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1 **Analysis of radiative properties and direct radiative forcing**
2 **estimates of dominant aerosol clusters over an urban-desert**
3 **region in West Africa**

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25 **Abstract**

26 The strategic location of the AERONET site (Ilorin) makes it possible to obtain information
27 on several aerosol types and their radiative effects. The strong reversal of wind direction
28 occasioned by the movement of the ITCZ during the West Africa Monsoon (WAM) plays a
29 major role in the variability of aerosol nature at this site. Aerosol optical depth (AOD) (675
30 nm) and Angstrom exponent (AE) (440-870 nm) with 1st and 99th percentile values of 0.08
31 and 2.16, and 0.11 and 1.47, respectively, confirms the highly varying nature of aerosol at
32 this site. Direct radiative forcing (DRF) and radiative forcing efficiency (RFE) of aerosol as
33 retrieved from the AERONET sun-photometer measurements are estimated using radiative
34 transfer calculations for the period 2005-2009 and 2011-2015. The DRF and RFE of
35 dominant aerosol classes - desert dust (DD), biomass burning (BB), urban (UB) and gas
36 flaring (GF) - have been estimated. Median (\pm standard deviation) values of DRF at top-of-
37 atmosphere (TOA) for the DD, BB, UB and GF aerosol classes are $-27.5 \pm 13.2 \text{ Wm}^{-2}$, -
38 $27.1 \pm 8.3 \text{ Wm}^{-2}$, $-11.5 \pm 13.2 \text{ Wm}^{-2}$ and $-9.6 \pm 8.0 \text{ Wm}^{-2}$, respectively. While that of RFE for
39 DD, BB, UB and GF aerosol classes are $-26.2 \pm 4.1 \text{ Wm}^{-2}\delta^{-1}$, $-35.2 \pm 4.6 \text{ Wm}^{-2}\delta^{-1}$, -31.0 ± 8.4
40 $\text{Wm}^{-2}\delta^{-1}$ and $-37.0 \pm 10.3 \text{ Wm}^{-2}\delta^{-1}$, respectively. The DD aerosol class showed the largest
41 DRF but the smallest RFE, arguably, due to the high SSA and asymmetry factor values for
42 this aerosol type. Its smallest AOD notwithstanding, the GF class could cause more
43 perturbation to the Earth-Atmosphere system in the sub-region both directly and indirectly
44 possibly due to the presence of black carbon and other co-emitted aerosol and the ageing of
45 the GF aerosols. This study presents the first estimate of DRF for aerosols of gas flaring
46 origin and shows that its radiative potential can be of similar magnitude to biomass burning,
47 and urban aerosol in West Africa.

48

49

50 **1 Introduction**

51 Atmospheric aerosols perturb the Earth's radiative energy balance both indirectly and directly
52 on regional and global scales (Charlson et al., 1992; Haywood and Shine, 1995; Rana et al.,
53 2009). The ability of the aerosols to alter the amount of radiation depends on their
54 concentration, composition, and particle size distribution (Verma et al., 2017). All of these
55 determining factors vary significantly with aerosol sources. Increased concentrations of
56 anthropogenic aerosols in the atmosphere since the pre-industrial times has been suggested to
57 be partly responsible for the onset of global warming (IPCC, 2013).

58 When perturbation of the radiative budget of the Earth-atmosphere system results from the
59 scattering and absorption of incoming solar radiation by atmospheric aerosol, the resulting
60 radiative forcing is termed Direct Radiative Forcing (DRF). When atmospheric aerosols
61 absorb radiation, they eventually dissipate such radiation, thereby altering the microphysical
62 properties and lifetime of clouds, which invariably affect precipitation. Forcing resulting
63 from such alterations is termed Indirect Radiative Forcing (IPCC, 2013). The contribution of
64 aerosol to the total forcing due to well-mixed greenhouse gases is still associated with large
65 uncertainties (Myhre, 2013).

66 The West African climate has a unique weather pattern due to the West African Monsoon
67 (WAM) which is characterised by large-scale seasonal reversals of wind regimes (Sultan and
68 Janicot, 2000; Barry and Chorley, 2009). The movement of the Intertropical Convergence
69 Zone (ITCZ) and Intertropical front (ITF) are responsible for the seasonal reversal of the
70 prevailing wind pattern in the region. Deep convection occurs in organised systems referred
71 to as Mesoscale Convective System (MCS) (Mathon and Laurent, 2001). MCS associated
72 with the ITCZ can lead to rapid uplift and large scale redistribution of aerosols (Reeves et al.,
73 2010).

74 In recent years, in the West Africa region, anthropogenic emissions of aerosols and gaseous
75 pollutants have increased substantially, largely due to increasing population and industrialisa-
76 tion; a trend expected to continue until 2030 (Liousse et al., 2014). Dominant anthropogenic
77 sources of aerosol at the study site are fossil fuel combustion, vehicular emission, biomass
78 burning, and industrial emission while the dominant natural aerosol source is desert dust.
79 Despite growing evidences in support of the impacts of anthropogenic aerosols on regional
80 radiative budget, strict regulations on emissions are still not available in major African cities
81 and, where they are available, they are very weak (Liousse et al., 2012).

82 In Nigeria, the various wind patterns and seasons are associated with different dominant
83 aerosol types. While the North-easterly Harmattan (NEH) wind, pre-dominant in the dry
84 season (November- February), brings desert dust and biomass burning aerosols into the
85 region. The South-westerly Monsoon (SWM) wind, associated with the onset of the WAM
86 (April-October), brings predominantly urban and industrial aerosol which are believed to
87 contain more carbonaceous aerosols (Knippertz et al., 2015). The properties and concentra-
88 tions of these aerosol types vary significantly with the wind pattern. Studies of atmospheric
89 aerosol and their radiative effects are very scarce in Nigeria.

90 In this study, the radiative properties of key aerosol types at this urban-desert station were
91 analysed and their DRF and RFE at the TOA was estimated to provide the first estimate of
92 climate forcing of gas-flaring aerosols in the region.

93 **2 Methodology**

94 **2.1 Description of the AERONET site and prevailing climatic condition**

95 The Ilorin AERONET site (8.32° N, 4.34° E) is located at a site between the densely
96 populated monsoonal forest region of the south and Sahel Savannah region of the north.
97 There is pronounced variation in the climatic conditions of the region governed by the
98 movement of the intertropical convergence zone (ITCZ) and intertropical front (ITF), which

99 are responsible for the seasonal reversal of the wind direction (the West African Monsoon
100 (WAM)).

