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# Origin and Evolution of Synoptic-Scale Vortices Initiated at Low Level Downwind of the Hoggar Mountains

Jean-Philippe Duvel

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1           **Origin and evolution of synoptic-scale vortices initiated at low-level**  
2                           **downwind of the Hoggar Mountains**

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5                           Jean-Philippe Duvel

6                           *Laboratoire de Météorologie Dynamique, CNRS, Paris, France*

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21           *Corresponding author address:*  
22           Laboratoire de Météorologie Dynamique,  
23           École Normale Supérieure,  
24           24, rue Lhomond, F75231 Paris, France  
25           [jpduvel@lmd.ipsl.fr](mailto:jpduvel@lmd.ipsl.fr)  
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## ABSTRACT

28 Numerous low-level vortices are initiated downwind of the Hoggar Mountains and  
29 progress towards the Atlantic coast on the northern path of African Easterly Waves (AEWs).  
30 These vortices occur mostly in July and August and more specifically when the northern  
31 position of the Saharan heat low (SHL) generates stronger and vertically expanded easterly  
32 winds over Hoggar mountains. At synoptic time-scales, a composite analysis reveals that  
33 vortex initiation and westward motion are also statistically triggered by a reinforcement of  
34 these easterly winds by a wide and persistent high-pressure anomaly developing around the  
35 Strait of Gibraltar and by a weak wave trough approaching from the east. The vortices are  
36 generated in the lee of the Hoggar, about 1000 km west of this approaching trough, and  
37 intensify rapidly. The evolution of the vortex perturbation is afterward comparable with the  
38 known evolution of the AEWs of the northern path and suggest a growth due to dry  
39 barotropic and baroclinic processes induced in particular by the strong cyclonic shear  
40 between the reinforced easterly winds and the monsoon flow. These results show that vortex  
41 genesis promoted by changes in orographic forcing due to the strengthening of easterly winds  
42 over Hoggar mountains is a source of intensification of the northern path of AEWs in July  
43 and August. These results also provide a possible mechanism to explain the role of the SHL  
44 and of particular mid-latitude intraseasonal disturbances on the intensity of these waves.

## 45 **1. Introduction**

46 Over West Africa, synoptic-scale vortices associated with the trough of African Easterly  
47 Waves (AEWs) are moving on two paths located roughly on either side of 15°N. Vortices on  
48 the northern path are dry and located mostly at low-level (i.e. ~850hPa), while vortices on the  
49 southern path are associated with deep convection and located mostly at mid-level (i.e.  
50 ~700hPa) near the African Easterly Jet (AEJ) (see e.g., Thorncroft and Hodges 2001, Chen et  
51 al. 2008, Hopsch et al. 2007, Duvel 2021). The initiation of a vortex at a given pressure level  
52 corresponds to a deepening of the wave trough and therefore to an intensification of the  
53 AEW. This deepening will sometimes persist for long distances over West Africa and the  
54 Atlantic Ocean where vortices of both paths are known sources of tropical storms and  
55 hurricanes (see e.g., Hopsch et al. 2007, Chen et al. 2008, Chen and Liu 2014, Russel et al.  
56 2017, Duvel 2021). In July and August, low-level vortices on the northern path are initiated  
57 mostly over a small region in the lee of Hoggar mountains Duvel (2021). The objective of  
58 this paper is to explore the conditions of formation of these “Hoggar vortices”, in relation to

59 seasonal and intraseasonal variations of the atmospheric circulation over the Hoggar  
60 orography.

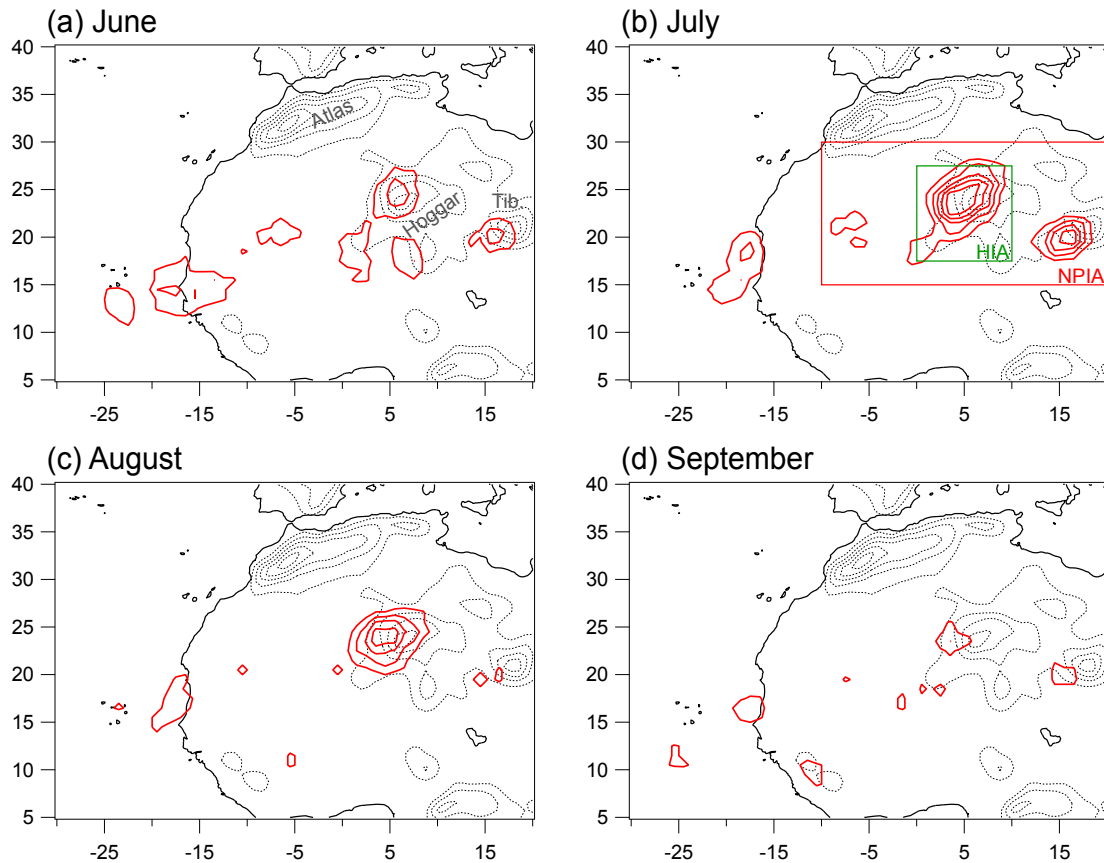
61 The role of orography in triggering or intensifying AEW on the northern (hereafter AEW-  
62 N) or the southern (AEW-S) path has been proposed by many studies since pioneer work  
63 based on radiosonde measurements. Carlson (1969) first suggested that the origin of AEW-S  
64 could be linked to moist convection over the topography of central and eastern Africa. This  
65 role of orographic convection has been later established in more detail by several studies  
66 using meteorological analyzes and satellite measurements (Reed et al. 1988, Thorncroft and  
67 Hodges 2001, Berry and Thorncroft 2005, Mekonnen et al. 2006, Mekonnen and Rossow  
68 2018). Dry orographic processes were also mentioned, like in Reed et al. (1988) who  
69 identified a cluster of AEW-N initiations located over the Sahara downwind of the Hoggar  
70 Mountains ( $\sim 5^{\circ}\text{E}$ ,  $24^{\circ}\text{N}$ ). Thorncroft and Hodges (2001) also found more frequent low-level  
71 vortex initiation downstream of the Hoggar Mountains and invoked the possible role of  
72 orography in the genesis of the low-level northern path disturbances. Duvel (2021) found that  
73 most of the low-level AEW-N vortices that reach the Atlantic Ocean, where they may trigger  
74 tropical cyclones, are initiated just west of the Hoggar Mountains in July and August. In  
75 addition to their potential role in cyclogenesis, Fiedler et al. (2014) found many of these low-  
76 level cyclonic disturbances are at the origin of Saharan dust lifting between the Hoggar and  
77 the Atlantic Ocean.

78 Studies based on numerical model simulations also suggested that the interaction between  
79 the Hoggar mountains and the large-scale flow could impact AEW-N. Thorncroft and Rowell  
80 (1998) showed that the strength of the low-level northeasterly flow over the Hoggar impact  
81 AEW-N amplitude. Hamilton et al. (2020) showed that the wave kinetic energy at low-level  
82 is reduced north of  $15^{\circ}\text{N}$  over West Africa when the topography is reduced or removed.  
83 White et al. (2021) showed more specifically a large reduction of the kinetic energy of AEW-  
84 N when the Hoggar and Tibesti Mountains are removed, due to reduction in baroclinic energy  
85 conversion related to reduced vertical wind shear. The initiation of vortices by a dry flow  
86 around a mountain was studied in Mozer and Zehnder (1996a, b) using an idealized  
87 numerical model. They showed that the blocking of an easterly flow by the Hoggar mountain  
88 may generate a barotropically unstable jet at low level which produces lee vortices  
89 downstream, being a possible source of AEW-N. Smaller meso-scale vortices are also  
90 initiated in the cyclonically sheared strip between the Harmattan and the monsoon flow. For

91 example, Bou Karam et al. (2009) studied short-lived and stationary vortices of smaller sizes  
92 initiated south of the Hoggar Mountains.

93 Previous studies have therefore suggested that the orographic forcing of the Hoggar plays  
94 a role in the initiation or intensification of AEW-N. In contrast with AEW-S, the deep  
95 convection must have little direct influence on wave initiation and growth for AEW-N path  
96 that is located north of 15°N over the Sahara Desert. Moreover, the AEW-N path is located  
97 more than 10° north of the AEJ core and near the surface, suggesting that physical sources  
98 other than AEJ instabilities may play a significant role on AEW-N activity. The literature  
99 review above shows that few papers have specifically addressed the observed seasonal and  
100 intraseasonal mechanisms behind the initiation of vortices downwind of the Hoggar and the  
101 impact of these vortices on AEW-N activity. These papers are based either on idealized  
102 simulations (Mozer and Zehnder, 1996b) or on sensitivity studies using numerical models  
103 with or without orography (see e.g., Hamilton et al. 2020, White et al. 2021). These  
104 sensitivity studies provide an estimate of the overall impact of orography on AEW-N, but as  
105 the removal of orography also significantly affects the mean dynamics and thermodynamics,  
106 it is difficult to isolate the specific orography processes that play a role in the AEW-N  
107 dynamics.

