Lavenu, F., Kaufman, Y. J., Vande Castle, J.,
 1062 Setzer, A., Markham, B., Clark, D., Frouin, R., Halthore, R., Karneli, A., O'Neill,
 1063 N. T., Pietras, C., Pinker, R. T., Voss, K., and Zibordi, G.: An emerging ground-
 1064 based aerosol climatology: Aerosol optical depth from AERONET, *J. Geophys.*
 1065 *Res.*, 106(D11), 12067-12097, 2001.

1066 Huffman, G. J., R. F. Adler, M. Morrissey, D. Bolvin, S. Curtis, R. Joyce, B. McGavock,
 1067 and J. Susskind
 1068 Huffman, G. J., Adler, R. F., Morrissey, M., Bolvin, D., Curtis, S., Joyce, R., McGavock,
 1069 B., and Susskind, J.: Global Precipitation at One-Degree Daily Resolution from
 1070 Multisatellite Observations. *J. Hydrometeor.*, 2, 36–50, 2001.

1071 Jickells, T. D., An, Z. S., Andersen, K. K., Baker, A. R., Bergametti, G., Brooks, N., Cao,
 1072 J. J., Boyd, P. W., Duce, R. A., Hunter, K. A., Kawahata, H., Kubilay, N.,
 1073 LaRoche, J., Liss, P. S., Mahowald, N., Prospero, J. M., Rigwell, A. J., Tegen, I.,
 1074 and Torres, R.: Global iron connections between desert dust, ocean
 1075 biogeochemistry, and climate, *Science*, 308(5718), 67-71, 2005.

1076 Karyampudi, V. M., Palm, S. P., Reagan, J. A., Fang, H., Grant, W. B., Hoff, R. M.,
 1077 Moulin, C., Pierce, H. F., Torres, O., Browell, E. V., and Melfi, S. H.: Validation
 1078 of the Saharan dust plume conceptual model using lidar, Meteosat, and ECMWF
 1079 data, *Bull. Am. Meteorol. Soc.*, 80, 1045-1075, 1999.

1080 Kaufman, Y. J., Koren, I., Remer, L. A., Tanré, D., Ginoux, P., and Fan, S.: Dust
 1081 transport and deposition observed from the Terra-Moderate Resolution Imaging
 1082 Spectroradiometer (MODIS) spacecraft over the Atlantic Ocean, *J. Geophys.*
 1083 *Res.*, 110, D10S12, doi:10.1029/2003JD004436, 2005.

1084 Kieber, R. J., Willey, J. D., and Avery, G. B.: Temporal variability of rainwater
 1085 iron speciation at the Bermuda Atlantic Time Series Station, *J. Geophys. Res.*,
 1086 108(C8), 3277, doi:10.1029/2001JC001031, 2003.

1087 Lau, W. K.M., and Kim, K.-M.: How Nature Foiled the 2006 Hurricane Forecasts,
 1088 *Eos Trans. AGU*, 88(9), doi:10.1029/2007EO090002, 2007.

1089 Levy, R. C., Remer, L.A., Kleidman, R. G., Mattoo, S., Ichoku, C., Kahn, R., and Eck, T.
 1090 F.: Global evaluation of the Collection 5 MODIS dark-target aerosol products
 1091 over land, *Atmos. Chem. Phys.*, 10, 10399-10420, doi:10.5194/acp-10-10399-
 1092 2010, 2010.

1093 Mahowald, N. M., Baker, A. R., Bergametti, G., Brooks, N., Duce, R. A., Jickells, T. J.,

1094 Kubilay, N., Prospero, J. M., and Tegen, I.: Atmospheric global dust cycle and
 1095 iron inputs to the ocean, *Global Biogeochem. Cycles*, 19, GB4025,
 1096 doi:10.1029/2004GB002402, 2005.

1097 McGill, M. J., Hlavka, D. L., Hart, W. D., Scott, V. S., Spinhirne, J. D., , and Schmid, B.:
 1098 The cloud physics lidar: Instrument description and initial measurement
 1099 results, *Appl. Opt.*, 41, 3725 – 3734, 2002.

1100 McGill, M. J., Hlavka, D. L., Hart, W. D., Welton, E. J., and Campbell, J. R.:
 1101 Airborne lidar measurements of aerosol optical properties during SAFARI-2000,
 1102 *J. Geophys. Res.*, 108(D13), 8493, doi:10.1029/2002JD002370, 2003.

1103 McGill, M. J., Li, L., Hart, W. D., Heymsfield, G. M., Hlavka, D. L., P. E., Tian, L.,
 1104 Vaughan, M. A., and Winker, D. M.: Combined lidar-radar remote sensing:
 1105 Initial results from CRYSTAL-FACE, *J. Geophys. Res.*, 109, D07203,
 1106 doi:10.1029/2003JD004030, 2004.

1107 Nowottnick, E., Colarco, P., Ferrare, R., Chen, G., Ismail, S., Anderson, B., and Browell,
 1108 E.: Online simulations of mineral dust aerosol distributions: Comparison to
 1109 NAMMA observations and sensitivity to dust emission parameterization, *J.*
 1110 *Geophys. Res.*, 115, D03202, doi:10.1029/2009JD012692, 2010.