101 The WAM is a coupled atmosphere-ocean-land system which is characterised by summer
102 rainfall and winter drought (Lafore et al., 2010). The rainfall in the West African sub-region
103 results essentially from the northward movement of the low-level monsoon airflow from
104 March to August and the southward retreat from September to November. At their northern-
105 most position, the humid monsoonal wind from the south meet drier and warmer air to form
106 the ITF (Cornforth, 2012). During the dry season (November - March), the West Africa sub-
107 region experiences strong emissions of pollutants resulting from extensive biomass burning
108 of vegetation often from land preparation for the incoming planting season. During the wet
109 season (May to October), the region is strongly influenced by mesoscale convective systems
110 (MCS), which affects the compositions of the atmosphere through several ways including
111 rapid vertical transport of aerosols to the upper troposphere (Law et al., 2010; Mari et al.,
112 2011).

113 Gas flaring is a prominent and persistent source of atmospheric aerosols which includes soot
114 (black carbon), SO₂, CO, NO_x (NO + NO₂), PAH and VOCs, especially in the oil-rich
115 regions of the world (Fawole et al., 2016a). There are over 300 active flare sites in the region
116 where an estimated 23.7 (44.4 metric tons of CO₂ equivalent) and 15.1 (28.3 metric ton of
117 CO₂ equivalent) billion cubic meters (bcm) of natural gas is flared in 2006 and 2008,
118 respectively (Elvidge et al., 2011; Fawole et al., 2016a). In 2012, of the 325 active flare sites
119 identified in the Nigeria oil field, 97 (~ 30 %) ranked among the top 1000 largest flares out
120 of the 7467 identified globally (Elvidge et al., 2015).

121 2.2 Trajectory calculation and classification

122 Seven-day (168 hours) back trajectories were calculated using the UK's Universities Global
123 Atmospheric Modelling Programme (UGAMP) offline trajectory model. The model is driven
124 by six-hourly ERA-Interim (European Centre for Medium-Range Weather Forecasts Interim
125 Re-Analysis) wind analyses data. The trajectories of particles are calculated backward in time
126 by interpolating these wind analysis to the current particle position. The position (latitude,
127 longitude) and pressure were output every trajectory time step of 0.6 hours. The choice of 7-
128 day back trajectory length is due to the atmospheric lifetime of between 5 and 9 days
129 estimated for black carbon (BC) and particulate organic matter (POM), respectively (Cooke
130 and Wilson, 1996; Cooke et al., 1997; Stier et al., 2006; Koch et al., 2009). Both BC and
131 POM are major constituent of aerosol in the study area.

132 As shown in several studies, for example, Bibi et al. (2016) and Alam et al. (2016), atmospheric
133 aerosols could be clustered using the inter-relationships between different pairs of their
134 microphysical and optical properties. Using similar techniques, prominent aerosol classes were
135 identified at the study site in Fawole et al. (2016b) as: Biomass burning (BB), Desert dust (DD),
136 Urban (UB) and Gas flaring (GF) aerosols. In terms of optical and microphysical properties, these
137 classes vary significantly (Fawole et al., 2016b); mixed classes (DD-BB, DD-UB, GF-UB and
138 GF-DD) were also identified. Using similar clustering technique of analysing aerosol optical and
139 microphysical properties, Bibi et al. (2016) classified aerosol in the Indo-Gangetic plains into
140 dust, biomass and urban/industrial aerosol classes.

141 In this study, the properties of the single-source dominant classes was analysed to estimate their
142 direct radiative forcing and forcing efficiency. For details of the trajectory classification and
143 analysis of the variation of the optical and microphysical properties of the identified aerosol
144 classes see Fawole et al. (2016b).

145 2.3 AERONET data analysis

146 The absolute magnitude of aerosol radiative forcing is determined, predominantly, by the
147 values of aerosol optical depth (AOD) and single scattering albedo (SSA), while its sign is
148 dependent on the SSA and surface albedo. Both AOD and SSA vary significantly with the
149 source of the aerosol (Pani et al., 2016). In this study, Version 3 Level 1.5 of AERONET data
150 released in January 2018 to which improved cloud screening and new quality controls have
151 been applied were used to estimate the DRF and RFE of anthropogenic and natural aerosol
152 classes in the West Africa sub-region. Adequate knowledge of aerosol SSA, hemispheric
153 backscatter fraction (b) and AOD can be used to calculate the mean TOA aerosol radiative
154 forcing for optically thin, partially absorbing aerosol (Haywood and Shine, 1995). For sites
155 like Ilorin, where differences in the diurnal variation of aerosol properties (extensive and
156 intensive) could be highly pronounced, the use of monthly averages of aerosol parameters
157 will only provide highly generalised estimates of the optical and microphysical properties of
158 the aerosol at such a site.

159 One of the key properties that determine the climate forcing ability of an aerosol is the
160 angular distribution of the light scattered by the aerosol particles (Marshall et al., 1995). The
161 angular distribution of scattered light intensity at a specific wavelength is referred to as the
162 phase function (P). The asymmetry parameter, g , an important intensive parameter of
163 aerosols for estimating its climate forcing ability could be derived from P . Values of g range
164 between -1 for entirely backscattered light to +1 for entirely forward scattered light (Andrews
165 et al., 2006). The fraction of backscattered light is the ratio of the integral of the volume
166 scattering function over the backward half solid angle divided by the integral of the volume
167 scattered function over the full solid angle (Horvath et al., 2016).

168 For AERONET retrievals, uncertainties in the direct sun measurements are within ± 0.01 for
 169 longer wavelengths greater than 440 nm and ± 0.02 for shorter wavelengths less than 440nm.
 170 AOD estimated uncertainty varies spectrally from ± 0.01 to ± 0.02 with the highest error in the
 171 ultraviolet wavelengths (Holben et al., 1998; Eck et al., 1999). For all sky radiance
 172 wavelengths (that is, 440, 675, 870, and 1020 nm), the uncertainty in SSA is expected to be
 173 ± 0.03 based on Version 1almucantar retrieval computations (Dubovik et al., 2000; Holben et
 174 al., 2006).

175 **2.3.1 Relationship between the asymmetry parameter and the backscatter fraction**

176 Several studies (e.g., (Wiscombe and Grams (1976); Marshall et al. (1995); Kokhanovsky
 177 and Zege (1997))) have attempted to parameterise the backscatter fraction (b) in terms of the
 178 asymmetry parameter (g). Studies estimating aerosol DRF have either adopted an
 179 approximate relation between b and g or look-up tables of parameterisation of aerosol optical
 180 properties such as those of Hess et al. (1998) and D'Almeida et al. (1991).

181 In this study, assuming spherical particles, approximate relations given in equation (1) as
 182 cited in Horvath et al. (2016) and equation (2) according to Delene and Ogren (2002) have
 183 been used to estimate backscatter fraction (b) and average upscatter fraction, β , respectively.