108 The objective of this paper is to study the origin of the observed initiation of low-level  
109 vortices in the lee of the Hoggar by relating this initiation to characteristics of the large-scale  
110 flow over this mountain. This paper considers two time-scales, the seasonal time scale to  
111 explore the average large-scale conditions that may explain why the Hoggar vortices are  
112 mostly initiated in July and August and the synoptic time scales that gives the particular  
113 conditions at time of vortex initiation. Section 2 presents the analysis approach and gives a  
114 statistic on the vortex tracks and on their impact on the AEW-N signal. The origin of the  
115 seasonal variation of the circulation over the Hoggar and its potential effect on vortex  
116 initiation and propagation is discussed in section 3. The conditions of formation of the  
117 vortices at synoptic time-scales (i.e. 2-10 days) are presented in section 4 and the  
118 characteristics of the vortices as they move toward the African coast are presented in section  
119 5. The impact of these results on the origin of AEW-N is discussed in section 6.



120  
 121 Figure 1: (**red contours**) Number of primary “Atlantic Vortex” initiations for a given month in 1° boxes for the  
 122 38 years. For a better legibility, initiation fields are smoothed by a 3°x3° running mean. The first contour is 0.5  
 123 and the contour increment is 0.5. (**black dotted contours**) Orography with a contours every 250m from 500m.  
 124 On panel b, the red rectangle defines the North Path Initiation Area (NPIA) for initiation of AEW-N vortices, the  
 125 green square defines the Hoggar Initiation Area (HIA) for “Hoggar vortices”

## 126 2. Analysis approaches and vortex statistics

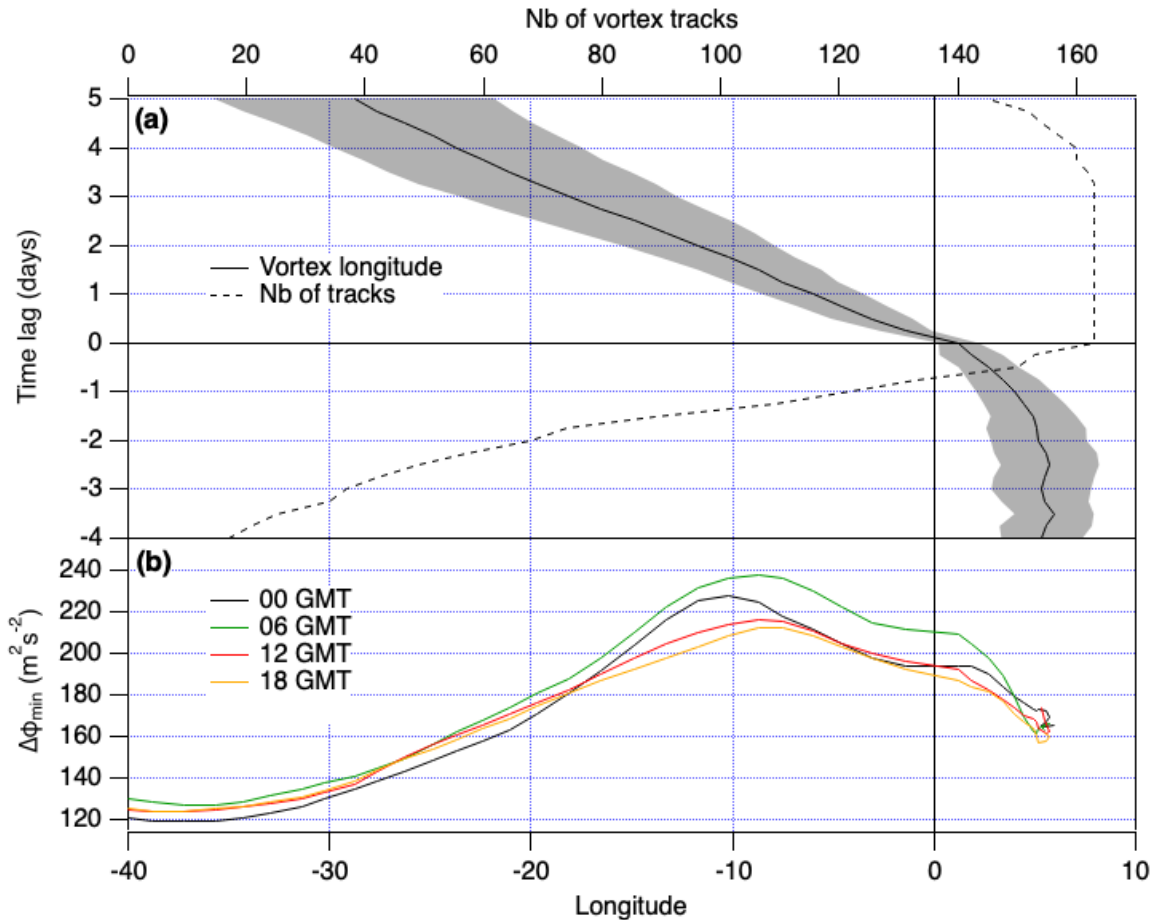
### 127 *Vortex tracking algorithm*

128 The analysis is based on ERA-Interim (Dee et al. 2011) meteorological re-analyses  
 129 between 1980 and 2017 with a horizontal resolution of 0.75° x 0.75° and a 6-hour time step.  
 130 In order to assess the proper representation of the vortices in ERA-I and to analyze possible  
 131 impact of these vortices on the cloudiness, we also use brightness temperatures of the  
 132 CLAUS (Cloud Archive User Service) dataset (Hodges et al., 2000) that are available with a  
 133 3-hour time step for years 1983 to 2009. The set of vortex tracks used in this study is the  
 134 same as that used in Duvel (2021) and is obtained using the objective tracking approach  
 135 described in Duvel (2015) and Duvel et al. (2017). This approach is based on geopotential  
 136 height anomaly  $\Delta\phi$  for a particular isobar (850 hPa here for AEW-N vortices). At a given  
 137 grid point,  $\Delta\phi$  is defined as the difference between the geopotential height and its average

138 over a region of  $\pm 7.5^\circ$  (i.e.,  $\pm 10$  ERA-I gridpoints) centered on the grid point. The “vortex  
139 area” is an ensemble of continuous model grid points with values of  $\Delta\phi$  lower than a negative  
140 threshold. This threshold is adjusted for each vortex and each time step to limit the vortex  
141 size to a maximum area of 100 model grid points (i.e., a radius of  $4.2^\circ$  for a circular shape).  
142 The intensity of a vortex is given by the minimum value  $\Delta\phi_{min}$  (i.e., maximum absolute  
143 value) of  $\Delta\phi$  in the vortex area. A given vortex is tracked over time by considering the  
144 overlap between the vortex areas for two consecutive time steps. A vortex track is therefore a  
145 time series of the successive position of the barycenter (i.e. the center weighted by  $\Delta\phi$ ) of  
146 these vortex areas, the first position being considered as the initiation location. As in Duvel  
147 (2021), we only consider vortex tracks that last more than two days and remain at least one  
148 day over the Atlantic Ocean. These vortices are called “Atlantic vortices” hereinafter. For the  
149 “Atlantic vortex” distribution maps reported in Figure 1, we consider only primary vortex  
150 initiations at 850 hPa. A primary initiation corresponds to a first vortex detection that does  
151 not result from the vertical extension on an already existing vortex track at another pressure  
152 level (700 hPa here). This distinction is mostly useful for the AEW-S path and near the coast  
153 where many low-level initiations are related to downward extension of mid-level vortices  
154 (Duvel 2021). The area of initiation of AEW-N vortices is defined as the area north of  $15^\circ\text{N}$   
155 and east of  $10^\circ\text{W}$ , which is the Northern Path Initiation Area (NPIA) outlined in red in Figure  
156 1b. Over the NPIA, most low-level AEW-N vortices initiations are primary initiations  
157 (Duvel, 2021) and this distinction is therefore not made in what follows. A vortex track is  
158 considered cyclogenetic if it is located within  $3^\circ$  of an IBTrACS system (Knapp et al. 2010)  
159 for at least one time step.

#### 160 *Atlantic Vortex initiation statistics*

161 Between June and September, there are about 3 Atlantic vortices initiated each month at  
162 850 hPa over the NPIA and about 12% of these vortices are cyclogenetic (Table 1). There is  
163 also a large seasonal variation of the number of initiations, with a maximum of 4.4 per month  
164 in July and less than 2 per month in September. The cyclogenetic efficiency of these vortices  
165 (i.e., the probability that their track matches an IBTrACS system) has a sharp maximum of  
166 29% in August. As discussed in Duvel (2021) the cyclogenesis efficiency of these dry AEW-  
167 N vortices is indeed as large as the AEW-S vortices in August when the cyclogenesis  
168 potential index of the Atlantic Ocean is a maximum. The cyclogenesis efficiency of AEW-N  
169 vortices is small on the average because they peak in July when this index is smaller.



170 Figure 2: (a) Average (solid) and standard deviation (gray part) of the Hoggar vortex longitude (bottom axis) as  
 171 a function of the time lag in regard to the crossing of the Greenwich meridian for vortex initiated in the main  
 172 development region (0°E to 10°E; 17.5°N to 27.5°N) in July and August; (dashed) Number of vortex tracks (top  
 173 axis) considered for the different time lags. (b) Corresponding longitudinal distribution of the average absolute  
 174 value of  $\Delta\phi_{min}$  for four hours of the day.  
 175

176 This paper focuses on Atlantic vortices initiated in the Hoggar Initiation Area (HIA: 0°E  
 177 to 10°E; 17.5°N to 27.5°N; green square in Fig. 1b). These vortices are hereafter called  
 178 "Hoggar vortices". For the 38 years, there are 166 Hoggar vortices that represent more than  
 179 60% of the Atlantic vortices of the AEW-N path in July and August. These Hoggar vortices  
 180 have a cyclogenetic efficiency comparable to that of all Atlantic Vortices of the AEW-N path  
 181 (Table 1).

	(NPIA) North Path Vortices		(HIA) Hoggar Vortices	
	Nb. Initiations	Cyclogenetic	Nb. Initiations	Cyclogenetic
June	2.7	1%	1.2	2%
July	<b>4.4</b>	5%	<b>2.4</b>	2%
August	3.4	<b>29%</b>	1.9	<b>29%</b>
September	1.9	15%	0.6	13%
Average	3.1	12%	1.5	12%

182 Table 1: Monthly average number of initiations (over 1980-2017) and average cyclogenetic probability of  
 183 Atlantic vortices of the north path initiated in the NPIA and in the HIA. Seasonal maxima are highlighted in  
 184 bold.



