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Identification of processes that control the stable isotope composition of rainwater in the humid tropical West-Central Africa

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Abstract: This study interprets 11 years (2006 to 2016) and 6 months (March to August in 2017) of respectively monthly and daily isotopic (δD and $\delta^{18}O$) monitoring of rain at Douala (Cameroon), a humid tropical station in Western Africa. The main scope is to analyze the climate controls on precipitation isotopes at different timescales. Firstly, we examine the annual cycles of $\delta^{18}O$. Over the 11 years of survey, the annual cycle exhibits a W shape that is quite reproducible from year to year, with two minima in spring and autumn periods. Based on back trajectory calculations and remote sensing observations of water vapor isotopic composition, we show that the observed depletion in spring and autumn is due to strong convective activity along air mass trajectories. The same effect of convective activity can be observed at the daily timescale. At seasonal and daily time scales, the isotopic composition is also strongly tied to the convective organization and cloud types. More depleted precipitation is associated with larger areas of high clouds. Very low to low clouds are observed in July-August, mid-level to high clouds are dominant in June and high to very high clouds characterize March-April-May, thus explaining the enriched (depleted) values in summer (spring). Finally, this paper highlights the importance of large scale meteorological conditions controls on precipitation stable isotope composition in the Gulf of Guinea.

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Abstract

This study interprets 11 years (2006 to 2016) and 6 months (March to August in 2017) of respectively monthly and daily isotopic (δD and $\delta^{18}O$) monitoring of rain at Douala (Cameroon), a humid tropical station in Western Africa. The main scope is to analyze the climate controls on precipitation isotopes at different timescales. Firstly, we examine the annual cycles of $\delta^{18}O$. Over the 11 years of survey, the annual cycle exhibits a W shape that is quite reproducible from year to year, with two minima in spring and autumn periods. Based on back trajectory calculations and remote sensing observations of water vapor isotopic composition, we show that the observed depletion in spring and autumn is due to strong convective activity along air mass trajectories. The same effect of convective activity can be observed at the daily timescale. At seasonal and daily time scales, the isotopic composition is also strongly tied to the convective organization and cloud types. More depleted precipitation is associated with larger areas of high clouds. Very low to low clouds are observed in July-August, mid-level to high clouds are dominant in June and high to very high clouds characterize March-April-May, thus explaining the enriched (depleted) values in summer (spring). Finally, this paper highlights the importance of large scale meteorological conditions controls on precipitation stable isotope composition in the Gulf of Guinea.

Keywords: $\delta^{18}O$, water vapor, convective activity, GPCP precipitation, air back trajectory.

Declaration of interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

Response to Revision Request

The authors would like to thank all the reviewers for their comments and suggestions for improvements which helped us to propose a new corrected version of the manuscript.

Here below is a point to point answer to comments.

Reviewer #2

1- I have some reservation about the explanation on the intra-seasonal time scale. For example, the authors have used the TES derived hydrogen vapor isotope data and compared their result of precipitation $\delta^2\text{H}$ variability. Though the absolute values of these two different datasets are not comparable as correctly mentioned by the authors, however a qualitative match is expected. According to Fig. 7 the seasonal characters are similar, but a considerable mismatch is observed during the later period of the year. For example, from Jan to Sep the precipitation isotope maintains a more or less constant positive offset relative to the vapor isotopic record, which is expected. But in the month of Oct the offset nearly vanishes and surprisingly in the month of Nov the offset value turns negative! This behavior defies the principle of isotopic fractionation. It may so happen that within the limits of analytical uncertainties, this negative offset is not significantly different from zero. But this aspect needs to be discussed in detail, especially why the offset values differ consistently in the months of Oct and Nov. Which atmospheric processes drive such kind of anomalous behavior?

In Figure 7, δD_V and δD_P do not have the same y-axis. For instance, in October (November) δD_V is approximately equal to -100‰ (-45‰) while δD_P equal to -30‰ (-10‰). So, there is always a positive offset between the isotopic composition of water vapor and precipitation despite the fact that we observe on Figure 7 that the two curves are bringing together. Moreover, once again, it is clearly mentioned in the text that we can't compare absolute values obtained from TES with observed precipitation isotopes. Concerning the processes involved in the variability of δD_V and δD_P , on Figure 7, the seasonal variability of $\delta\text{D}_\text{P} - \delta\text{D}_\text{V}$ provides information on rainwater – water vapor interactions and it is demonstrated that post-condensational effects are negligible.

However, regarding this comment from the reviewer, to avoid misunderstanding, **we added in the title of Figure 7 a precision concerning the y-axis and precision in the text (lines 364-365) about the positive offset between δD_V and δD_P .**

Other issues

2- Line: 372: It is known that the process of condensation is an equilibrium process; if so, then why 'there is no reason for the precipitation to be in equilibrium with the vapor'?

We removed this sentence to avoid misunderstanding and confusion.

3- Line 376: Pls provide a complete description of the relationship between dD_p and dD_v . It has been reported that the slope for this kind of relation is closer to 1 (Conroy et al. 2016). But, the authors get a very high value of slope, 1.68. The reason for such a large deviation should be explained. In this context, it is also suggested to use the q-dD diagram (Conroy et al. 2016) in order to characterize the moisture sources.

Many thanks for bringing this to our attention and for the reference which has been taken in account (see lines 378 – 379). **We corrected it** and putted now the right value of the slope (see line 375).

4- In order to study the effect of continental recycling, the authors have calculated "the percentage of time of the air parcel over the continent during the last 3 days along the trajectory" (Line 404-405), termed as F_{land} . Two issues are important in this context. Firstly, the method of calculation of F_{land} is not provided. Secondly, and more importantly, at what height the trajectories have been calculated? If the trajectories have been calculated for a single height, then they do not necessarily represent a major amount of moisture transport to the sampling site, since moisture is typically transported at a spectrum of heights. A detailed quantitative calculation of moisture transport for different heights is expected to be presented here. Without this exercise the conclusion of 'poor continental recycling' could be misleading.

The method of calculation of F_{land} was added in the text; see lines 405-408. About the second issue, we thank the reviewer for his comment which helped us to improve this section of the manuscript. Indeed, **we calculated air back trajectories at different heights: 800, 850, 900, 950 and 980 hPa**. We have not performed investigation at more than 800 hPa because it is the boundary layer that feeds the convection and it is with the water vapor from the low layers that the rain is re-evaporated. By calculating trajectories at these different heights, we observed that there is no difference between them, especially since we focus only on air back trajectory at 3 days before Douala. Therefore, our conclusion of "poor continental recycling" in the study region can be considered as correct. **Detailed precision on this issue were added in section 4.3.2 (see lines 408 to 414)**.

5- Line 457: how do you define the parameter "cloud surface"? How it is measured? How do you take care of the uncertainties in measurement?

Information concerning this is well detailed in section 3.3, line 238 and lines 240-241. Nevertheless thanks to this comment, to ensure the methodology described in the text is more understandable and robust, **we added more precision in lines 238, 240-241**.

6- One of the highlights "Local site precipitation $\Delta^{18}O$ of Douala was found to represent large-scale meteorology" may be confusing. I suggest rewording as 'Local site precipitation $\Delta^{18}O$ of Douala appears to respond to large-scale metrology".

Done.

7- The title of the MS is slightly misleading. Use of the word "and" in 'Western and Central Africa' may mean that the study sites are situated at both the western and central Africa. It

may be slightly modified as "Identification of processes that control the stable isotope composition of rainwater in the humid tropical West-Central Africa".

Done. Thank you for this suggestion.

Minor issues

8- Line 179: Ratios of d18O/d16O or d2D/D1H is not technically correct. Better use "ratios of 18O/16O or 2H/1H".

Done.

9- Line 212: Lekshmy et al. 2014 used TRMM derived rainfall data, not the GPCP products.

Corrected as suggested.

10- Line 270: Pls replace 'delta' by 'precipitation isotopes'.

Done. It is now line 269.

11- The MS is full of grammatical mistakes; use of the article is poorly done. I have mentioned some of them, but this is only an indicative list. It is suggested to get the MS examined by a native English speaker, or take the help of a professional service provider on the English language.

Done. Syntax errors have been corrected. In general, the English of the document has been improved and we thank the reviewer who encouraged us.

12- Line 282: 'regional scale' should be followed by "on a".

Done. It is now line 281

13- Line 296-97: please check for typos in the citations.

Done.

14- Line 311: English problem: "trajectories come..." is awkward. Trajectories are not object, but some kind of attribute. In this context, it may be replaced by moisture. Similar kind of errors is there in Line 313, 316, 318.

Corrected.

15- Line: 318: typo

Corrected.

16 - Line 321: "monsoon winds become less influent.." is not a good English. May be written as "strength of monsoon winds weaken.

Done.

17 - Line 327: The wording like "GPCP precipitation increases" may be replaced by 'the GPCP precipitation data show an increasing trend...!.

Done.

18 - Line 333: 'rainfall is most depleted' may mean that rainfall intensity is less. Better write "isotopic value of rainfall is depleted in spring....!.

Corrected as suggested. It is now line 332.

19 - Line 358: 'given' should be replaced by "provided".

Done. It is now line 357.

20 - Line 360: trajectories should be followed by 'the'.

Done. It is now line 359

Reviewer #3

1- Please avoid lumped references. Please talk about the specific points of each reference.

This remark has been considered and we would like to reassure the editors and the reviewers that each reference in the manuscript concerns a specific point or aspect.

2- How is the reproducibility of $\delta^{18}O$ seasonal cycle from year to year?

This fact is clearly detailed and explained in section 4.1. The Figure 2 also discuss about it.

3- what is the correlation between monsoon and the transition to the dry season?

General comments on climate dynamic in the study area are given in section 2. There is no correlation between monsoon and the transition toward the dry season period, except that the second occurs when the first one ends. The monsoon is a SW wet wind which is most predominant at Douala from April to August. As the ITCZ begins its southward withdrawal in September until November the monsoon wind become less efficient, giving way to the Harmattan (NE dry wind) and the transition toward the dry period often correspond to the end of October – November. This is stated in section 2 and in section 4.2.2 by looking the pattern of air back trajectories.

Reviewer #4: Manuscript ID: HYDROL31370

Manuscript Title: Identification of processes that control the stable isotope composition of rainwater in the humid tropical Western and Central Africa

The paper is in order and can be recommended for publishing in present format.

Thank you very much for your positive comment.

Highlights

- $\delta^{18}\text{O}$ was measured in precipitation in Douala (West Central Africa) at daily and monthly scales.
- key factors controlling stable isotope ratios of meteoric waters in humid tropical areas of Africa were identified thus contributing to the understanding of atmospheric processes
- Temporal variations of $\delta^{18}\text{O}$ are controlled by the intensity of upstream convection, the size of convective systems and clouds elevation
- Local site precipitation $\delta^{18}\text{O}$ of Douala appears to respond to large-scale meteorology

1 **Identification of processes that control the stable isotope composition of**
2 **rainwater in the humid tropical West-Central Africa**

3

4

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33

34 **Abstract**

35

36 This study interprets 11 years (2006 to 2016) and 6 months (March to August in 2017) of
37 respectively monthly and daily isotopic (δD and $\delta^{18}O$) monitoring of rain at Douala
38 (Cameroon), a humid tropical station in Western Africa. The main scope is to analyze the
39 climate controls on precipitation isotopes at different timescales. Firstly, we examine the
40 annual cycles of $\delta^{18}O$. Over the 11 years of survey, the annual cycle exhibits a W shape that is
41 quite reproducible from year to year, with two minima in spring and autumn periods. Based
42 on back trajectory calculations and remote sensing observations of water vapor isotopic
43 composition, we show that the observed depletion in spring and autumn is due to strong
44 convective activity along air mass trajectories. The same effect of convective activity can be
45 observed at the daily timescale. At seasonal and daily time scales, the isotopic composition is
46 also strongly tied to the convective organization and cloud types. More depleted precipitation
47 is associated with larger areas of high clouds. Very low to low clouds are observed in July-
48 August, mid-level to high clouds are dominant in June and high to very high clouds
49 characterize March-April-May, thus explaining the enriched (depleted) values in summer
50 (spring). Finally, this paper highlights the importance of large scale meteorological conditions
51 controls on precipitation stable isotope composition in the Gulf of Guinea.

52

53 Keywords: $\delta^{18}O$, water vapor, convective activity, GPCP precipitation, air back trajectory.

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59 **1. Introduction**

60 The atmosphere is an essential hydrological environment. It contains all the water vapor
61 (0.001% of all the water of the Earth; Delmas et al. 2005) that forms the clouds by
62 condensation which can then generate precipitations (liquid/solid) as a function of air
63 temperature. The atmosphere is also an essential place of transfer and exchange for the global
64 functioning of the Earth system. Its multiple interactions with the oceans, the continent and
65 the biosphere make it an important study environment for understanding global changes.
66 According to IPCC (1998), these global changes must affect tropical regions which are among
67 the most vulnerable to possible anthropogenically induced climatic changes.

68 During the last decade, the scientific community, environmental institutions, governments,
69 and local communities have increased their awareness of the importance of current tropical
70 climate variability, based on the premise that, under increasing anthropogenic influence on the
71 climate, changes in regional and global circulation may lead to an intensification of extreme
72 events (i.e. floods or severe droughts: e.g., Pohl et al. 2017). In parallel, the use of stable
73 isotopes of water, both δD and $\delta^{18}O$ has provided insights in the study of atmospheric water
74 cycle (e.g., Rozanski et al. 1993, Araguas - Araguas et al. 2000, Celle-Jeanton et al. 2004); the
75 key to understand the future climate changes or global changes. Relationship between δD and
76 $\delta^{18}O$ in natural meteoric waters serves as a foundational reference to determine regional and
77 local deviations from equilibrium processes and the potential origin of the water vapor.
78 Moreover, water losses due to evaporation, incorporation of recycled atmospheric moisture,
79 and/or mixing between isotopically distinct reservoirs leave a unique water fingerprint that
80 can be used for climate-reconstructions (Moerman et al. 2013). Recently, isotopic
81 composition of tropical meteoric water has also proven its usefulness as an indicator of
82 modern climate variability (Vuille and Werner 2005; Ishizaki et al. 2012; Sanchez-Murillo
83 2015).

84 While in extra-tropical climates, stable isotope variations in meteoric
85 waters have been successfully explained by air temperature variability (Dansgaard 1964;
86 Rozanski et al. 1993), the case of tropical humid regions proved to be very much complex
87 since temperature variability is much weaker. The amount effect occurs as a result of
88 convective precipitation and can be nonlocal (Vimeux et al. 2005; He et al. 2015). The degree
89 of organization of convective systems has been shown to impact the atmospheric conditions at
90 the large-scale (Tobin et al. 2012; Wing et al. 2017). Consistently, it also impacts the isotopic
91 composition of water vapor and precipitation. Many additional factors can potentially play a
92 role, such as orographic effects, continental recycling or moisture origin combined with
93 complex microclimates (Rozanski et al. 1993, Lachniet and Paterson 2009).

94 These problematics are of major importance for the paper. What are the key factors
95 controlling stable isotope ratios of meteoric waters in humid tropical areas of West and
96 Central Africa? Answers to this question will contribute to the understanding of atmospheric
97 processes in the study region. Unlike arid and semi-arid African areas, where isotopic
98 variability has already been assessed in previous work (e.g; Taupin et al. 1997, Celle-Jeanton
99 et al. 2001; Risi et al. 2008b, 2010; Lutz et al. 2011, Tremoy 2012; Tremoy et al. 2014), the
100 Gulf of Guinea (GOG) region and tropical humid areas in Central Africa (Figure 1) are still
101 under-documented in isotopic data. There, the atmospheric cycle of water remains poorly
102 documented and isotopic data may help improve their knowledge. Indirectly, the paper will
103 help to understand rainfall variability in Douala and the origin of precipitated water through
104 isotopic analysis. In this objective, we took advantage of the long term monthly monitoring
105 set up in Douala, Cameroon (Figure 1) as part of the Global Network for Isotopes in
106 Precipitation (GNIP) framework (IAEA/WMO 2018) from 2006. In parallel, a daily sampling
107 has been carried out in 2017. This article represents the first valuation of precipitation isotope
108 data in the region. Datasets of other GNIP stations in the GOG and Central Africa (Figure 1)

109 are also integrated in this study, together with remote sensing data that document water vapor
110 isotopic composition, convective activity and cloud properties.

