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# On Vortices Initiated over West Africa and Their Impact on North Atlantic Tropical Cyclones

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1 On vortices initiated over West Africa and their impact on  
2 North Atlantic tropical cyclones

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## Abstract

Using 38 years of the ERA Interim dataset, an objective tracking approach is used to analyze the origin, characteristics and cyclogenesis efficiency (CE) of synoptic-scale vortices initiated over West Africa and the Atlantic Ocean. Many vortex tracks initiated near the coast or over the ocean result from a vertical expansion of a “primary” vortex track that was initiated earlier over West Africa. Low-level (850hPa) primary vortices are initiated mainly in July near the Hoggar mountains (5°E, 24°N), while mid-level (700 hPa) primary vortices are initiated mainly in August-September near Guinea highlands (10°W, 10°N). The CE of all these vortices is about 10% in July and 30% in August. The average CE is however smaller for low-level “Hoggar” vortices because they peak in July when the cyclogenesis potential index of the Atlantic Ocean is weak. Seasonal and interannual modulations of the cyclogenesis is related more to this index than to the *number* of vortices crossing the West African coast. Cyclogenesis is nearly equally distributed between the coast and 60°W, but the part of the cyclogenesis due to vortices initiated over West Africa decreases from 80% near the coast to about 30% at 60°W. The most probable delay between the vortex vertical expansion and cyclogenesis is 2 days, but it can be up to 10 days. This analysis also confirms previous results, such as the larger CE for vortices extending at low-levels over the continent at 10°N, or the delayed and therefore west-shifted cyclogenesis of low-level “Hoggar” vortices.

## 41 1. Introduction

42 Vortices initiated over West Africa are known sources of tropical storms and hurricanes  
43 over the Northern Atlantic Ocean and are mostly associated with African Easterly Waves  
44 (AEWs) (Erickson 1963, Carlson 1969, Burpee 1972, Landsea 1993, Avila et al. 2000, Hop-  
45 sch et al. 2007, Russel et al. 2017). This relation between AEWs and cyclogenesis moti-  
46 vated several observational studies on the structure of these waves and on their life cycle  
47 (Reed et al. 1977, Reed et al. 1988a, Duvel 1990, Thorncroft and Hodges 2001), as well as  
48 theoretical studies on their physical origin related to instabilities of the African Easterly  
49 Jet (AEJ) (e.g. Thorncroft and Hoskins 1994). Vortices associated with AEW troughs follow  
50 two main paths over West Africa that generally merge into a single path over the Atlantic  
51 Ocean (Reed et al. 1988a and 1988b, Diedhiou et al. 1999, Thorncroft and Hodges 2001,  
52 Fink and Reiner 2003, Hodges et al. 2003). The north path coincides with the low-level  
53 monsoon trough around 20°N and consists of dry low-level vortices. The south path coin-  
54 cides with the rainfall maximum around 10°N and consists of moist mid-level vortices  
55 associated with deep convection. Previous analysis (e.g., Reed et al. 1977, Diedhiou et al.  
56 2002) showed that the south path draws more energy from barotropic energy conversion  
57 at the jet level, while the north path draws more energy from baroclinic energy conver-  
58 sion at lower levels. Vortices of the north and the south paths are often coherent features  
59 over the African Continent (Carlson 1969, Nitta and Takayabu 1985, Pytharoulis and  
60 Thorncroft 1999, Fink and Reiner 2003) (see example in Fig.1).

61 Some ambiguities remain on the impact of north and the south path vortices on cyclogen-  
62 esis. Thorncroft and Hodges (2001), Hopsch et al. (2007) and Dieng et al. (2017) esti-  
63 mated that vortices from the north path generally dissipate shortly after they leave the  
64 African coast, the south path being therefore the main source of cyclogenesis. On the other

65 hand, Chen et al. (2008) found that among cyclogenetic AEWs, 55% are from the north  
66 path and 45% from the south path. They also noticed that cyclogenesis tend to form fur-  
67 ther west on the north path compared to the south path. With some agreement, Ross and  
68 Krishnamurti (2007) showed that both paths may lead to cyclogenesis, especially when  
69 they interact near the coast. More recently, Chen and Liu (2014) considered the influence  
70 of merging dry and wet vortices on MDR cyclogenesis. They found that among tropical  
71 cyclones (TCs) initiated from these vortices, 70% are associated with a dry vortex merg-  
72 ing with a wet vortex.

73 The main objectives of this study are: (i) to clarify the origin of low- and mid-level vortices  
74 of the north and the south path and, (ii) to determine the role of these vortices in cyclo-  
75 genesis depending on their origin. This study does not consider all AEWs perturbations,  
76 but persistent vortices mostly associated with the trough of developed AEWs. The initia-  
77 tion of a vortex (see Table 1 for the definition of different terms used in this article) can  
78 therefore be distinct from the initiation of the corresponding AEW. For example, the ini-  
79 tiation of a persistent vortex (i.e. when it becomes identifiable in the geopotential height,  
80 see section 2.1) may result from the intensification of a relatively weak AEW initiated fur-  
81 ther east. This intensification may be due to mesoscale convective systems (MCS) (e.g.,  
82 Ritchie and Holland 1997; Montgomery et al. 2006; Cecelski and Zhang 2013) and  
83 mesoscale convective vortex (MCV) (e.g. Zhang et al. 2018) developing in the trough of  
84 the wave. A persistent vortex at a given pressure level may also result from an AEW reach-  
85 ing a critical latitude (where the wave phase speed equals the zonal wind) favoring the  
86 creation of a “pouch” reinforcing the role of MCS in further vortex intensification and cy-  
87 clogenesis (Dunkerton et al. 2009, Cecelski and Zhang 2013). To determine the origin of  
88 these vortices, a particular emphasis is put on the conjunction, or merging, of low- and  
89 mid-level vortex tracks. This does not concern only the “confluence” of low-level vortices

90 of the north path and mid-level vortices of the south path (Ross and Krishnamurti 2007,  
91 Hanks et al. 2015) generated independently or associated with a common AEW. The in-  
92 itiation of a vortex track at a given pressure level may also result from the vertical expan-  
93 sion of an already existing “primary” vortex. In this case, it is not really an initiation but  
94 just the point at which the vortex stretches vertically and remains vertically expanded for  
95 the following days. This “secondary” vortex initiation may for example result from verti-  
96 cal stretching related to the formation of MCSs and MCVs within a “primary” vortex. It  
97 may also result from the perturbation, possibly linked to orography, of the synoptic cir-  
98 culation near the primary vortex just before the actual vortex merging.

99 As shown in Chen and Liu (2014), the merger of a low-level dry vortex with a more intense  
100 mid-level moist vortex near the coast favors the cyclogenesis. The development of a mid-  
101 level vortex down to the surface generates frictional drag that may reinforce the low-level  
102 moisture convergence associated with the deep convection and favor cyclogenesis (Gray  
103 1998). In addition, the vertical expansion of the Lagrangian recirculation (the “pouch”) in  
104 a vortex from the mid-troposphere to the boundary layer is important for cyclogenesis  
105 (Wang et al. 2012, Hanks et al. 2015). This vertical expansion may also appear well be-  
106 fore cyclogenesis, as discussed in Thorncroft and Hodges (2001) and Hopsch et al. (2010)  
107 who showed that cyclogenetic vortices tend to extend at low-levels near the Fouta Djallon  
108 mountains (or Guinea highlands). This shows that the interannual cyclonic activity of the  
109 south path could be related in part to AEW characteristics near the coast, and not only to  
110 environmental conditions over the ocean (Gray 1968; DeMaria et al. 2001).

111 The orography may influence AEW initiation, either by its impact on the convective activ-  
112 ity (see, e.g. Carlson 1969, Burpee 1972, Albigat and Reed 1980, Berry and Thorncroft  
113 2005, Mekonnen et al. 2006, Mekonnen and Rossow 2018, Hamilton et al. 2020) or by its

114 impact in the setting up of a mean state favorable to their development (Wu et al. 2009,  
115 Hamilton et al. 2020). Orography can also promote vortex initiation by amplifying AEWs.  
116 Indeed, the deep convection triggered by the orography not only promotes the genesis of  
117 AEWs in East Africa, but also their intensification over West Africa (Hamilton et al. 2020).  
118 The role of orography on the dynamical initiation of dry vortices was explored by Mozer  
119 and Zehnder (1996a) using numerical simulations of a dry adiabatic flow. They showed  
120 that barotropic instabilities resulting from the blocking of the easterly flow by a mountain  
121 range can produce lee vortices that propagate downstream, in particular over West Africa  
122 (Mozer and Zehnder 1996b). A strengthening of low-level easterly winds (Harmattan)  
123 over the Hoggar mountains by an AEW could therefore favor the appearance of a vortex  
124 over West Africa. The potential role of the Hoggar mountains on the genesis of low-level  
125 vortices for the north path was also mentioned in Reed et al. (1988b) and in Thorncroft  
126 and Hodges (2001). Dry low-level vortices studied by Bou Karam et al. (2009) and that  
127 develop in the lee of Hoggar and Aïr mountains are also certainly related in part to AEWs  
128 of the north path.

129 A detailed description of the objective vortex tracking approach and of the datasets is  
130 given in section 2. Different aspects of the origin, the merging and the cyclogenesis of the  
131 vortices are analyzed in the following sections: initiation and track density in section 3,  
132 seasonal and interannual variations in section 4, and longitudinal variations in section 5.  
133 Section 6 explores different types of vortex merging (confluence, expansion, local devel-  
134 opment) and their impact on cyclogenesis. The results are summarized and discussed in  
135 section 7.

## 136 2. Data and analysis

### 137 2.1. Vortex detection

138 The vortex tracking is done for 38 seasons (June to October) between 1980 and 2017 us-  
139 ing ERA-Interim (ERA-I, Dee et al. 2011). This dataset has a horizontal resolution of  $0.75^\circ$   
140  $\times 0.75^\circ$  and a 6-hour time step. The vortex tracking is based on the geopotential height  
141 ( $\phi$ ) of a given pressure level (Duvel 2015, Duvel et al. 2017). The first step of this approach  
142 is to compute a geopotential height anomaly defined as the difference between  $\phi$  and its  
143 average over a region of  $\pm 7.5^\circ$  (i.e.  $\pm 10$  ERA-I gridpoints). The “vortex area” is the ensem-  
144 ble of continuous grid points with a negative geopotential height anomaly and with an  
145 absolute value  $\Delta\phi$  of this anomaly larger than a threshold  $\Delta\phi_s$ . To detect both weak and  
146 strong vortices while conserving a reasonable vortex size,  $\Delta\phi_s$  varies between a minimum  
147 value  $\Delta\phi_{\min}$  adapted to weak depression, and a maximum value  $\Delta\phi_{\max}$  adapted to strong  
148 cyclonic depression. Technically, the threshold is increased from  $\Delta\phi_{\min}$  until it gives a vor-  
149 tex area with an equivalent radius lower than  $R_{\max} = 4.2^\circ$ . This value is a reasonable vortex  
150 size that allows a good overlap of the vortices between two timesteps. For the weakest  
151 vortices, the equivalent radius at  $\Delta\phi_{\min}$  must be larger than  $1.2^\circ$ . The threshold is chosen  
152 among an ensemble  $\Delta\phi_s = \Delta\phi_{\min} + s^2$  (where  $s$  is an integer  $0 \leq s \leq 40$ ). The  $s^2$  progression of  
153 the thresholds gives a better size adjustment capability for weak vortices.

### 154 2.2. Vortex track

155 A given vortex is tracked over time by considering the overlap (weighted by  $\Delta\phi$ ) between  
156 vortex areas for two consecutive time steps. If several vortices overlap, the tracking uses  
157 the vortex with the largest overlap. The remaining vortices either start a new track gen-  
158 erated by splitting or stop by “fusion”. The positions of the vortex track are the bary-  
159 centers of the vortex area computed considering all its gridpoints weighted by  $\Delta\phi$ . We



160 consider only vortices initiated in the domain 60°W-50°E and 0°N-30°N (Fig.1), but the  
161 tracking is done for a domain extending westward to 130°W and northward to 60°N. Since  
162 we are mostly interested in the cyclogenesis resulting from these vortices, we consider  
163 only vortices with a barycenter that spends at least one day over the Atlantic Ocean and  
164 that spends at least one timestep south of 20°N (in order to remove vortices that do not  
165 enter the MDR). Vortices that fulfill all these conditions are called Atlantic Vortices (AV)  
166 in the following sections.

### 167 2.3. Cyclogenetic vortices

168 If the barycenter of a vortex is located within 3° of an IBTrACS (Knapp et al. 2010) system  
169 (regardless of its intensity) for at least one timestep, it is considered a Cyclogenetic Atlan-  
170 tic Vortex (CAV), the others being therefore Non-CAV (NCAV). With this criterion, the  
171 match between the CAV and its IBTrACS system is longer than 1 day for 97% of the CAV  
172 and the average match is about 7 days. The first match generally corresponds to the first  
173 time-step of the IBTrACS system and is considered here as the CAV cyclogenesis point  
174 (see the example in Fig.1).

175 The CAV statistics was also used to refine the definition of  $\Delta\phi_{\min}$ . CAV are indeed always  
176 detected when the corresponding TC appears over the ocean, whatever the value of  $\Delta\phi_{\min}$ ,  
177 but the initiation of the CAV may shift westward for large  $\Delta\phi_{\min}$ . This is because a weak  
178 depression may break the vortex track and gives a new vortex downstream (i.e. west-  
179 ward). On the other hand, a too small  $\Delta\phi_{\min}$  produces many weak tracks of little interest,  
180 or even inconsistent structures in the geopotential anomaly field that reduces the number  
181 of AV. For this study, we use the two thresholds ( $\Delta\phi_{\min} = 40 \text{ m}^2\text{s}^{-2}$  at 850 hPa and  $50 \text{ m}^2\text{s}^{-2}$   
182 at 700 hPa) that maximize the number of CAV initiated over the continent. Using a larger  
183  $\Delta\phi_{\min}$  (up to  $80 \text{ m}^2\text{s}^{-2}$ ) decreases this number but this has no significant impact on the

184 conclusions given below, in particular on the average vortex distributions (spatial, sea-  
185 sonal) or on the relative impact of the North and the South path on cyclogenesis.