185 Figure 2 represents the average longitude and strength attained by the Hoggar vortices at  
186 different time lags after they cross the Greenwich meridian. Hoggar vortices move westward  
187 with an average speed of about  $7.5 \text{ ms}^{-1}$ . The beginning of their westward progression is  
188 associated with a strengthening of the vortex, as revealed by the increasing absolute value of  
189  $\Delta\phi_{min}$  before the passage on the Greenwich meridian (Fig.2b). Before that, due to the  
190 persistent depression in the lee of the Hoggar, many vortices first detected by the tracking  
191 algorithm in the HIA remain motionless and weak for a variable number of days (Fig.2a). In  
192 the following, only diurnal averages will be considered, but it is worth noting that there is a  
193 diurnal variation of the dynamic and thermodynamic structure of the boundary layer over  
194 West Africa (Parker et al. 2005, Abdou et al. 2010). This is due to the weak turbulent mixing  
195 in the boundary layer during nighttime that decreases the effect of the surface friction and  
196 generates a low-level vertical stratification. This favors the development of nighttime low-  
197 level jets that are maximal at sunrise and that may impact Hoggar vortex intensity. This  
198 diurnal cycle has indeed a consistent signature on the vortex intensity over continental  
199 regions with maximum absolute value of  $\Delta\phi_{min}$  at 0600 GMT and minimum at 1800 GMT  
200 (Fig.2b).

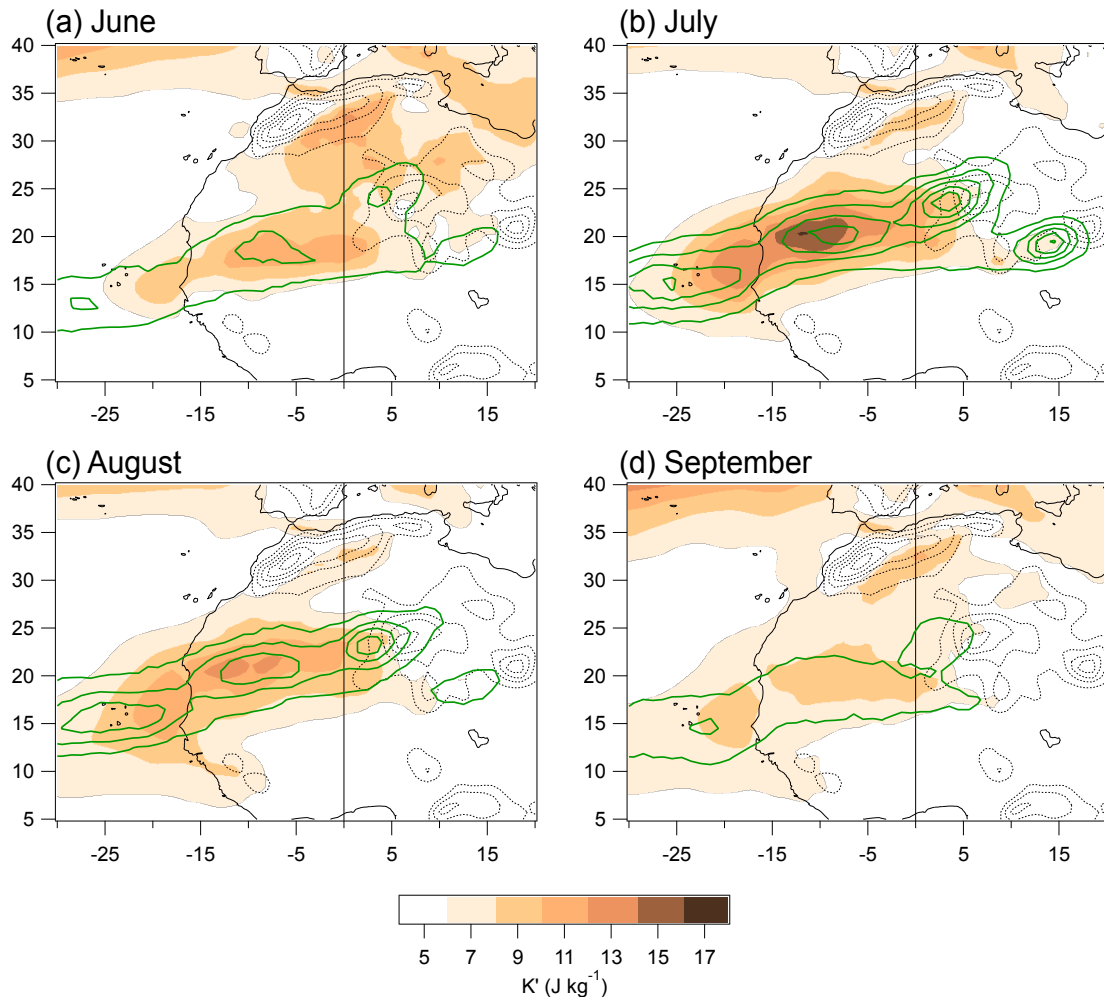
#### 201 *Atlantic Vortex Occurrence and AEW-N amplitude*

202 The AEW-N vortex occurrence, i.e. the number of hours during which a vortex  
203 barycenter is present in a given area, is reported in Figure 3. These maps reveal the most  
204 probable trajectory of the low-level Atlantic vortices of the north path and show the  
205 predominance of vortices initiated near the Hoggar and the Tibesti in July and August. These  
206 vortices propagate southwestward to the Atlantic Ocean roughly along the cyclonically  
207 sheared strip between the Harmattan wind and the monsoon flow. The relation between  
208 Atlantic Vortices and the AEW-N is studied by comparing the vortex occurrence to the  
209 AEW-N activity. This activity may be estimated by computing the perturbation kinetic  
210 energy  $K'$  (in  $\text{J kg}^{-1}$ ) of the horizontal wind at low-level (see e.g., Rydbeck and Maloney  
211 2014, Hamilton et al. 2020, White et al. 2021):

$$215 \quad K' = \frac{1}{2}(u'^2 + v'^2) \quad (1)$$

212 where  $u'$  and  $v'$  are the 2-10-day band-filtered zonal and meridional winds at 850hPa. The  
213 distribution of  $K'$  shows that regions of high AEW amplitude are also region of large vortex  
214 occurrence (Fig.3). In addition, the seasonal variation of  $K'$  is also in phase with that of the

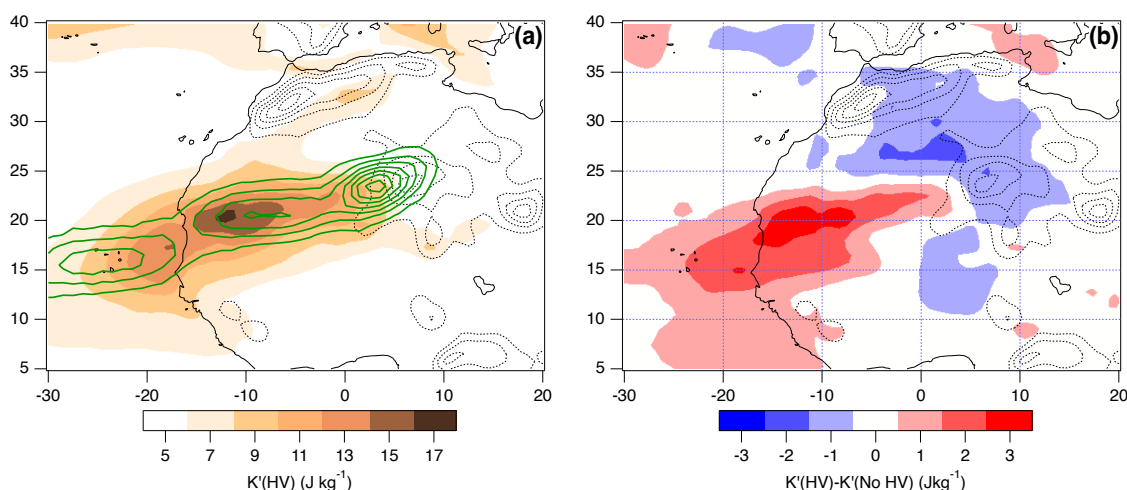
216 vortex occurrence with maximum values in July. The maximum value of  $K'$  in July and  
 217 August (Fig.3b and 3c) is obtained around  $10^\circ\text{W}$ , that is consistent with the longitude of  
 218 maximum vortex intensity reported in figure 2b. Over the Atlantic, the decrease in  $K'$  despite  
 219 the large vortex occurrence is due to the regular weakening of the average vortex amplitude  
 220 as they move westward over the Atlantic (Fig.2b).



221 Figure 3: **(green contours)** Occurrence of AEW-N vortices initiated over the NPIA, expressed as the number of  
 222 hours by month a vortex barycenter is present in a  $1^\circ$  box. For a better legibility, occurrence fields are  
 223 smoothed by a  $3^\circ \times 3^\circ$  running mean. The first contour is 0.5 and the contour increment is 1. **(colors)** Monthly  
 224 average of the perturbation kinetic energy  $K'$  (Eq.1) ( $\text{J kg}^{-1}$ ) of the horizontal wind at 850 hPa in the 2–10-day  
 225 band. **(dotted black contours)** Orography with a contours every 250m from 500m.

227 The specific role of the Hoggar vortices on the AEW-N activity is estimated by  
 228 comparing  $K'$  for periods with (HV) and without (No HV) active Hoggar vortices. Active  
 229 Hoggar vortex periods are defined as all time-steps between  $d_0-1$  day and  $d_0+4$  days,  $d_0$  being  
 230 the ensemble of days at which each of the 166 Hoggar vortices crosses the Greenwich  
 231 meridian. Active Hoggar vortex periods represent about 1/3 of the time for July and August.  
 232 As for the average  $K'$ , the distribution of  $K'$ (HV) for these active periods well corresponds to

233 the specific Hoggar vortex occurrence in July and August (Fig.4a). Compared to  $K'$ (No HV),  
 234  $K'$ (HV) is augmented by about 20% over the most active region around 20°N and 10°W  
 235 (Fig.4b). This confirms that active Hoggar vortex periods correspond to enhanced AEW-N  
 236 activity over West Africa. In contrast,  $K'$ (HV) is slightly reduced between Hoggar and Atlas  
 237 Mountains around 27.5°N, possibly due to the relation between Hoggar vortex initiations and  
 238 relatively stable (and thus giving smaller  $K'$ ) northeasterly winds associated with  
 239 intraseasonal mid-latitudes perturbations (see section 4). Note that adding vortices initiated  
 240 downwind of the Tibesti mountains (those around 20°N-15°E in Fig.1b), about 220 vortices  
 241 cross the Greenwich meridian and  $K'$ (HV) is reinforced by about 30% around 20°N and  
 242 10°W compared to  $K'$ (No HV) (not shown). This highlights the fact that the impact of  
 243 vortices initiated near the orography on the AEW amplitude is broader than the impact of the  
 244 Hoggar vortices alone. Nevertheless, as the conditions of vortex initiation over the Tibesti  
 245 and the Hoggar are slightly different, the following analyses will focus on Hoggar vortices  
 246 only.