111

112 Figure 1.

113

114 **2. Regional climate: the West African Monsoon (WAM) dynamic**

115 Seasonality in the tropics is mostly determined by the seasonal migrations of the Inter
116 Tropical Convergence Zone (ITCZ, Preston-White and Tyson 1988; Schott et al. 2003), a
117 highly energetic feature of earth climate that is associated with deep convection. The ITCZ or
118 meteorological equator in Africa is the result of convergence between the Harmattan
119 (northeasterly dry wind) and Monsoon (southwesterly wet wind) in low levels of the
120 atmosphere (Sultan and Janicot 2000, Fink et al. 2018). Nicholson and Grist (2003) have
121 shown that the rain belt over Western and Central Africa is positively correlated with the
122 migration of the ITCZ. During the boreal spring (Mar-Apr-May), ITCZ holds a position of
123 about 5°N; half of the rain belt is located over the continent and the other one on the ocean
124 (Waliser and Gautier 1993). The mean rainfall amount calculated for this period, according to
125 GNIP measurements, is 232 mm for Douala, 138.4 mm for Cotonou, 137.7 mm for Sao-Tome
126 and 114.1 mm for Bangui. The relative humidity increases significantly over the continent
127 and sea surface temperatures (SST) present their annual peak (>27°C) during this period
128 (Dezfuli and Nicholson 2013). Then, ITCZ abruptly shifts from 5°N to 10°N in June-July
129 (Sultan and Janicot 2003). The so-called monsoon jump causes a subsidence in the southern
130 part of the GOG and the development of a cold tongue complex (Gu and Adler 2004). Low
131 SSTs suppress then partly or totally rainfall as e.g. in Sao-Tome (Figure 3). However, in the

132 Gulf of Biafra (Figure 1), SSTs remain high enough ($>26^{\circ}\text{C}$) to partly maintain convection
133 (Odekunle and Eludoyin 2008). Moisture increases over land and enhances rainfall there.
134 Simultaneously, convective activity migrates to the North, and the monsoon flow crosses the
135 Equator from South (GOG) to North (Guinean and then Sahelian Africa). At this time of the
136 year (Jun-Jul-Aug), a rainfall peak is observed (Figure 3) in Douala (618.3 mm) and Bangui
137 (145.5 mm). In Cotonou, Ogu et al. (2016) have also shown a good correlation between SSTs
138 and precipitation, connecting rainfall amount in June to anomalously warm waters of the Gulf
139 of Benin.

140 The abruptness of the northward progression of the ITCZ is in sharp contrast to its
141 withdrawal, which appears as a more progressive southward progression. Like most regions in
142 the equatorial latitudes, the Guinean Coast/Central Africa thus benefits from a double passage
143 of the ITCZ in spring and autumn.

144 The unimodal distribution and the highest rainfall amount (~ 4000 mm/year) observed in
145 Douala (Figure 3) is very particular as the whole region is generally characterized by a
146 bimodal distribution with two precipitation maxima in April-May (or May-June) and October-
147 November (or September-October), with a total amount lower than 2000 mm/year. The
148 concave shape of the Gulf of Biafra may induce a convergence of southwesterly winds and
149 topographic ascents forced by the relief of Mount Cameroon (> 4000 m, Figure. 1). This
150 mesoscale convergence and orographic influence (Vondou et al. 2017) are all the more
151 marked if the monsoon flow is intense and thick, which is the case in the core of the summer
152 monsoon, when the ITCZ is far to the North. These regional features can explain the
153 unimodal regime observed at Douala while further west (in Cotonou), and in southern
154 hemisphere (Sao-Tome), the annual cycle is bimodal.

155 **3. Data and methods**

156 ***3.1. Rainfall sampling and isotopes analyses***

157 This paper makes use of the 106 monthly samples of rainwater that have been collected from
158 July 2006 to December 2016 at the GNIP station of Douala (see details of the GNIP stations
159 in [Table 1](#) and [Figure 1](#)). The sampling followed the standard protocols ([IAEA 2012](#)).
160 Samples were collected by using a rain gauge which consists of a plastic funnel (diameter =
161 10 cm) coupled with a filter mesh to prevent contaminations by debris. A 5 cm layer of
162 mineral oil has been systematically added into the rain collector to avoid fractionation of the
163 collected rainwater. For each month, total rainfall was collected. Samples were taken
164 regularly, stored in a totalizer and kept at 4°C before being transferred in 50 ml amber glass
165 bottles, tightly capped and sent to the International Atomic Energy Agency (IAEA) laboratory
166 in Vienna, Austria for stable isotopes determinations. Stable isotopes of hydrogen and oxygen
167 were then analyzed by laser absorption spectroscopy following the method described by
168 [Penna et al. \(2010\)](#).

169 **Table 1**

170
171 The rainwater daily survey was conducted from March to August 2017 at the campus of the
172 University of Douala (X= 9.7461; Y= 4.062; Z= 17 m.a.s.l.), approximately at 7 km from the
173 GNIP station. Seventy samples were collected using a Palmex rain gauge that presents the
174 advantage to avoid evaporation without using medicinal paraffin oil ([Gröning et al. 2012](#)).
175 Daily samples were stored using the same protocols as for monthly rainfall. Isotopes analyses
176 were performed at the Hydrogeology Department of the University of Corsica, France, by

177 using a liquid-water stable isotope analyzer DLT-100 Los Gatos Research (Aggarwal et al.
178 2006; Penna et al. 2010).

179 Ratios of $^{18}\text{O}/\delta^{16}\text{O}$ and $^2\text{H}/^1\text{H}$ are expressed in delta units (‰, parts per mil) relative to Vienna
180 Standard Mean Ocean Water (V-SMOW). The analytical precision is $\pm 0.1\text{‰}$ for oxygen-18
181 and $\pm 1\text{‰}$ for deuterium.

182 Local meteorological settings (precipitation amount and duration, air temperatures, vapor
183 pressure) were provided by the National Weather Direction of Cameroon.

184

185 ***3.2. Tropospheric Emission Spectrometer (TES) data***

186 With the advent of new technology in stable water isotopes, such as spectrometer on-board
187 satellite, it has become easier to analyze isotopic composition of water vapor based on
188 indirect laser measurement (Aemisegger et al. 2012). TES instrument on board on the Aura
189 satellite is a nadir-viewing infrared Fourier transform spectrometer from which the deuterium
190 content of water vapor (δD_v) can be retrieved (Worden et al. 2006; Worden et al 2007). The
191 sensitivity of the retrieval is typically larger between 900 hPa and 400 hPa with a peak at 700
192 hPa. On average, δD_v retrieved over these levels has a precision of 1.5% or about 15 parts per
193 thousand (per mile) relative to Standard Mean Ocean Water. Uncertainties are reduced by
194 averaging several measurements (Worden et al. 2006; Risi et al. 2013). Then the precision is
195 sufficient for characterizing the global distribution of evaporation and condensation processes
196 (Worden et al. 2006).

197 There is on average 1.8 degrees of freedom for δD_v retrievals in the tropics (Worden et al.
198 2012), meaning that vertical profiles bear information on more than one level. To ensure good
199 data quality, we selected only the measurements for which the quality flag is set to unity and
200 for which the degree of freedom of the signal is higher than 0.5 (Risi et al. 2013). Here, to
201 document the water vapor composition as close as possible to the surface, we use the δD

202 values retrieved by TES at 900 hPa from 2004-2008 at a monthly scale and we focus on a
203 multi-year mean seasonal cycle and temporal variations rather than absolute values.

204

205 ***3.3. Convection and cloud datasets***

206 Convective activity associated with the West African Monsoon (WAM) was analyzed using
207 the Global Precipitation Climate Project one degree daily (GPCP-1dd; Huffman et al. 2001)
208 data. The same dataset is used for both monthly and daily analyses. Data were retrieved from
209 the National Oceanic and Atmospheric Administration (NOAA) website
210 (<https://www.esrl.noaa.gov/psd/cgi-bin/data/composites/printpage.pl>). The robustness of
211 GPCP products has been demonstrated in many studies (e.g.; Huffman et al. 1995, 1997;
212 Adler et al. 2017) through a comparison with other proxies of convection or by multi-proxy
213 studies: OLR (Outgoing Longwave Radiation) and TRMM (Tropical Rainfall Measurement
214 Mission). GPCP products are a combination of precipitation data provided by a multiple
215 sources of satellite-gauge (SG). They are obtained by optimally merging precipitation
216 estimates computed from microwave, infrared, sounder data observed by the international
217 constellation of precipitation-related satellites, and rain gauge analyses, taking advantage of
218 the strengths of each data type. Mean SG products are computed by combining multi-satellite
219 estimates with rain gauge analysis (Huffman et al. 2001). In this study, we use the GPCP-1dd
220 (1-degree grid over the entire globe at 1-day) both for monthly and daily analyses.

221 To document the cloud types and convection organization in Western Central Africa, we use a
222 cloud type (CT) product issued from geostationary MSG (Meteosat Second generation)
223 satellite data and developed by SAFNWC (Satellite application facilities in support to
224 nowcasting) /MSG algorithms (for more information see
225 <http://www.nwcsaf.org/web/guest/scientific-documentation>). Clouds types are determined
226 from their top temperature or pressure and, for high level clouds from their opacity (Dommo

227 et al. 2018; Seze 2015). The SAFNWC CT offers a classification of clouds into 12 classes:
228 free land, free sea, very low clouds, low clouds, medium clouds, high clouds, very high
229 clouds, thin cirrus, medium cirrus, thick cirrus, fractional clouds and semi-transparent above
230 low or medium clouds, at 3-km spatial resolution for regions close to MSG sub-satellite point
231 (0N, 0E) and a time step of 15-min. For our study purposes, focus is given to the altitude of
232 clouds (low, medium, high, etc.) and their organization (as measured by the spatial extent of
233 connected cloud pixels) to seek links with the isotopic contents of rainfall. Data are extracted
234 for the region bounded by latitudes 10S–10N and longitudes 0E–20E, for the period from
235 March to June 2017.

236 Based on these images, very high clouds areas (corresponding to very organized convective
237 system) around Douala were calculated for each event using ArcGis mapping software,
238 through the tool “measure”. Indeed, since the contours of cloud masses **or pixels** are
239 approximated to geometric forms (circle, rectangular, trapezoidal shape, etc.), the areas were
240 calculated based on these forms. **This calculation was performed after a zoom of the image at**
241 **100% in order to limit uncertainties or errors by using the tool of the software.** For events
242 with duration of more than 15 min, we have taken the maximum cloud area throughout the
243 duration of rainfall.

244 **3.4. Back trajectories**

245 In order to assess the influence of air masses pathways on the isotopic composition of
246 precipitation, and the importance of the location of convective activity in the region, we
247 compute air back trajectories at 6h time steps, 10 days prior to arrival in the sampling site.
248 This operation was performed for monthly and daily scales when isotopic data is available.
249 Winds (at 900 hPa) were simulated by using the general circulation model LMDZ5A
250 (Hourdin et al. 2013) guided by reanalysis products (ERA-Interim, Dee et al. 2011) of the

251 European Centre for Medium-Range Weather Forecasts (ECMWF). Back trajectories (speed
252 and direction) were then computed with a 2D algorithm (similar to Vimeux et al. 2005) in
253 order to approximate the moisture transport near the ground surface.

254

255 **4. Results and discussion**

256 *4.1. Annual cycles of isotopes (δ) in precipitation of Douala*

257 **Figure 2** presents the seasonal variations of $\delta^{18}\text{O}$ from 2006 to 2016. The annual cycle of δ is
258 quite reproducible from year to year. It exhibits a W shape most of the time, with minima in
259 spring and autumn periods. Because of this rather good reproducibility of $\delta^{18}\text{O}$ seasonal cycle
260 from year to year; we further consider a multi-year mean seasonal cycle in the section 4.2, and
261 investigate the factors associated with this W cycle of δ .

262 However, the magnitude varies widely from one year to another for a given month. Such
263 inter-annual variability could either reflect variations (i) in the climate seasonal background,
264 or (ii) in synoptic / intra-seasonal variability, which could modify seasonal mean fields
265 through upscaling processes, or eventually affect isotopic variations. These hypotheses are
266 tested and discussed below.

267  Figure 2.

268

269 **4.2. Factors responsible for precipitation isotopes variations at seasonal scale**

270 *4.2.1. Highlighting of a regional context*

271 Seasonal variations of $\delta^{18}\text{O}$ and precipitation amount for GNIP stations in GOG and Central
272 Africa region are presented in **Figure 3**.

273

274  Figure 3.

275
276 Similarities can be observed in the seasonal evolution of $\delta^{18}\text{O}$ at Cotonou, Bangui, Douala
277 and Sao-Tome (Figure 3): (i) from January-February to April-May, $\delta^{18}\text{O}$ decreases; (ii) an
278 enrichment in isotopic content is observed in summer (in June, July or August depending on
279 the station); (iii) isotope contents then decrease until September or October and (iv) a new
280 increase occurs until December. This relative homogeneity suggests that isotopic seasonality
281 is controlled by mechanisms of, at least, on regional scale.
282 Correlation coefficients between $\delta^{18}\text{O}$, precipitation and temperature are very low for all the
283 stations (Table 2). The weak correlation between $\delta^{18}\text{O}$ and air temperature (Table 2)
284 highlights the lack of temperature effect in the GOG and Central Africa regions. Despite the
285 fact that, in tropical maritime stations, a local amount effect is often observed (Rozanski et al.
286 1993), the poor correlation between $\delta^{18}\text{O}$ and local precipitation amount shows that this effect
287 does not dominate the isotopic seasonality in Central Western Africa. Therefore, it is clear
288 that, the local climate parameters do not control the seasonal variation of δ in precipitation.
289 Thus we hypothesize that convective activity at the regional scale could be involved. This
290 hypothesis is tested in the next section for the Douala station.

291 Table 2.

292

293 4.2.2. Influence of regional convective activity

294 Many studies in China (e.g., Gao et al. 2013; He et al. 2015 ; Yu et al. 2016; Guo et al. 2017;
295 Shao et al. 2017; Gao et al. 2018), India and Indo-Pacific region (e.g., Chakraborty et al.
296 2015; Rahul and Gosh 2016; Cai et al. 2016; He et al. 2018), South America (e.g., Hoffmann
297 et al. 2003; Vimeux et al. 2005; Villacis et al. 2008, Samuels-Crow et al. 2014) and Sahelian
298 Africa (e.g., Risi et al. 2008b; Risi et al. 2010b; Tremoy et al. 2012) based on both

299 observations and models, have shown that convective activity, that occurs upstream the
300 pathway of air parcels, is a major control of rainwater isotope composition in the tropics at
301 daily, seasonal and inter-annual timescales.

302 Convective activity is known to deplete water vapor through 3 main processes: i) precipitating
303 downdrafts, either at the convective-scale or at the meso-scale (Risi et al. 2008a, Kurita
304 2013), bring down depleted water vapor from the mid-troposphere to the boundary layer; ii)
305 rain evaporation, when concerning a small proportion of each raindrop, adds depleted water
306 vapor to the lower troposphere (Worden et al. 2007, Risi et al. 2010); iii) rain-vapor diffusive
307 exchanges in a saturated atmosphere can also deplete the water vapor (Lawrence et al. 2004).

308 GPCP precipitations are used to examine convective activity upstream the sampling site of
309 Douala. Figure 4 presents monthly mean precipitation and back trajectories calculated for
310 each month. As expected, moisture comes from the Atlantic Ocean most of the time and
311 precipitation, at this timescale, is entirely controlled by the seasonal migration of the ITCZ.

312 In November and December, moisture comes from the North-East without undergoing any
313 convection. From December to January, there is an abrupt shift of air parcel from North-
314 Easterly to South-Westerly, in line with the beginning of long northward migration of the
315 ITCZ. From March to May (spring season), air masses undergo strong convective activity
316 over the Gulf of Guinea. During the period June - August, air parcels still travel over the Gulf
317 of Guinea, but air masses undergo a weak convective activity because the ITCZ has shifted
318 further north over the Sahel region. From September to October (autumn period), the ITCZ
319 retreats to the South so that air masses undergo strong convective activity once again. The
320 shift between October and November trajectories marks the transition to the dry season when
321 the strength of monsoon winds weakens.