#### 186 *2.4. Merging between low-level and mid-level vortex tracks*

187 A vortex track generally merges over the ocean with a vortex track of the other pressure  
188 level (e.g. Fig.1). The common parts of the tracks are detected by looking at pairs of 700  
189 hPa and 850 hPa vortices having their barycenters located within 3° of each other. The  
190 first time this happens is the *merging time* between the two vortices and corresponds to  
191 the beginning of a vertically expanded vortex track that generally persists until vortex  
192 dissipation. It is thus possible to define a primary and a secondary vortex before merging,  
193 the primary being the older at merging time. If the secondary vortex is produced shortly  
194 before merging (e.g. <1 day), the merging corresponds to a *vertical expansion* of the pri-  
195 mary vortex. In this case, the primary vortex initiation can be considered as the onset of  
196 the disturbance, e.g. the time at which a persistent closed circulation is formed near an  
197 AEW trough. If the secondary vortex is produced long before merging (e.g. ≥1 day), we  
198 may consider that the two vortices are formed “independently”, while possibly related to  
199 a common synoptic or large-scale feature, and the merging corresponds to the *confluence*  
200 of the two vortex tracks (e.g. Fig.1). In this case, which roughly corresponds to the path  
201 merger studied in Hanks et al. (2015), the primary vortex is just the older one. AV initi-  
202 ated at the same time for both levels are counted as 700 hPa primary vortices.

#### 203 *2.5. Advantages and shortcomings of the tracking and merging algorithms*

204 AEWs and their associated vortices are complex and intermittent features with perturba-  
205 tion patterns depending in particular on geographical location, pressure level and season.  
206 Different tracking algorithms based on different physical parameters, pressure levels and  
207 processing approach are thus susceptible to extract a different ensemble of cases. It is not

208 necessarily a problem if the results are interpreted considering the particularities of each  
209 algorithm. The vortex tracking approach used in this study is no exception to the rule. This  
210 algorithm is not designed to track all AEWs, in particular weak AEWs which are better  
211 targeted by dedicated algorithms such as those of Berry et al. (2007), Agudelo et al.  
212 (2011), Brammer and Thorncroft (2015) and Belanger et al. (2016). This algorithm de-  
213 tects the trough of the wave as soon as it is sufficiently developed to generate a persistent  
214 closed circulation (i.e. a detectable geopotential minimum). This is consistent with the  
215 approach of Thorncroft and Hodges (2001) but perhaps with a more precise vortex loca-  
216 tion because the geopotential field is naturally smoother than the vorticity field and does  
217 not require the use of low resolution. The adaptive threshold makes it possible to detect  
218 the vortex initiation at an early stage. However, vortex initiations are probably statisti-  
219 cally shifted to the west compared to the initiation of the associated AEW since, as dis-  
220 cussed above, a vortex is initiated only when the wave reaches a sufficient amplitude. This  
221 must be kept in mind when interpreting the results.

222 Regarding vortex merging, it should be noted that a vortex of the north path may coexist  
223 but not merge with a vortex of the south path. These vortices are classified as non-mer-  
224 gers by the algorithm while they are theoretically part of the same perturbation. Some  
225 low-level vortices on the north path are therefore not considered to be cyclogenetic  
226 whereas they can influence cyclogenetic vortices of the south path via a modification of  
227 the environmental flow (e.g., Brammer and Thorncroft 2015, 2017). These cases are prob-  
228 ably not very frequent as the two paths generally merge over the ocean (e.g. Reed et al  
229 1988a, 1988b, Diedhiou et al. 1999, Thorncroft and Hodges 2001, Ross and Krishnamurti  
230 2007).

231 For some mature AEWs, the mid-level vortex of the south path may first expand down-  
232 ward and then merge with a low-level north path vortex, temporally forming a system  
233 with one mid-level vortex and two low-level vortices (e.g., Carlson 1969; Berry and Thorn-  
234 croft 2005). The couple of low-level vortices generally join over the ocean west of the  
235 Guinea highlands. The algorithm considers those cases, but associates the mid-level track  
236 with either the north or the south low-level track (quite randomly since it is based on the  
237 vortex overlap at “fusion” time, see above). The other low-level track becomes either a  
238 non-merging AV or is eliminated if it does not fulfil the AV criteria before fusion. As shown  
239 in Ross and Krishnamuri (2007), such fusion between *low-level* vortex tracks near the  
240 coast is the exception rather than the rule.

### 241 3. Vortex initiation and track density

#### 242 3.1. All Atlantic vortices

243 Over West Africa, AV are initiated mostly north of 15°N at 850 hPa and south of 15°N at  
244 700 hPa (Fig.2a and 2b). At 700 hPa (Fig.2a), AV initiations are maximal at 10°N and west  
245 of the Fouta Djallon mountains in agreement with previous results (e.g., Thorncroft and  
246 Hodges 2001, Berry and Thorncroft 2005, Chen 2006, Hopsch et al. 2010). This is likely  
247 associated with AEW reinforcement by latent heat release in organized deep convection  
248 that generally develops in the lee side of high terrain (Laing et al. 2008). There are two  
249 secondary maxima at 5°E and 30°E that are located east of the Cameroon and the Ethio-  
250 pian highlands (only results west of 30°E are shown to focus on regions with large AV  
251 density). At 850 hPa (Fig.2b), AV initiations are clearly related to the orography with a  
252 large maximum centered west of Hoggar mountains (5°E, 24°N) and a secondary maxi-  
253 mum centered west of Tibesti mountains (16°E, 20°N). This relation with orography is

254 clearer than in previous studies, probably because of the larger and more persistent im-  
255 pact of orography on geopotential compared to vorticity. The precise physical origin of  
256 these AV initiations is not studied in this paper. They are not necessarily the signature of  
257 AEW initiations, but they do show the influence of orography in forming persistent low-  
258 level closed circulations that propagates westward to the ocean. For the 38 seasons  
259 (JJASO), the number of AV initiated in our domain (60°W to 50°E) is comparable for both  
260 pressure levels (~1700 or about 45 by season). More details are given below for initiation  
261 density maps of non-merging AVs (Fig.3), merging non-cyclogenetic AVs (NCAV, Fig.4),  
262 and cyclogenetic AVs (CAV, Fig.5).

### 263 *3.2. Non-Merging Atlantic vortices*

264 Non-merging AVs are disturbances with no persistent vertical expansion and they are  
265 generally not cyclogenetic. Initiations of non-merging AV (Fig.3) are more evenly distrib-  
266 uted over the domain compared to the AV ensemble, with a relatively higher density over  
267 the Atlantic Ocean. In fact, vortices initiated over land and reaching and staying over the  
268 ocean for at least one day have certainly more chance to be stronger vortices since the  
269 weaker ones will dissipate before reaching the ocean. On the contrary, AV generated over  
270 the ocean can be short-lived systems (i.e. 1 day). This fact certainly contributes to increase  
271 the proportion of weak non-merging and non-cyclogenetic AVs over the ocean. At 700 hPa  
272 (Fig.3a), a particular pattern is underlined in the northwest sector of the domain with AV  
273 generated north of 15°N at mid-level and that propagate southwestward over the ocean.  
274 These mid-level vortices are probably linked to the Moroccan vortex (Cheng et al. 2019)  
275 generated by the interaction between the AEW and the basic-state potential vorticity gra-  
276 dient.

### 277 3.3. Non-cyclogenetic Atlantic vortices (NCAV)

278 Mid-level primary NCAVs (Fig.4a) are mainly initiated around 10°N over the ocean west  
279 of Guinea highlands and over land west of Cameroon highlands and Jos Plateau. Only a  
280 few low-level secondary NCAVs are initiated at 10°N and may correspond to local down-  
281 ward expansion of mid-level vortices. Low-level secondary NCAVs (Fig.4c) are mainly in-  
282 itiated around 15°N over the ocean and thus north of the mid-level vortex track (see Fig.6)  
283 and the merger occurs mostly near 20°W (Fig.4e). These mergers may correspond to dry  
284 vortex mergers studied in Chen and Liu (2014). Low-level primary NCAVs (Fig.4b) are  
285 mainly initiated near the Hoggar mountains and have thus a totally different density pat-  
286 tern compared to low-level secondary NCAVs (Fig.4c). These primary vortices do not dis-  
287 sipate over the ocean, but rather merge west of 20°W (Fig.4f) with mid-level vortices ini-  
288 tiated mostly to the west of the Guinea highlands (Fig.4d). The maximum merging density  
289 (Fig.4e and 4f) is shifted west by only a few degrees compared to the maximum initiation  
290 density of secondary NCAV. This shows that the merging occurs generally shortly after  
291 the initiation of the secondary vortex and may therefore be considered as a vortex deep-  
292 ening, even if the secondary vortex is initiated at some distance from the primary vortex  
293 track. However, among the secondary vortices initiated over the continent east of 10°W,  
294 some are initiated south of 15°N at 700 hPa (Fig.4d) and north of 15°N at 850 hPa (Fig.4c)  
295 while there is no *local* primary vortex counterpart. This reveals initiations of independent  
296 north and south vortex tracks that will later merge by *confluence* (see section 6).

### 297 3.4. Cyclogenetic Atlantic Vortices (CAV)

298 Figure 5a suggests that there are relatively more CAV initiated over the continent. The  
299 ratio between initiations occurring east and west of 10°W is indeed 1.5 times greater for  
300 the CAV compared to NCAV. This shows that AEWs that develop a vortex further east at

301 the jet level have more chance to be cyclogenetic. Many secondary low-level CAV are also  
302 initiated (Fig.5c) and merge (Fig.5e) over the continent, in good agreement with previous  
303 studies showing that cyclogenetic AEWs generally extend at low-levels over the continent  
304 (see section 5.2). For CAV, maximum initiation and merging are both on the coast near  
305 10°N, suggesting a large influence of deep convection on the vertical expansion. This  
306 downward expansion is also more distributed in longitude for CAV compared to NCAV.  
307 Some secondary vortices appear west of 30°W over the ocean (Fig.5c), corresponding to  
308 late downward expansion and cyclogenesis. A more detailed analysis (not shown) reveals  
309 that : (i) the maximum cyclogenesis west of 20°W (Fig.5g) is related to downward expan-  
310 sion near the coast and over West Africa along 10°N and (ii) that the smaller cyclogenesis  
311 maximum west of 30°W is related to secondary CAV initiated (Fig.5c) north of 15°N near  
312 20°W.

313 Low-level primary CAV initiations (Fig.5b) are more frequent near the Hoggar mountains,  
314 but there are also some initiations near 10°N over West Africa showing that cyclogenetic  
315 low-level vortices may lead slightly mid-level ones on the south-path. Mid-level secondary  
316 CAV are initiated mainly near the coast, around 35°W and west of 50°W (Fig.5d). Those  
317 formed over the ocean merge rapidly and correspond to a local upward expansion of the  
318 low-level vortices (Fig.5f) shifted westward for CAV compared to NCAV. Cyclogenesis of  
319 vortices initiated near Hoggar and Tibesti mountains is also shifted westward compared  
320 to the cyclogenesis of vortices initiated at the jet level, with a weak cyclogenesis near  
321 25°W (Fig.5h). This agrees with the results of Chen et al. (2008) showing that disturb-  
322 ances of the north path take longer to initiate a TC.

323 Compared to the NCAV ensemble, CAV have a more variable initiation distribution for  
324 both primary and secondary vortices. While most NCAV mergers occur near the coast just

325 west of 20°W (Fig.4e and 4f), CAV mergers occur either further west over the ocean or  
326 over land and at the coast for downward expansion of mid-level vortices (Fig.5e). This  
327 suggests that there are specific or favorable processes for the vertical expansion of the  
328 vortices before cyclogenesis over these regions, related for example to pouch formation  
329 over the ocean and to the development of deep convection over land.

### 330 *3.5. Initiation and cyclogenesis statistics*

331 A summary of the distribution of primary AV initiations and of their influence on cyclo-  
332 genesis is reported in table 2. Summing all primary AVs initiated over the north and the  
333 south path domains, there are 938 initiations and thus about 25 by season (JJASO) (Table  
334 2a). As expected, because we detect only developed AEWs, this is small compared to the  
335 43 AEWs initiated over West Africa per season (JJAS) in Chen (2006) and the 61 AEWs  
336 per season (MJJASON) in Avila et al. (2000) that suggest a quite continuous wave activity  
337 with more than one trough crossing the coast every 3 or 3.5 days. Compared to the pre-  
338 sent analysis, Chen (2006) found a more comparable number of initiations for the south  
339 path (~13 by seasons), but a larger number of initiations for the north path (~30 by sea-  
340 son), suggesting that a larger proportion of weak AEWs are not selected by our approach  
341 for the north path.

342 About 23% of primary AV initiated on the south path are cyclogenetic compared to only  
343 10% on the north path (Table 2b). For the 38 hurricane seasons, there are 305 IBTrACS  
344 systems (about 57% of the total number of North Atlantic systems) initiated south of  
345 20°N, 217 in the MDR and 88 west of 60°W. Over the MDR, 16% of the cyclogenesis is due  
346 to low-level primary AV of the north path and 38% to mid-level primary AV of the south  
347 path (Table 2c). In total, 62% of the MDR cyclogenesis is due to primary AV initiated east  
348 of 10°W at both pressure levels. However, this percentage drops to 30% of the entire



349 North Atlantic cyclogenesis. This is half of the cyclogenesis due to AEWs found in previous  
350 studies (e.g. Avila et al. 2000, Chen et al. 2008, Russel et al. 2017) using a subjective AEW  
351 detection. This suggests that about half of these cyclogenetic AEWs are associated with a  
352 vortex initiated east of 10°W, the remainder being mainly initiated near the coast (Fig.  
353 5a).