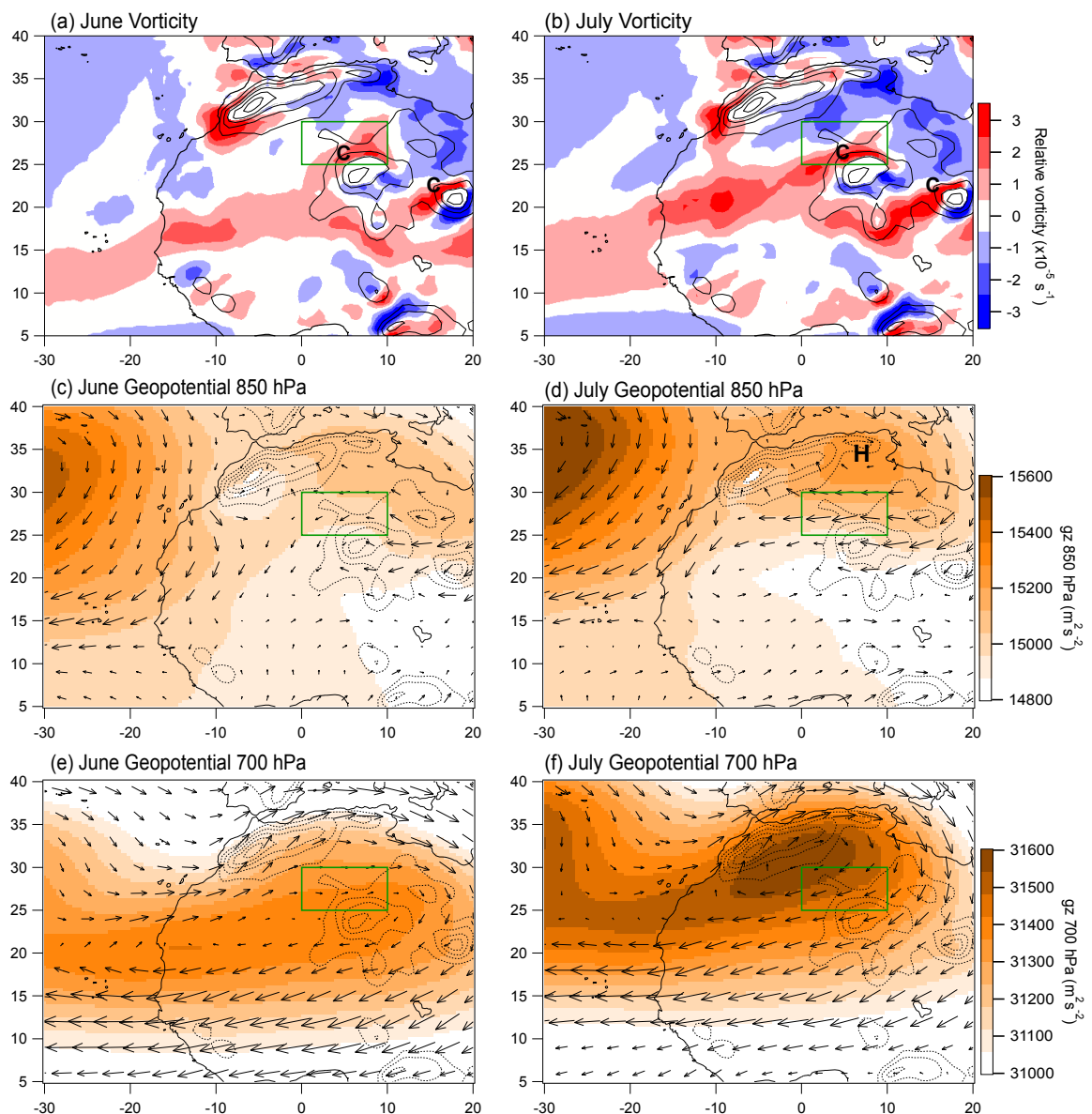


247 Figure 4: (a) (colors) Average perturbation kinetic energy  $K'$  ( $J kg^{-1}$ ) of the horizontal wind at 850 hPa in the 2-  
 248 10-day band in July and August for periods with active Hoggar vortices. (green contours) Occurrence of  
 249 “Hoggar Vortices” initiated in the HIA, expressed as the number of hours a vortex barycenter is present in a 1°  
 250 box. For a better legibility, occurrence fields are smoothed by a 3°x3° running mean. The first contour is 1 and  
 251 the contour increment is 1. (b) (colors) Average  $K'$  difference between periods with (HV) and without (No HV)  
 252 Hoggar vortices. All differences in red or blue are significant at more than 99%. (dotted black contours)  
 253 Orography with a contours every 250m from 500m.  
 254

### 255 *Composite analysis*

256 In section 4, synoptic time-scale perturbations corresponding to Hoggar vortex initiation  
 257 are studied using a composite analysis. The reference days  $d_0$  of the composite are days when  
 258 the barycenter of a Hoggar vortex crosses the Greenwich meridian. This meridian is the  
 259 western boundary of the HIA and is crossed by Hoggar vortices while they strengthen and

260 begin their progression toward the coast (Fig. 2). This is therefore the relevant criteria to  
 261 define the reference day  $d_0$  of the Hoggar vortex composites. Composite fields are computed  
 262 by averaging anomalies over the ensemble of days  $d_0$ . Anomalies of a given ERA-I field for  
 263 each  $d_0$  is computed as a difference between the daily mean field (4 timesteps beginning at 0  
 264 GMT) and a climatological value for this day of the year. This daily climatological value is  
 265 obtained by linear interpolation between two monthly averages attributed to the 15th day of  
 266 each month. The evolution of the average atmospheric state prior and after Hoggar vortex  
 267 initiations is computed following the same procedure between  $d_0-4$  days and  $d_0+4$  days.



268  
 269 Figure 5: Average fields for June and July for: (a and b) Average relative vorticity at 900 hPa; (c and d) Average  
 270 geopotential height and wind at 850hPa; arrows length is  $4^\circ$  for a wind of  $10\text{ms}^{-1}$ ; and (e and f) idem at 700  
 271 hPa. (black dotted contours) Orography with a contours every 250m from 500m. The region used in Figure 6 is  
 272 highlighted in green.

273

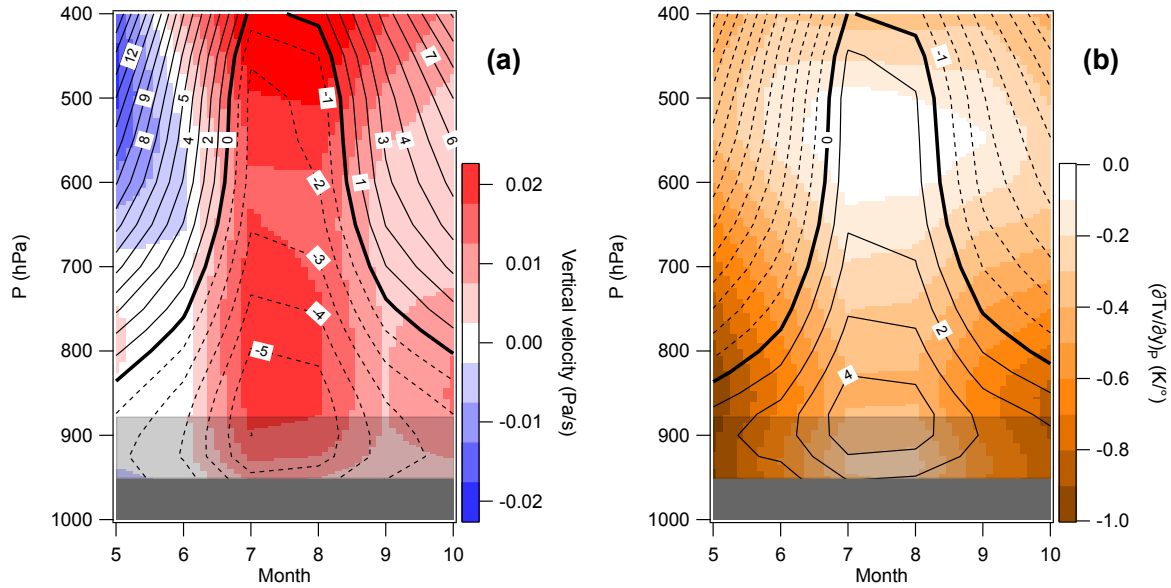
274

### 275 **3. Seasonal evolution of the large-scale environment**

276 The objective of this section is to extract specificities of the large-scale circulation in July and  
277 August and to understand how they can favor vortex initiation in the lee side of the Hoggar  
278 mountains. As shown by Sultan and Janicot (2003), there is a sudden onset of the monsoon at  
279 the end of June due to the migration of Saharan Heat Low (SHL) and to different positive  
280 feedbacks partly linked to orography (Semazzi and Sun 1997). The resulting difference  
281 between June and July in low-level wind and vorticity (Fig. 5) reveals two important features  
282 that potentially have a significant impact on the increased frequency of Hoggar vortices in  
283 July. The first feature is the cyclonic vorticity north of Hoggar and Tibesti mountains (points  
284 C in Fig.5) which may be attributed to increased drag closer to the mountain that generates a  
285 cyclonic shear in the flow having the mountain to its left. This cyclonic vorticity is larger and  
286 extends further west in July. The second feature is the cyclonic strip between the Hoggar and  
287 the Atlantic coast due to the horizontal shear between the Harmattan and the monsoon flow.  
288 This cyclonic strip is stronger and displaced northward in July in good agreement with the  
289 AEW-N activity and vortex occurrence (Fig.3a and 3b). Note that part of the increase in  
290 mean vorticity, particularly near 10°W-20°N, must be due to the higher frequency of the  
291 vortex itself, but the increase in cyclonic shear is also due to the strengthening of the  
292 Harmattan and of the monsoon flow. These two points lead to a continuous and reinforced  
293 cyclonic strip between the leeward side of the Hoggar and the coast in July.