184

185
$$b = \left[1.1 \left(\frac{1}{1-g} \right)^{1.85} + 1 \right]^{-1} \dots\dots\dots (1)$$

186
$$\beta = 0.0817 + 1.8495b + 2.9682b^2 \dots\dots\dots (2)$$

187 **2.4 Aerosol Radiative Forcing**

188 **2.4.1 Estimating direct radiative forcing**

189 The direct radiative forcing (DRF), ΔF , of aerosol at the top of the atmosphere (TOA) is
 190 estimated using the expression derived by Charlson et al. (1992). According to Haywood and

191 Shine (1995) the radiative transfer equation proposed by (Charlson et al. (1992)) is simplified
192 as given in equation (3).

$$193 \quad \Delta F \approx -DS_oT_{at}^2(1 - A_c)\omega\bar{\beta}\bar{\delta} \times \left((1 - R_s)^2 - \frac{2R_s}{\beta} \left(\frac{1}{\omega} - 1 \right) \right) \dots\dots\dots (3)$$

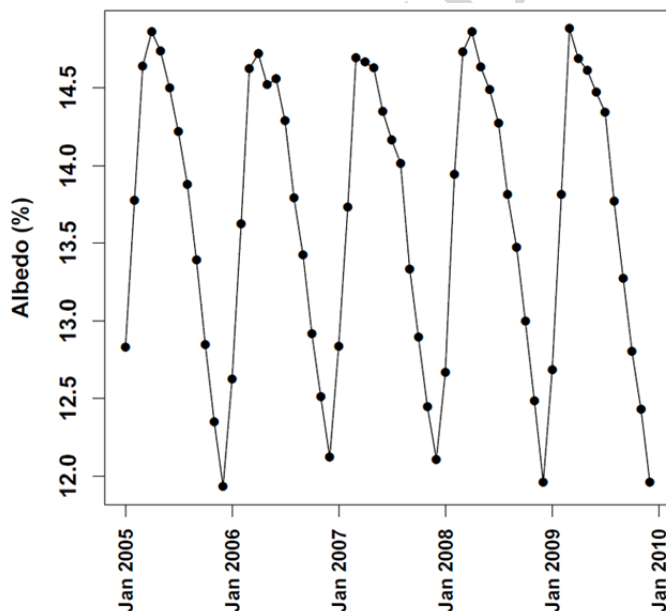
194 where D is the fractional day length, ω is the spectrally weighted single scattering albedo, S_o
195 is the Solar constant, T_{at} is the atmospheric transmission, A_c is the fractional cloud amount, R_s
196 is the surface reflectance, $\bar{\beta}$ is the spectrally weighted backscattered fraction and, $\bar{\delta}$ is the
197 spectrally weighted AOD. The critical value of SSA at which the DRF shifts from positive to
198 negative is dependent on the surface albedo and asymmetric parameter, g (Haywood and
199 Boucher, 2000; Kassianov et al., 2007). One advantage of the analytical solution for the
200 radiative transfer equation as stated above (equation (3)) over a radiative transfer model is an
201 explicit dependence on individual parameters determining the radiative forcing (Chylek and
202 Wong, 1995). As cloud cover (A_c) is a parameter in equation (3), to use the expression, the
203 assumption is that the cloud cover is above the aerosol layer which is a typical atmospheric
204 condition in the region considered in this study.

205 Schemes of wavelength-dependent aerosol parameters are time-consuming and quite complex
206 to be incorporated into radiative forcing calculations and radiative transfer codes that can
207 produce representative and accurate estimates of radiative forcing with one or two
208 wavelength regions (Blanchet, 1982). In their study to examine the possibility of replacing
209 aerosol parameters by wavelength-independent parameters and the accuracy and
210 representativeness of such average parameters for the complete solar spectrum, Blanchet
211 (1982) found out that results of calculations with average parameter are in close agreement
212 with corresponding terms at a wavelength (λ) of 700 nm. Haywood (1995), using detailed
213 radiative transfer codes, tested the representativeness of average aerosol parameter and found
214 that results at around $\lambda=700$ nm were quite similar to those of using the entire solar spectrum.

215 Hence, the use of aerosol parameters at $\lambda=675$ nm, which is the nearest to 700 nm in the
216 range of wavelengths at which aerosol parameters are measured by AERONET sun-
217 photometers, in our estimations.

218 Fractional day-length, solar constant and atmospheric transmittance are assumed to be 0.5,
219 1370 W m^{-2} and 0.76, respectively (Haywood and Shine, 1995). To estimate DRF and RFE
220 of the different aerosol classes identified in Fawole et al. (2016b), monthly mean values of
221 cloud amount (A_c) was obtained from the ASOS-AWOS-METAR dataset (NOAA, 1998;
222 Yang et al., 2016) for the nearest airport (Cotonou) to the site, where sufficient amount of
223 cloud cover data are available. The surface reflectance data used are model output for albedo
224 simulations for Ilorin (2005 – 2009) (R.T Pinker, personal communication, August 2016).
225 Figure 1 presents the time series for the mean monthly surface albedo for Ilorin during 2005
226 – 2009. The yearly pattern of the variation of surface albedo (reflectance) is quite similar for
227 the five-year period model output available.

228



229

230 **Figure 1:** Time series for monthly mean surface albedo at Ilorin for the period (01/2005 –
231 12/2009).

232 **2.4.2 Radiative forcing efficiency**

233 SSA and backscatter fraction of the particle can be used to calculate the TOA aerosol forcing
234 (ΔF) per unit aerosol optical depth (AOD); this is called aerosol forcing efficiency (Sheridan
235 et al., 2002; Kaufman et al., 2005). In this study, to compare the forcing potential of the
236 various aerosol classes, we estimated the forcing efficiency using equation (4). Forcing
237 efficiency, ($\Delta F/\delta$), is the aerosol radiative forcing per unit AOD. AOD is a major extensive
238 property of the aerosol, which determines the magnitude of its radiative forcing. Forcing
239 efficiency depend only on the nature and composition of the aerosol rather than its amount
240 (Sheridan and Ogren, 1999).

241
$$\frac{\Delta F}{\delta} = -DS_o T_{at}^2 (1 - A_c) \omega \bar{\beta} \times \left((1 - R_s)^2 - \frac{2R_s}{\beta} \left(\frac{1}{\omega} - 1 \right) \right) \dots\dots\dots (4)$$

242 Virkkula et al. (2014), in their study to assess the effect of aerosol from different phases of
243 biomass burning - flaming and smouldering - on the chemical and physical properties of
244 airborne aerosols in the Boreal forest, used the expression in equation (4) to estimate the
245 radiative forcing efficiency of the biomass burning aerosols. Rizzo et al. (2013), using
246 equation (4), estimated aerosol forcing efficiency over a primary forest site in Amazonia.