354 Only 38% of the MDR cyclogenesis is due to primary AV initiated locally over the Atlantic  
355 Ocean and coastal regions at either 700 hPa (20%) or 850 hPa (18%) (Table 2c). These  
356 primary AV are more numerous compared to continental primary AV, especially at 700  
357 hPa where there is a large concentration near the coast (Fig.4a and 5a), but they have a  
358 weak cyclogenesis efficiency (9% at 700 hPa and 11% at 850hPa) (Table 2b). West of  
359 60°W and south of 20°N (Table 2d), only 37% of the cyclogenesis is associated with AV  
360 coming from eastern regions: 12% from the continent and 25% from the Atlantic Ocean  
361 and coastal regions. This shows that cyclogenesis occurring far from the West African  
362 coast have logically little chance to be associated with a persistent vortex initiated over  
363 West Africa or near the coast (see also section 5).

### 364 *3.6. Track density*

365 A track density map gives the probability of presence of a vortex of a given type. The track  
366 density is maximal around 25°W for both levels and for all NCAV types (Fig.6). This is due  
367 to the superposition of AVs just initiated near the coast and those initiated over the con-  
368 tinent and not yet dissipated over the ocean. The speed of the vortex barycenters also  
369 tends to slow down and thus to increase the vortex occurrence at these longitudes (not  
370 shown). For non-merging NCAV (Fig.6a and 6b) this probability is smaller compared to  
371 primary merging NCAV (Fig.6c and 6d), mostly because of their shorter duration. There  
372 is no clear gap in the NCAV track density near the coast. Low-level vortices initiated near

373 the Hoggar mountains follow quite continuously a west-southwestward trajectory and  
374 mid-level vortices initiated south of the AEJ follow quite continuously a west-northwest-  
375 ward trajectory. Secondary NCAV (Fig.6e and 6f) have a smaller track density, not only  
376 over West Africa, but also over the ocean because of their later development.

377 At 700 hPa, primary CAV track density (Fig.7a) grows east-northeastward from West Af-  
378 rica with a maximum density at 25°W and a large track density until the western bound-  
379 ary. As expected, associated secondary track density at 850 hPa is smaller over the conti-  
380 nent and is maximal further west over the ocean (Fig.7c). The track density of primary  
381 CAV at 850 hPa (Fig.7b) is very different with a maximum shifted southwestward near  
382 40°W over the ocean. There is a weak south-path track associated with low-level primary  
383 CAV initiated over West Africa. At 15°W, the maximum track density jumps from 20°N to  
384 15°N, suggesting a discontinuity of AV tracks near the coast for CAV. This is not due only  
385 to primary vortices initiated near the coast. Low-level CAV initiated near the Hoggar  
386 mountains have indeed a similar track density pattern (not shown) and are thus not dis-  
387 sipated near the coast. This discontinuity may be associated with the feature described in  
388 Chen and Liu (2014) that intense AEWs tend to shift dry low-level vortices southwest-  
389 ward before favoring cyclogenetic merge with a mid-level vortex.

#### 390 4. Seasonal and interannual variation

391 Previous analyses (e.g., Duvel 1990; Thorncroft and Hodges 2001; Hopsch et al. 2007;  
392 Ross and Krishnamurti 2007) generally found that north path activity is larger at the be-  
393 ginning of the season and south path activity at the end of the season (August and Sep-  
394 tember). As shown in figure 8, this seasonal variation is mostly related to the number of  
395 primary merging AV initiations ( $N_{AV}$ ) over West Africa (Fig.8b) which is maximum in July

396 at 850 hPa and in August at 700 hPa. Since primary AV at 850 hPa are mostly initiated  
397 near Hoggar mountains (Fig.4b and 5b), this illustrates the large efficiency of these oro-  
398 graphic processes at the start of the hurricane season. For both pressure levels, the num-  
399 ber of primary CAV initiations ( $N_{CAV}$ ) over West Africa is small in July, maximal in August,  
400 and decreases for the following months. AVs initiated near the coast (Fig.8a) and non-  
401 merging AV (Fig.8c and 8d) have a weak seasonal variation. For coastal regions (Fig.8a),  
402  $N_{AV}$  is more evenly distributed along the season with values of about 1 per month at 700  
403 hPa and 0.5 per month at 850 hPa.

404 The cyclogenesis efficiency ( $CE = N_{CAV}/N_{AV}$ ) of primary AV initiated over West Africa is  
405 maximal in August ( $\sim 30\%$ ) and small in July ( $< 10\%$ ) for both paths (Fig.9a). Both paths  
406 have thus an equivalent CE for July and August. This is consistent with Agudelo et al.  
407 (2011) who showed that about 80% of the seasonal cycle of CE is determined by the large-  
408 scale environment over the ocean that is optimal in August and September (Emanuel and  
409 Nolan 2004, Tippet et al. 2011, Menkes et al. 2011). The low average CE for the north path  
410 (10%) in table 2b is thus mostly due to its unfavorable seasonal phase with maximum  
411 vortex initiations in July. According to previous results (e.g. Duvel 1990, Ross and Krish-  
412 namurti 2007), AEW amplitude and low-level vorticity near the coast is also maximal in  
413 August-September, that could explain part of the relation between south path vortices  
414 intensity and cyclogenesis.

415 The Merging Efficiency (ME) index that is just the proportion of merging AV represents  
416 the ability of the vortices to deepen and could be also related to the cyclogenesis. The ME  
417 for primary AV initiated over West Africa (Fig.9b) has however high values for the north  
418 path between June and August with no correlation with CE. Nevertheless, the smaller ME  
419 in September at 850 hPa may be at the origin of the corresponding smaller CE. This shows

420 that the vertical deepening of a vortex is a necessary but not a sufficient condition for  
421 cyclogenesis and that a larger ME does not necessarily lead to a larger CE at the seasonal  
422 time scale.

423 Primary AVs initiated in August and September over West Africa have an averaged CE of  
424 26% that agrees well with the value found by Agudelo et al. (2011). This CE is however  
425 very variable from one year to another (Fig.9c). This suggests that the cyclogenesis does  
426 not depend on the *number* of AVs initiated over the continent during the active season.  
427 According to Agudelo et al. (2011), about half of the interannual variability of the cyclo-  
428 genesis may be attributed to variations of the large-scale environment. Our result sug-  
429 gests that the remaining part is not due to the *number* of AV generated over West Africa,  
430 but rather to vortex intensification due to more local and stochastic process (such as the  
431 development of MCSs and MCVs).

## 432 5. Longitudinal variations of cyclogenesis

### 433 *5.1. Impact of the CAV origin on the cyclogenesis longitude*

434 The average number of cyclogenesis is quite evenly distributed with longitude (Fig.10).  
435 However, the CAV origin (West Africa, coast or open ocean) strongly depends on the cy-  
436 clogenesis longitude. Vortices initiated over West Africa are the main contributor for the  
437 MDR cyclogenesis (62% on average, Table 2c) with a cyclogenesis fraction that gradually  
438 decreases from 80% near the coast to about 30% at 60°W. The large fraction near the  
439 coast may be due to the fact that many AV initiated over West Africa are already strong  
440 and vertically developed when arriving over the ocean. These vortices increase low-level  
441 convergence and possibly create a "pouch" effect (Dunkerton et al. 2009) that may pro-  
442 mote deep convection and cyclogenesis near the coast. The comparison of figures 5g and

443 5h shows that most of the cyclogenesis near the coast is due to the moist mid-level CAVs  
444 of the south path and not to the dry low-level CAVs of the north path that take more time  
445 to trigger cyclogenesis. The cyclogenesis fraction due to CAV initiated over the ocean in-  
446 creases from zero at 20°W to about 40% at 60°W. Such CAV initiation and intensification  
447 over the ocean may be related to local development of organized deep convection, possi-  
448 bly triggered by weak AEWs formed further east over West Africa. By contrast, the cyclo-  
449 genesis fraction due to primary AV initiated near the coast is quite constant with longi-  
450 tude (~20%), probably because it mixes the process of the two other regional sources of  
451 CAV.

## 452 *5.2. Vortex intensity and vortex merging*

453 The impact of vortex intensity on CE may be studied by comparing average CAV and NCAV  
454 intensity as a function of longitude or relative to merging time. For the south path at 700  
455 hPa (Fig.11a), the primary NCAV intensity is relatively constant with longitude with a  
456 slight increase west of 0°E and a slight decrease over the ocean. This agrees with the av-  
457 erage growth rate reported in Thorncroft and Hodges (2001). As expected, the CAV inten-  
458 sity is larger and increases westward over the ocean, mostly because of the increasing TC  
459 proportion and TC strength. An interesting point is that CAV intensity is also significantly  
460 larger between 5°W and the coast (Fig.11a), showing that AV having a large intensity over  
461 the continent are more likely to be cyclogenetic. This is compatible with Hopsch et al.  
462 (2007) who showed that AEW variance near the coast is positively correlated with the  
463 number of TCs at interannual time-scale. This also agrees with previous studies (Thorn-  
464 croft and Hodges 2001, Hopsch et al, 2010, Agudelo et al. 2011, Arnault and Roux 2011,  
465 Brammer and Thorncroft 2015) showing that characteristics of the AEW trough (low-  
466 level vorticity, column humidity, vertical velocity and other parameters related to deep  
467 convection) between 0°E and the coast have an influence on further wave development

468 and cyclogenesis. Other features such as the effect of consecutive AEW troughs (Dieng et  
469 al. 2017) may also have an influence on AEW development near the coast. The intensity  
470 of north path NCAV (Fig.11b) is larger over land at 850 hPa compared to 700 hPa. While  
471 this intensity decays strongly near the coast, as noticed by Thorncroft and Hodges (2001),  
472 it remains stronger or equal to the NCAV intensity of the south path for all longitudes.  
473 Although less significant than at 700 hPa, there is also a tendency for a larger CAV inten-  
474 sity near the coast at 850 hPa.

475 The downward deepening of 700 hPa vortices is associated with a transitory intensifica-  
476 tion for both CAV and NCAV (Fig.11c). The upward deepening of 850 hPa vortices corre-  
477 sponds to a pause in an intensity decrease over the previous 2 days (Fig.11d). The CAV  
478 intensity becomes significantly larger than the NCAV intensity about 36 hours after vortex  
479 deepening, partly because of the progressively greater TC occurrence westward. Cyclo-  
480 genesis generally occurs after vortex deepening (Fig.11c and 11d) with most probable  
481 delay around two days, but cyclogenesis can occur up to 10 days after vortex deepening.  
482 The consistent timing between the deepening and the cyclogenesis demonstrates the re-  
483 markable agreement between IBTrACS and the vortex dynamics in ERA-I.

## 484 6. Different types of vortex merging

485 The previous sections present the difference between primary and secondary vortices,  
486 but with no consideration of the delay between initiation and merging. However, the na-  
487 ture of the merging can be interpreted differently depending on this delay. Four merging  
488 categories are illustrated in Fig.12 based on the merging types described in section 2.4.  
489 The first category (Fig.12a) corresponds to the *confluence* of two vortex tracks. The sec-  
490 ond category (Fig.12b) corresponds to a local development of the depression, related for

491 example to MCSs, with delay  $\leq 1$  day between initiation and merging for both levels. The  
492 next two categories (Fig.12c and 12d) correspond to a delay between initiation and merg-  
493 ing  $> 1$  day for the primary vortex and  $\leq 1$  day for the secondary vortex. This is interpreted  
494 as a vertical expansion of the primary vortex. The 1-day threshold is somewhat arbitrary  
495 among reasonable values (say between 12 h and 2 days). It is chosen because it gives a  
496 relatively homogeneous distribution of the four merging categories (between 21% and  
497 30%) and it makes it possible to identify quite clearly the source of the vortices for each  
498 category (Fig.13).

499 For merging happening more than one day after vortex initiations at both levels, initiation  
500 density maps clearly reflect the north and the south paths over the continent and near the  
501 coast (Fig.13a). This “confluence” category represents 25% of the mergers, with 15% ini-  
502 tiated east of  $10^\circ\text{W}$  and 10% initiated near the coast. The shift of the merging densities  
503 between the two pressure levels (Fig.14a) shows that the low-level vortex is located  
504 northwest and therefore slightly ahead of the mid-level vortex at merging time. This  
505 merging category is associated with 23% of the cyclogenesis, mostly east of  $30^\circ\text{W}$ . These  
506 mergers peak in July and August (Fig.15a) and give a CE of about 20% in August and Sep-  
507 tember that agrees with the value found by Hanks et al. (2015) for a similar definition of  
508 the confluence of north and south path vortices (but not identical since they consider only  
509 cyclogenesis east of  $40^\circ\text{E}$ ).