294 The development of this strong and continuous cyclonic strip in July is mostly due to the  
295 reinforcement of the low-level easterly circulation north of 20°N in association with the  
296 reinforced latitudinal pressure gradient at low-level (Fig.5c and 5d). Note that the  
297 reinforcement of the anticyclonic circulation of the Libyan High (Point H in Fig.5d) may be  
298 favored by orographic effect due to the northeastern edge of Atlas Mountains (Fig.5b). At  
299 higher levels, the high pressure generated mostly by the presence of the SHL is centered on  
300 the Hoggar in June, giving westerly winds to the northern side of the Hoggar, and reinforced  
301 and centered south of the Atlas in July, giving strong easterly winds to the northern side of  
302 the Hoggar (Fig.5e and 5f). The seasonal evolution of the zonal wind profiles north of the

303 Hoggar (0°E-10°E; 25°N-30°N; green rectangle in Figure 5) between May and October  
 304 shows a clear reinforcement and a vertical extension of easterly winds from the surface up to



305  
 306 Figure 6: Evolution of monthly mean profiles for the region (0°E to 10°E; 25°N to 30°N; green region in Figure 5)  
 307 between May and October for: (a) the zonal wind (contours, ms<sup>-1</sup>) and vertical motion (colors in Pa s<sup>-1</sup>,  
 308 positive values for downward motion); (b) latitudinal gradient of the geopotential height (contours, m/°) and  
 309 latitudinal gradient of the virtual temperature (colors in K/°). Maximum and average orography for the region  
 310 is shown in light and dark grey respectively.

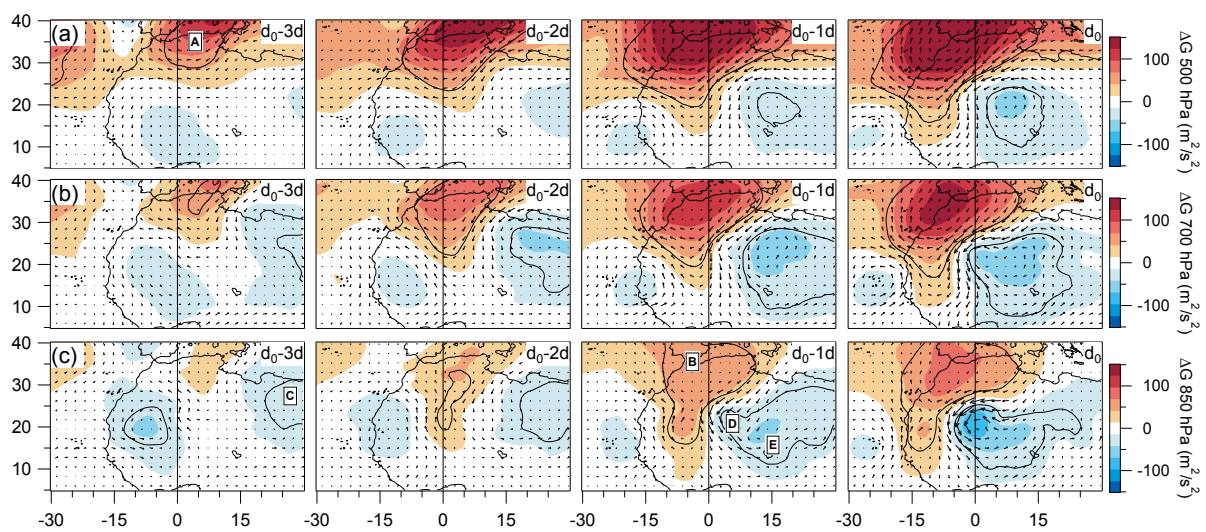
311 400 hPa in July and August, associated with reinforced subsidence (Fig.6a). This can be  
 312 mostly attributed to the northward migration of the SHL that places the area north of the  
 313 Hoggar under the southeastern descending branch of the Libyan High. Figure 6b shows that  
 314 this can be also understood in term of local thermal wind structure for which the zonal  
 315 geostrophic wind  $U_g$  on an isobar  $P_n$  is given by:

$$325 \quad U_g(P_n) = -\frac{1}{f} \left( \frac{\partial \Phi(P_0)}{\partial y} \right)_P - \frac{R}{f} \sum_{i=1}^n \left( \frac{\partial \overline{T_{vi}}}{\partial y} \right)_P \ln(P_{i-1}/P_i) \quad (2)$$

316 where  $f$  is the Coriolis parameter,  $\phi(P_n)$  the geopotential height of the isobar  $P_n$ ,  $\overline{T_{vi}}$  is the  
 317 average virtual temperature between the two pressure levels  $P_i$  and  $P_{i-1}$ , and  $R$  the gas  
 318 constant for dry air. Figure 6b is computed using monthly means  $T_v$  and  $\phi(P_0)$  taken in ERA-  
 319 I. The stronger easterly wind near the surface in July is due to larger low-level latitudinal  
 320 pressure gradient (first RHS term of Eq.2; contours in Fig.6b) between the Hoggar and the  
 321 Mediterranean Sea (Fig.5c and d) that mostly results from the northward migration of the  
 322 SHL, as described in Sultan and Janicot (2003). Second, this northward migration of the SHL  
 323 reduces the magnitude of the latitudinal gradient of the virtual temperature (second RHS term  
 324 of Eq.2; colors in Fig.6b) in July and August and maintains these easterlies up to 400 hPa.

326 Before July and after August, when the SHL is shifted southward, the larger absolute value of  
 327 this gradient causes the easterlies to weaken and turn west rapidly with height.

328 In summary, this section shows that the stronger easterly winds over the Hoggar in July  
 329 reinforce cyclonic vorticity north and west of the orography and could be at the origin of  
 330 more frequent Hoggar vortex initiation during these months. In addition, the northward  
 331 displacement and the strengthening of the cyclonic strip between the Hoggar and the coast in  
 332 July may promote the development and maintenance of the vortex disturbance through  
 333 barotropic and baroclinic processes, as shown by previous analyses of the energy source of  
 334 AEW-N (see e.g., Norquist et al. 1977, Diedhiou et al. 2002, and section 5).

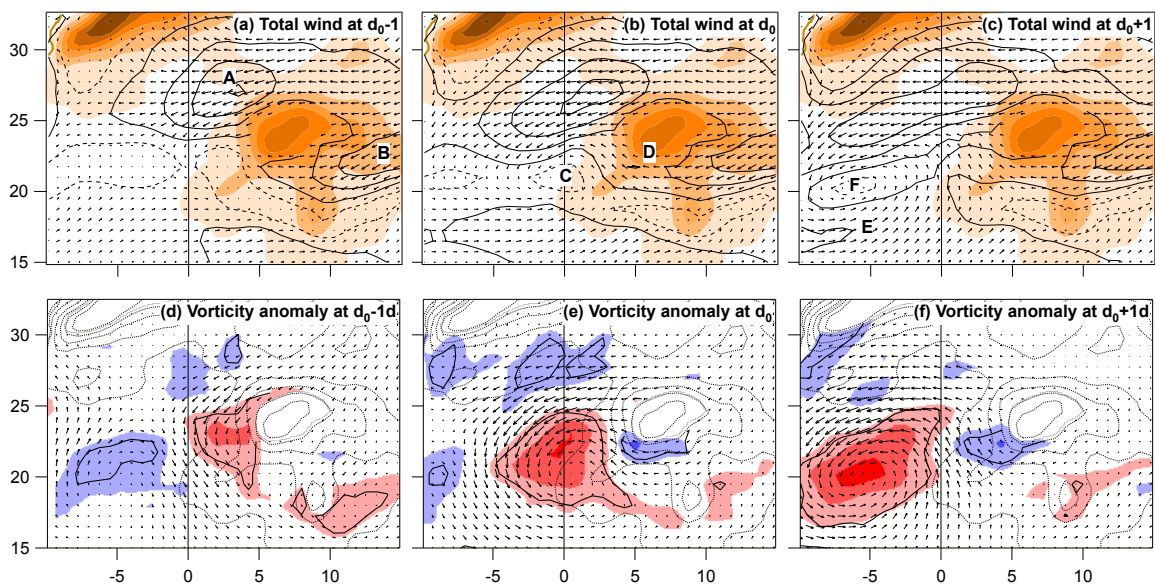


335  
 336 Figure 7: Composite of the horizontal wind anomaly (arrows, length of  $5^\circ$  for  $5\text{ms}^{-1}$ ) and of the geopotential  
 337 height anomaly at (a) 500 hPa, (b) 700 hPa and (c) 850 hPa for the period  $[d_0 - 3 \text{ days}, d_0]$ . The day  $d_0$  is the day  
 338 when the barycenter of the Hoggar vortex at 850 hPa crosses the Greenwich meridian. The black contour  
 339 delineates regions for which the composite geopotential perturbation is significant at the 99% level.

#### 340 4. Hoggar vortex genesis

341 The objective of this section is to determine large-scale and local conditions leading to  
 342 Hoggar vortex initiations. To this end, composites of the 166 Hoggar vortices are computed  
 343 for the dates of initiation  $d_0$  and for each of the three days before and after  $d_0$  (see section 2).  
 344 Three days before Hoggar vortex initiation, there is a high-pressure anomaly significant at the  
 345 99% level above 700 hPa over a region centered near the Strait of Gibraltar (point A in  
 346 Fig.7a). This anomaly amplifies, extends downward and spread horizontally, giving  
 347 northeasterly winds blowing at 850 hPa over the Hoggar one day before  $d_0$ . A weak but  
 348 statistically significant depression anomaly is also initiated at low-levels three days before  $d_0$   
 349 near the Mediterranean Sea around  $30^\circ\text{N}$  and  $30^\circ\text{E}$  (point C in Fig.7c). The following days,  
 350 this “easterly low” moves westward, extends up to 500 hPa, strengthens and contributes to

351 increase the northeasterly flow over the Hoggar. At  $d_0-1d$ , a secondary minimum in the 850  
 352 hPa geopotential height anomaly appears in the lee side of the Hoggar (point D in Fig.7c)  
 353 more than 1000 km west of the “easterly low” which is still centered at  $20^\circ N-15^\circ W$  and  
 354 shifted southward compared to previous days (point E in Fig.7c). The secondary minimum  
 355 may be considered as the initiation of the Hoggar vortex which then strongly intensifies  
 356 between  $d_0-1d$  and  $d_0$  in association with a horizontal expansion and a strengthening of the  
 357 “Gibraltar high”. At  $d_0$ , the vortex is centered on the Greenwich meridian and therefore at  $10^\circ$   
 358 west of the “easterly low” at 500hPa which continues its slow westward progression.