247 **3 Results and discussions**

248 **3.1 Climatology of aerosol properties**

249 Significant variation of aerosol optical and microphysical properties in the multiyear analysis
250 of aerosol properties at the Ilorin AERONET site is a strong indication of the varying sources
251 of aerosols at the site. The range of values, at $\lambda=675$ nm, for aerosol optical depth (AOD),
252 single scattering albedo (SSA) and asymmetry parameter (g) are 0.04 – 3.71, 0.68 – 0.99, and
253 0.58 – 0.8, respectively. The variation of aerosol properties is more pronounced between the
254 non-WAM and WAM months due to seasonal reversal of the prevailing wind direction, and
255 hence, a change of dominant sources of aerosols.

256 **3.1.1 Temporal variability of Aerosol Optical Depth, Fine Mode Fraction and**
257 **Angström exponent**

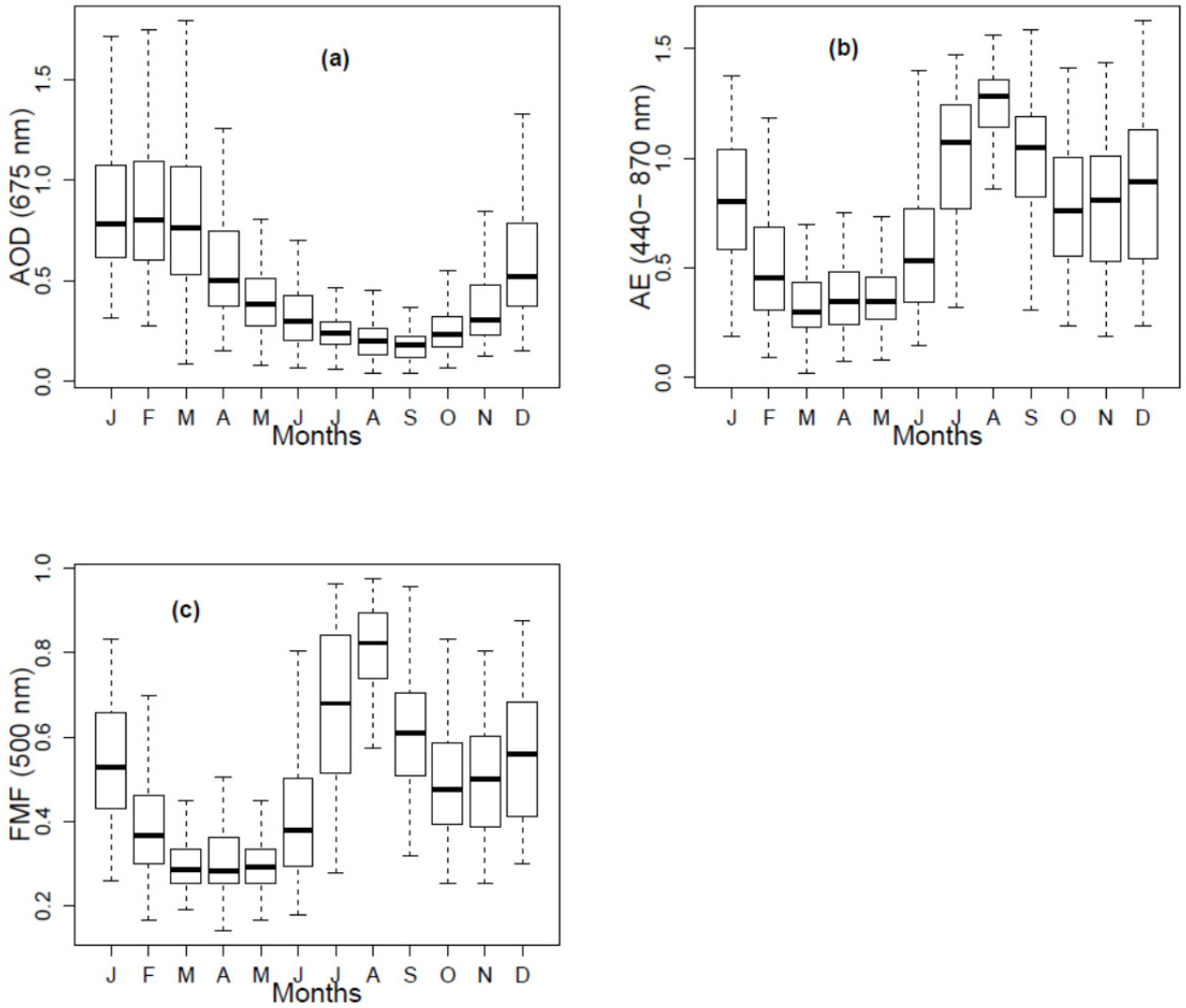
258 During the NEH months (November - February), the values of the aerosol optical depth
259 (AOD_{675}) and Angström Exponent ($AE_{440-870}$) are 1.22 ± 0.17 and 0.35 ± 0.06 , respectively.
260 These are months of intense biomass burning in the West African sub-region as well as
261 intrusion of dust from the Sahara and Sahel regions. Aerosol loading in the SWM months
262 (April - October), are characterised by lower AOD (675 nm) and high AE (440-870 nm) with
263 median values of 0.58 ± 0.23 and 1.02 ± 0.19 , respectively (see Figure 2). The monthly plots in
264 Figure 2 are obtained from average daily data of aerosol parameters. Compared to AE values
265 for similar dust sites, the relatively high average AE value of the dust aerosol in the NEH
266 months is probably due to contributions of biomass burning aerosol at that time of the year
267 (Fawole et al., 2016b).

268
269
270 The significant seasonal pattern in the AE and fine mode fraction (FMF) of aerosol at the site,
271 as seen in Figure 2(b) and 2(c), is due to varying aerosol sources and/or changes in
272 atmospheric transport. There is the dominance of fine mode aerosol fraction during the West
273 Africa Monsoon months when the prevailing wind is the moist SWM. The influx of urban-
274 industrial air is expected to predominate during the WAM months, between April and
275 October (Fawole et al., 2016b). The value of AE peaks between July and September and is
276 lowest between February and March. The lower AE values during the peak of the dry season
277 show the strong intrusion of dust in this region at that period of the year. During the WAM
278 months, the peak AE values between July and September coincide with the peak values of
279 backscatter fraction, b (see Figures 2(b) and 3(b)). This strongly suggests an increase in fine
280 particle fraction, which is attributable to inflow of urban and industrial emissions from the
281 south of the AERONET site. The variation of aerosol parameter with prevailing Monsoonal
282 wind as observed in this study has been observed in similar studies over Karachi, Pakistan

283 during the period 2006-2008 (Bibi et al., 2017) and Ahmedabad, India (Ramachandran and
284 Kedia, 2010).