322 Figure 5 by presenting the variability of monthly mean GPCP precipitation along the back
323 trajectory over time (in hours), towards Douala station for May (spring period), August

324 (summer period), October (autumn period) and January (winter period) shows a quantitative
325 analysis. It reveals where, along the trajectories, convective activity becomes higher and thus
326 impacts $\delta^{18}\text{O}$ of Douala rainfall. Overall, the GPCP precipitation data show an increasing
327 trend from the Southern Atlantic to Douala. Along the air back trajectories, at 7 days (168
328 hours) to the sampling site, precipitation is higher in August and May. First significant
329 changes are observable at 120 hours, when a decrease (increase) of GPCP precipitation occurs
330 in January (October). It substantially increases in May and August. At 3 days (72 hours)
331 before Douala, convective activity is stronger in October (and to a lesser extent in May) than
332 August and January. Since isotopic value of rainfall is depleted in spring and autumn (Figures
333 2-3), $\delta^{18}\text{O}$ seems to be mostly sensitive to convective activity only in the past few days before
334 reaching Douala. This is consistent with the “memory” of convection in the isotopic
335 composition of water as discussed by Risi et al (2008a), Tuinenburg et al (2015) and Gao et al
336 (2013). Stable isotopes are imprint of convection along air parcel trajectories, and in the case
337 of Douala, precipitation seems to acquire its signature on average 72h before reaching the
338 station. Figure 6 shows the precipitation averaged over the past 72h (3 days) along the
339 backward trajectories, for each month. The precipitation $\delta^{18}\text{O}$ is significantly anti-correlated
340 with this average precipitation ($r^2=0.60$). This supports our hypothesis that convective activity
341 along trajectories significantly controls the seasonality of precipitation $\delta^{18}\text{O}$ in Douala.
342 In summary, the moisture “source” at Douala is most of the year in the GOG. Two seasonal
343 precipitation maxima along the trajectories (Figure 6) are related to the seasonal migration of
344 the ITCZ. Thus, even if precipitation in Douala shows a unimodal regime with a rainfall peak
345 in August, $\delta^{18}\text{O}$ records bimodal cycle as in the “source” area. This suggests that the regional
346 convective activity is the main control of the isotopic composition of precipitation in Douala.

347 In the following section, we use the isotopic information of the water vapor to provide
348 additional insights about the main processes controlling the isotopic composition of
349 precipitation.

350

351 Figure 4.

352

353 Figure 5.

354

355 Figure 6.

356 ***4.2.3. Information from the isotopic composition of water vapor (δv)***

357 The isotopic composition of the water vapor **provided** by the TES instrument can help to
358 understand the variation of δ in precipitation: it allows separating the relative effect of
359 processes acting along **the trajectories of the water vapor** (δD_v) and local post-condensational
360 processes ($\delta D_p - \delta D_v$) following the equation below:

$$361 \quad \delta D_p = \delta D_v + (\delta D_p - \delta D_v) \quad (1).$$

362 δD_v and δD_p (**Figure 7a**) show a similar seasonality ($r^2 = 0.57$). Both δD_v and δD_p present
363 depleted values in April-May and September-October (**Figure 7a**), corresponding to more
364 active convection in the GOG. **In addition, there is always a positive offset between the**
365 **isotopic composition of water vapor and precipitation.** Thus, the isotopic information on the
366 advected water vapor is preserved in the isotopic composition of the rain.

367

368 Figure 7.

369 However, absolute values of δD_p and TES δD_v should not be directly compared. TES was
370 calibrated using in-situ measurements on local sites (Worden et al 2012) but not specifically
371 in West Africa. In addition, TES data at 900hPa represent an average over several vertical
372 levels in the lower troposphere. Therefore, focus is given instead to temporal variability. Yet,
373 assuming that the vertical δD gradient between 1000 and 900hPa is constant, $\delta D_p - \delta D_v$
374 variations can be interpreted as variations in the rain-water vapor interaction processes.

375 The regression between δD_p and δD_v gives a slope of $a_1 = 1$ whereas the values for δD_p and
376 $\delta D_p - \delta D_v$ are $a_2 = -0.68$ and $r^2 = 0.18$. This insignificant r^2 attests that the rain-water vapor
377 interaction processes do not control the δD_p variations. Based on the slope a_1 , δD_v accounts
378 for 100% of the δD_p variability. A similar case has been reported in western tropical Pacific
379 (Conroy et al. 2016). This confirms that the seasonal variability of the isotopic composition of
380 rainfall is predominantly influenced by δD_v , i.e. by the processes that affect water vapor along
381 the trajectories. This means that the variability of $\delta D_p - \delta D_v$ acts only to dampen and to blur the
382 variability of δD_p . For instance, the evolution of $\delta D_p - \delta D_v$ presents a maximum in spring
383 (Figure 7b) and autumn, when δD_v is most depleted.

384 4.3. Main controls of precipitation isotopic composition at daily scale

385 4.3.1. Temporal evolution of δ and link with upstream convection

386 Daily $\delta^{18}O$ in Douala varies from -0.3‰ to -7.4‰ , with a mean value of -3.0‰ close to the
387 monthly weighted mean of -2.8‰ calculated for the 2006-2016 period. These daily data
388 appear then to be representative of the 2006-2016 mean seasonal cycles. Moreover, as for the
389 seasonal scale, daily rainwater (Figure 8) is most depleted in spring (April-May) and more
390 enriched in summer (July-August).

391

392

Figure 8.

393 In order to investigate upstream convection effects, we calculated the correlation between the
394 $\delta^{18}\text{O}$ of rainfall and the precipitation recorded at 1 to 3 days earlier, along back trajectories
395 (Table 3). In March, April and August, the correlations are significant ($\geq 95\%$), suggesting
396 that, like for the seasonal timescale, the daily evolution is driven by convection along
397 trajectories. Yet, this mechanism is insufficient to explain completely the variability of δ in
398 daily precipitation, especially in May, June and July.

399 Table 3.

400 *4.3.2. Factors besides the upstream convection intensity: continental recycling and* 401 *organization of convective systems*

402 *Continental recycling?*

403 The goal of this section is to test whether continental recycling has significant influence on
404 $\delta^{18}\text{O}$ at the daily time scale. To that end, we calculated the percentage of time of the air parcel
405 over the continent during the last 3 days (72h) along the trajectory, hereafter F_{land} . This
406 calculation was relatively simple since air back trajectories have been computed at 6h time
407 steps and then for instance, the travel time of 36h of air mass over the continent will
408 correspond to F_{land} equal to 50%. However, for this issue, we calculated air back trajectories
409 at different heights: 800, 850, 900, 950 and 980 hPa since moisture is typically transported at
410 a spectrum of heights. As it is the boundary layer that feeds the convection and it is with the
411 water vapor from low layers that the rain is re-evaporated, investigation was not performed
412 above 800 hPa. Results from this quantitative calculation (not shown here) of moisture
413 transport for different heights show that there is no difference between them, especially since
414 we focus only on air back trajectory at 3 days before Douala. In June, July and August, F_{land} is
415 0% for all rainy days (except on June 11th when $F_{\text{land}} = 25\%$). Table 4 presents the detailed

416 results for spring period and shows an anti-correlation between $\delta^{18}\text{O}$ and F_{land} . Precipitation
417 are thus more depleted, when the transit time of air parcels over the continent is long, during
418 the last 3 days along the back trajectory This is in contradiction with the expected effect of a
419 continental recycling.

420 Table 4.

421
422 Since δ is controlled in spring by a convection upstream the sampling site, the fact that $\delta^{18}\text{O}$ is
423 more depleted with the increase of F_{land} , suggests a significant influence of convection
424 intensity over the continent. Indeed, [Xu and Zipser \(2012\)](#) demonstrated that convection is
425 generally more intense over the land than over the ocean. The more intense the convection
426 along trajectories, the more depleted the water vapor. Therefore, moisture origin has only an
427 indirect effect on isotopic composition of precipitation, depending on whether air mass goes
428 through regions of a strong convection. The positive and significant correlation between F_{land}
429 and GPCP precipitation confirms this mechanism.

430 Therefore, we suggest that, δ variations at a daily timescale are partly controlled by the
431 intensity of the convection along air parcel trajectories and that continental recycling is not
432 involved.

433

434 ***Influence of the organization of convective systems?***

435 Several studies have shown that convective systems deplete the low-level water vapor more
436 efficiently when they are more organized ([Lawrence et al. 2004](#), [Risi et al. 2008](#), [Tremoy et al](#)
437 [2014](#)), probably because moister air in larger systems allows for more efficient rain-vapour
438 diffusive exchanges ([Risi et al. 2008a](#)). In addition, convective systems deplete more vapor as
439 they extend to the upper troposphere ([Lacour et al 2018](#)) and as the extent of their anvils

440 (measured by the fraction of stratiform clouds) is large (Aggarwal et al. 2016). These features
441 are typically associated with higher degrees of organization.

442 Here, we test whether the type of convective organization has a significant impact on the $\delta^{18}\text{O}$
443 observed at Douala. By using satellite image from the SAF classification, we define 3 classes
444 of clouds in the region from March to August according to their altitude: (i) very low to low;
445 (ii) medium to high and (ii) very high clouds. The low cloud class is mostly observed in July-
446 August (Table 5 and Figure 9). The mid-level to high cloud class is dominant in June and the
447 very high clouds are mostly present in spring (Figure 9). The area of cloud systems increases
448 throughout spring, reaching a maximum in May, and decreases again from June. In July-
449 August, the areas are too small to be calculated.

450

451  Figure 9.

452 Spatial organization of convective systems also determines rainfall event duration. The latter
453 increases globally from March to August (Figure 10), ranging from 12 minutes to 1278 min
454 (i.e. more than 21 hours).

455

456  Figure 10.

457 Combining this information on cloud altitude, cloud area and rainfall duration; we can infer
458 that in March-April, most convective systems correspond to deep mesoscale convective
459 systems (MCS; Mapes and Houze 1993; Laing and Fritsch 1997; Mathon et al. 2001; Fink et
460 al. 2006; Tremoy et al. 2014). In May-June, convective systems are MCS that become even
461 larger and organized, but extend less deep in altitude (in June). In contrast, in July-August,

462 convective systems are isolated; characterized by small cumulonimbus in the middle of low-
463 to-medium clouds (Figure 9b) that follow one another throughout the day, explaining the
464 apparent long durations of rain events.

465 The type of convective organization emerges then as an important control on the precipitation
466 isotopic composition. Cloud surface shows a significant negative correlation with
467 precipitation $\delta^{18}\text{O}$ from March to June (Table 5). The larger the cloud area, the more depleted
468 the precipitation. In addition, the more the convective system is organized, the longer is the
469 duration of the event (Laurent and Machado 2002, Fiolleau et al. 2013). Consistently, event-
470 scale $\delta^{18}\text{O}$ is anti-correlated with event duration in March ($r^2 = 0.44$), April ($r^2 = 0.54$), May
471 ($r^2 = 0.64$) and more weakly in June ($r^2 = 0.34$). The longer is the convective system, the more
472 depleted is the precipitation.

473 These different types of convective organization can also explain the isotopic evolution at the
474 seasonal scale. In spring, the organized, deep MCS deplete the water vapor efficiently,
475 leading to the observed precipitation $\delta^{18}\text{O}$ minimum. In contrast, in July-August, the
476 shallower, small isolated cumulonimbus clouds deplete the water vapor less efficiently,
477 leading to the observed precipitation $\delta^{18}\text{O}$ maximum in spite of the local precipitation
478 maximum observed at Douala.

479 Surprisingly, the r^2 between cloud area and $\delta^{18}\text{O}$ are close and even higher (in May and June)
480 than that observed between $\delta^{18}\text{O}$ and upstream precipitation (Table 3). After verifying that
481 upstream convection (GPCP precipitation recorded at 72h to Douala along the trajectory) and
482 cloud area are uncorrelated (Table 5), we can assert that two independent parameters control
483 $\delta^{18}\text{O}$: upstream convection intensity and the size of the convective system (both at local scale
484 and at upstream, see the Table 5).

485

486

Table 5.

487

488 **5. Conclusion and outlook**

489 This study aims at investigating the processes controlling the year-to-year (inter-annual),
490 month-to-month (seasonality) and day-to-day (intra-seasonal) variability of rainfall isotopic
491 composition in the Gulf of Guinea (GOG) region and especially in Douala (Cameroon). We
492 observed that the annual cycle of δ is quite reproducible. Most years presents a W shape with
493 minima in spring and autumn. $\delta^{18}\text{O}$ and δD_v appear to be mainly controlled by upstream
494 convection and by the size of convective systems. We identified that the continental recycling
495 doesn't impact the rainwater isotopes in the humid tropical area of the GOG. In particular, the
496 $\delta^{18}\text{O}$ minima in spring and autumn are associated with strong convective activity in the GOG
497 and large, long-lived and deep mesoscale convective systems, whereas the $\delta^{18}\text{O}$ maximum in
498 July-August is associated with reduced convective activity in the GOG and isolated shallow
499 convective systems. The importance of upstream convective activity in controlling the $\delta^{18}\text{O}$ of
500 precipitation at various time scales is in line with a large body of recent research in different
501 tropical regions influenced by a monsoon system. The importance of the type of convective
502 system (size, organization, vertical extension) is also consistent with a growing number of
503 recent studies but this is the first time that it is demonstrated through a detailed analysis of
504 such a large number of individual convective systems. The findings of this study (obtained by
505 integrating in situ and satellite measurements) advance our understanding of the temporal
506 variation of precipitation stable isotopes in humid tropical area such as Douala, and shed a

507 new light on the importance of large scale meteorological conditions controls on precipitation
508 stable isotope composition in the GOG.
509 Notwithstanding, numerical climate modelling could be a useful complementary approach to
510 further analyse the factors controlling rainfall isotopic composition in and around Douala
511 Moreover, the climate in Douala being representative of a small area with monomodal rainfall
512 regime surrounded by area with bimodal rainfall, high resolution modelling is necessary to
513 correctly capture such particularity. More robust conclusions could be obtained with larger
514 samples, which could be obtained by extending the length of the record. To better understand
515 the role of convective organization on the isotopic composition, sampling at the intra-event
516 scale is necessary. Finally, measuring the isotopic composition in the water vapor in addition
517 to precipitation would be useful to isolate the post-condensation effects.

518

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527

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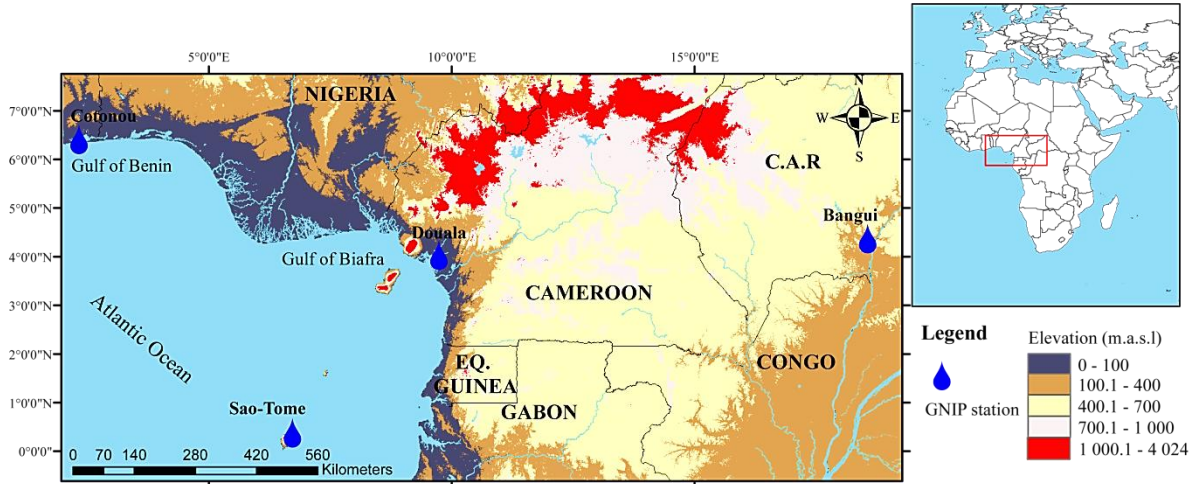
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946 **Figures and captions.**