359  
 360 Figure 8: (a, b, c) Composite of the total wind field at 900 hPa at  $d_0-1$  day,  $d_0$ , and  $d_0+1$  day. Arrows length is  $1^\circ$   
 361 for  $10ms^{-1}$  and black contours are the wind strength (dotted line for  $2ms^{-1}$  and increment of  $2ms^{-1}$ ). (d, e, f)  
 362 Composites of relative vorticity (colors as for Fig.5, black contour for statistical significance at 99%) and wind  
 363 (arrows length is  $1^\circ$  for  $5ms^{-1}$ ) anomalies at 900 hPa. The orography with a height step of 200m is represented  
 364 by colors for a, b and c and by dotted contours for d, e, and f.

365 Figure 8 highlights the low-level wind and vorticity evolution near the Hoggar at higher  
 366 spatial resolution during vortex initiation. The wind speed is stronger northwest of the  
 367 Hoggar, over the Tademait Plateau (point A in Fig.8a), and between the Hoggar and the  
 368 Tibesti (point B). However, only the wind west of the Hoggar is perturbed during these 3  
 369 days, showing the preponderance of dynamical processes downwind of the Hoggar on vortex  
 370 initiation. One day before initiation, a significant relative vorticity anomaly extends the  
 371 average vorticity strip (Fig.5b) in the lee of the Hoggar (Fig.8d). The vortex is then  
 372 asymmetric with maximum vorticity and wind anomalies north of the vortex center (point C  
 373 in Fig.8b). The northeast side of the vortex is alimented by an easterly flow deviated around  
 374 the southern edge of the Hoggar mountain (point D). The next day, the center of the vortex



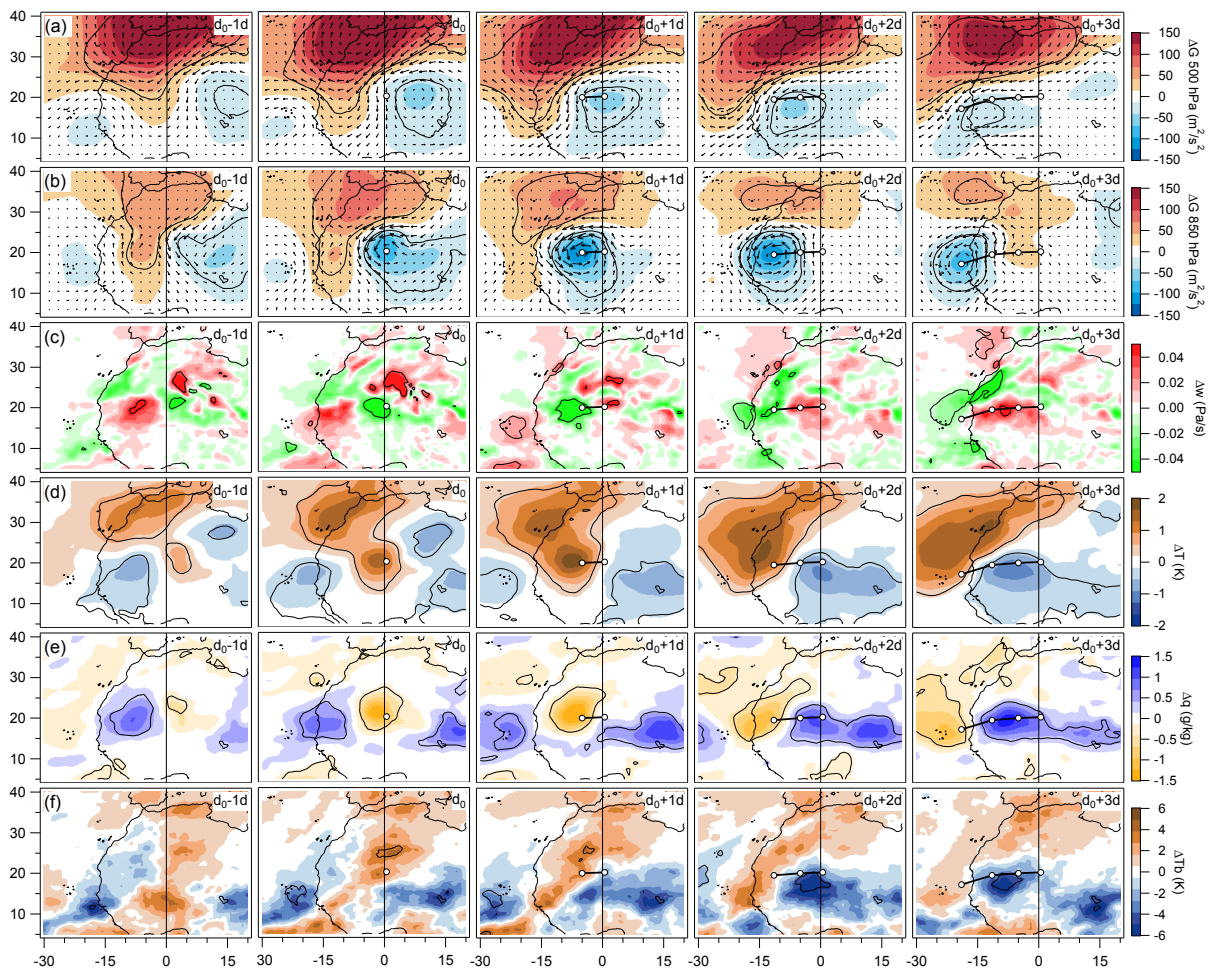
375 (point F in Fig. 8c) is located about  $6^\circ$  further west, giving a speed of about  $7.7 \text{ ms}^{-1}$  that  
376 corresponds to the north-east wind speed to its north. The southwesterly monsoon flow (point  
377 E in Fig.8c) is reinforced and recurves to form the eastern side of the vortex which is then  
378 more axisymmetric.

379 The dynamical perturbation due to the mountain is complex and involves different  
380 processes that may lead to the formation of vortices downwind. As discussed in Mozer and  
381 Zehnder (1996a and b) (hereinafter MZa and MZb), the conservation of the potential vorticity  
382 for a dry flow passing south of an isolated mountain (without column depth change) may  
383 result in a low-level jet that is barotropically unstable, leading to the production of synoptic  
384 vortices that separate from the mountains and move downstream. However, the jet south of  
385 the Hoggar (point D) is weak here and the Hoggar vortex initiation seems more in agreement  
386 with the vortex generated during the transient period at the beginning of the MZ simulations.  
387 This transient period is associated with the formation of a so-called starting vortex attributed  
388 to column stretching as the air initially at the top of the mountain is forced downstream  
389 (Huppert and Brian 1976). In MZ simulations, the transient period results from the  
390 instantaneous incorporation of orography in the flow, but it could result here from the rapid  
391 intensification of the easterly flow over the mountain. A starting vortex indeed also appears  
392 for wind modulations due to planetary Rossby waves in an easterly flow over orography  
393 (Zehnder 1991). The evolution shown in figures 7 and 8 resembles the Zehnder results with a  
394 cyclonic vortex forming in the lee of the mountain while a wave trough is still quite far to the  
395 east, displaced southward (as point E in Fig.7c) and reinforced because of the conservation of  
396 the total vorticity. The fact that the vortex appears while the easterly flow is reinforced by the  
397 “Gibraltar high” and the approaching trough suggests that Hoggar vortex initiation process  
398 could be understood more as a starting vortex rather than a vortex due to barotropic  
399 instability of the jet south of the Hoggar. Nevertheless, such barotropic and baroclinic  
400 instabilities due to the strong cyclonic shear existing between the Greenwich meridian and  
401 the coast certainly plays a role in intensifying and maintaining the vortex, this is analyzed in  
402 the next section.

## 403 **5. Hoggar vortex evolution**

404 After its initiation, the vortex moves westward along  $20^\circ\text{N}$  and reaches a maximum  
405 strength around  $10^\circ\text{W}$ . The “Gibraltar high” remains quite stationary between  $d_0-1d$  and  
406  $d_0+3d$  and provides strong easterly winds on the north side of the vortex during its genesis