1111 Pawson, S., Stolarski, R. S., Douglass, A. R., Newman, P. A., Nielsen, J. E., Frith, S. M.,
 1112 and Gupta, M. L.: Goddard Earth Observing System chemistry-climate
 1113 model simulations of stratospheric ozone-temperature coupling between 1950 and
 1114 2005, *J. Geophys. Res.*, 113, D12103, doi:10.1029/2007JD009511, 2008.

1115 Petit, R. H., Legrand, M., Jankowiak, J., Molinié, J., Asselin de Beauville, C., Marion, G.,
 1116 and Mansot, J. L.: Transport of Saharan dust over the Caribbean Islands:

1117 Study of an event, J. Geophys. Res., 110, D18S09, doi:10.1029/2004JD004748,
 1118 2005.

1119 Pfister, L., Selkirk, H. B., Starr, D. O., Rosenlof, K., and Newman P. A.: A
 1120 meteorological overview of the TC4 mission, J. Geophys. Res., 115, D00J12,
 1121 doi:10.1029/2009JD013316, 2010.

1122 Prospero, J. M., Ginoux, P., Torres, O., Nicholson, S. E., and Gill, T. E.:
 1123 Environmental characterization of global sources of atmospheric soil dust
 1124 identified with the Nimbus 7 Total Ozone Mapping Spectrometer (TOMS)
 1125 absorbing aerosol product, Reviews of Geophysics, 40(1), 1002,
 1126 doi:10.1029/2000RG000095, 2002.

1127 Prospero, J., and Lamb, P.: African droughts and dust transport to the Caribbean:
 1128 climate change implications, Science, 302, 1024 – 1027, 2003.

1129 Remer, L. A., Kaufman, Y. J., Tanré, D., Mattoo, S., Chu, D. A., Martins, J. V., Li, R.-R.,
 1130 Ichoku, C., Levy, R. C., Kleidman, R. G., Eck, T. F., Vermote, E., and Holben, B.
 1131 N.: The MODIS aerosol algorithm, products, and validation, J. Atmos. Sci.,
 1132 62(4), 947-973, 2005.

1133 Rienecker, M. M., Suarez, M. J., Todling, R., Bacmeister, J., Takacs, L., Liu, H.-C., Gu,
 1134 W., Sienkiewicz, M., Koster, R. D., Gelaro, R., Stajner, I., and Nielsen, J. E.: The
 1135 GEOS- 5 Data Assimilation System -Documentation of Versions 5.0.1, 5.1.0, and
 1136 5.2.0. Technical Report Series on Global Modeling and Data Assimilation 104606,
 1137 27, 2008.

1138 Rienecker, M., et al., MERRA – NASA’s Modern-Era Retrospective Analysis for
 1139 Research and Applications. J. Climate, Submitted, 2011.

1140 Rosenfeld, D., Rudich, Y., and Lahav, R.: Desert dust suppressing precipitation: A
 1141 possible desertification loop, *Proceedings of the National Academy of Sciences*,
 1142 98(11), 5975-5980, 2001.

1143 Smirnov, A, Holben, B. N., Eck, T. F., Dubovik, O., and Slutsker, I.: Cloud-
 1144 screening and quality control algorithms for the AERONET database, *Remote*
 1145 *Sens. Environ.*, 73(3), 337-349, 2000.

1146 Sokolik, I. N., and Toon, O. B.: Direct radiative forcing by anthropogenic airborne
 1147 mineral aerosols, *Nature*, 381(6584), 681-683, 1996.

1148 Toon, O. B., Starr, D. O., Jensen, E. J., Newman, P. A., Platnick, S., Schoeberl, M. R.,
 1149 Wennberg, P. O., Wofsy, S. C., Kurylo, M. J., Maring, H., Jucks, K. W., Craig,
 1150 M. S., Vasques, M. F., Pfister, L., Rosenlof, K. H., Selkirk, H. B., Colarco, P. R.,
 1151 Kawa, S. R., Mace, G. G., Minnis, P., and Pickering, K. E.: Planning,
 1152 implementation, and first results of the Tropical Composition, Cloud and Climate
 1153 Coupling Experiment (TC4), *J. Geophys. Res.*, 115, D00J04,
 1154 doi:10.1029/2009JD013073, 2010.

1155 Vaughan, M. A. , Young, S., Winker, D., Powell, K., Omar, A., Liu, Z., Hu, Y., and
 1156 Hostetler, C.: Fully automated analysis of space-based lidar data: An overview of
 1157 the CALIPSO retrieval algorithms and data products, *Proc. SPIE Int. Soc. Opt.*
 1158 *Eng.*, 5575, 16-30, 2005.

1159 Zhu, X. R., Prospero, J. M., and Millero, F. J.: Diel variability of soluble Fe(II) and
 1160 soluble total Fe in North African dust in the trade winds at Barbados, *J. Geophys.*
 1161 *Res.*, 102(D17), 21,297–21,305, doi:10.1029/97JD01313, 1997.

1162 Zhu, A., Ramanathan, V., Li, F., and Kim, D.: Dust plumes over the Pacific, Indian, and
1163 Atlantic oceans: Climatology and radiative impact, J. Geophys. Res., 112(D16),
1164 D16208, doi:10.1029/2007JD008427, 2007.
1165
1166