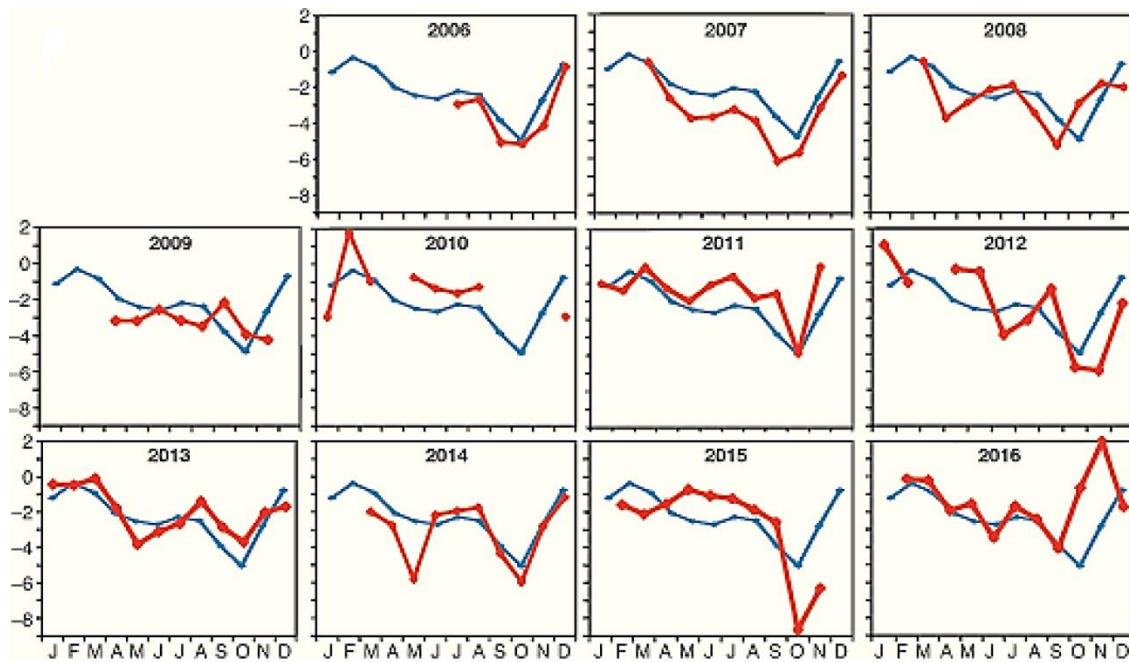
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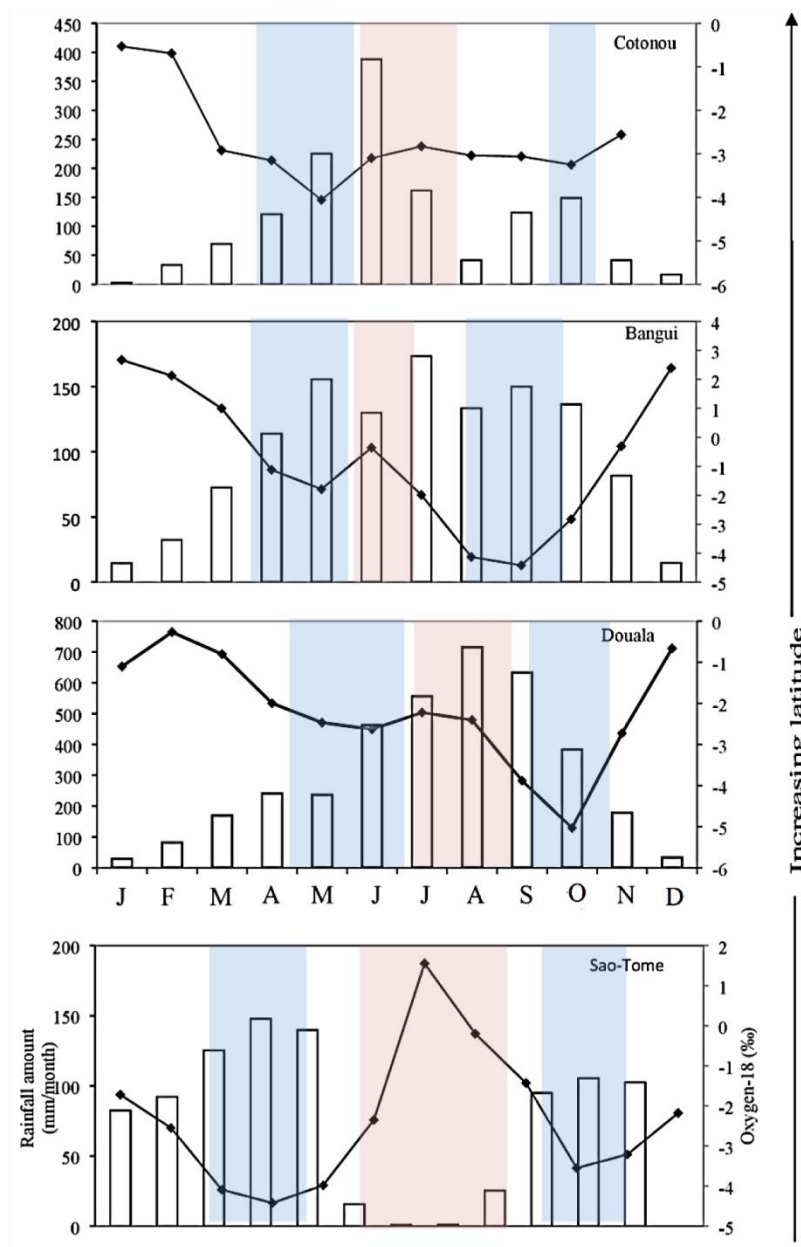


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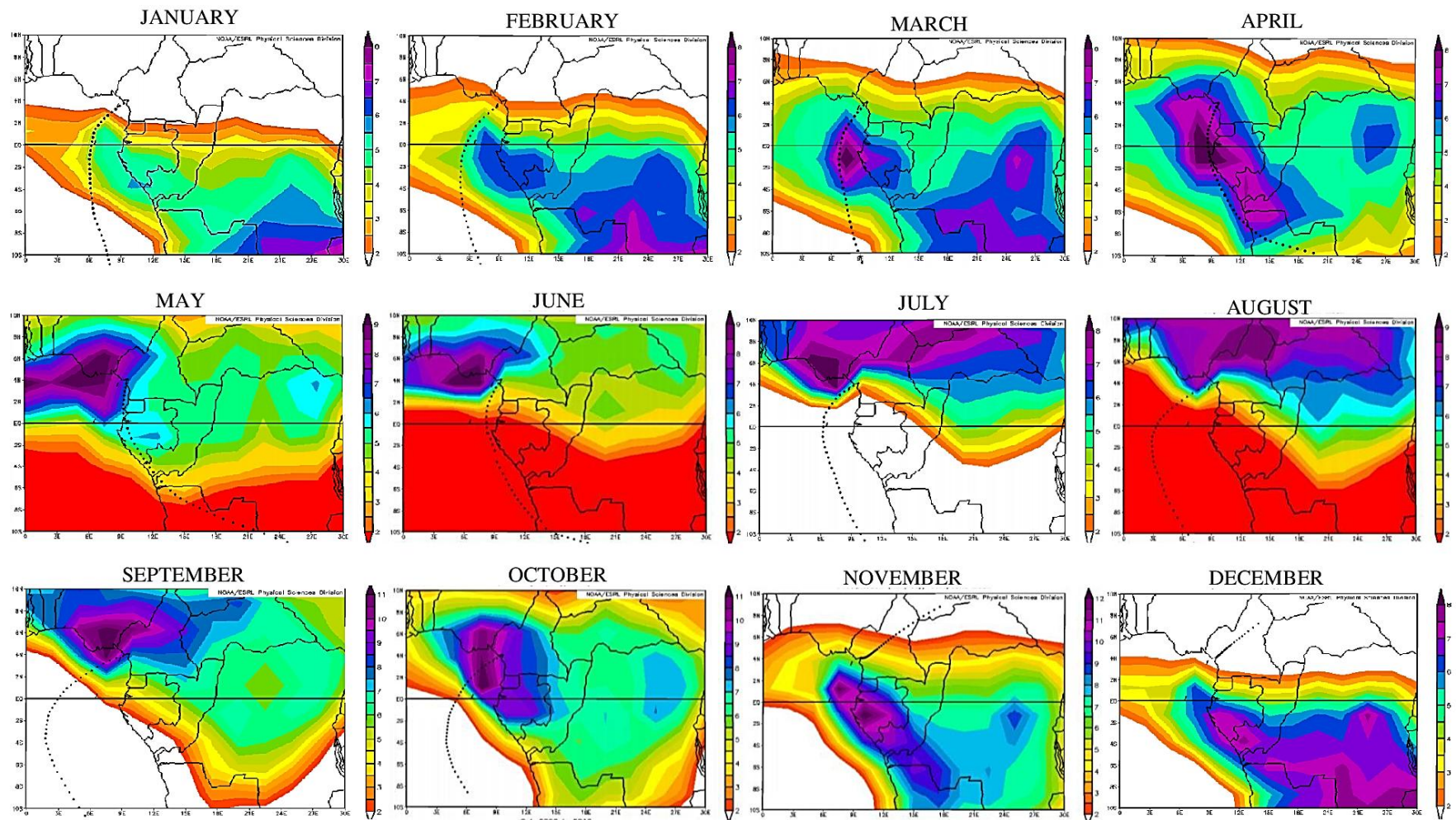
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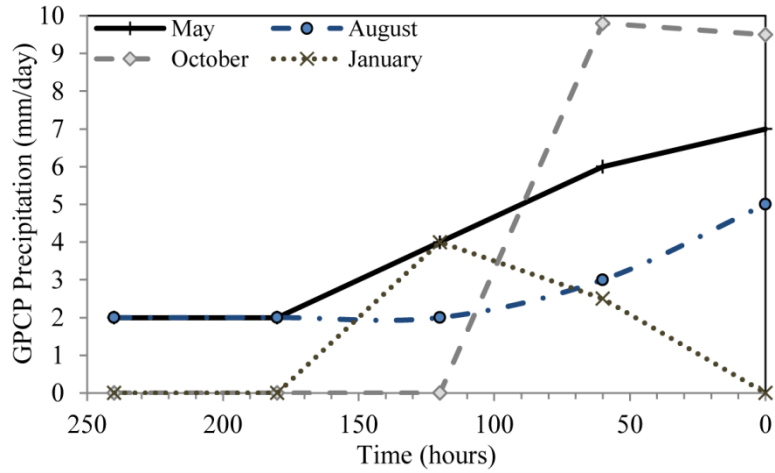


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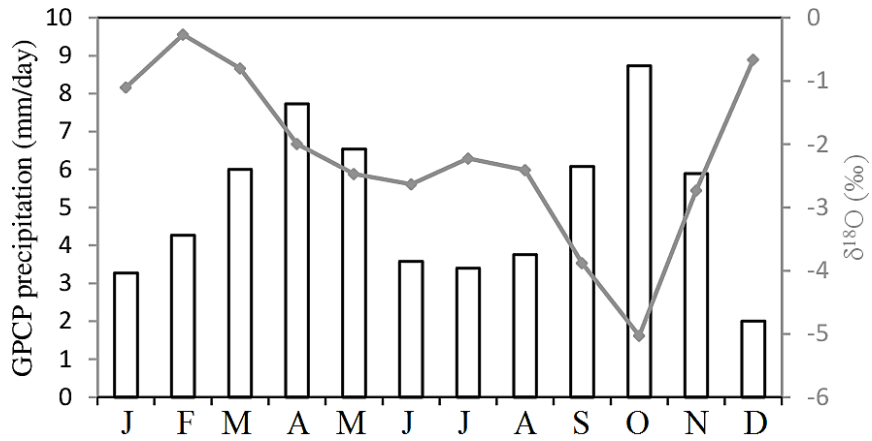


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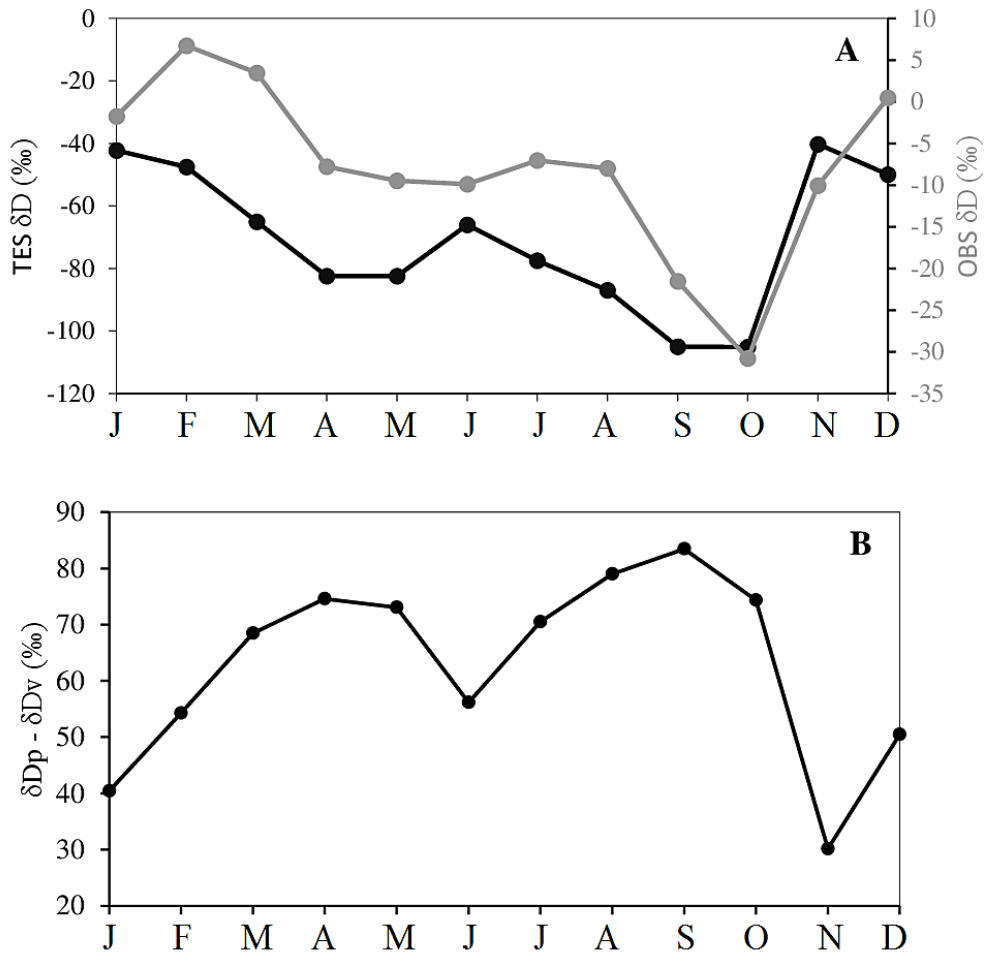


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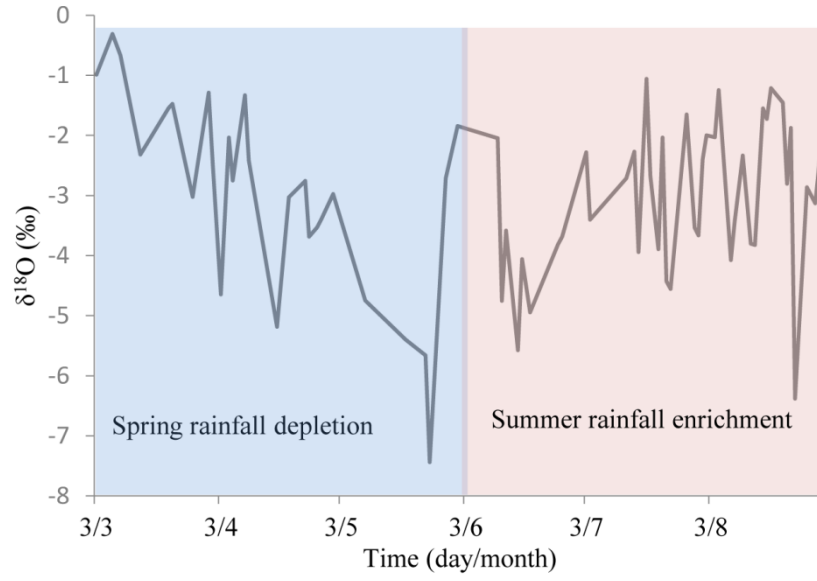


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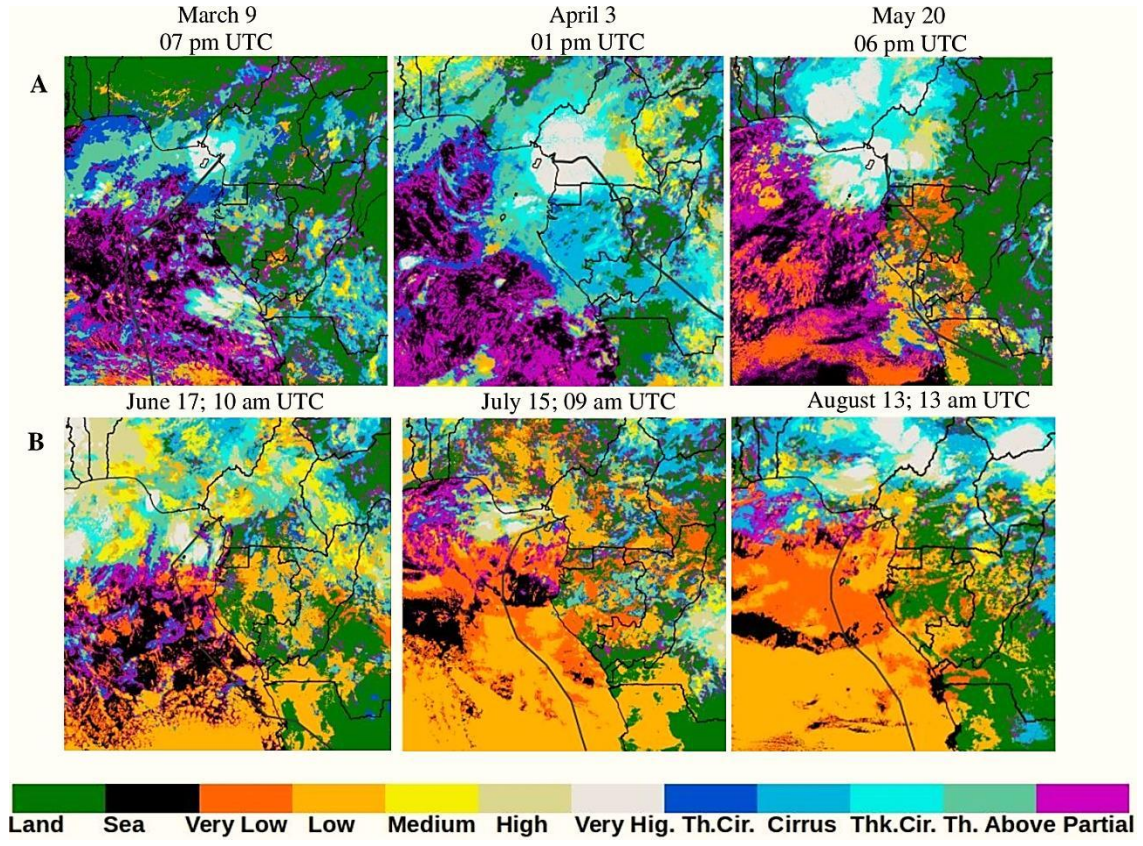