510 Initiation density maps for fast merging vortices (Fig.13b) is concentrated offshore, west  
511 of the Guinean coast, for both levels. As expected for these fast mergers that corresponds  
512 to a local vortex development, initiation (Fig.13b) and merging (Fig.14b) density maps  
513 are quite similar. These fast merging vortices are associated with 24% of the cyclogenesis,

514 with a clear maximum for occurrence and CE in September (Fig.15b). Downward deepening  
515 ing of primary 700 hPa vortices (Fig.13c) concerns mostly mid-level vortices initiated  
516 over the continent, those initiated offshore west of the Guinean coast being fast mergers  
517 (Fig.13b). Only a few mid-level vortices expand downward over the continent, mostly  
518 near the Guinea highlands. The most part expands downward by merging with low-level  
519 vortices initiated over the ocean around 15°N (Fig.13c and 14c). This category represents  
520 the largest amount of cyclogenesis (37%), mostly near the coast, with a peak activity in  
521 August and September (Fig.15c). Upward deepening of primary 850 hPa vortices  
522 (Fig.13d) concerns mostly low-level vortices initiated west of the Hoggar and Tibesti  
523 mountains. A few of these vortices expand early over the continent, but most of them ex-  
524 pand over the ocean near 20°W by merging with mid-level vortices initiated west of the  
525 Guinea highlands (Fig.13d and 14d). The mid-level vortices are clearly located south of  
526 the low-level vortices at merging time (Fig.14d), suggesting that these mid-level vortices  
527 are induced by the synoptic dynamical perturbation of a low-level vortex and not by deep  
528 convection developing into the low-level vortex. This category is relatively frequent  
529 (24%), but represents only 16% of the cyclogenesis. This is mostly due to the fact that  
530 these mergers peak in July (Fig.15d) when the MDR cyclogenesis potential index related  
531 to the large-scale environment is weak. Note that CE is large in September for this merging  
532 category compared to the ensemble of low-level primary AV (Fig.9a) that includes non-  
533 mergers.



## 534 7. Conclusions

### 535 7.1. Importance of distinguishing primary and secondary vortices

536 This study shows the importance of considering at least two pressure levels, one in the  
537 low-level monsoon and Harmattan flows (850 hPa) and one in the AEJ (700 hPa), in order  
538 to take into account the complexity of vortex initiation processes over West Africa. Fig-  
539 ures 4 and 5 shows that many vortex tracks initiated near the coast or over the Atlantic  
540 Ocean result from a vertical expansion of primary vortices initiated over West Africa  
541 where they follow very distinct paths depending on the pressure level. As illustrated in  
542 figures 13a and 14a, about 25% of mergers are confluences between north and south path  
543 vortices initiated independently over West Africa (such as the case shown in figure 1) or  
544 near the coast. For the "expansion" categories c and d and even for the "local develop-  
545 ment" category b, the merging does not generally occur at the same place as the initiation  
546 of the secondary vortex (Fig. 13 and 14). This suggests that some secondary vortices do  
547 not develop vertically from the primary vortex, but nearby due to the dynamic influence  
548 of this primary vortex. This can be for example the case for the dry low-level vortices that  
549 are associated with mid-level vortices described in Chen and Liu (2014) and that have a  
550 signature near 15°N in figures 4c, 5c, 13a and 13c.

551 In agreement with Chen (2008), merging categories involving the north path tend to delay  
552 the cyclogenesis to the west. The vortex deepening associated with merging is a necessary  
553 but not a sufficient condition for cyclogenesis. The deepening tends to increase the vortex  
554 strength, especially for mid-level primary vortices (Fig.11c), but this is generally not suf-  
555 ficient for cyclogenesis. The cyclogenesis efficiency increases for vortices that deepen

556 over West Africa or near the coast, but it also largely depends on environmental condi-  
557 tions, as discussed above. Cyclogenesis occurs more often about two days after deepening,  
558 but it can be delayed for up to 10 days.

#### 559 *7.2. Seasonal and Interannual variations of continental primary vortices*

560 The number of primary vortices produced on the north path is maximal in July and grad-  
561 ually decreases in August and September. For the south path, this number is maximal in  
562 August and September. Since the cyclogenetic potential index of the MDR is far larger in  
563 August and September compared to July, the south path is more efficient for cyclogenesis  
564 (23% of CAV) compared to the north path (10% of CAV). However, the cyclogenetic effi-  
565 ciency is about 30% for both paths in August, suggesting that low-level north path vorti-  
566 ces are nearly as efficient as mid-level south path vortices when the cyclogenesis potential  
567 index is high. The interannual variation of the total *number* of AV produced in August and  
568 September over West Africa is not correlated with the cyclogenesis. This suggests that  
569 seasonal and interannual modulations of the cyclogenesis over the North Atlantic is re-  
570 lated more to large-scale environmental conditions than to the *number* of vortices cross-  
571 ing the West African coast. This is not a common tendency for other basins and time-  
572 scales. For example, the MJO modulation of cyclogenesis over the Indian Ocean is more  
573 linked to the number of vortices than to the intensification of these vortices (Liebman et  
574 al. 1994, Duvel 2015). For the Atlantic Ocean, the weak impact of the number of vortices  
575 can be tentatively attributed to the large and sustained number of vortices produced over  
576 West Africa and near the coast during the whole season. These numerous vortices may  
577 seed cyclogenesis over the MDR as soon as local environmental conditions are favorable.  
578 As shown in figure 10, this process is however mostly valid near the coast and is progres-  
579 sively replaced toward the west by local vortex initiations due for example to MCS organ-

580 ization or to AEWs reaching a critical layer. Patricola et al. (2018) showed that the cyclo-  
581 genesis frequency is maintained in a climate model if the AEWs are suppressed. This is  
582 because the model generates TCs by other mechanisms. In nature, however, it is possible  
583 that the cyclogenesis will be reduced, especially near the coast, if the AEW activity de-  
584 creases significantly. This will be the case if the development of a local vortex is not as  
585 efficient for cyclogenesis as the triggering by an already closed circulation coming from  
586 West Africa. In such a case, the excess of convective instability (and ocean heat content)  
587 could be released gradually by a succession of MCSs rather than rapidly by a single TC. In  
588 global warming simulations, current GCMs indicate an increase in the north path intensity  
589 and a decrease of the south path intensity (Skinner and Diffenbaugh 2014, Hannah and  
590 Aiyyer 2017, Kebe et al. 2020). Brannan and Martin (2020) also showed a shift of the  
591 south path amplitude toward the end of the season. These changes may influence the lon-  
592 gitudinal distribution of the cyclogenesis and the average cyclogenesis by modifying the  
593 seasonal phase of the AEW activity.

### 594 *7.3. North path and orography*

595 This analysis clearly shows the fundamental influence of the Hoggar mountains on the  
596 initiation of persistent dry vortices of the north path. This agrees with the recent study of  
597 Hamilton et al. (2020) who showed, using a high-resolution model, that the wave kinetic  
598 energy at low-level is reduced north of 15°N over West Africa when the orography is re-  
599 duced or removed. Previous results of Mozer and Zehnder (1996b), while based on a  
600 coarser resolution and simplified dry model, give interesting additional information on  
601 this vortex genesis process. In their model the easterly flow is blocked by the Hoggar  
602 mountains and forms a low-level jet primarily to the south. When the simulation starts, a  
603 vortex develops west of the Hoggar mountains as a transient disturbance that moves

604 downstream. The dry vortex genesis region west of the Hoggar mountains could thus re-  
605 sult from episodic reinforcements of the northeastern wind (Harmattan), related for ex-  
606 ample to intraseasonal pulsation of the West African heat low (Lavaysse et al. 2010), that  
607 would cause the development of such transient disturbances. This agrees with a hypoth-  
608 esis of Chen (2006) that the baroclinic instability at low-levels may be triggered by the  
609 intrusion of dry northerlies and transported westward by the Harmattan. This also agrees  
610 with Thorncroft and Rowell (1998) who showed a positive correlation between the  
611 strength of the easterly low-level wind and the AEW activity in a GCM, possibly due to the  
612 resulting stronger interaction with the orography. Further analysis using the vortex da-  
613 taset obtained here and ECMWF re-analysis could provide an interesting assessment of  
614 the role of these orographic processes on the origin of the dry north path vortices.

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- 628 Agudelo, P., C. D. Hoyos, J. Curry, and P. J. Webster, 2011: Probabilistic discrimination be-  
629 tween large-scale environ- ments of intensifying and decaying African easterly waves. *Cli-*  
630 *mate Dynamics*, **36**, 1379–1401.
- 631 Albignat, J. P., and R. J. Reed, 1980: The origin of African wave disturbances during Phase  
632 III of GATE. *Mon. Wea. Rev.*, **108**, 1827–1839.
- 633 Arnault, J., and F. Roux, 2011: Characteristics of African easterly waves associated with  
634 tropical cyclogenesis in the Cape Verde Islands region in July–August–September of  
635 2004–2008. *Atmos. Res.*, **100**, 61–82.
- 636 Avila, L. A., R. J. Pasch, and J.-G. Jiing, 2000: Atlantic tropical systems of 1996 and 1997:  
637 Years of contrasts, *Mon. Wea. Rev.*, **128**, 3695 – 3706.
- 638 Belanger, J. I., M. T. Jelinek, J. A. Curry, 2016: A climatology of easterly waves in the tropical  
639 Western Hemisphere. *Geosci. Data J.*, **3**, 40–49.
- 640 Berry, G., and C. D. Thorncroft, 2005: Case study of an intense African easterly wave. *Mon.*  
641 *Wea. Rev.*, **133**, 752–766.
- 642 Berry, G., C. D. Thorncroft and T. Hewson, 2007: African easterly waves during 2004 -  
643 Analysis using objective techniques. *Mon. Wea. Rev.*, **135**, 1251–1267.
- 644 Bou Karam, D., C. Flamant, P. Tulet, M. C. Todd, J. Pelon, and E. Williams, 2009: Dry cyclo-  
645 genesis and dust mobilization in the intertropical discontinuity of the West African Mon-  
646 soon: A case study, *J. Geophys. Res.*, **114**, D05115.
- 647 Brammer, A., and C. D. Thorncroft, 2015: Variability and evolution of African easterly  
648 wave structures and their relationship with tropical cyclogenesis over the eastern Atlan-  
649 tic. *Mon. Wea. Rev.*, **143**, 4975–4995.

650 Brammer, A., and C. D. Thorncroft, 2017: Spatial and temporal variability of the three-  
651 dimensional flow around African easterly waves. *Mon. Wea. Rev.*, **145**, 2879–2895.

652 Brannan, A. L., and E. R. Martin, 2019: Future characteristics of African Easterly Wave  
653 tracks. *Climate Dynamics*, **52**, 5567–5584.

654 Burpee, R. W., 1972: The origin and structure of easterly waves in the lower troposphere  
655 of North Africa. *J. Atmos. Sci.*, **29**, 77– 90.

656 Carlson, T. N., 1969: Synoptic histories of three African disturbances that developed into  
657 Atlantic hurricanes. *Mon. Wea. Rev.*, **97**, 256–276.

658 Cecelski, S. F., and D.-L. Zhang, 2013: Genesis of Hurricane Julia (2010) within an African  
659 easterly wave: Low-level vortices and upper-level warming. *J. Atmos. Sci.*, **70**, 3799–3817.

660 Chen T.-C., 2006: Characteristics of African easterly waves depicted by ECMWF reanalyses  
661 for 1991–2000. *Mon. Wea. Rev.*, **134**, 3539–3566.

662 Chen T. C., S.-Y. Wang, and A. J. Clark, 2008. North Atlantic hurricanes contributed by Afri-  
663 can easterly waves north and south of the African easterly jet. *J. Climate*, **21**, 6767–6776.

664 Chen, S.-H., and Y.-C. Liu, 2014: The relation between dry vortex merger and tropical cy-  
665 clone genesis over the Atlantic Ocean, *J. Geophys. Res. Atmos.*, **119**, 11,641–11,661.

666 Cheng, Y. M., C. D. Thorncroft, and G. N. Kiladis, 2019: Two contrasting African easterly  
667 wave behaviors. *J. Atmos. Sci.*, **76**, 1753-1768.

668 Dee, D.P., and co-authors, 2011: The ERA-Interim reanalysis: Configuration and perfor-  
669 mance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597.

670 DeMaria, M., J. A. Knaff, and B. H. Conell, 2001: A tropical cyclone genesis parameter for  
671 the tropical Atlantic. *Wea. Forecasting*, **16**, 219–233.

672 Diedhiou, A., S. Janicot, A. Viltard, P. de Félice, and H. Laurent, 1999 : Easterly waves re-  
673 gimes and associated convection over West Africa and the tropical Atlantic: results from  
674 the NCEP/NCAR and ECMWF reanalyses. *Climate Dynamics*, **15**, 795 - 822.

675 Diedhiou, A., Janicot, S., Viltard, A., and P. de Felice, 2002: Energetics of easterly wave dis-  
676 turbances over West Africa and the tropical Atlantic: A climatology from 1979-95  
677 NCEP/NCAR reanalyses. *Climate Dynamics*, **18**, 487–500.

678 Dieng, A. L., S. M. Sall, L. Eymard, M. Leduc-Leballeur, and A. Lazar, 2017: Trains of African  
679 easterly waves and their relationship to tropical cyclone genesis in the eastern Atlantic.  
680 *Mon. Wea. Rev.*, **145**, 599 – 616.

681 Dunkerton, T. J., M. T. Montgomery, and Z. Wang, 2009: Tropical cyclogenesis in a tropical  
682 wave critical layer: Easterly waves. *Atmos. Chem. Phys.*, **9**, 5587–5646.

683 Duvel, J. P., 1990: Convection over tropical Africa and the Atlantic Ocean during northern  
684 summer. Part II: Modulation by easterly waves. *Mon. Wea. Rev.*, **118**, 1855–1868.

685 Duvel, J. P., 2015: Initiation and Intensification of Tropical Depressions over the Southern  
686 Indian Ocean: Influence of the MJO. *Mon. Wea. Rev.*, **143**, 2170–2191.

687 Duvel, J.P., S.J. Camargo and A.H. Sobel, 2017: Role of the Convection Scheme in Modeling  
688 Initiation and Intensification of Tropical Depressions over the North Atlantic. *Mon. Wea.*  
689 *Rev.*, **145**, 1495-1509.

690 Emanuel, K.A., and D.S. Nolan, 2004: Tropical cyclone activity and global climate. In: Pro-  
691 ceedings of 26th conference on hurricanes and tropical meteorology. *American Meteoro-*  
692 *logical Society*, Miami, FL, pp 240–241.