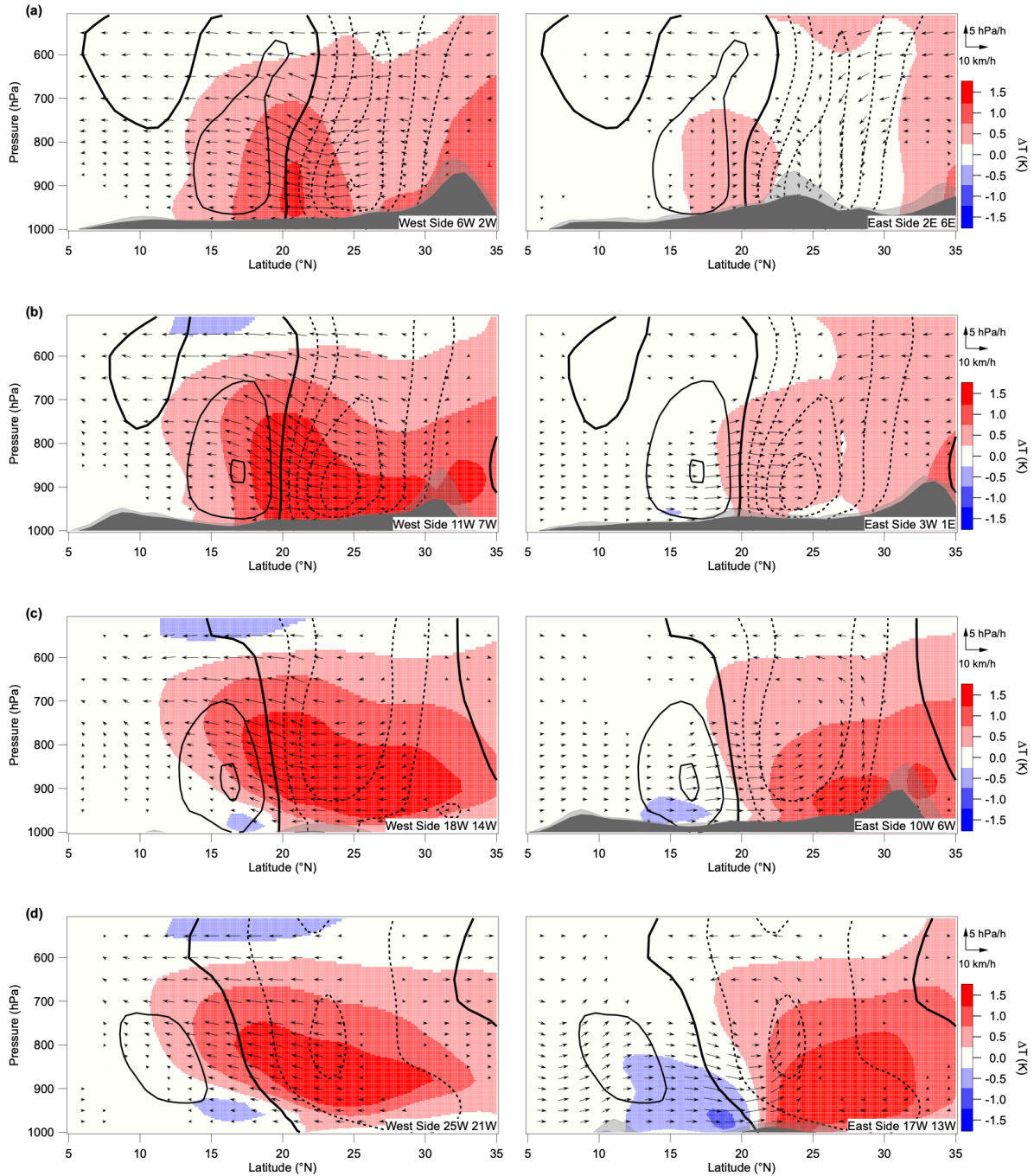
407 and during its progression toward the coast (Fig.9a and 9b). Between  $d_0-1d$  and  $d_0+1d$ , these  
 408 easterly winds are associated with a significant subsidence anomaly (Fig.9c) above the  
 409 Hoggar associated with warm (Fig.9d) and dry (Fig.9a) anomalies in the vortex. These dry  
 410 processes during the vortex genesis are consistent with the band of positive  $T_b$  anomaly  
 411 measured from space east of the Greenwich meridian at  $d_0-1d$  (Fig. 9f) and associated with a  
 412 northerly wind anomaly. The positive  $T_b$  anomaly is large and significant near  $25^\circ N$  at  $d_0$  and  
 413 consistent with subsiding warm and dry air north and west of the vortex center. During the  
 414 vortex progression over the continent, warm and dry Saharan air is advected southward in the  
 415 west side of the vortex. The resulting warm anomaly near the center of the vortex (Fig. 9d)  
 416 gives a low-level warm-core structure that decreases the vortex cyclonic circulation above  
 417 and confines the vortex circulation at low-levels.



418  
 419 Figure 9: As in figure 7, but for the period [ $d_0-1$ ,  $d_0+3$  days] for the anomaly of different dynamical and  
 420 thermodynamical parameters from top to bottom: (a) geopotential height  $\Delta G$  at 500hPa and (b) 850 hPa, (c)  
 421 vertical velocity  $\Delta \omega$  (positive downward), (d) temperature  $\Delta T$  and (e) specific humidity  $\Delta q$  at 850 hPa, and (f)  
 422 infrared window brightness temperature  $\Delta T_b$  measured by the geosynchronous satellite Meteosat. The white  
 423 markers represent the evolution of the position of the dynamical center of the 850 hPa vortex from  $d_0$ .

424 On the other hand, the warm anomaly over the Atlantic Ocean West of Morocco tends to  
425 maintain the anticyclonic geostrophic circulation around the “Gibraltar high” at 500 hPa (Fig.  
426 9a). When the vortex reaches the coast, there is a large band of warm and dry anomaly over  
427 the ocean and a band of cold and moist anomaly over the Sahelian zone between 15°N and  
428 20°N. This evolution of the low-level temperature and humidity (Fig. 9d and 9e) is consistent  
429 with the evolution of the observed anomaly of infrared brightness temperature (Fig. 9f). In  
430 particular, negative  $T_b$  anomalies, which correspond to enhanced mid and high cloud cover  
431 computed using thresholds at 230K and 210K (not shown) are associated with colder and  
432 moister air temperature at low-levels, especially at  $d_0+2d$  and  $d_0+3d$ . At this time, the high  
433 cloud cover north of 15°N is maximal in the east side of the vortex that is equivalent to the  
434 southerly wind sector of an AEW-N. This is in agreement with previous studies (see e.g.,  
435 Duvel 1990, Gu et al. 2004, Kiladis et al. 2006) showing maximum convection in the  
436 southerly wind sector of the wave north of 15°N. This is in contrast with the maximum  
437 convection and mesoscale convective systems found in the wave trough around 10°N for  
438 AEW-S between the Greenwich meridian and the coast (see e.g., Kiladis et al. 2006, Núñez  
439 Ocasio et al. 2020). This moist anomaly over Sahelian regions is probably similar to the  
440 moisture surges discussed in detail in Couvreux et al. (2009) for June 2006 and to the  
441 northward burst of the West African monsoon studied in Cuesta et al. (2009) for the end of  
442 July 2006. On the opposite, the west side of the vortex with positive  $T_b$  anomalies indicates a  
443 region of suppressed convection ahead of the vortex due to dry and warm northerlies. As  
444 stated in section 2, some of these dry vortices can lead to cyclogenesis, either near the coast  
445 or later over the Atlantic (e.g. Chen et al. 2008, Chen and Liu 2014, Duvel 2021). They are  
446 however poorly cyclogenetic compared with vortices of the AEW-S partly because they tend  
447 to occur before the heart of the hurricane season, but also, as stated by Hopsch et al. (2010),  
448 because warm and dry conditions west of the AEW trough, caused by advection of Saharan  
449 air, inhibit the development of deep convection and further deepening of the wave trough.

450 Figure 10 shows the evolution of the anomalies of the three-dimensional dynamical  
451 structure of the vortex as it moves toward the coast. At  $d_0$ , the vortex is strongly asymmetric  
452 with an easterly wind anomaly of  $4 \text{ m s}^{-1}$  at 25°N and a westerly wind anomaly of only  
453  $1.5 \text{ m s}^{-1}$  at 17°N (Fig. 11a). The southerly wind perturbation on its east side is weak and  
454 vanishes above 800 hPa. This strong asymmetry of the wind anomaly is consistent with the  
455 transient period discussed above and suggests that the main driving force of the vortex  
456 formation at this early stage is the acceleration of the northeasterlies near the surface



457  
 458 Figure 10: July and August latitudinal distribution of the anomaly of the temperature profiles (colors) and  
 459 meridional and vertical wind (arrows) averaged over the west side (left) and the east side (right) of the  
 460 composite vortex for (a)  $d_0$ , (b)  $d_0+1$  day, (c)  $d_0+2$  days, and (d)  $d_0+3$  days. The longitudes  
 461 defining the west and east sides are indicated on the lower right corner of each panel. The corresponding  
 462 zonal wind anomalies (contours,  $\Delta u = \pm 1 \text{ ms}^{-1}$  starting from 0 with bold contour) is averaged for the central part  
 463 of the vortex located between the west and east sides. The average (dark grey) and the maximum (light grey)  
 464 orography for the west and the east side of the vortex are also displayed.

465 northwest of the Hoggar. The associated perturbation of northeasterly and subsiding winds up  
 466 to 500 hPa is consistent with a reinforcement of the Libyan high circulation associated with  
 467 the “Gibraltar high” anomaly. A warming of more than 1K extends roughly between the

468 surface and 700 hPa on the west side of the vortex (Fig.10a) and certainly contributes to the  
469 deepening of the vortex depression for the following day.

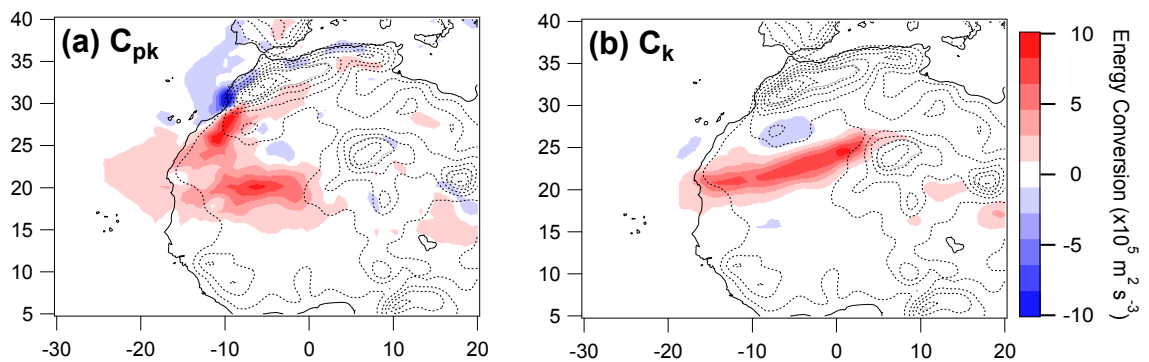
470 At  $d_0+1$  day around  $5^\circ\text{W}$  (Fig.10b), the vortex is more axisymmetric and stronger with larger  
471 warming on the west side for air rising over the monsoon flow. On the east side, a colder and  
472 moister southerly wind perturbation penetrates further north. At  $d_0+2$  days around  $12^\circ\text{W}$   
473 (Fig.10c), the vortex is shifted upward with maximum easterlies around 800 hPa on the north  
474 side. On the west side, the warming is maximal around 800 hPa at  $15^\circ\text{N}$  and located above a  
475 cold anomaly at the surface. Both the northerly wind uplift on the west side and the southerly  
476 wind uplift on the east side (Fig.9c and 10c) participate in the vortex uplift. This tendency is  
477 reinforced at  $d_0+3$  days (Fig.10d) with wind perturbations maximal around 800 hPa at  $19^\circ\text{W}$ .  
478 This vortex ascent during its travel between the Hoggar and the coast may be attributed to the  
479 lift of the Saharan air above the monsoon flow on its west side and to the lift possibly related  
480 to orography on its east side after  $d_0+2$  days. This is consistent with the lifting process  
481 analyzed by Drame et al. (2011) for a Saharan Air Layer (SAL) episode that occurred in July  
482 2010 in association with a westward moving thermal low that is indeed one of the 166  
483 Hoggar vortices considered here.