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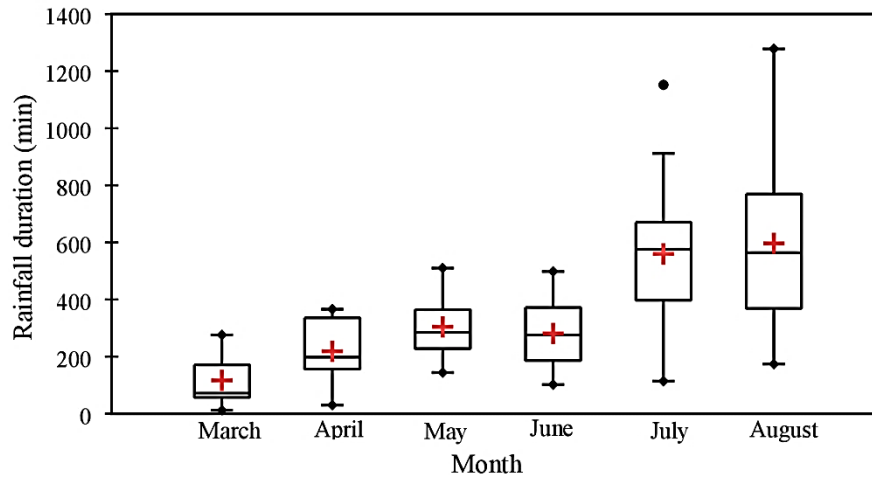


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Table 1. Characteristics of GNIP sampling sites in GOG and Central African regions. Data from Cotonou, Bangui and Sao-Tome have been obtained online from the GNIP database (AIEA/OMM 2018).

Site	Country	Sampling period	Elevation (m.a.s.l)	Mean annual precip*(mm)	Vapor Pressure (hPa)	Mean annual T* (°C)
Cotonou	Benin	2005 - 2012	14	1395	28.3	27.1
Bangui	CAR*	2009 - 2015	363	1208	25.7	26.3
Douala	Cameroon	2006 - 2016	18	3720	30.5	27
Sao-Tome	Sao-Tome and Principe	1962- 1976	8	933.7	26.2	25.2

* Precip. = Precipitation, T = temperatures; CAR= Central African Republic; dec. deg = decimal degrees

Table 2. Coefficient of determination (r^2) between $\delta^{18}\text{O}$ and local meteorological settings (precipitation (P) and temperature (T)) for monthly GNIP stations of Cotonou, Bangui, Douala and Sao-Tome, at different sampling periods.

Stations	Sampling period	$r^2 \delta^{18}\text{O}/\text{P}$	$r^2 \delta^{18}\text{O}/\text{T}$
Cotonou (n=95)	2005 - 2012	0.05	0.13
Bangui (n=81)	2009 - 2016	0.24	0.04
Douala (n=106)	2006 - 2016	0.12	0.09
Sao-Tome (n=123)	1962 - 1976	0.19	0.20

Table3. Coefficient of determination (r^2) between $\delta^{18}\text{O}$ and average GPCP precipitation in average over the first day, two first days and three first days along the air parcel trajectory before rainfall in Douala. Correlation coefficients $\geq 95\%$ are highlighted in bold and underlined.

		March (n=08)	April (n=12)	May (n=07)	June (n=09)	July (n=14)	August (n=20)
$\delta^{18}\text{O}$ Vs GPCP Precipitation	One day before	<u>0.53</u>	<u>0.43</u>	0.40	0.03	0.14	0.15
	Two days before	<u>0.55</u>	0.23	0.28	0.14	0.03	<u>0.23</u>
	Three days before	<u>0.64</u>	0.16	0.44	0.00	0.11	<u>0.23</u>

Table 4. F_{land} statistics during the last 3 days along the air parcels trajectories for the months of March, April and May. Relationship (r^2) between F_{land} , $\delta^{18}\text{O}$, and upstream GPCP precipitation. For the summer period, F_{land} corresponds to 0% for almost all rainy days. Correlation coefficients $\geq 95\%$ are in bold and underlined.

	March	April	May
F_{land} (%)	0	0	25
<i>Minimum</i>			
F_{land} (%)	100	100	100
<i>Maximum</i>			
F_{land} (%)	25	69.45	70.83
<i>Mean</i>			
F_{land} (%)	0	83.34	79.17
<i>Median</i>			
r^2 (F_{land} vs $\delta^{18}\text{O}$)	0.23	<u>0.33</u>	0.28
r^2 (F_{land} vs GPCP precip.)	0.51	0.42	0.85

Table 5. Information on the organization of convection (cloud types, cloud area and local/upstream) and its influence on the isotopic composition of daily rainfall over the period from March 2017 to August 2017. Cloud areas were only calculated for very high classes. Correlation coefficients significant $\geq 95\%$ are in bold and underlined.

	March	April	May	June	July	August
Very low to low clouds (%)	0	0	0	0	71.43	54
Medium to high clouds (%)	0	0	18.18	72.73	21.43	40.5
Very high clouds (%)	100	100	81.82	27.27	7.14	5.5
Local without upstream convection (%)	75	27.27	0	22.29	7.14	0
Upstream and local convection (%)	25	72.73	100	77.71	92.86	100
Minimum cloud area (km ²)	5300 ± 100	4500 ± 100	63400 ± 100	2400 ± 100		-
Maximum cloud area (km ²)	35300 ± 100	158100 ± 100	285600 ± 100	280000 ± 100	3077 ± 100	615 ± 100
Mean cloud area (km ²)	17800 ± 100	73700 ± 100	225100 ± 100	138500 ± 100	-	-
r ² Clouds area Vs rainfall duration	<u>0.54</u>	<u>0.63</u>	<u>0.51</u>	0.34	-	-
r ² Clouds area Vs δ ¹⁸ O	<u>0.44</u>	<u>0.55</u>	<u>0.97</u>	<u>0.90</u>	-	-
r ² Clouds area Vs GPCP precipitation	0.30**	0.10**	0.01**	0.46**		
r ² Rainfall duration Vs δ ¹⁸ O	<u>0.44</u>	<u>0.54</u>	<u>0.64</u>	0.24	0.00	0.07

**Anti-correlation

Cover letter to Editor for revised manuscript

Ref: HYDROL31370

Bertil NLEND

Université de Bourgogne Franche-Comté
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December 09, 2019

Dear Editor,

Thank you for the opportunity to revise our manuscript entitled *Identification of processes that control the stable isotope composition of rainwater in the humid tropical West-Central Africa*. We appreciate your interest, the careful review and constructive suggestions made by reviewers. It is our belief that the manuscript is substantially improved after making the suggested edits.

Following this letter, are the reviewers comments with our responses in blue colour, including how and where the text was modified. The revision has been developed in consultation with all co-authors, and each author has given approval to the final form of this revision. The changes or/and added text are marked in the text by a yellow highlight.

We very much hope that the revised manuscript will be accepted for publication in Journal of Hydrology.

Thank you for your consideration.

Sincerely yours,

Bertil NLEND

on behalf of the authors

Response to Revision Request

The authors would like to thank all the reviewers for their comments and suggestions for improvements which helped us to propose a new corrected version of the manuscript.

Here below is a point to point answer to comments.

Reviewer #2

1- I have some reservation about the explanation on the intra-seasonal time scale. For example, the authors have used the TES derived hydrogen vapor isotope data and compared their result of precipitation $\delta^2\text{H}$ variability. Though the absolute values of these two different datasets are not comparable as correctly mentioned by the authors, however a qualitative match is expected. According to Fig. 7 the seasonal characters are similar, but a considerable mismatch is observed during the later period of the year. For example, from Jan to Sep the precipitation isotope maintains a more or less constant positive offset relative to the vapor isotopic record, which is expected. But in the month of Oct the offset nearly vanishes and surprisingly in the month of Nov the offset value turns negative! This behavior defies the principle of isotopic fractionation. It may so happen that within the limits of analytical uncertainties, this negative offset is not significantly different from zero. But this aspect needs to be discussed in detail, especially why the offset values differ consistently in the months of Oct and Nov. Which atmospheric processes drive such kind of anomalous behavior?

In Figure 7, δD_V and δD_P do not have the same y-axis. For instance, in October (November) δD_V is approximately equal to -100‰ (-45‰) while δD_P equal to -30‰ (-10‰). So, there is always a positive offset between the isotopic composition of water vapor and precipitation despite the fact that we observe on Figure 7 that the two curves are bringing together. Moreover, once again, it is clearly mentioned in the text that we can't compare absolute values obtained from TES with observed precipitation isotopes. Concerning the processes involved in the variability of δD_V and δD_P , on Figure 7, the seasonal variability of $\delta\text{D}_\text{P} - \delta\text{D}_\text{V}$ provides information on rainwater – water vapor interactions and it is demonstrated that post-condensational effects are negligible.

However, regarding this comment from the reviewer, to avoid misunderstanding, **we added in the title of Figure 7 a precision concerning the y-axis and precision in the text (lines 364-365) about the positive offset between δD_V and δD_P .**

Other issues

2- Line: 372: It is known that the process of condensation is an equilibrium process; if so, then why 'there is no reason for the precipitation to be in equilibrium with the vapor'?

We removed this sentence to avoid misunderstanding and confusion.

3- Line 376: Pls provide a complete description of the relationship between dD_p and dD_v . It has been reported that the slope for this kind of relation is closer to 1 (Conroy et al. 2016). But, the authors get a very high value of slope, 1.68. The reason for such a large deviation should be explained. In this context, it is also suggested to use the q-dD diagram (Conroy et al. 2016) in order to characterize the moisture sources.

Many thanks for bringing this to our attention and for the reference which has been taken in account (see lines 378 – 379). **We corrected it** and putted now the right value of the slope (see line 375).

4- In order to study the effect of continental recycling, the authors have calculated "the percentage of time of the air parcel over the continent during the last 3 days along the trajectory" (Line 404-405), termed as F_{land} . Two issues are important in this context. Firstly, the method of calculation of F_{land} is not provided. Secondly, and more importantly, at what height the trajectories have been calculated? If the trajectories have been calculated for a single height, then they do not necessarily represent a major amount of moisture transport to the sampling site, since moisture is typically transported at a spectrum of heights. A detailed quantitative calculation of moisture transport for different heights is expected to be presented here. Without this exercise the conclusion of 'poor continental recycling' could be misleading.

The method of calculation of F_{land} was added in the text; see lines 405-408. About the second issue, we thank the reviewer for his comment which helped us to improve this section of the manuscript. Indeed, **we calculated air back trajectories at different heights: 800, 850, 900, 950 and 980 hPa**. We have not performed investigation at more than 800 hPa because it is the boundary layer that feeds the convection and it is with the water vapor from the low layers that the rain is re-evaporated. By calculating trajectories at these different heights, we observed that there is no difference between them, especially since we focus only on air back trajectory at 3 days before Douala. Therefore, our conclusion of "poor continental recycling" in the study region can be considered as correct. **Detailed precision on this issue were added in section 4.3.2 (see lines 408 to 414).**

5- Line 457: how do you define the parameter "cloud surface"? How it is measured? How do you take care of the uncertainties in measurement?

Information concerning this is well detailed in section 3.3, line 238 and lines 240-241. Nevertheless thanks to this comment, to ensure the methodology described in the text is more understandable and robust, **we added more precision in lines 238, 240-241.**

6- One of the highlights "Local site precipitation $\Delta^{18}O$ of Douala was found to represent large-scale meteorology" may be confusing. I suggest rewording as 'Local site precipitation $\Delta^{18}O$ of Douala appears to respond to large-scale metrology".

Done.

7- The title of the MS is slightly misleading. Use of the word "and" in 'Western and Central Africa' may mean that the study sites are situated at both the western and central Africa. It

may be slightly modified as "Identification of processes that control the stable isotope composition of rainwater in the humid tropical West-Central Africa".

Done. Thank you for this suggestion.

Minor issues

8- Line 179: Ratios of d18O/d16O or d2D/D1H is not technically correct. Better use "ratios of 18O/16O or 2H/1H".

Done.

9- Line 212: Lekshmy et al. 2014 used TRMM derived rainfall data, not the GPCP products.

Corrected as suggested.

10- Line 270: Pls replace 'delta' by 'precipitation isotopes'.

Done. It is now line 269.

11- The MS is full of grammatical mistakes; use of the article is poorly done. I have mentioned some of them, but this is only an indicative list. It is suggested to get the MS examined by a native English speaker, or take the help of a professional service provider on the English language.

Done. Syntax errors have been corrected. In general, the English of the document has been improved and we thank the reviewer who encouraged us.

12- Line 282: 'regional scale' should be followed by "on a".

Done. It is now line 281

13- Line 296-97: please check for typos in the citations.

Done.

14- Line 311: English problem: "trajectories come..." is awkward. Trajectories are not object, but some kind of attribute. In this context, it may be replaced by moisture. Similar kind of errors is there in Line 313, 316, 318.

Corrected.

15- Line: 318: typo

Corrected.

16 - Line 321: "monsoon winds become less influent.." is not a good English. May be written as "strength of monsoon winds weaken.

Done.

17 - Line 327: The wording like "GPCP precipitation increases" may be replaced by 'the GPCP precipitation data show an increasing trend...!.

Done.

18 - Line 333: 'rainfall is most depleted' may mean that rainfall intensity is less. Better write "isotopic value of rainfall is depleted in spring....!.

Corrected as suggested. It is now line 332.

19 - Line 358: 'given' should be replaced by "provided".

Done. It is now line 357.

20 - Line 360: trajectories should be followed by 'the'.

Done. It is now line 359

Reviewer #3

1- Please avoid lumped references. Please talk about the specific points of each reference.

This remark has been considered and we would like to reassure the editors and the reviewers that each reference in the manuscript concerns a specific point or aspect.

2- How is the reproducibility of $\delta^{18}O$ seasonal cycle from year to year?

This fact is clearly detailed and explained in section 4.1. The Figure 2 also discuss about it.

3- what is the correlation between monsoon and the transition to the dry season?

General comments on climate dynamic in the study area are given in section 2. There is no correlation between monsoon and the transition toward the dry season period, except that the second occurs when the first one ends. The monsoon is a SW wet wind which is most predominant at Douala from April to August. As the ITCZ begins its southward withdrawal in September until November the monsoon wind become less efficient, giving way to the Harmattan (NE dry wind) and the transition toward the dry period often correspond to the end of October – November. This is stated in section 2 and in section 4.2.2 by looking the pattern of air back trajectories.

Reviewer #4: Manuscript ID: HYDROL31370

Manuscript Title: Identification of processes that control the stable isotope composition of rainwater in the humid tropical Western and Central Africa

The paper is in order and can be recommended for publishing in present format.

Thank you very much for your positive comment.