693 Erickson, C. O., 1963: An incipient hurricane near the West African coast. *Mon. Wea. Rev.*,  
694 **91**, 61-68.

695 Fink, A. H., and A. Reiner, 2003: Spatio-temporal variability of the relation between Afri-  
696 can easterly waves and West African squall lines in 1998 and 1999. *J. Geophys. Res.*, **108**,  
697 4332.

698 Gray, W. M., 1968: Global view of the origin of tropical disturbances and storms. *Mon. Wea.*  
699 *Rev.*, **96**, 669–700.

700 Gray, W. M., 1998: The formation of tropical cyclones. *Meteor. Atmos. Phys.*, **67**, 37–69.

701 Hamilton, H. L., Núñez Ocasio, K. M., Evans, J. L., Young, G. S., and J. D. Fuentes, 2020: Topo-  
702 graphic influence on the African Easterly Jet and African Easterly Wave energetics. *J. Ge-*  
703 *ophys. Res.: Atmospheres*, **125**, e2019JD032138.

704 Hannah, W. M., and A. Aiyyer, 2017: Reduced African Easterly Wave Activity with Quad-  
705 rupled CO<sub>2</sub> in the Superparameterized CESM. *Journal of Climate*, **30**(20), 8253– 8274.

706 Hanks, I., Z. Wang, G. Zhang, and C. L. Fritz, 2015: Merger of African easterly waves and  
707 formation of Cape Verde storms. *Quart. J. Roy. Meteor. Soc.*, **141**, 1306–1319.

708 Hodges, K. I., 1995: Feature tracking on the unit sphere. *Mon. Wea. Rev.*, **123**, 3458–3465.

709 Hodges, K. I., 1999: Adaptive constraints for feature tracking. *Mon. Wea. Rev.*, **127**, 1362–  
710 1373.

711 Hodges, K. I., B. J. Hoskins, J. Boyle, and C. Thorncroft, 2003: A comparison of recent rea-  
712 nalysis datasets using objective feature tracking: Storm tracks and tropical easterly  
713 waves. *Mon. Wea. Rev.*, **131**, 2012–2037.

714 Hopsch S. B., C.D. Thorncroft, K. Hodges and A. Aiyyer, 2007: West African storm tracks  
715 and their relationship to Atlantic tropical cyclones. *J. Climate* **20**: 2468 – 2483.

716 Hopsch S. B., Thorncroft C. D. and K. R. Tyle, 2010: Analysis of African Easterly wave struc-  
717 tures and their role in influencing tropical cyclogenesis. *Mon. Wea. Rev.*, **138**, 1399–1419.



718 Kebe, I., I. Diallo, M. B. Sylla, F. De Sales, and A. Diedhiou, 2020: Late 21st Century Projected  
719 Changes in the Relationship between Precipitation, African Easterly Jet, and African East-  
720 erly Waves. *Atmosphere*, **11**, 353.

721 Knapp, K.R., M.C. Kruk, D.H. Levinson, H.J. Diamond, and C.J. Neumann, 2010: The Interna-  
722 tional Best Track Archive for Climate Stewardship (IBTrACS): Unifying tropical cyclone  
723 data. *Bull. Amer. Meteor. Soc.*, **91**, 363–376.

724 Laing A.G., Carbone R., Levizzani V. and J. Tuttle, 2008: The propagation and diurnal cycles  
725 of deep convection in northern tropical Africa. *Q. J. Roy. Meteorol. Soc.*, **134**, 93–109.

726 Landsea, C. W., 1993: A climatology of intense (or major) Atlantic hurricanes. *Mon. Wea.*  
727 *Rev.*, **121**, 1703–1713.

728 Lavaysse, C., C. Flamant, S. Janicot, and P. Knippertz, 2010: Links between African easterly  
729 waves, midlatitude circulation and intraseasonal pulsations of the West African heat low.  
730 *Q. J. R. Meteorol. Soc.*, **136**, 141–158.

731 Liebman, B., H. H. Hendon, and J. D. Glick, 1994: The relationship between tropical cy-  
732 clones of the western Pacific and Indian Oceans and the Madden–Julian oscillation. *J. Me-*  
733 *teor. Soc. Japan*, **72**, 401– 411.

734 Mekonnen, A., C. D. Thorncroft, and A. R. Aiyyer, 2006: Analysis of convection and its as-  
735 sociation with African easterly waves. *J. Climate*, **19**, 5405–5421.

736 Mekonnen, A., and W.B. Rossow 2018: The interaction between deep convection and east-  
737 erly wave activity over Africa: convective transitions and mechanisms. *Mon. Wea. Rev.*,  
738 **146**, 1945– 1961.

739 Menkes, C. E., M. Lengaigne, P. Marchesiello, N. C. Jourdain, E. M. Vincent, J. Lefèvre, F.  
740 Chauvin, and J.-F. Royer, 2011: Comparison of tropical cyclogenesis indices on seasonal to  
741 interannual timescales, *Climate Dyn.*, **38**, 301– 321.

742 Montgomery, M. T., M. E. Nicholls, T. A. Cram, and A. B. Saunders, 2006: A vortical hot  
743 tower route to tropical cyclogenesis. *J. Atmos. Sci.*, **63**, 355–386.

744 Mozer, J. B., and J. A. Zehnder, 1996a: Lee vorticity production by large-scale tropical  
745 mountain ranges. Part I: Eastern North Pacific tropical cyclogenesis. *J. Atmos. Sci.*, **53**, 521–  
746 538

747 Mozer, J. B., and J. A. Zehnder, 1996b: Lee vorticity production by large-scale tropical  
748 mountain ranges. Part II: A Mechanism for the Production of African Waves. *J. Atmos. Sci.*,  
749 **53**, 539–549.

750 Nitta, T., and Y. Takayabu, 1985: Global analysis of the lower tropospheric disturbances  
751 in the Tropics during the northern summer of the FGGE year. Part II: Regional character-  
752 istics of the disturbances. *Pure Appl. Geophys.*, **123**, 272–292.

753 Patricola C.M., R. Saravanan, and P. Chang, 2018: The response of Atlantic tropical cy-  
754 clones to suppression of African easterly waves. *Geophys. Res. Lett.*, **45**, 471–479.

755 Pytharoulis, I., and C. Thorncroft, 1999: The low-level structure of African easterly waves  
756 in 1995. *Mon. Wea. Rev.*, **127**, 2266–2280.

757 Reed, R. J. , D. C. Norquist, and E. E. Recker, 1977: The structure and properties of African  
758 wave disturbances as observed during Phase III of GATE. *Mon. Wea. Rev.*, **105**, 317–333.

759 Reed, R. J., E. Klinker, and A. Hollingsworth, 1988a: The structure and characteristics of  
760 African easterly wave disturbances as determined from the ECMWF operational analy-  
761 sis/forecast system. *Meteor. Atmos. Phys.*, **38**, 22–33.

762 Reed, R. J., A. Hollingsworth, W. A. Heckley, and F. Delsol, 1988b: An evaluation of the per-  
763 formance of the ECMWF operational forecasting system in analyzing and forecasting trop-  
764 ical easterly wave disturbances over Africa and the tropical Atlantic. *Mon. Wea. Rev.*, **116**,  
765 824–865.

766 Ritchie, E. A., and G. J. Holland, 1997: Scale interactions during the formation of Typhoon  
767 Irving. *Mon. Wea. Rev.*, **125**, 1377– 1396.

768 Ross, R. S., and T. N. Krishnamurti, 2007: Low-level African easterly wave activity and its  
769 relation to Atlantic tropical cyclogenesis in 2001. *Mon. Wea. Rev.* **135**: 3950 – 3964.

770 Russell, J. O., A. Aiyyer, J. D. White, and W. Hannah, 2017: Revisiting the connection be-  
771 tween African Easterly Waves and Atlantic tropical cyclogenesis. *Geophys. Res. Lett.*, **44**,  
772 587–595.

773 Thorncroft, C. D., and B. J. Hoskins, 1994: An idealized study of African easterly waves.  
774 Part II: A nonlinear view. *Quart. J. Roy. Meteor. Soc.*, **120**, 983–1015.

775 Thorncroft, C. D., and K. Hodges, 2001: African easterly wave variability and its relation-  
776 ship to Atlantic tropical cyclone activity. *J. Climate*, **14**, 1166–1179.

777 Tippett, M. K., Camargo, S. J. & Sobel, A. H., 2011: A Poisson regression index for tropical  
778 cyclone genesis and the role of large-scale vorticity in genesis. *J. Climate*, **24**, 2335–2357.

779 Wang, Z., M. T. Montgomery, and C. Fritz, 2012: A first look at the structure of the wave  
780 pouch during the 2009 PREDICT-GRIP dry runs over the Atlantic. *Mon. Wea. Rev.*, **4**, 1144–  
781 1163.

782 Wu, M. C., O. Reale, S. D. Schubert, M. J. Suarez, R. D. Koster, and P. J. Pegion, 2009: African  
783 Easterly Jet: Structure and Maintenance, *J. Climate*, **22**, 4459–4480.

784 Zhang, Y. C., F. Q. Zhang, C. A. Davis, and J. H. Sun, 2018: Diurnal evolution and structure of  
785 long-lived mesoscale convective vortices along the Mei-yu front over the East China  
786 plains. *J. Atmos. Sci.*, **75**, 1005–1025.

787

Acronyms Specific terms	Definition
Vortex	<p>A vortex is defined here as a synoptic-scale circulation around a local minimum in the geopotential height of a given pressure level. This is equivalent to a closed circulation, or a minimum of the streamfunction, in the geostrophic approximation. We consider that this approximation is valid in most cases even if it is not strictly the case for some of the smaller vortices.</p> <p>The term vortex also refers to the entire life cycle of a vortex disturbance moving over West Africa and the Atlantic Ocean.</p>
Vortex track	<p>A vortex track is a time series of consecutive barycenter positions of a given moving vortex. We consider only vortex tracks initiated over the domain 60°W-50°E and 0°N-30°N. The tracking is done for a larger domain 130°W-50°E and 0°N-60°N.</p>
Atlantic Vortex (AV)	<p>An AV is a vortex that spend at least one day over the Atlantic Ocean and enters the domain south of 20°N.</p>
North path	<p>The north path is the ensemble of low-level (850 hPa) AV initiated over West Africa north of 15°N and east of 10°W.</p>
South path	<p>The south path is the ensemble of mid-level (700 hPa) AV initiated over West Africa south of 15°N and east of 10°W.</p>
Main Development Region (MDR)	<p>The MDR is the domain extending from 10°W to 60°W and from 5°N to 20°N.</p>
Cyclogenetic Atlantic Vortex (CAV)	<p>A CAV is an AV that has at least one barycenter position within 3° of an IBTrACS system. More than 97% of CAVs have more than 4 consecutive positions (1 day) in common with their IBTrACS system.</p>
Non-Cyclogenetic Atlantic Vortex (NCAV)	<p>An NCAV is an AV with all barycenter positions at more than 3° from any IBTrACS system.</p>
AV Merging	<p>A merging of a low-level (850 hPa) and a mid-level (700 hPa) AV occurs when they are within 3° of each other for the first time. After merging, the two AV tracks are nearly identical in most cases (e.g. figure 1).</p>
Primary AV	<p>When two AVs merge, the primary AV is the older at merging time.</p>
Secondary AV	<p>When two AVs merge, the secondary AV is the younger at merging time.</p>
Confluence	<p>If both AVs are initiated long before merging (e.g. <math>\geq 1</math> day), the merging is called a “confluence” of the two vortex tracks.</p>
Vertical expansion (deepening)	<p>If the secondary AV is initiated shortly before merging (e.g. <math>&lt; 1</math> day), the merging corresponds more to a vertical expansion (or deepening) of the primary AV.</p>

789 Table 1: Definition of different acronyms and specific terms.

791

Region of Initiation	Primary AV Initiations				Impact on Cyclogenesis			
	(a) Number of AV initiations		(b) % of CAV		(c) % cyclogenesis MDR		(d) % cyclogenesis Western Regions	
	850 hPa	700 hPa	850 hPa	700 hPa	850 hPa	700 hPa	850 hPa	700 hPa
North Path	<b>428</b>	32	<b>10</b>	16	<b>16</b>	2	<b>5</b>	0
South Path	54	<b>424</b>	26	<b>23</b>	6	<b>38</b>	1	<b>6</b>
Continent, east of 10°W	482	456	12	23	22	40	6	6
	938		17		62		12	
Atl. and coast, west of 10°W	562	770	11	9	18	20	11	14
	1332		10		38		25	

792

793 Table 2 : Statistics on (a) the number of primary AV initiations and (b) the percentage of  
794 cyclogenetic AV (CAV) for different regions of initiation and contribution of AVs coming  
795 from these initiation regions to cyclogenesis (c) over the MDR and (d) over regions west  
796 of 60°W and south of 20°N. North and south path initiation regions are located east of  
797 10°W and north and south of 15°N, respectively (Fig.1). The “continent” statistics corre-  
798 sponds to the summation of north and south path statistics. The statistics are shown for  
799 each pressure level (850 hPa and 700 hPa) and on average for both pressure levels for the  
800 continent and for the Atlantic and coastal regions. Numbers in bold represent the main  
801 contribution for the north and the south path.

802

803 **Figure captions**

804 Figure 1: Domain of vortex initiations (thin black) and more specifically for the north  
805 (blue) and the south (red) path vortices. The limits of the Main Development Region  
806 (MDR) (magenta) include longitudes between 10°W and 60°W. Also shown is an example  
807 for two vortex tracks, merging near the coast, and at the origin of tropical cyclone Edouard  
808 in September 2014 (markers are drawn one day apart). “Atlantic and coast” and  
809 “Continent” refer to areas west and east of 10°W, respectively.

810 Figure 2: Initiation density of the barycenter of all Atlantic Vortices (AV) detected at (a)  
811 700 hPa and (b) 850 hPa between June and October 1980-2017. AV are vortices spending  
812 at least one day over the Atlantic Ocean. Fields are smoothed by a 5°x5° running mean.  
813 The first contour (dotted line) is 0.5 and the increment is 1. The total number of AV initi-  
814 ations is reported in the upper right corners.