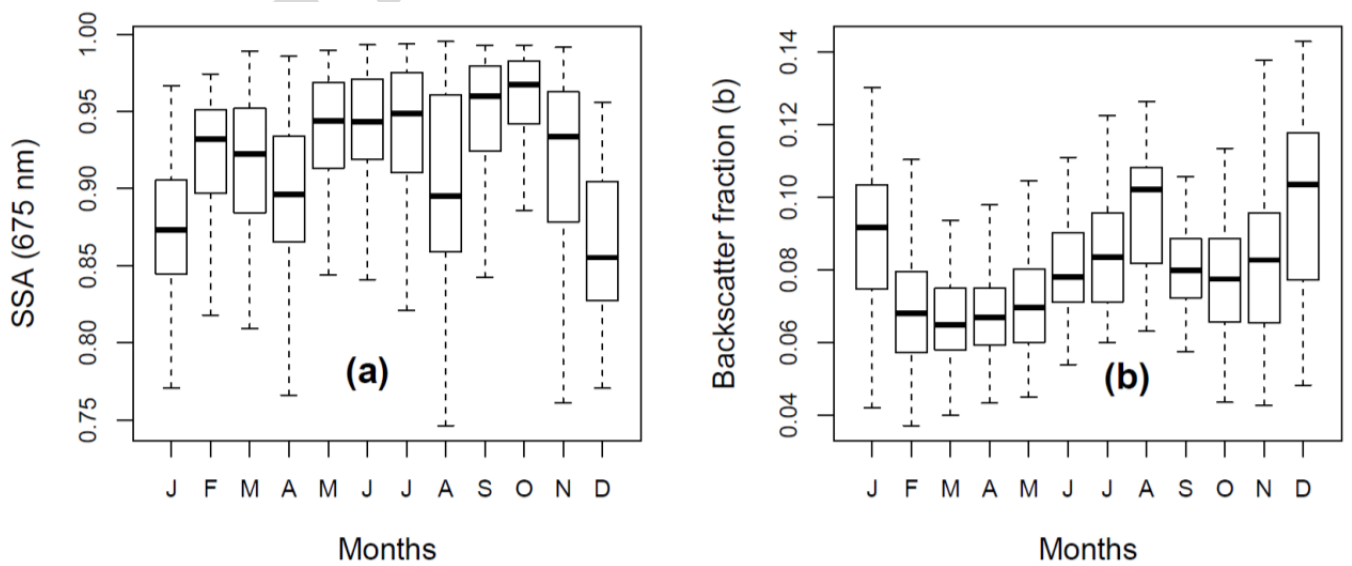
285 **3.1.2 Temporal variability of single scattering albedo (SSA) and backscatter fraction**

286 In the West Africa sub-region, there are significant differences in the relative amount of
287 scattering and absorption of aerosols at different periods of the year. These differences result
288 in the variation of SSA during the years as shown Figures 3(a). During the SWM months, as
289 shown in the multiyear mean monthly SSA values in Figures 3(a), inland flow of south-
290 westerly monsoon winds are rich in partially absorbing aerosols from the urban and industrial
291 site including gas flaring emissions from the intense gas flaring activities in the Niger Delta
292 region. In Figure 3(a), the WAM months (April –August) exhibit the widest range of SSA
293 (0.66-0.98) which could result in a wider range of DRF. This wide range of SSA could be
294 attributable to the diverse nature of aerosol in the urban and industrial emissions from the
295 south of Ilorin.



296
297
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Figure 2: Multiyear monthly variations of (a) AOD, (b) Angstrom exponent (AE), and (c) Fine Mode Fraction (FMF) of aerosol during the year 2005-2015



300
301
302

Figure 3: Multiyear average monthly (a) variation of SSA (675 nm) and (b) variation of backscatter fraction during the year 2005-2015

303 The boxplots in Figure 3(b) show a steady increase in the backscatter fraction from the lowest
304 average values of 0.07 ± 0.01 during March (peak of the NEH months) to the highest average
305 values of 0.1 ± 0.02 during the peak of the WAM months. The median values of the
306 backscatter fraction of the non-WAM and WAM months correspond to asymmetric
307 parameter (g) values of 0.72 ± 0.1 and 0.61 ± 0.1 , respectively. This arguably suggests a steady
308 increase in the concentration of fine-mode aerosol fraction during the WAM months which is
309 attributable to increased inflow of combustion aerosols from urban and industrial emissions.
310 Mie theory predicts a higher backscatter fraction for fine-mode spherical aerosol particles
311 (Andrews et al., 2011). The wide range of backscatter fraction during the non-WAM months
312 (NDJF) is due to mixture of biomass burning and intrusion of desert dust which are intense
313 during the Harmattan haze period in the region. The mean monthly value of backscatter
314 fraction in Figure 3(b) shows a bi-modal distribution with peaks during the intense biomass
315 burning season (NDJ) and the peak of the WAM months (July-August) when the ITCZ is
316 northernmost allowing enhanced inland flow of aerosol from south of the AERONET site.

317 **3.2 Variability of Angström exponent and (AOD) for the aerosol classes**

318 The median (\pm standard deviation) values of the optical and microphysical properties of the
319 identified aerosol classes are presented in Table 1. Unless otherwise stated, average values of
320 aerosol parameters are reported at 675 nm. The Angström exponents (AE) discussed for the
321 aerosol classes were estimated using the 440 nm and 870 nm wavelength pair. The highly
322 varying range of AOD_{440} and AE values; $0.07 - 3.87$ and $0.01 - 1.74$, respectively, strongly
323 suggests a broad range of contributing sources to the aerosol loading at the study site. As the
324 distribution of most of the aerosol parameters for the classes are non-Gaussian, the median
325 values are reported with the standard deviations given in brackets.

326 **3.2.1 Desert dust (DD)**

327 The DD aerosol class consists of 209 days of aerosol signals, which are predominant in the
328 NEH months and the early days of the onset of WAM months. The major source of desert
329 dust considered in this class classification is the Sahara and Sahel dust regions (13 – 18° N; 6
330 – 17° E). The median values of AOD₄₄₀ and AE₄₄₀₋₈₇₀ for this aerosol class is 1.13(±0.54) and
331 0.3(±0.12), respectively. These values agree well with those from studies for similar sites in
332 the Bodélé depression of Northern Chad (Todd et al., 2007), Indo-Gangetic plains (Bibi et al.,
333 2016) and dust regions of China (Wang et al., 2004). The average value for AOD is highest
334 for the desert dust class while AE is the least. With a median SSA value of 0.95(±0.02), this
335 class is the least absorbing.