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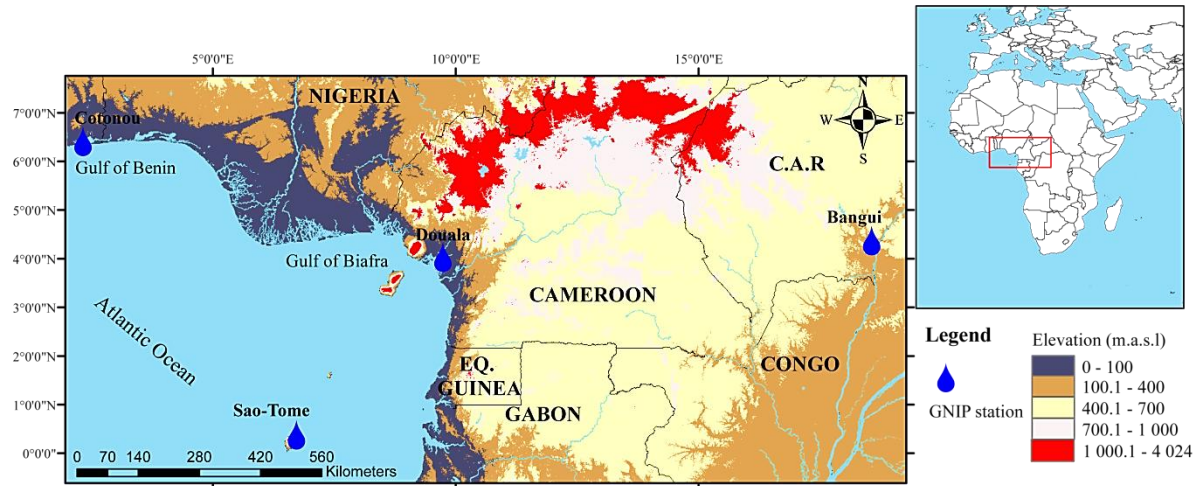


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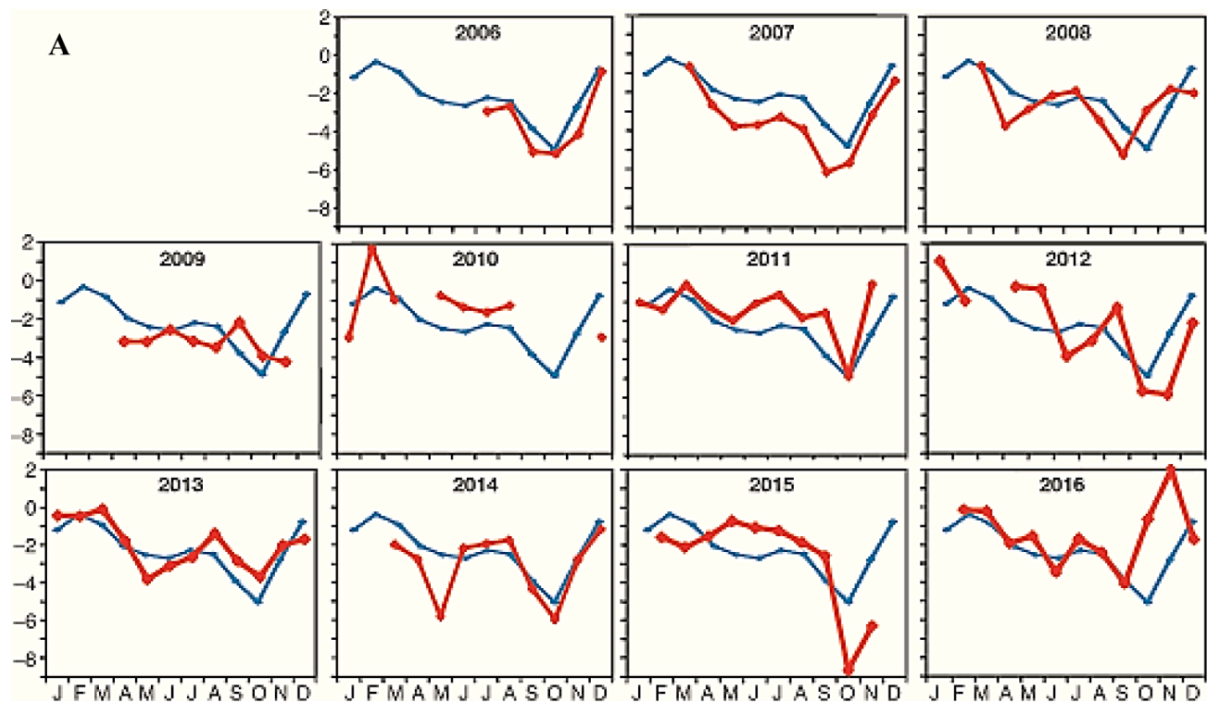


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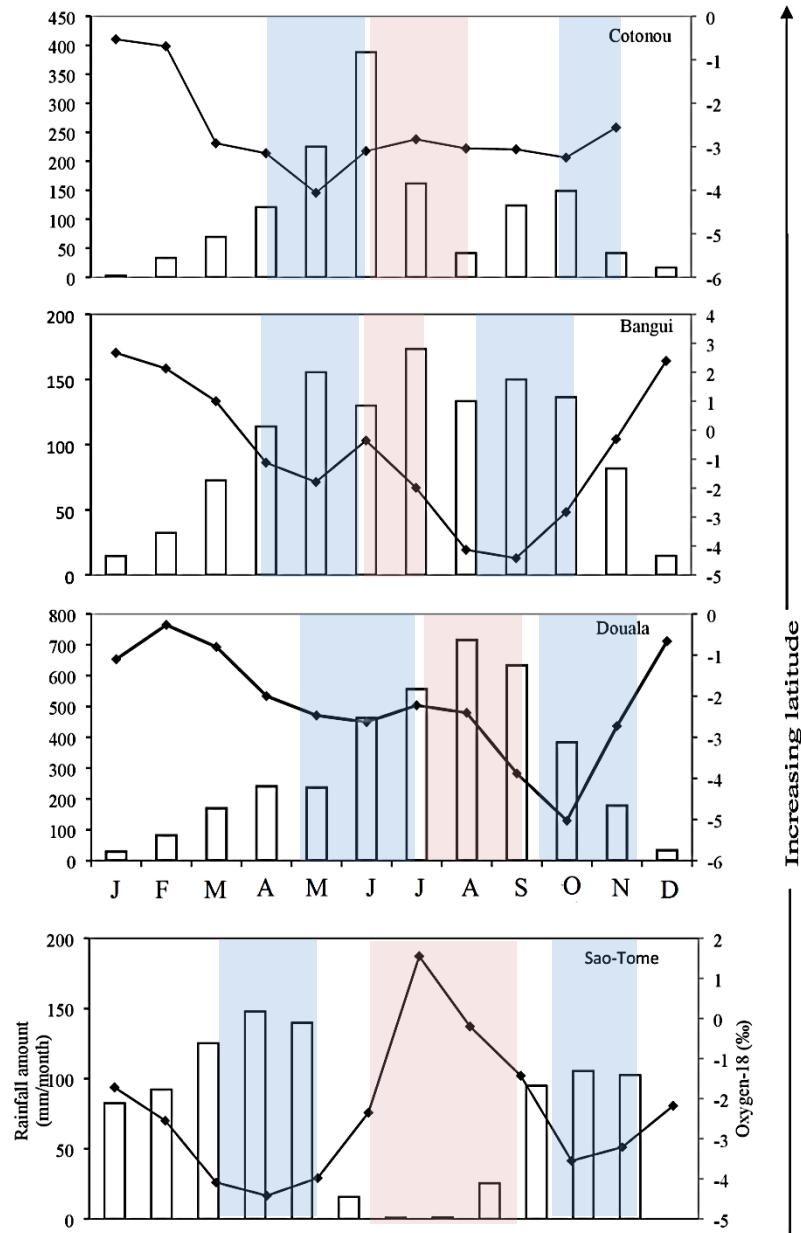


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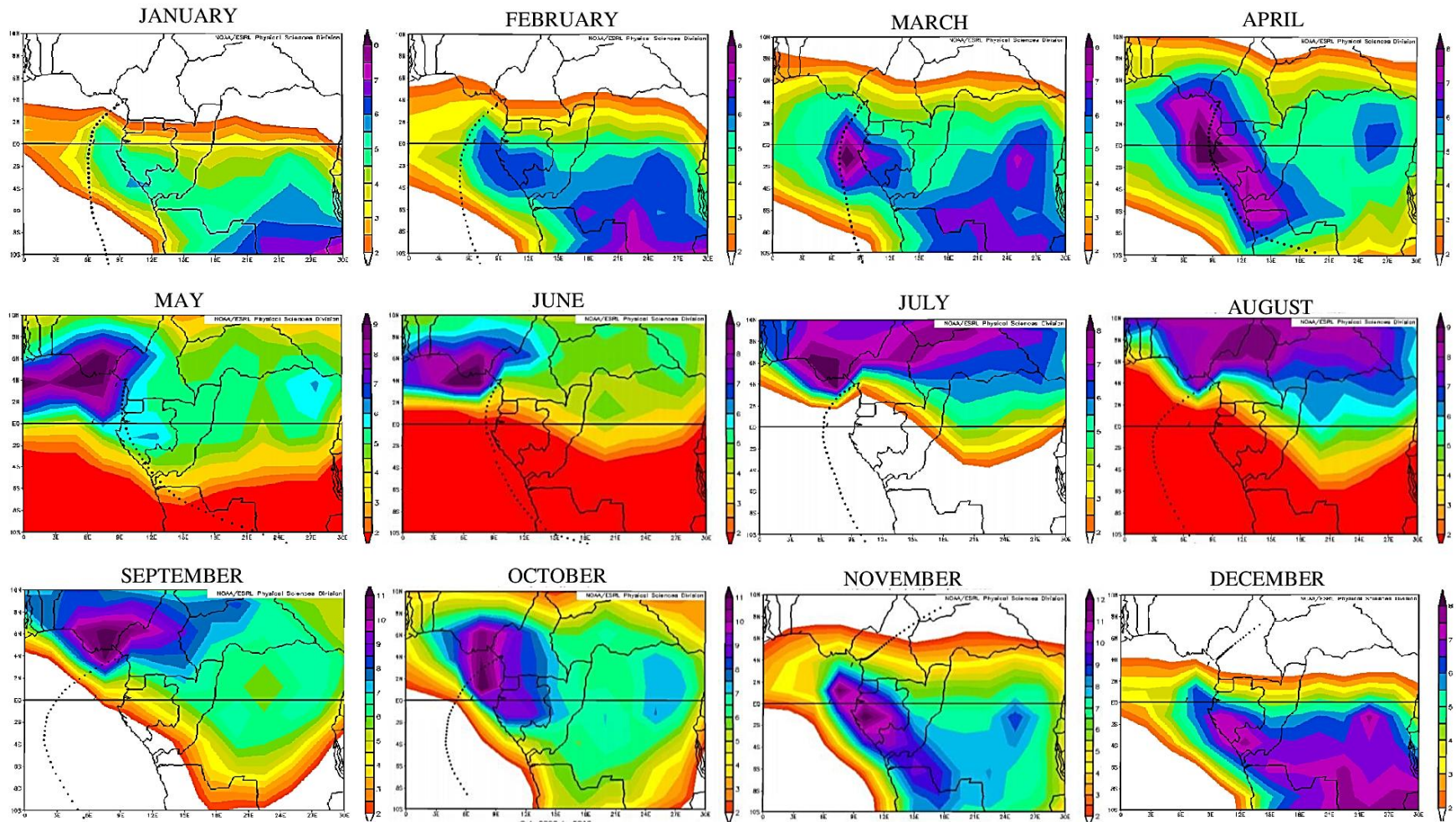


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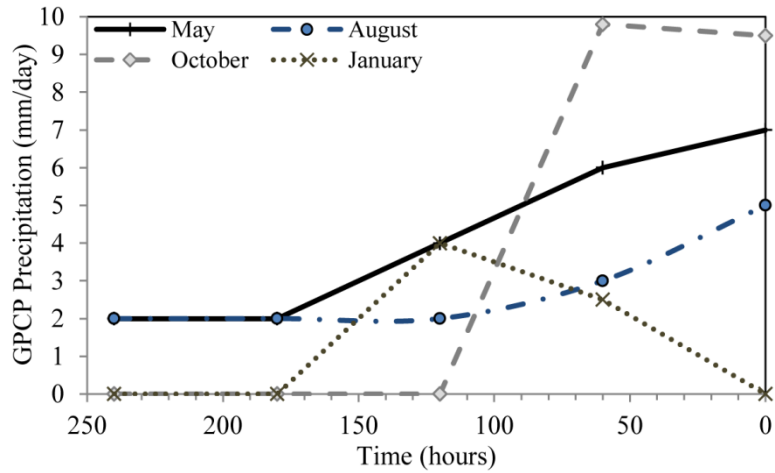


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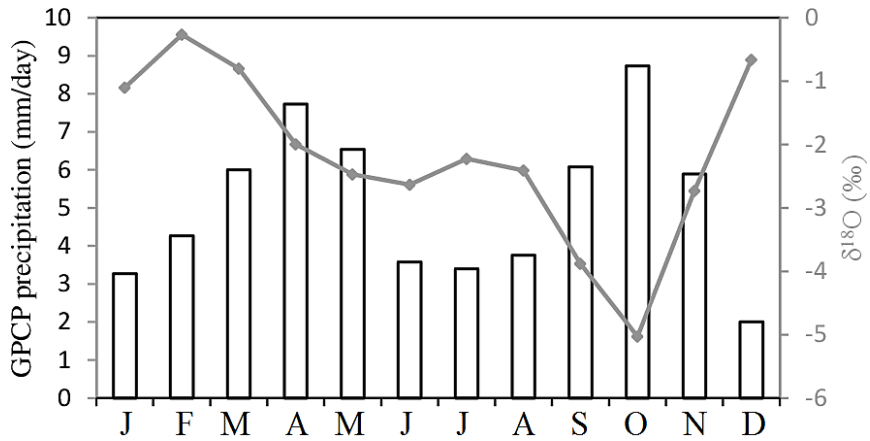


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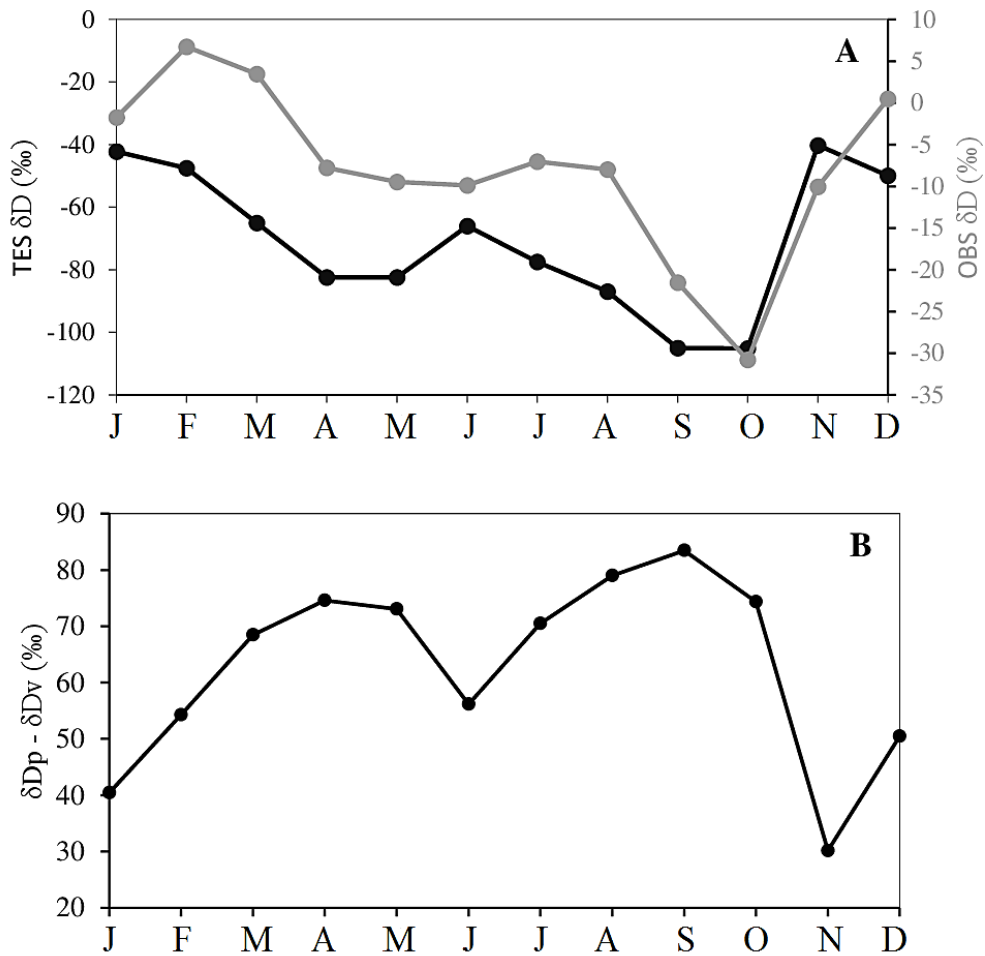