815 Figure 3: As in figure 2 but for AV tracks that do not merge with an AV track of the other  
816 pressure level.

817 Figure 4: As in figure 2, but for initiation density of the barycenter of Not Cyclogenetic  
818 Atlantic Vortices (NCAV) detected at (a) 700 hPa and (b) 850 hPa and that merge with a  
819 secondary AV track of the other pressure level. (c) Initiation density of secondary vortices  
820 at 850 hPa associated with a primary vortex at 700 hPa. (d) Initiation density of secondary  
821 vortices at 700 hPa associated with a primary vortex at 850 hPa. Merging position density  
822 for the barycenter of the primary vortices at (e) 700 hPa and (f) 850 hPa. The first contour  
823 (dotted line) is 0.5 and the increment is 1.

824 Figure 5: As in figure 4, but for initiation density of Cyclogenetic Atlantic Vortices (CAV).  
825 The bottom panel represents the density of the first match between a CAV and an IBTrACS

826 system for (g) 700 hPa and (h) 850 hPa primary vortices. This match generally corre-  
827 sponds to the first time-step of an IBTrACS system and is considered here as a cyclogene-  
828 sis. The first contour (dotted line) is 0.125 and the increment is 0.25.

829 Figure 6: Track density for AVs that do not merge (a, b), primary merging NCAV (c, d), and  
830 corresponding secondary NCAV at the other level (e, f). The first contour (dotted line) is  
831 0.5 and the increment is 1.

832 Figure 7: As in figure 6 for primary CAVs (a, b) and corresponding secondary CAVs (c, d).  
833 The first contour (dotted line) is 0.125 and the increment is 0.25.

834 Figure 8: Seasonal variation of the average monthly number of primary AV ( $N_{AV}$ : filled  
835 circles) and CAV ( $N_{CAV}$ : open circles) initiations at 700 hPa (red) and 850 hPa (blue) for  
836 (a) merging AV initiated near the coast (between  $10^{\circ}W$  and  $20^{\circ}W$ ); (b) merging AV initi-  
837 ated over West Africa (east of  $10^{\circ}W$ ); (c) and (d), as for (a) and (b), but for non-merging  
838 AV.

839 Figure 9: Seasonal variations of (a) cyclogenetic efficiency (CE) and (b) merging efficiency  
840 (ME) for primary AV initiated east of  $10^{\circ}W$  and (red) south of  $15^{\circ}N$  at 700 hPa, (blue)  
841 north of  $15^{\circ}N$  at 850 hPa. (black) average for both paths. (c) as in (a) but for interannual  
842 variations of the mean CE for August and September.

843 Figure 10: Longitudinal distribution of the proportion of cyclogenesis as a function of the  
844 easternmost initiation region of the CAV at either 700 hPa or 850 hPa: (red) West Africa,  
845 (green) Coast and (blue) Ocean. The dotted curve represents the longitudinal distribution  
846 of the season-average number of cyclogenesis.

847 Figure 11: Longitudinal distribution of the average vortex intensity (maximum  $\Delta g_z$  in the  
848 “vortex area”) for primary CAV and NCAV tracks initiated east of  $10^{\circ}W$  and between (a)  
849  $5^{\circ}N$  and  $15^{\circ}N$  at 700 hPa and (b)  $15^{\circ}N$  and  $25^{\circ}N$  at 850 hPa. (c) and (d), as for (a) and (b),



850 but for the average vortex intensity as a function of the delay relative to the vortex deep-  
851 ening for deepening occurring west of 10°W (i.e. for already formed continental AVs). The  
852 shading represents the uncertainty on the average values for a 99% significance level. In  
853 (c) and (d), the bar graph represents the number of vortices with a given delay between  
854 vortex deepening and cyclogenesis.

855 Figure 12: Schematic illustrating the four merging categories of the 700 hPa (red) and the  
856 850 hPa (blue) vortex tracks: (a) merging of vortices of the north and the south path, (b)  
857 local vortex development, (c) downward expansion of a vortex of the south path and, (d)  
858 upward expansion of a vortex of the north path. The proportion of each merging category  
859 is reported in the lower left.

860 Figure 13: Initiation density (#/season/10° box) for (red) 700 hPa and (blue) 850 hPa  
861 Atlantic Vortices belonging to the four merging categories a, b, c and d of figure 12. The  
862 first contour (dotted line) is 1 and the increment is 1. The black dots correspond to the  
863 location of cyclogenesis (first match with an IBTrACS system). The proportion of merge  
864 and of cyclogenesis are reported in the upper right boxes for each category. Fields are  
865 smoothed by a 5°x5° running mean.

866 Figure 14: As in figure 13, but for the merging density. The first contour (dotted line) is 1  
867 and the increment is 1.

868 Figure 15: Seasonal variation of the number of merges due to AV (red) and to CAV (black)  
869 for the four merging categories of figure 12. The grey filled curves represent the cyclogen-  
870 esis efficiency.

871

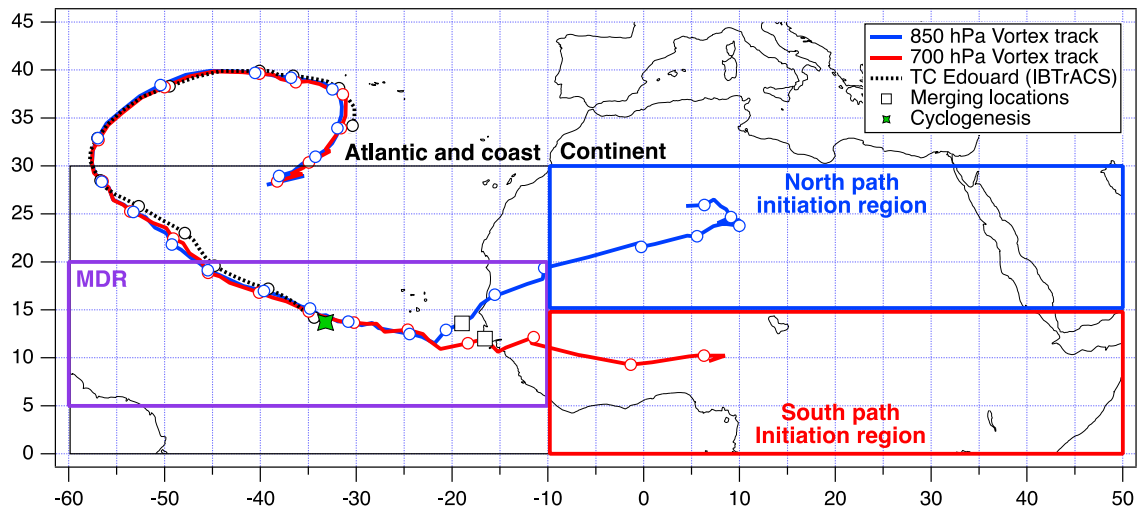


Figure 1: Domain of vortex initiations (thin black) and more specifically for the north (blue) and the south (red) path vortices. The limits of the Main Development Region (MDR) (magenta) include longitudes between 10°W and 60°W. Also shown is an example for two vortex tracks, merging near the coast, and at the origin of tropical cyclone Edouard in September 2014 (markers are drawn for 00GMT). “Atlantic and coast” and “Continent” refer to areas west and east of 10°W, respectively.

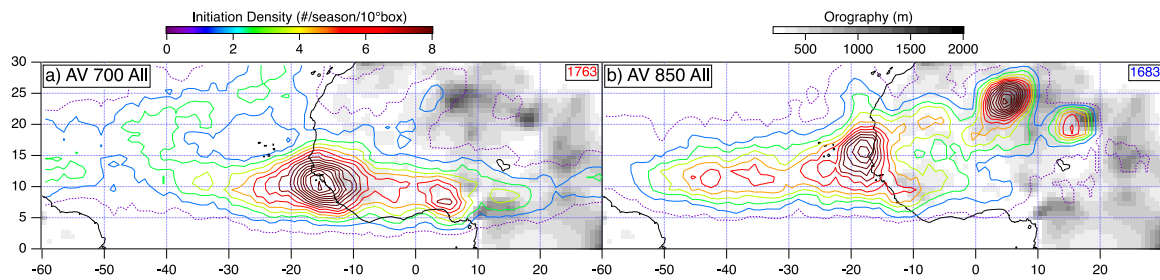


Figure 2: Initiation density of the barycenter of all Atlantic Vortices (AV) detected at (a) 700 hPa and (b) 850 hPa between June and October 1980-2017. AV are vortices spending at least one day over the Atlantic Ocean. Fields are smoothed by a  $5^\circ \times 5^\circ$  running mean. The first contour (dotted line) is 0.5 and the increment is 1. The total number of AV initiations is reported in the upper right corners.

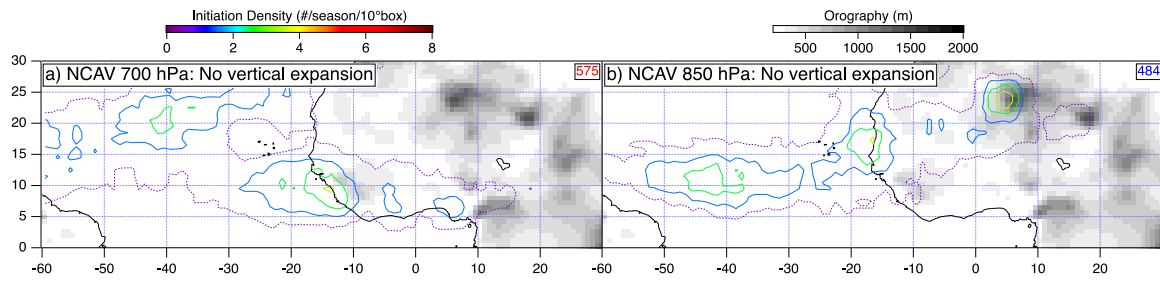


Figure 3: As in figure 2 but for AV tracks that do not merge with an AV track of the other pressure level.

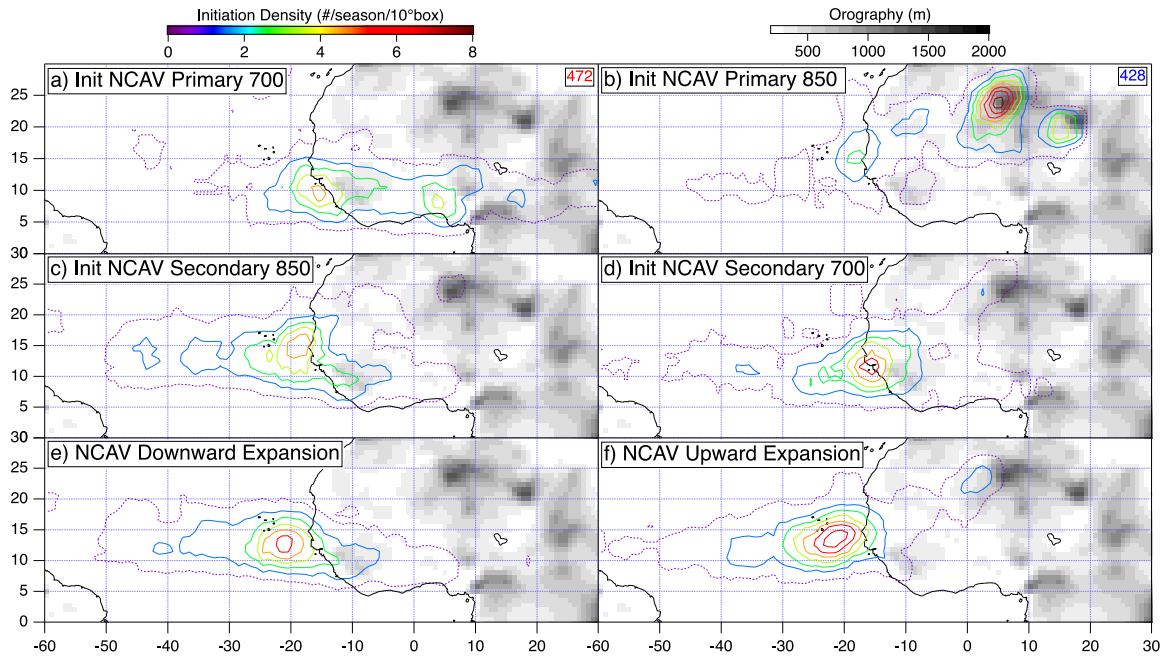


Figure 4: As in figure 2, but for initiation density of the barycenter of Not Cyclogenetic Atlantic Vortices (NCAV) detected at (a) 700 hPa and (b) 850 hPa and that merge with a secondary AV track of the other pressure level. (c) Initiation density of secondary vortices at 850 hPa associated with a primary vortex at 700 hPa. (d) Initiation density of secondary vortices at 700 hPa associated with a primary vortex at 850 hPa. Merging position density for the barycenter of the primary vortices at (e) 700 hPa and (f) 850 hPa. The first contour (dotted line) is 0.5 and the increment is 1.