336 **3.2.2 Urban aerosol (UB)**

337 Aerosol signature in the urban aerosol class is prominent in the WAM months when the
338 south-westerly moist monsoon wind is prevalent in the region. For this class, the median
339 AOD₄₄₀ and AE values are 0.53(±0.35) and 0.52(±0.34), respectively. Even though this AE
340 value is low, it is still higher than that for the DD aerosol class. The DD aerosol class is
341 expected to contain a higher fraction of coarse aerosol. This class of aerosol (Urban), with a
342 median value of SSA of 0.93(±0.04), is partially absorbing arguably due to increased
343 carbonaceous particle content from anthropogenic sources in the urban area.

344 **3.2.3 Gas flaring aerosol (GF)**

345 This class is similar to the urban class but has a lower median value of AOD₄₄₀ and an
346 average AE value, which is higher than that of the urban aerosol by a factor of ~2. For this
347 class, the median values of AOD₄₄₀ and AE values are 0.41(±0.26) and 1.16(±0.29),
348 respectively. A median value of SSA of 0.9(0.06) makes it more absorbing than the urban
349 aerosol class, which is attributable to it having a relatively higher carbonaceous particulate

350 content. This class is estimated to have an average Absorption Angstrom Exponent (AAE) of
351 0.98(\pm 0.25) in contrast to urban aerosols, which has an AAE value of 1.2(\pm 0.38) (Fawole et
352 al., 2016b). The gas-flaring region, south of the AERONET site, contains more than 300
353 active flares (Elvidge et al., 2015), where it is estimated that more than 25 % of the annual
354 natural gas production is flared (Elvidge et al., 2009; Ite and Ibok, 2013; Anejionu et al.,
355 2015).

356 It should, however, be noted that gas flaring, a prominent source of soot (BC), also emit other
357 aerosol including volatile organic compounds (VOCs), SO₂ and NO_x, some of which exert a
358 cooling effect on the climate (USEPA, 2012).

359 **3.2.4 Biomass burning**

360 Similar to findings from the studies by Bibi et al. (2016) and (Tiwari et al., 2016), the BB
361 aerosol class is characterised by high AOD and high AE, which is typical of biomass burning
362 sites. Although, it is prevalent almost at the same time as the desert dust season during the
363 NEH months, it can be distinguish by its lower SSA and higher AE values. For this class, the
364 median values of AOD₄₄₀ and AE are 0.93(\pm 0.3) and 1.0(\pm 0.25), respectively. The range of
365 values for AOD₄₄₀ and AE are in agreement with values reported by Ogunjobi et al. (2008)
366 and (Pace et al., 2006) for similar biomass burning sites in West Africa and around Central
367 Mediterranean, respectively. The region of biomass burning considered in the classification
368 of this class are (i) 6.5 – 11.5° N; 3° W – 3° E and (ii) 6.5 – 11.5° N; 13.6° E – 22.5° E (see
369 Fawole et al. (2016b)). The choice of these BB regions are based on data obtained from
370 MODIS active fire detection over Africa as reported by Roberts et al. (2009). In agreement
371 with reports from previous studies from similar sites, this class, with median SSA value of
372 0.86 (\pm 0.04), is the most absorbing class presumably due to its enhanced organic carbon (OC)
373 content.

374 Table 2 presents the mean monthly surface reflectance, R_s , and cloud amount, A_C used in the
 375 DRF estimation. Sources of these parameters are stated in section 2.4.1.

376

377 **Table 1:** Summary of parameter for aerosol classes used to estimate their DRF

| | Backscatter fraction (b) | AOD (675 nm) | Asymmetry parameter (675 nm) | SSA (675 nm) | DRF (Wm^{-2}) | Forcing efficiency (FE) ($\text{Wm}^{-2}\delta^{-1}$) |
|-----------|---------------------------------|---------------------|-------------------------------------|---------------------|--|---|
| DD | 0.06±0.01 | 0.91±0.44 | 0.74±0.03 | 0.97±0.02 | -30.3±13.4 | -31.0±3.3 |
| BB | 0.1±0.02 | 0.61±0.26 | 0.66±0.03 | 0.87±0.03 | -23.6±8.9 | -39.0±4.0 |
| UB | 0.08±0.02 | 0.38±0.23 | 0.70±0.03 | 0.96±0.04 | -11.7±7.5 | -32.4±5.4 |
| GF | 0.1±0.02 | 0.29±0.21 | 0.66±0.04 | 0.91±0.06 | -8.2±5.8 | -36.0±7.8 |

378

379

380 **Table 2:** Mean monthly surface reflectance and cloud amount for study site

| | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sept | Oct | Nov | Dec |
|-------------------------|-------------|-------------|-------------|-------------|-------------|-------------|-------------|-------------|-------------|-------------|-------------|-------------|
| R_s | 0.13 | 0.14 | 0.15 | 0.15 | 0.15 | 0.14 | 0.14 | 0.14 | 0.13 | 0.13 | 0.12 | 0.12 |
| Cloud amount | 0.22 | 0.35 | 0.35 | 0.35 | 0.35 | 0.35 | 0.35 | 0.35 | 0.35 | 0.35 | 0.35 | 0.15 |