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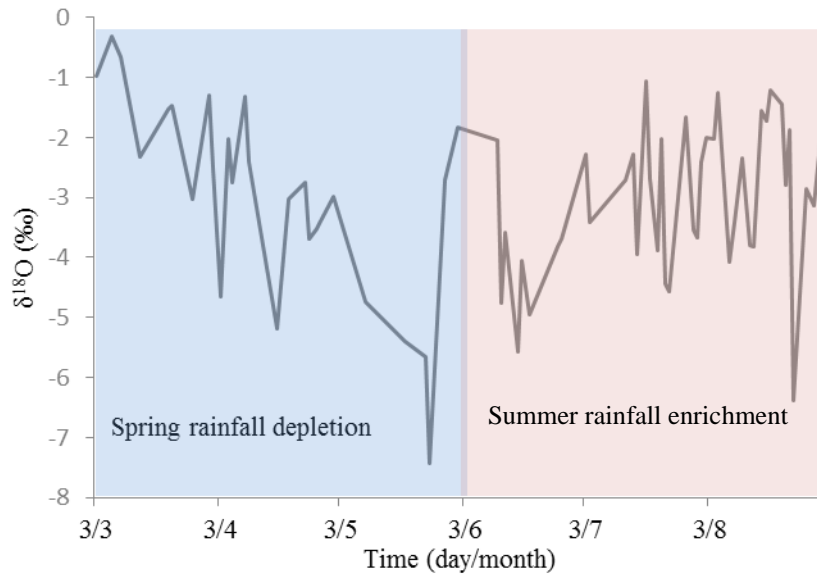


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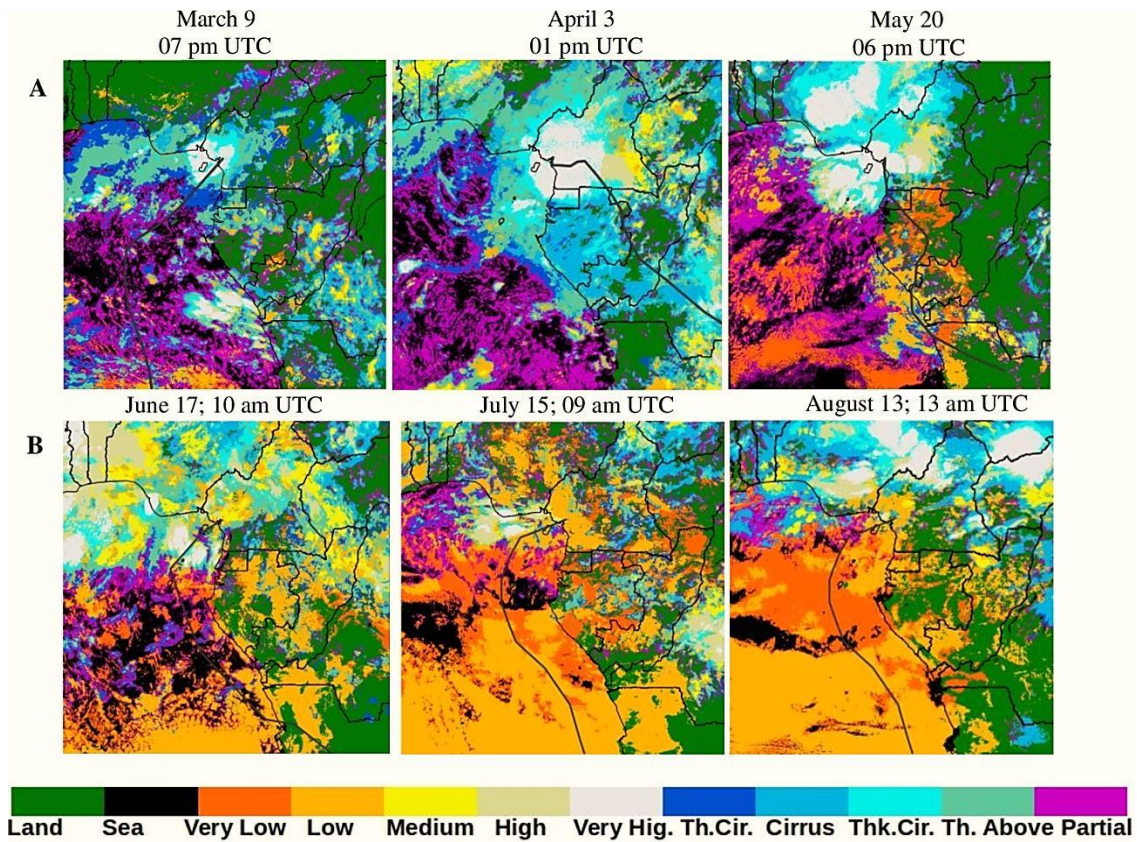


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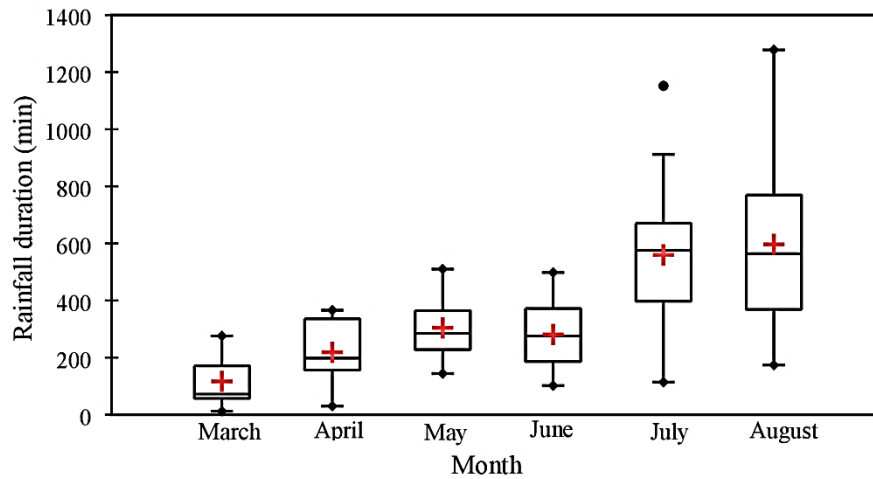


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Table 1. Characteristics of GNIP sampling sites in GOG and Central African regions. Data from Cotonou, Bangui and Sao-Tome have been obtained online from the GNIP database (AIEA/OMM 2018).

Site	Country	Sampling period	Elevation (m.a.s.l)	Mean annual precip*(mm)	Vapor Pressure (hPa)	Mean annual T* (°C)
Cotonou	Benin	2005 - 2012	14	1395	28.3	27.1
Bangui	CAR*	2009 - 2015	363	1208	25.7	26.3
Douala	Cameroon	2006 - 2016	18	3720	30.5	27
Sao-Tome	Sao-Tome and Principe	1962- 1976	8	933.7	26.2	25.2

* Precip. = Precipitation, T = temperatures; CAR= Central African Republic; dec. deg = decimal degrees

Table 2. Coefficient of determination (r^2) between $\delta^{18}\text{O}$ and local meteorological settings (precipitation (P) and temperature (T)) for monthly GNIP stations of Cotonou, Bangui, Douala and Sao-Tome, at different sampling periods.

Stations	Sampling period	$r^2 \delta^{18}\text{O}/\text{P}$	$r^2 \delta^{18}\text{O}/\text{T}$
Cotonou (n=95)	2005 - 2012	0.05	0.13
Bangui (n=81)	2009 - 2016	0.24	0.04
Douala (n=106)	2006 - 2016	0.12	0.09
Sao-Tome (n=123)	1962 - 1976	0.19	0.20

Table3. Coefficient of determination (r^2) between $\delta^{18}\text{O}$ and average GPCP precipitation in average over the first day, two first days and three first days along the air parcel trajectory before rainfall in Douala. Correlation coefficients $\geq 95\%$ are highlighted in bold and underlined.

		March (n=08)	April (n=12)	May (n=07)	June (n=09)	July (n=14)	August (n=20)
$\delta^{18}\text{O}$ Vs GPCP Precipitation	One day before	<u>0.53</u>	<u>0.43</u>	0.40	0.03	0.14	0.15
	Two days before	<u>0.55</u>	0.23	0.28	0.14	0.03	<u>0.23</u>
	Three days before	<u>0.64</u>	0.16	0.44	0.00	0.11	<u>0.23</u>

Table 4. F_{land} statistics during the last 3 days along the air parcels trajectories for the months of March, April and May. Relationship (r^2) between F_{land} , $\delta^{18}\text{O}$, and upstream GPCP precipitation. For the summer period, F_{land} corresponds to 0% for almost all rainy days. Correlation coefficients $\geq 95\%$ are in bold and underlined.

	March	April	May
F_{land} (%)	0	0	25
<i>Minimum</i>			
F_{land} (%)	100	100	100
<i>Maximum</i>			
F_{land} (%)	25	69.45	70.83
<i>Mean</i>			
F_{land} (%)	0	83.34	79.17
<i>Median</i>			
r^2 (F_{land} vs $\delta^{18}\text{O}$)	0.23	<u>0.33</u>	0.28
r^2 (F_{land} vs GPCP precip.)	0.51	0.42	0.85

Table 5. Information on the organization of convection (cloud types, cloud area and local/upstream) and its influence on the isotopic composition of daily rainfall over the period from March 2017 to August 2017. Cloud areas were only calculated for very high classes. Correlation coefficients significant $\geq 95\%$ are in bold and underlined.

	March	April	May	June	July	August
Very low to low clouds (%)	0	0	0	0	71.43	54
Medium to high clouds (%)	0	0	18.18	72.73	21.43	40.5
Very high clouds (%)	100	100	81.82	27.27	7.14	5.5
Local without upstream convection (%)	75	27.27	0	22.29	7.14	0
Upstream and local convection (%)	25	72.73	100	77.71	92.86	100
Minimum cloud area (km²)	5300 \pm 100	4500 \pm 100	63400 \pm 100	2400 \pm 100		-
Maximum cloud area (km²)	35300 \pm 100	158100 \pm 100	285600 \pm 100	280000 \pm 100	3077 \pm 100	615 \pm 100
Mean cloud area (km²)	17800 \pm 100	73700 \pm 100	225100 \pm 100	138500 \pm 100	-	-
r² Clouds area Vs rainfall duration	<u>0.54</u>	<u>0.63</u>	<u>0.51</u>	0.34	-	-
r² Clouds area Vs $\delta^{18}\text{O}$	<u>0.44</u>	<u>0.55</u>	<u>0.97</u>	<u>0.90</u>	-	-
r² Clouds area Vs GPCP precipitation	0.30**	0.10**	0.01**	0.46**		
r² Rainfall duration Vs $\delta^{18}\text{O}$	<u>0.44</u>	<u>0.54</u>	<u>0.64</u>	0.24	0.00	0.07

**Anti-correlation

1 **Identification of processes that control the stable isotope composition of**
2 **rainwater in the humid tropical West-Central Africa**

3

4

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33

34 **Abstract**

35

36 This study interprets 11 years (2006 to 2016) and 6 months (March to August in 2017) of
37 respectively monthly and daily isotopic (δD and $\delta^{18}O$) monitoring of rain at Douala
38 (Cameroon), a humid tropical station in Western Africa. The main scope is to analyze the
39 climate controls on precipitation isotopes at different timescales. Firstly, we examine the
40 annual cycles of $\delta^{18}O$. Over the 11 years of survey, the annual cycle exhibits a W shape that is
41 quite reproducible from year to year, with two minima in spring and autumn periods. Based
42 on back trajectory calculations and remote sensing observations of water vapor isotopic
43 composition, we show that the observed depletion in spring and autumn is due to strong
44 convective activity along air mass trajectories. The same effect of convective activity can be
45 observed at the daily timescale. At seasonal and daily time scales, the isotopic composition is
46 also strongly tied to the convective organization and cloud types. More depleted precipitation
47 is associated with larger areas of high clouds. Very low to low clouds are observed in July-
48 August, mid-level to high clouds are dominant in June and high to very high clouds
49 characterize March-April-May, thus explaining the enriched (depleted) values in summer
50 (spring). Finally, this paper highlights the importance of large scale meteorological conditions
51 controls on precipitation stable isotope composition in the Gulf of Guinea.

52

53 Keywords: $\delta^{18}O$, water vapor, convective activity, GPCP precipitation, air back trajectory.

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59 **1. Introduction**

60 The atmosphere is an essential hydrological environment. It contains all the water vapor
61 (0.001% of all the water of the Earth; Delmas et al. 2005) that forms the clouds by
62 condensation which can then generate precipitations (liquid/solid) as a function of air
63 temperature. The atmosphere is also an essential place of transfer and exchange for the global
64 functioning of the Earth system. Its multiple interactions with the oceans, the continent and
65 the biosphere make it an important study environment for understanding global changes.
66 According to IPCC (1998), these global changes must affect tropical regions which are among
67 the most vulnerable to possible anthropogenically induced climatic changes.

68 During the last decade, the scientific community, environmental institutions, governments,
69 and local communities have increased their awareness of the importance of current tropical
70 climate variability, based on the premise that, under increasing anthropogenic influence on the
71 climate, changes in regional and global circulation may lead to an intensification of extreme
72 events (i.e. floods or severe droughts: e.g., Pohl et al. 2017). In parallel, the use of stable
73 isotopes of water, both δD and $\delta^{18}O$ has provided insights in the study of atmospheric water
74 cycle (e.g., Rozanski et al. 1993, Araguas - Araguas et al. 2000, Celle-Jeanton et al. 2004); the
75 key to understand the future climate changes or global changes. Relationship between δD and
76 $\delta^{18}O$ in natural meteoric waters serves as a foundational reference to determine regional and
77 local deviations from equilibrium processes and the potential origin of the water vapor.
78 Moreover, water losses due to evaporation, incorporation of recycled atmospheric moisture,
79 and/or mixing between isotopically distinct reservoirs leave a unique water fingerprint that
80 can be used for climate-reconstructions (Moerman et al. 2013). Recently, isotopic
81 composition of tropical meteoric water has also proven its usefulness as an indicator of
82 modern climate variability (Vuille and Werner 2005; Ishizaki et al. 2012; Sanchez-Murillo
83 2015).

84 While in extra-tropical climates, stable isotope variations in meteoric
85 waters have been successfully explained by air temperature variability (Dansgaard 1964;
86 Rozanski et al. 1993), the case of tropical humid regions proved to be very much complex
87 since temperature variability is much weaker. The amount effect occurs as a result of
88 convective precipitation and can be nonlocal (Vimeux et al. 2005; He et al. 2015). The degree
89 of organization of convective systems has been shown to impact the atmospheric conditions at
90 the large-scale (Tobin et al. 2012; Wing et al. 2017). Consistently, it also impacts the isotopic
91 composition of water vapor and precipitation. Many additional factors can potentially play a
92 role, such as orographic effects, continental recycling or moisture origin combined with
93 complex microclimates (Rozanski et al. 1993, Lachniet and Paterson 2009).

94 These problematics are of major importance for the paper. What are the key factors
95 controlling stable isotope ratios of meteoric waters in humid tropical areas of West and
96 Central Africa? Answers to this question will contribute to the understanding of atmospheric
97 processes in the study region. Unlike arid and semi-arid African areas, where isotopic
98 variability has already been assessed in previous work (e.g; Taupin et al. 1997, Celle-Jeanton
99 et al. 2001; Risi et al. 2008b, 2010; Lutz et al. 2011, Tremoy 2012; Tremoy et al. 2014), the
100 Gulf of Guinea (GOG) region and tropical humid areas in Central Africa (Figure 1) are still
101 under-documented in isotopic data. There, the atmospheric cycle of water remains poorly
102 documented and isotopic data may help improve their knowledge. Indirectly, the paper will
103 help to understand rainfall variability in Douala and the origin of precipitated water through
104 isotopic analysis. In this objective, we took advantage of the long term monthly monitoring
105 set up in Douala, Cameroon (Figure 1) as part of the Global Network for Isotopes in
106 Precipitation (GNIP) framework (IAEA/WMO 2018) from 2006. In parallel, a daily sampling
107 has been carried out in 2017. This article represents the first valuation of precipitation isotope
108 data in the region. Datasets of other GNIP stations in the GOG and Central Africa (Figure 1)

109 are also integrated in this study, together with remote sensing data that document water vapor
110 isotopic composition, convective activity and cloud properties.