484 An exhaustive computation of the energy budget of the vortices is outside the scope of  
485 this study. However, two important kinetic energy conversion parameters may be estimated  
486 on the basis of the composite perturbations in order to estimate the consistency with the  
487 energy budget of AEW-N at low-level. Previous results on AEWs (e.g., Norquist et al. 1977,  
488 Lau and Lau 1992, Diedhiou et al. 2002, Alaka and Maloney 2014, Hamilton et al. 2020,  
489 White et al. 2021) found large positive values of the baroclinic overturning term and of the  
490 barotropic conversion term at low levels around  $20^\circ\text{N}$  and west of the Hoggar. These terms  
491 represent respectively the conversion of eddy available potential energy to eddy kinetic  
492 energy ( $C_{pk}$ ) and the transfer of mean to eddy kinetic energy ( $C_k$ ) and are defined as:

493 
$$C_{pk} = -\frac{R}{P} \overline{\omega' T'} ; \quad C_k = -\overline{[\mathbf{V}'_H \cdot (\mathbf{V}' \cdot \nabla) \mathbf{V}'_H]}$$

494 where  $\mathbf{V}$  is the three-dimensional wind,  $\mathbf{V}_H$  is horizontal wind ( $\text{m s}^{-1}$ ),  $\omega$  is pressure velocity  
495 ( $\text{Pa s}^{-1}$ ),  $T$  is air temperature (K). The prime is used to represents composite anomalies (i.e.,  
496 eddy perturbations) and the overbar represents an average between  $d_0-4$  days and  $d_0+3$  days.  
497 As expected, the warm northerly wind anomaly rising on the west side of the vortex and the  
498 cold southerly wind anomaly subsiding on the east side of the vortex (Fig.9c and 9d) give a

499 large positive  $C_{pk}$  at low-levels around  $20^{\circ}\text{N}$  between the Hoggar and the coast (Fig.11a) in  
 500 agreement with previous results on AEW-N. There is also large  $C_{pk}$  south of the Atlas  
 501 Mountains, due to rising warm air in the north side of the vortex (Fig.10c and 10d). However,  
 502 the nearly null  $C_{pk}$  near the Hoggar shows that baroclinic energy conversion plays no role in  
 503 vortex genesis. Figure 11b shows large positive barotropic conversion  $C_k$  due to the large  
 504 average cyclonic shear and the asymmetry of the vortex dominated by the northeasterly  
 505 winds. This suggests that the kinetic energy of the vortex during its genesis and for its  
 506 progression comes mainly from the mean northeasterly winds accelerated by the "Gibraltar  
 507 High". This large  $C_k$  at low-level is also in agreement in location and amplitude with previous  
 508 studies on AEW-N.



509  
 510 Figure 11: (a) baroclinic overturning ( $C_{pk}$ ) and (b) barotropic energy conversion ( $C_k$ ) between  $d_0-4$  days and  
 511  $d_0+3$  days at 900hPa.

## 512 6. Summary and discussion

513 Most of the low-level synoptic vortices formed over West Africa and propagating to the  
 514 Atlantic Ocean (i.e., Atlantic vortices) on the AEW-N track are initiated downwind of the  
 515 Hoggar Mountains in July and August. The main specificity in the local circulation for these  
 516 months compared to June and September is the reinforcement of the low-level easterly winds  
 517 over the Hoggar and the vertical expansion of subsiding easterly winds up to 400 hPa. This  
 518 vertical expansion favors the development of a low-level easterly jet that may be hampered in  
 519 June and September due to westerly winds above 800 hPa. The reinforcement and the larger  
 520 vertical expansion of these easterlies in July and August are associated with the northward  
 521 migration of the SHL. At the synoptic time-scales, the vortex initiation is associated with an  
 522 additional strengthening of these easterlies over the Hoggar. This extra strengthening is  
 523 associated with a high-pressure anomaly that develops first at mid-level around the Strait of  
 524 Gibraltar three days before Hoggar vortex initiations and then extends downward. Locally,

525 this "Gibraltar high" anomaly corresponds to an amplification of the Libyan anticyclone  
526 which is also a characteristic of July. Hoggar vortex initiations are also statistically associated  
527 with a depression coming from the east and located at around 30°E at  $d_0-3$  days. This  
528 "easterly low" appears first at low-level and then strengthens and extends up to 500 hPa  
529 before Hoggar vortex initiation. The precise origin of the "Gibraltar high" anomaly is  
530 probably multifactorial and deserves further studies. It could be related in particular to the  
531 SHL intraseasonal variability mode studied by Chauvin et al. (2010). This mode is linked to  
532 polar and subtropical jet fluctuations over the North Atlantic with a characteristic time scale  
533 of about 15 days and has some resemblance with the persistent mid-level wind anomaly  
534 around the "Gibraltar high".

535 The composite analysis thus objectively reveals that Hoggar vortices are statistically  
536 associated with the evolution of two features, the "Gibraltar high" and an "easterly low", both  
537 of which appear more than three days before vortex initiation. However, this composite  
538 initiation scenario being statistical, Hoggar vortices are certainly developing with various  
539 combination of these two features that are basically disconnected. The "Gibraltar high"  
540 indeed evolves over a much longer time-scale, as shown by its persistence in Figures 7 and 9,  
541 compared to the more frequent and faster "easterly low". The wave pattern evident in Figure  
542 7, especially in Figure 7c at  $d_0-2$  days, suggests that the "easterly low" statistically  
543 corresponds to the eastern trough of an initially weak easterly wave. The "Gibraltar high"  
544 clearly extends southward and downward as the ridge of this wave crosses the Greenwich  
545 meridian. Statistically, the Hoggar vortex initiation correspond therefore to an intensification  
546 of this easterly wave by the formation of an orographic vortex while transient northeasterly  
547 winds blow over the Hoggar between this reinforced ridge and the eastern trough. As  
548 suggested in figure 11b, the source of intensification of this vortex could be mostly kinetic  
549 energy transfer from the persistent northeasterly flow provided by the southward and  
550 downward extension of the "Gibraltar high" perturbation.

551 An important point is that both features lead to reinforced northeasterly wind over the  
552 Hoggar before vortex initiation southwest of the Hoggar about 1000 km east of the center of  
553 the "easterly low". The flow pattern around the Hoggar near initiation time shows some  
554 analogy with the transient period in the simulations analyzed in MZa, MZb and Zehnder  
555 (1991) which leads to a so-called starting vortex attributed to column stretching as the air  
556 initially at the top of the mountain is forced downstream. For the observed Hoggar vortices,

557 the transient character could result from the rapid intensification of the easterly flow due to  
558 the Gibraltar high development. One day before vortex initiation, the reinforcement of the  
559 northeasterly flow northwest of the Hoggar leads to a cyclonic vorticity anomaly in the lee of  
560 the mountain. The vortex then amplifies asymmetrically with a reinforcement of the wind and  
561 of the cyclonic vorticity north of its center and becomes afterward more symmetric with an  
562 amplification of the monsoon flow on its south side. The present analysis is concerned mostly  
563 with the origin of the vortices and their impact on the AEW-N amplitude. As might be  
564 expected, the vortex characteristics after their initiation resemble those of the AEW-N  
565 reported in the literature since the pioneering work of Carlson (1969) and Burpee (1972).  
566 Among the 166 Hoggar vortices, there are about 20% which are following a previous one and  
567 forming therefore a sort of wave packet of larger amplitude. In the composite Hoggar vortices  
568 shown in Figures 7 and 9, the trough which forms statistically near 30°E at d<sub>0</sub>-3d is located at  
569 approximately 35° east of the previous vortex (Fig.7c) and takes about 4-5 days to reach the  
570 position of this previous vortex, which is within the typically observed wavelength and period  
571 of AEWs. The maintenance of this vortex up to the coast may be attributed to low-level  
572 barotropic and baroclinic energy conversions resulting mostly from the strong cyclonic shear  
573 between the northeasterlies and the monsoon flow, in agreement with previous studies on  
574 AEW-N.

575 The two paths of the AEWs are well known, but they are often considered as the  
576 expression of the same phenomenon having its origin in the instability of the AEJ. For  
577 example, in Kiladis et al. (2006) and Hall et al. (2006), differences in the nature of AEWs are  
578 mostly attributed to differences in the basic-state AEJ depending in particular on the season.  
579 Hall et al. (2006) also highlight the fact that the modal growth in a dry model is not sufficient  
580 to account for the presence of AEWs and that a triggering of the wave is necessary. This  
581 triggering is generally attributed to convective warming in the heart or at the root of the AEJ  
582 (see e.g., Thorncroft et al. 2008). It is interesting to note that Thorncroft et al (2008) found  
583 that the maximum triggering efficiency is obtained for a shallow convective warming at  
584 20°N-15°E, that is the statistical position of approaching trough one day before initiation  
585 (Fig.7c). While the model used in Thorncroft et al. (2008) has no orography, it still has the  
586 temperature structure of the SHL and the associated large-scale barotropic and baroclinic  
587 instabilities. As shown in previous studies (see e.g., Grogan et al. 2016, Nathan et al. 2017),  
588 this triggering could also be due to the warming resulting from the radiative forcing of  
589 Saharan mineral dust. The role of mid-latitudes in triggering AEWs has also been highlighted



590 in Leroux et al. (2011) who showed using an idealized model that AEW packets can be  
591 associated with a slow eastward moving high pressure over the North Atlantic that presents  
592 similitudes with the “Gibraltar high” and with the perturbations at the origin of the SHL  
593 variability in Chauvin et al. (2010).

594 The results presented above offer another possibility for the triggering or intensification  
595 of AEW-N by invoking the impact of orographic disturbances caused by enhanced easterly  
596 winds over the Hoggar. This assumption does not contradict that of White et al (2021) who  
597 attribute the marked decrease in AEW-N energy in a model where the Hoggar and Tibesti  
598 mountains have been removed to the reduction in baroclinic energy conversion due to  
599 reduced vertical wind shear. This weaker vertical shear results from enhanced low-level  
600 easterlies to the west of the Hoggar and to a weaker AEJ due to reduced meridional surface  
601 temperature gradient (see also Hamilton et al. 2017). In fact, both processes may explain the  
602 high sensitivity of the AEW-N amplitude to the removing of the Hoggar and Tibesti  
603 orography in the White et al. (2021) sensitivity test. The “vertical shear hypothesis” considers  
604 that reinforced easterlies west of the flattened Hoggar region inhibit AEW-N, while the  
605 "orographic perturbation hypothesis” considers that reinforced easterlies over the orography  
606 is a source of intensification of AEW-N. To test more specifically the "orographic  
607 perturbation hypothesis” proposed here, additional sensitivity tests could be performed by  
608 varying the intensity and the vertical profile of the wind over the Hoggar orography, for  
609 example by imposing different latitudinal positions of the SHL.

610

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### 617 *Data Availability Statement*

618 ERA-Interim data used in this study are openly available at <https://www.ecmwf.int/>.

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