381

382

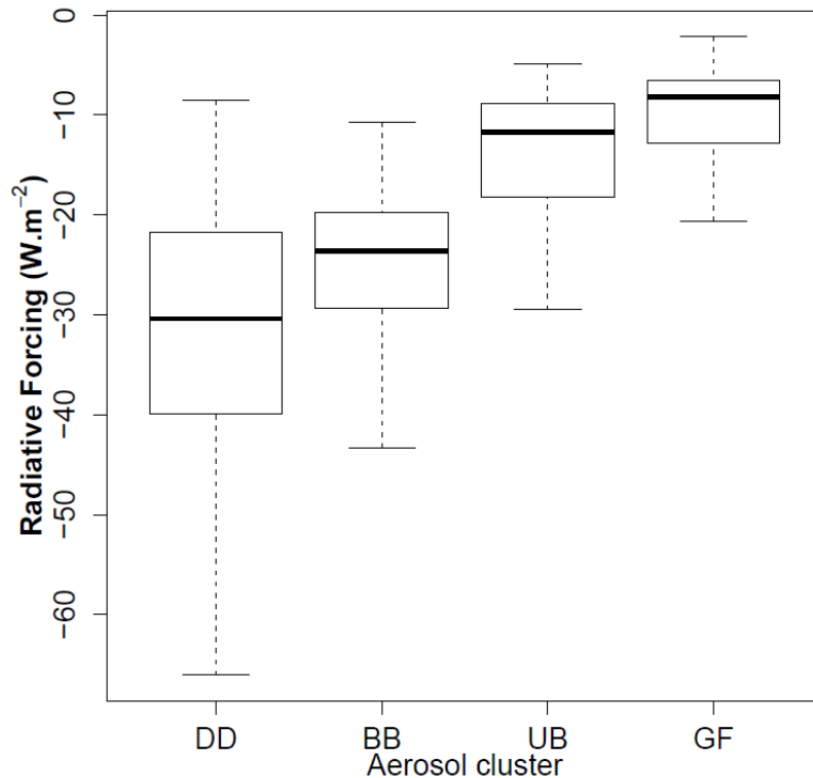
383 3.3 Aerosol radiative forcing

384 The direct radiative forcing (DRF) for each aerosol class at the TOA was estimated using the
 385 relationship in equation (3). The values of surface reflectance, R_s , and cloud amount, A_C used
 386 in the estimations range between 0.12 – 0.15 and 0.15 – 0.39, respectively. Table 1 presents,
 387 for each cluster, the range of values of AOD (τ), SSA (ω), backscatter fraction (b) used in the
 388 DRF estimations. The daily average values of these aerosol parameters were used for the
 389 DRF estimations of the four classes.

390 Figure 4 presents the variation of the DRF estimates at TOA for each aerosol cluster. The DD
 391 aerosol class with median AOD₆₇₅ value of 0.91(±0.44) has the highest DRF of -30.3±13.4

392 Wm^{-2} at TOA. Of the four classes identified, this class has the highest mean AOD. The
393 median DRF value for this class, as shown in Figure 4, is highest. It is believed that this high
394 DRF value is due to the large SSA (average value of 0.97 ± 0.02), which brings about less
395 absorption and more scattering, and the largest asymmetry factor, g , of 0.74 ± 0.03 , which
396 causes more forward scattering of incoming radiation, compared to the other aerosol classes.
397 Thus, consistent with the findings of García et al. (2012), this class has the most effective
398 cooling effect on the earth-atmosphere system at the TOA in the region.

399 The biomass burning (BB) aerosols class has an estimated DRF of $-23.6\pm 8.9 \text{ Wm}^{-2}$ at TOA
400 with average AOD_{675} value of $0.61(\pm 0.26)$. Compared to the DD class, this class has a
401 relatively shorter range of DRF. This value of DRF is comparable to the mean DRF obtained
402 by García et al. (2012) and Yoon et al. (2005) for similar biomass burning site in South
403 America and South Africa, respectively. Compared to the DD and BB aerosol classes, the
404 urban (UB) class exerts a smaller cooling effect. This class (urban) with average AOD_{675} of
405 $0.38(\pm 0.23)$, rich in anthropogenic urban aerosol is estimated to have a DRF value of -
406 $11.7(\pm 7.5) \text{ Wm}^{-2}$. In a study by Yoon et al. (2005), similar values of DRF was obtained for
407 US East Coast (Goddard Space Flight Center (GSFC)), a heavily populated urban area. With
408 median DRF value of $-8.2\pm 5.8 \text{ Wm}^{-2}$ ($\text{AOD}_{675} = 0.29\pm 0.21$), the GF class has the least
409 cooling effect at TOA. The GF class is believe to be rich in fossil fuel combustion emissions
410 including sulfate and black carbon (Fawole et al., 2016a)



411
412 **Figure 4:** Direct radiative forcing (DRF) at the TOA for the different aerosol classes

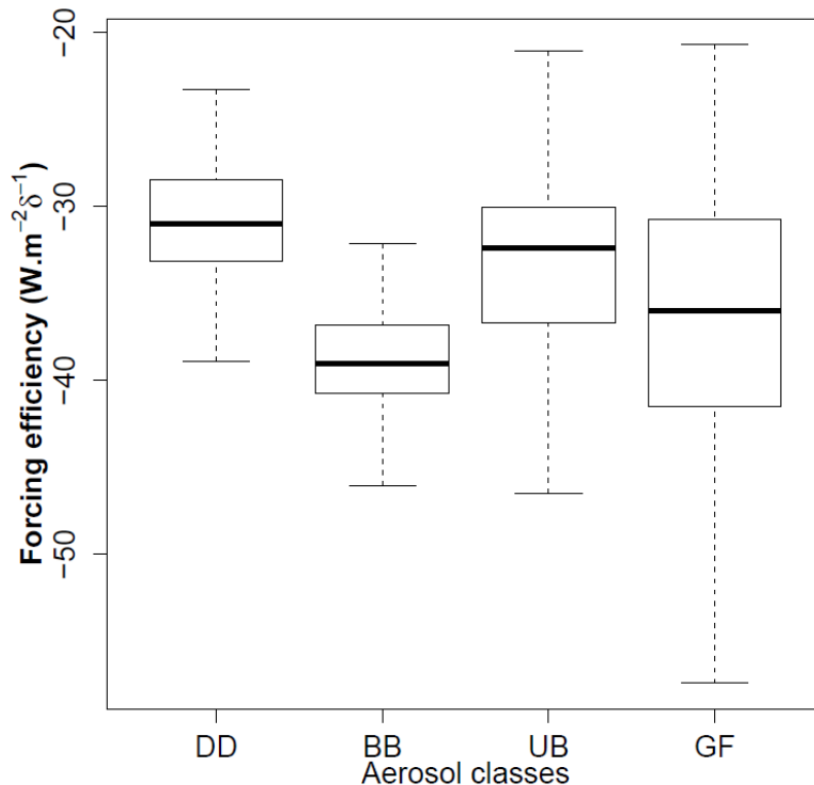
413

414 3.4 Radiative forcing efficiency (RFE)

415 The absolute magnitude of the DRF is dependent not only on the amount of radiation entering
416 the atmosphere but also on the quantity of aerosol perturbing the atmosphere (Bush and
417 Valero, 2003). For a better understanding of the impact of aerosol optical depth (AOD) on the
418 estimation of aerosol DRF, the radiative forcing efficiency (RFE) of the different classes was
419 estimated using equation (4). Since RFE is independent of AOD, it is a useful tool to compare
420 the forcing abilities of different aerosol types. As such, the influences of other variables, such
421 as SSA, absorption and scattering properties and, surface albedo might become more evident
422 (García et al., 2012).