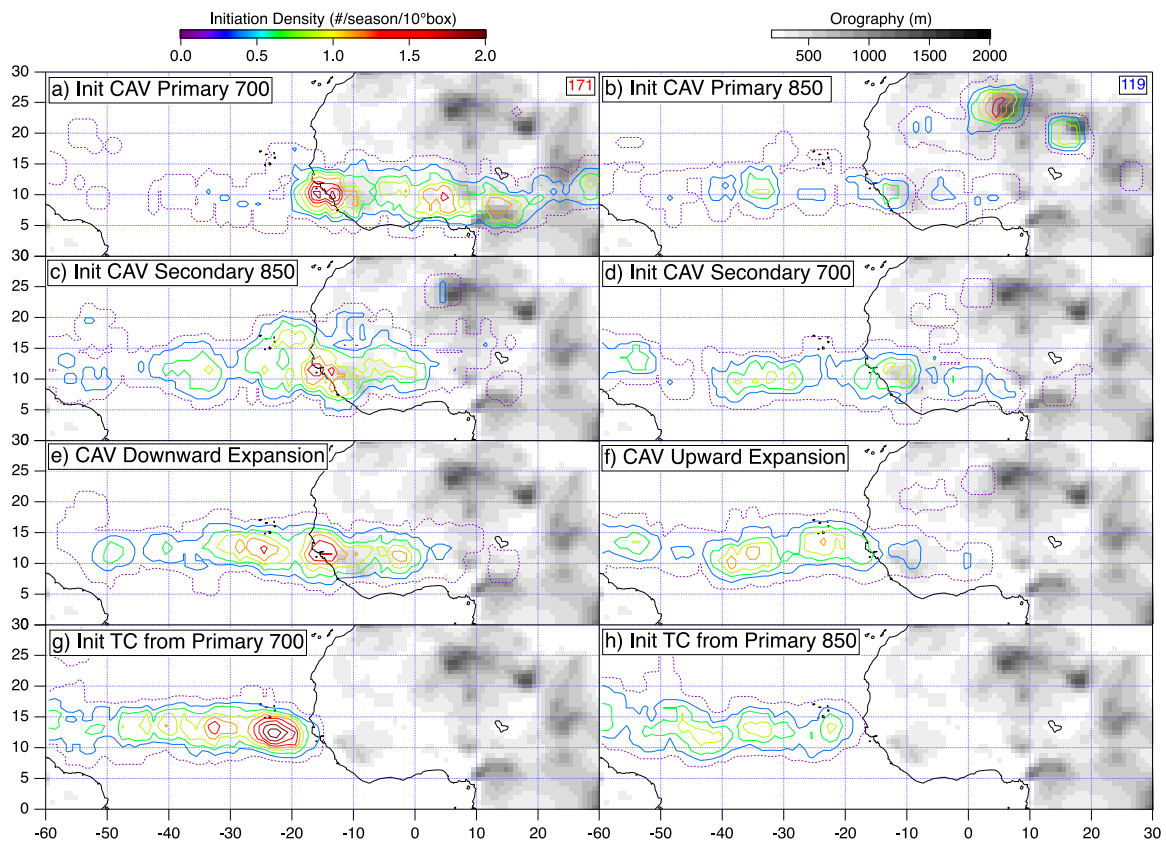


Figure 5: As in figure 4, but for initiation density of Cyclogenetic Atlantic Vortices (CAV). The bottom panel represents the density of the first match between a CAV and an IBTrACS system for (g) 700 hPa and (h) 850 hPa primary vortices. This match generally corresponds to the first time-step of an IBTrACS system and is considered here as a cyclogenesis. The first contour (dotted line) is 0.125 and the increment is 0.25.

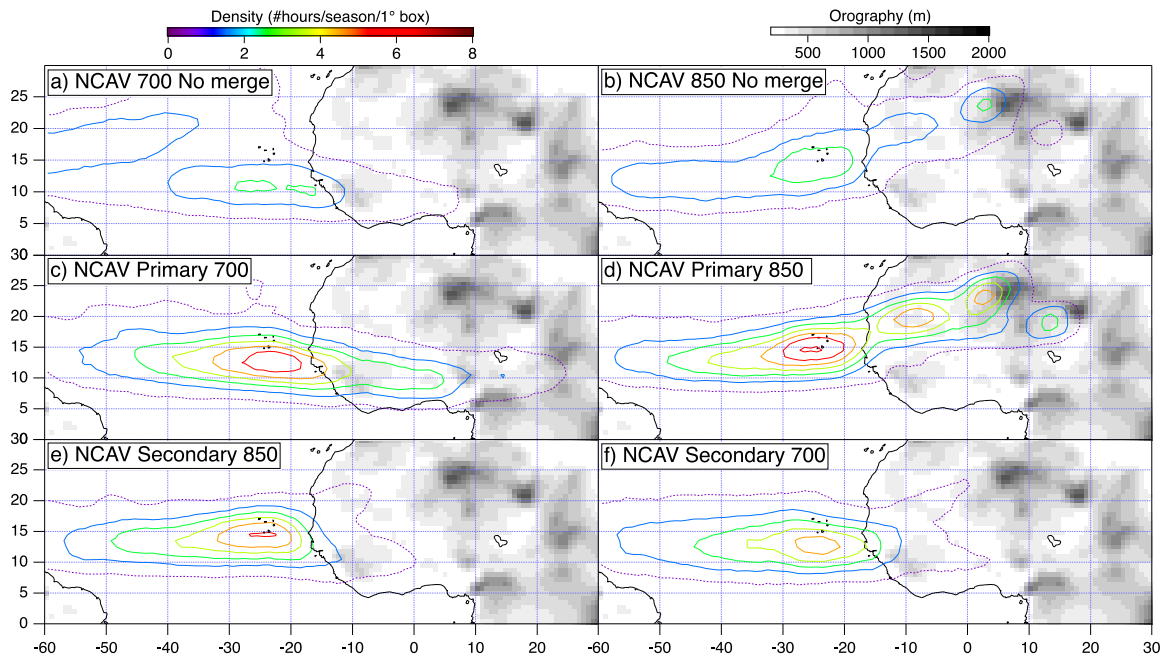


Figure 6: Track density for AVs that do not merge (a, b), primary merging NCAV (c, d), and corresponding secondary NCAV at the other level (e, f). The first contour (dotted line) is 0.5 and the increment is 1.

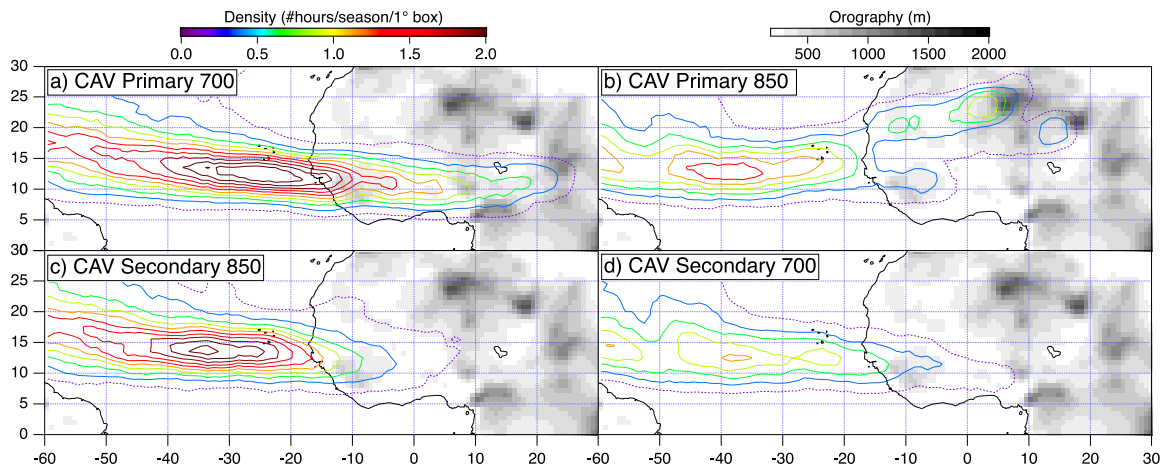


Figure 7: As in figure 6 for primary CAVs (a, b) and corresponding secondary CAVs (c, d).

The first contour (dotted line) is 0.125 and the increment is 0.25.



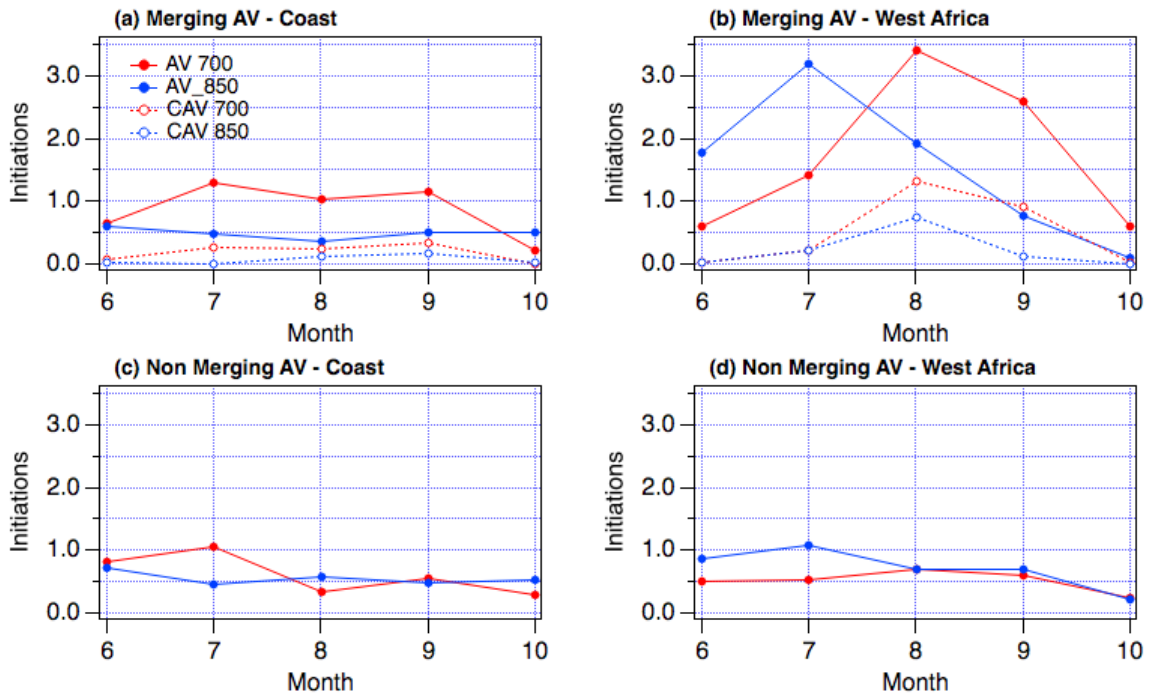


Figure 8: Seasonal variation of the average monthly number of primary AV ( $N_{AV}$ : filled circles) and CAV ( $N_{CAV}$ : open circles) initiations at 700 hPa (red) and 850 hPa (blue) for (a) merging AV initiated near the coast (between  $10^{\circ}W$  and  $20^{\circ}W$ ); (b) merging AV initiated over West Africa (east of  $10^{\circ}W$ ); (c) and (d), as for (a) and (b), but for non-merging AV.

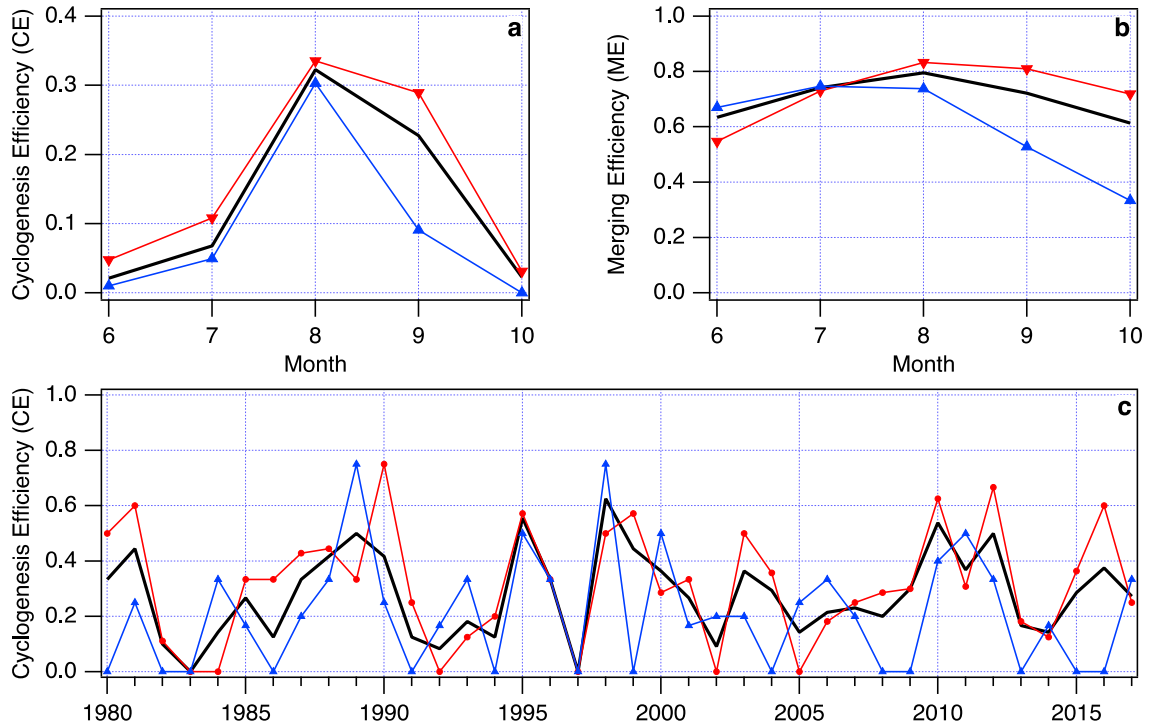


Figure 9: Seasonal variations of (a) cyclogenetic efficiency (CE) and (b) merging efficiency (ME) for primary AV initiated east of  $10^{\circ}\text{W}$  and (red) south of  $15^{\circ}\text{N}$  at 700 hPa, (blue) north of  $15^{\circ}\text{N}$  at 850 hPa. (black) average for both paths. (c) as in (a) but for interannual variations of the mean CE for August and September.