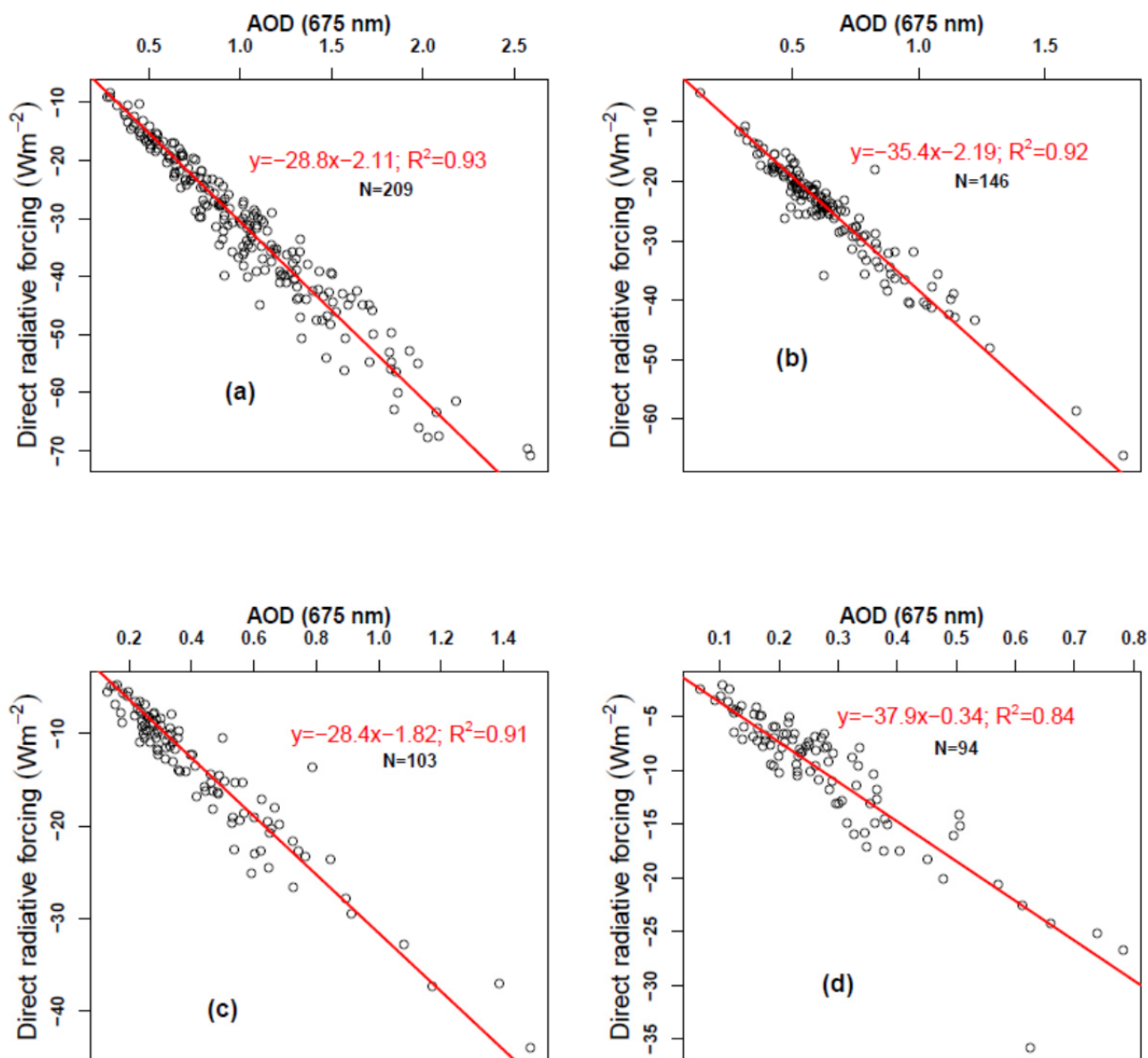
423 Figure 5 shows the variation of the RFE for the different aerosol classes. The natural aerosol,
424 desert dust (DD), has the least average RFE of $-31.0 \pm 3.3 \text{ Wm}^{-2}\delta^{-1}$. The BB aerosol class, like
425 the DD class, has a short range of RFE, but a higher mean RFE value of $-39.0 \pm 4.0 \text{ Wm}^{-2}\delta^{-1}$.

426 The UB and GF aerosol classes have average RFE values of -32.4 ± 5.4 and -36.0 ± 7.8 ,
 427 respectively. The BB aerosol class has the highest median RFE value. These two classes (UB
 428 and GF), compared to the DD and BB aerosol classes, have relatively wider ranges of RFE.
 429 As such, aerosols in these two classes have the ability to perturb the Earth-atmosphere system
 430 more in this region.



431
 432 **Figure 5:** Radiative forcing efficiency (RFE) for the different classes
 433

434 Figure 6 presents the relationship between DRF and aerosol optical depth AOD_{675} for the
 435 different aerosol types. The slope of best-fit line, shown in red, gives the average forcing
 436 efficiency, as estimated by equation (4). The regression equation for the plot and correlation
 437 of DRF and AOD values are also shown in red on Figure 6. In Figure 6, N is the number of
 438 days clustered into each aerosol class.



439
 440 **Figure 6:** The relationship between DRF and AOD (675 nm) for (a) Desert dust cluster (b)
 441 Biomass burning cluster, (c) Urban cluster and (d) Gas flaring cluster.

442
 443

444 4 Conclusion

445 The variations of the aerosol optical and microphysical parameters – AOD, SSA, asym-
 446 metric parameter, Angstrom exponent and backscatter fraction – were studied for the West
 447 African sub-region using AERONET retrievals from Ilorin, Nigeria. The DD aerosol class is
 448 characterised by high AOD and low AE. The BB aerosol class is characterised by high AOD
 449 and high AE while the GF class is characterised by low AOD and high AE. The direct
 450 radiative forcing of the various dominant aerosol types has been estimated using aerosol

451 parameters from AERONET retrievals as inputs in a simplified radiative transfer equation
452 proposed by Haywood and Shine (1995). Due to differences in methodologies and varying
453 aerosol sources/nature, it is difficult to directly compare results (average DRF values) from
454 literature. Desert dust (DD) and biomass burning aerosols were found to be the most effective
455 cooling aerosol at the TOA in the region. UB and GF aerosol classes which are suggested to
456 be rich in emissions from the combustion of fossil fuel (i.e. black carbon and sulphate) have
457 less cooling effects. The more absorbing aerosols (GF and BB) show the higher forcing
458 efficiency; and, GF aerosol class, the largest variability in RFE. These results suggest the
459 need for concerted efforts to adequately characterise and quantify emissions from real-world
460 gas flares as they make significant contributions to the radiative transfer in the Earth-
461 atmosphere system, particularly in oil-rich regions, where gas flaring is persistent, continuous
462 and substantial. To the best of our knowledge, this is the first estimate of DRF for gas-flaring
463 dominant aerosol class.

464 Findings from this study, especially as it relates to the GF cluster, suggest the need for an
465 adequate understanding of the behaviour and transformation of atmospheric aerosol of gas
466 flaring origin. A chemistry transport model with adequate schemes to simulate the behaviour
467 of aerosols will be very appropriate for this proposed study.

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474 5 References

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