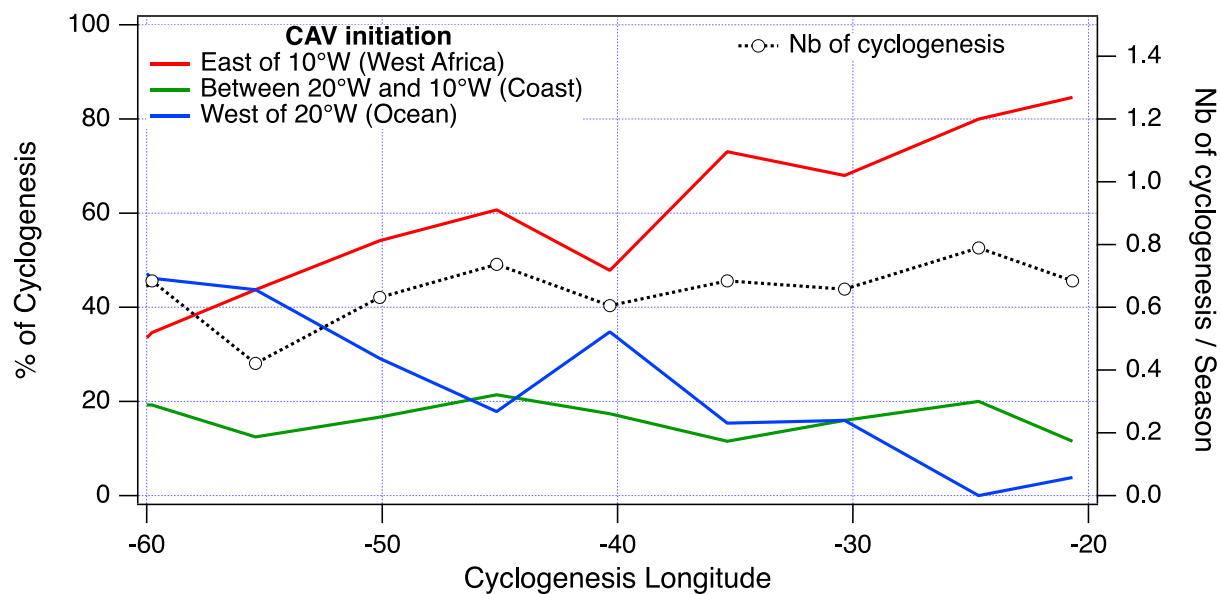


Figure 10: Longitudinal distribution of the proportion of cyclogenesis as a function of the easternmost initiation region of the CAV at either 700 hPa or 850 hPa: (red) West Africa, (green) Coast and (blue) Ocean. The dotted curve represents the longitudinal distribution of the season-average number of cyclogenesis.

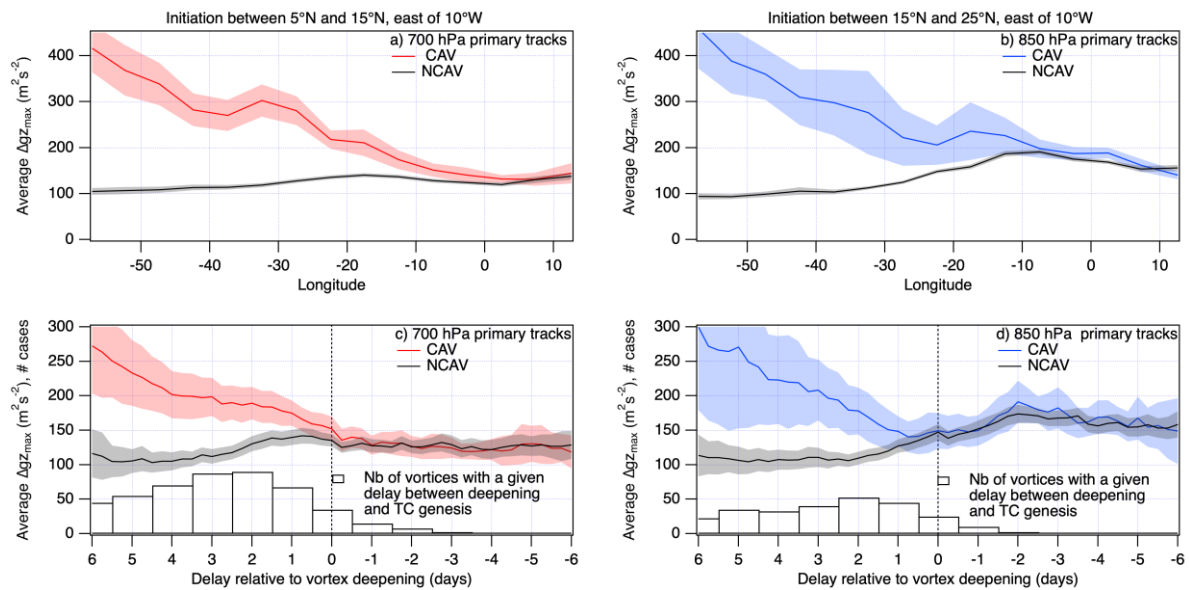


Figure 11: Longitudinal distribution of the average vortex intensity (maximum  $\Delta g_z$  in the “vortex area”) for primary CAV and NCAV tracks initiated east of 10°W and between (a) 5°N and 15°N at 700 hPa and (b) 15°N and 25°N at 850 hPa. (c) and (d), as for (a) and (b), but for the average vortex intensity as a function of the delay relative to the vortex deepening for deepening occurring west of 10°W (i.e. for already formed continental AVs). The shading represents the uncertainty on the average values for a 99% significance level. In (c) and (d), the bar graph represents the number of vortices with a given delay between vortex deepening and cyclogenesis.

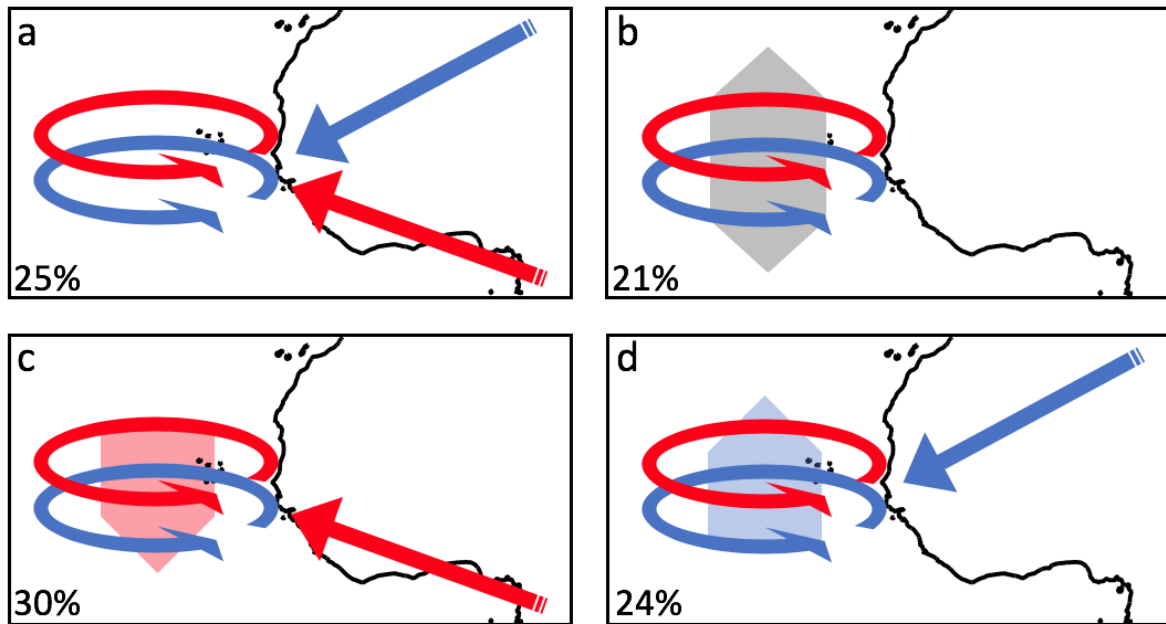


Figure 12: Schematic illustrating the four merging categories of the 700 hPa (red) and the 850 hPa (blue) vortex tracks: (a) merging of vortices of the north and the south path, (b) local vortex development, (c) downward expansion of a vortex of the south path and, (d) upward expansion of a vortex of the north path. The proportion of each merging category is reported in the lower left.

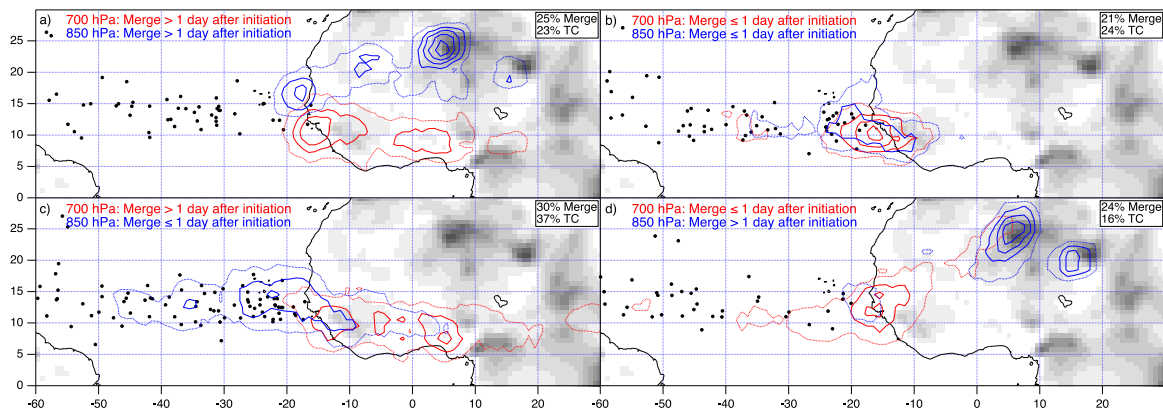


Figure 13: Initiation density ( $\#/season/10^\circ$  box) for (red) 700 hPa and (blue) 850 hPa Atlantic Vortices belonging to the four merging categories a, b, c and d of figure 12. The first contour (dotted line) is 1 and the increment is 1. The black dots correspond to the location of cyclogenesis (first match with an IBTrACS system). The proportion of merge and of cyclogenesis are reported in the upper right boxes for each category. Fields are smoothed by a  $5^\circ \times 5^\circ$  running mean.

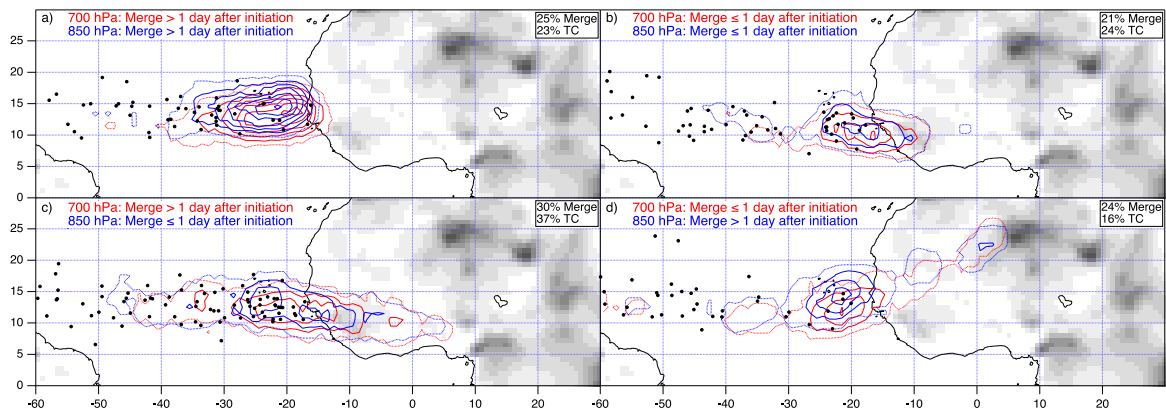


Figure 14: As in figure 13, but for the merging density. The first contour (dotted line) is 1 and the increment is 1.

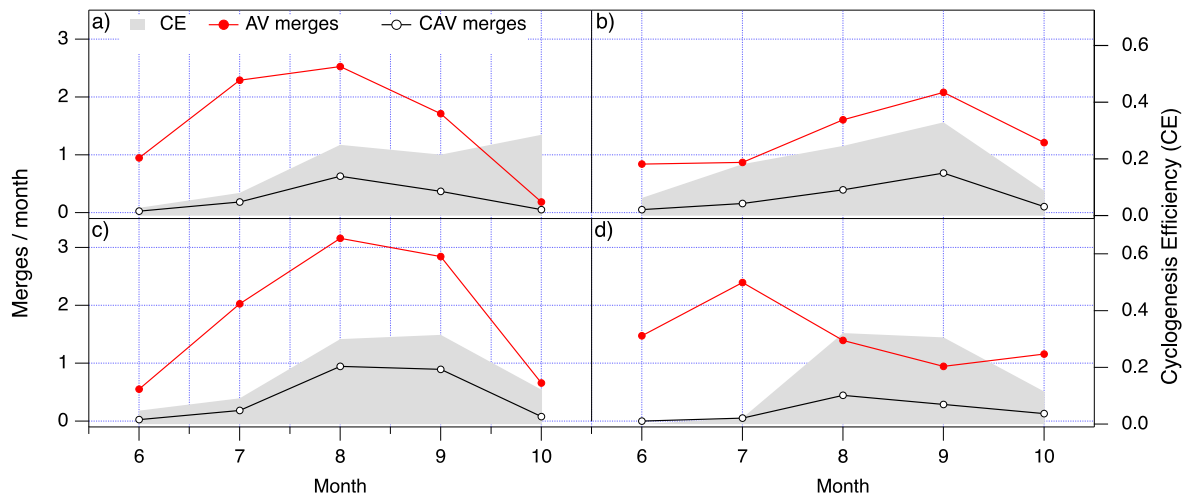


Figure 15: Seasonal variation of the number of merges due to AV (red) and to CAV (black) for the four merging categories of figure 12. The grey filled curves represent the cyclogenesis efficiency.