111

112 Figure 1.

113

114 **2. Regional climate: the West African Monsoon (WAM) dynamic**

115 Seasonality in the tropics is mostly determined by the seasonal migrations of the Inter
116 Tropical Convergence Zone (ITCZ, Preston-White and Tyson 1988; Schott et al. 2003), a
117 highly energetic feature of earth climate that is associated with deep convection. The ITCZ or
118 meteorological equator in Africa is the result of convergence between the Harmattan
119 (northeasterly dry wind) and Monsoon (southwesterly wet wind) in low levels of the
120 atmosphere (Sultan and Janicot 2000, Fink et al. 2018). Nicholson and Grist (2003) have
121 shown that the rain belt over Western and Central Africa is positively correlated with the
122 migration of the ITCZ. During the boreal spring (Mar-Apr-May), ITCZ holds a position of
123 about 5°N; half of the rain belt is located over the continent and the other one on the ocean
124 (Waliser and Gautier 1993). The mean rainfall amount calculated for this period, according to
125 GNIP measurements, is 232 mm for Douala, 138.4 mm for Cotonou, 137.7 mm for Sao-Tome
126 and 114.1 mm for Bangui. The relative humidity increases significantly over the continent
127 and sea surface temperatures (SST) present their annual peak (>27°C) during this period
128 (Dezfuli and Nicholson 2013). Then, ITCZ abruptly shifts from 5°N to 10°N in June-July
129 (Sultan and Janicot 2003). The so-called monsoon jump causes a subsidence in the southern
130 part of the GOG and the development of a cold tongue complex (Gu and Adler 2004). Low
131 SSTs suppress then partly or totally rainfall as e.g. in Sao-Tome (Figure 3). However, in the

132 Gulf of Biafra (Figure 1), SSTs remain high enough ($>26^{\circ}\text{C}$) to partly maintain convection
133 (Odekunle and Eludoyin 2008). Moisture increases over land and enhances rainfall there.
134 Simultaneously, convective activity migrates to the North, and the monsoon flow crosses the
135 Equator from South (GOG) to North (Guinean and then Sahelian Africa). At this time of the
136 year (Jun-Jul-Aug), a rainfall peak is observed (Figure 3) in Douala (618.3 mm) and Bangui
137 (145.5 mm). In Cotonou, Ogu et al. (2016) have also shown a good correlation between SSTs
138 and precipitation, connecting rainfall amount in June to anomalously warm waters of the Gulf
139 of Benin.

140 The abruptness of the northward progression of the ITCZ is in sharp contrast to its
141 withdrawal, which appears as a more progressive southward progression. Like most regions in
142 the equatorial latitudes, the Guinean Coast/Central Africa thus benefits from a double passage
143 of the ITCZ in spring and autumn.

144 The unimodal distribution and the highest rainfall amount (~ 4000 mm/year) observed in
145 Douala (Figure 3) is very particular as the whole region is generally characterized by a
146 bimodal distribution with two precipitation maxima in April-May (or May-June) and October-
147 November (or September-October), with a total amount lower than 2000 mm/year. The
148 concave shape of the Gulf of Biafra may induce a convergence of southwesterly winds and
149 topographic ascents forced by the relief of Mount Cameroon (> 4000 m, Figure. 1). This
150 mesoscale convergence and orographic influence (Vondou et al. 2017) are all the more
151 marked if the monsoon flow is intense and thick, which is the case in the core of the summer
152 monsoon, when the ITCZ is far to the North. These regional features can explain the
153 unimodal regime observed at Douala while further west (in Cotonou), and in southern
154 hemisphere (Sao-Tome), the annual cycle is bimodal.

155 **3. Data and methods**

156 ***3.1. Rainfall sampling and isotopes analyses***

157 This paper makes use of the 106 monthly samples of rainwater that have been collected from
158 July 2006 to December 2016 at the GNIP station of Douala (see details of the GNIP stations
159 in [Table 1](#) and [Figure 1](#)). The sampling followed the standard protocols ([IAEA 2012](#)).
160 Samples were collected by using a rain gauge which consists of a plastic funnel (diameter =
161 10 cm) coupled with a filter mesh to prevent contaminations by debris. A 5 cm layer of
162 mineral oil has been systematically added into the rain collector to avoid fractionation of the
163 collected rainwater. For each month, total rainfall was collected. Samples were taken
164 regularly, stored in a totalizer and kept at 4°C before being transferred in 50 ml amber glass
165 bottles, tightly capped and sent to the International Atomic Energy Agency (IAEA) laboratory
166 in Vienna, Austria for stable isotopes determinations. Stable isotopes of hydrogen and oxygen
167 were then analyzed by laser absorption spectroscopy following the method described by
168 [Penna et al. \(2010\)](#).

169 **Table 1**

170
171 The rainwater daily survey was conducted from March to August 2017 at the campus of the
172 University of Douala (X= 9.7461; Y= 4.062; Z= 17 m.a.s.l.), approximately at 7 km from the
173 GNIP station. Seventy samples were collected using a Palmex rain gauge that presents the
174 advantage to avoid evaporation without using medicinal paraffin oil ([Gröning et al. 2012](#)).
175 Daily samples were stored using the same protocols as for monthly rainfall. Isotopes analyses
176 were performed at the Hydrogeology Department of the University of Corsica, France, by

177 using a liquid-water stable isotope analyzer DLT-100 Los Gatos Research (Aggarwal et al.
178 2006; Penna et al. 2010).

179 Ratios of $^{18}\text{O}/\delta^{16}\text{O}$ and $^2\text{H}/^1\text{H}$ are expressed in delta units (‰, parts per mil) relative to Vienna
180 Standard Mean Ocean Water (V-SMOW). The analytical precision is $\pm 0.1\text{‰}$ for oxygen-18
181 and $\pm 1\text{‰}$ for deuterium.

182 Local meteorological settings (precipitation amount and duration, air temperatures, vapor
183 pressure) were provided by the National Weather Direction of Cameroon.

184

185 ***3.2. Tropospheric Emission Spectrometer (TES) data***

186 With the advent of new technology in stable water isotopes, such as spectrometer on-board
187 satellite, it has become easier to analyze isotopic composition of water vapor based on
188 indirect laser measurement (Aemisegger et al. 2012). TES instrument on board on the Aura
189 satellite is a nadir-viewing infrared Fourier transform spectrometer from which the deuterium
190 content of water vapor (δD_v) can be retrieved (Worden et al. 2006; Worden et al 2007). The
191 sensitivity of the retrieval is typically larger between 900 hPa and 400 hPa with a peak at 700
192 hPa. On average, δD_v retrieved over these levels has a precision of 1.5% or about 15 parts per
193 thousand (per mile) relative to Standard Mean Ocean Water. Uncertainties are reduced by
194 averaging several measurements (Worden et al. 2006; Risi et al. 2013). Then the precision is
195 sufficient for characterizing the global distribution of evaporation and condensation processes
196 (Worden et al. 2006).

197 There is on average 1.8 degrees of freedom for δD_v retrievals in the tropics (Worden et al.
198 2012), meaning that vertical profiles bear information on more than one level. To ensure good
199 data quality, we selected only the measurements for which the quality flag is set to unity and
200 for which the degree of freedom of the signal is higher than 0.5 (Risi et al. 2013). Here, to
201 document the water vapor composition as close as possible to the surface, we use the δD

202 values retrieved by TES at 900 hPa from 2004-2008 at a monthly scale and we focus on a
203 multi-year mean seasonal cycle and temporal variations rather than absolute values.

204

205 ***3.3. Convection and cloud datasets***

206 Convective activity associated with the West African Monsoon (WAM) was analyzed using
207 the Global Precipitation Climate Project one degree daily (GPCP-1dd; Huffman et al. 2001)
208 data. The same dataset is used for both monthly and daily analyses. Data were retrieved from
209 the National Oceanic and Atmospheric Administration (NOAA) website
210 (<https://www.esrl.noaa.gov/psd/cgi-bin/data/composites/printpage.pl>). The robustness of
211 GPCP products has been demonstrated in many studies (e.g.; Huffman et al. 1995, 1997;
212 Adler et al. 2017) through a comparison with other proxies of convection or by multi-proxy
213 studies: OLR (Outgoing Longwave Radiation) and TRMM (Tropical Rainfall Measurement
214 Mission). GPCP products are a combination of precipitation data provided by a multiple
215 sources of satellite-gauge (SG). They are obtained by optimally merging precipitation
216 estimates computed from microwave, infrared, sounder data observed by the international
217 constellation of precipitation-related satellites, and rain gauge analyses, taking advantage of
218 the strengths of each data type. Mean SG products are computed by combining multi-satellite
219 estimates with rain gauge analysis (Huffman et al. 2001). In this study, we use the GPCP-1dd
220 (1-degree grid over the entire globe at 1-day) both for monthly and daily analyses.

221 To document the cloud types and convection organization in Western Central Africa, we use a
222 cloud type (CT) product issued from geostationary MSG (Meteosat Second generation)
223 satellite data and developed by SAFNWC (Satellite application facilities in support to
224 nowcasting) /MSG algorithms (for more information see
225 <http://www.nwcsaf.org/web/guest/scientific-documentation>). Clouds types are determined
226 from their top temperature or pressure and, for high level clouds from their opacity (Dommo

227 et al. 2018; Seze 2015). The SAFNWC CT offers a classification of clouds into 12 classes:
228 free land, free sea, very low clouds, low clouds, medium clouds, high clouds, very high
229 clouds, thin cirrus, medium cirrus, thick cirrus, fractional clouds and semi-transparent above
230 low or medium clouds, at 3-km spatial resolution for regions close to MSG sub-satellite point
231 (0N, 0E) and a time step of 15-min. For our study purposes, focus is given to the altitude of
232 clouds (low, medium, high, etc.) and their organization (as measured by the spatial extent of
233 connected cloud pixels) to seek links with the isotopic contents of rainfall. Data are extracted
234 for the region bounded by latitudes 10S–10N and longitudes 0E–20E, for the period from
235 March to June 2017.

236 Based on these images, very high clouds areas (corresponding to very organized convective
237 system) around Douala were calculated for each event using ArcGis mapping software,
238 through the tool “measure”. Indeed, since the contours of cloud masses or pixels are
239 approximated to geometric forms (circle, rectangular, trapezoidal shape, etc.), the areas were
240 calculated based on these forms. This calculation was performed after a zoom of the image at
241 100% in order to limit uncertainties or errors by using the tool of the software. For events
242 with duration of more than 15 min, we have taken the maximum cloud area throughout the
243 duration of rainfall.

244 **3.4. Back trajectories**

245 In order to assess the influence of air masses pathways on the isotopic composition of
246 precipitation, and the importance of the location of convective activity in the region, we
247 compute air back trajectories at 6h time steps, 10 days prior to arrival in the sampling site.
248 This operation was performed for monthly and daily scales when isotopic data is available.
249 Winds (at 900 hPa) were simulated by using the general circulation model LMDZ5A
250 (Hourdin et al. 2013) guided by reanalysis products (ERA-Interim, Dee et al. 2011) of the

251 European Centre for Medium-Range Weather Forecasts (ECMWF). Back trajectories (speed
252 and direction) were then computed with a 2D algorithm (similar to Vimeux et al. 2005) in
253 order to approximate the moisture transport near the ground surface.

254

255 **4. Results and discussion**

256 *4.1. Annual cycles of isotopes (δ) in precipitation of Douala*

257 **Figure 2** presents the seasonal variations of $\delta^{18}\text{O}$ from 2006 to 2016. The annual cycle of δ is
258 quite reproducible from year to year. It exhibits a W shape most of the time, with minima in
259 spring and autumn periods. Because of this rather good reproducibility of $\delta^{18}\text{O}$ seasonal cycle
260 from year to year; we further consider a multi-year mean seasonal cycle in the section 4.2, and
261 investigate the factors associated with this W cycle of δ .

262 However, the magnitude varies widely from one year to another for a given month. Such
263 inter-annual variability could either reflect variations (i) in the climate seasonal background,
264 or (ii) in synoptic / intra-seasonal variability, which could modify seasonal mean fields
265 through upscaling processes, or eventually affect isotopic variations. These hypotheses are
266 tested and discussed below.

267  Figure 2.

268

269 **4.2. Factors responsible for precipitation isotopes variations at seasonal scale**

270 *4.2.1. Highlighting of a regional context*

271 Seasonal variations of $\delta^{18}\text{O}$ and precipitation amount for GNIP stations in GOG and Central
272 Africa region are presented in **Figure 3**.

273

274  Figure 3.

275
276 Similarities can be observed in the seasonal evolution of $\delta^{18}\text{O}$ at Cotonou, Bangui, Douala
277 and Sao-Tome (Figure 3): (i) from January-February to April-May, $\delta^{18}\text{O}$ decreases; (ii) an
278 enrichment in isotopic content is observed in summer (in June, July or August depending on
279 the station); (iii) isotope contents then decrease until September or October and (iv) a new
280 increase occurs until December. This relative homogeneity suggests that isotopic seasonality
281 is controlled by mechanisms of, at least, on regional scale.
282 Correlation coefficients between $\delta^{18}\text{O}$, precipitation and temperature are very low for all the
283 stations (Table 2). The weak correlation between $\delta^{18}\text{O}$ and air temperature (Table 2)
284 highlights the lack of temperature effect in the GOG and Central Africa regions. Despite the
285 fact that, in tropical maritime stations, a local amount effect is often observed (Rozanski et al.
286 1993), the poor correlation between $\delta^{18}\text{O}$ and local precipitation amount shows that this effect
287 does not dominate the isotopic seasonality in Central Western Africa. Therefore, it is clear
288 that, the local climate parameters do not control the seasonal variation of δ in precipitation.
289 Thus we hypothesize that convective activity at the regional scale could be involved. This
290 hypothesis is tested in the next section for the Douala station.

291 Table 2.

292

293 **4.2.2. Influence of regional convective activity**

294 Many studies in China (e.g., Gao et al. 2013; He et al. 2015 ; Yu et al. 2016; Guo et al. 2017;
295 Shao et al. 2017; Gao et al. 2018), India and Indo-Pacific region (e.g., Chakraborty et al.
296 2015; Rahul and Gosh 2016; Cai et al. 2016; He et al. 2018), South America (e.g., Hoffmann
297 et al. 2003; Vimeux et al. 2005; Villacis et al. 2008, Samuels-Crow et al. 2014) and Sahelian
298 Africa (e.g., Risi et al. 2008b; Risi et al. 2010b; Tremoy et al. 2012) based on both

299 observations and models, have shown that convective activity, that occurs upstream the
300 pathway of air parcels, is a major control of rainwater isotope composition in the tropics at
301 daily, seasonal and inter-annual timescales.

302 Convective activity is known to deplete water vapor through 3 main processes: i) precipitating
303 downdrafts, either at the convective-scale or at the meso-scale (Risi et al. 2008a, Kurita
304 2013), bring down depleted water vapor from the mid-troposphere to the boundary layer; ii)
305 rain evaporation, when concerning a small proportion of each raindrop, adds depleted water
306 vapor to the lower troposphere (Worden et al. 2007, Risi et al. 2010); iii) rain-vapor diffusive
307 exchanges in a saturated atmosphere can also deplete the water vapor (Lawrence et al. 2004).

308 GPCP precipitations are used to examine convective activity upstream the sampling site of
309 Douala. Figure 4 presents monthly mean precipitation and back trajectories calculated for
310 each month. As expected, moisture comes from the Atlantic Ocean most of the time and
311 precipitation, at this timescale, is entirely controlled by the seasonal migration of the ITCZ.

312 In November and December, moisture comes from the North-East without undergoing any
313 convection. From December to January, there is an abrupt shift of air parcel from North-
314 Easterly to South-Westerly, in line with the beginning of long northward migration of the
315 ITCZ. From March to May (spring season), air masses undergo strong convective activity
316 over the Gulf of Guinea. During the period June - August, air parcels still travel over the Gulf
317 of Guinea, but air masses undergo a weak convective activity because the ITCZ has shifted
318 further north over the Sahel region. From September to October (autumn period), the ITCZ
319 retreats to the South so that air masses undergo strong convective activity once again. The
320 shift between October and November trajectories marks the transition to the dry season when
321 the strength of monsoon winds weakens.

322 Figure 5 by presenting the variability of monthly mean GPCP precipitation along the back
323 trajectory over time (in hours), towards Douala station for May (spring period), August

324 (summer period), October (autumn period) and January (winter period) shows a quantitative
325 analysis. It reveals where, along the trajectories, convective activity becomes higher and thus
326 impacts $\delta^{18}\text{O}$ of Douala rainfall. Overall, the GPCP precipitation data show an increasing
327 trend from the Southern Atlantic to Douala. Along the air back trajectories, at 7 days (168
328 hours) to the sampling site, precipitation is higher in August and May. First significant
329 changes are observable at 120 hours, when a decrease (increase) of GPCP precipitation occurs
330 in January (October). It substantially increases in May and August. At 3 days (72 hours)
331 before Douala, convective activity is stronger in October (and to a lesser extent in May) than
332 August and January. Since isotopic value of rainfall is depleted in spring and autumn (Figures
333 2-3), $\delta^{18}\text{O}$ seems to be mostly sensitive to convective activity only in the past few days before
334 reaching Douala. This is consistent with the “memory” of convection in the isotopic
335 composition of water as discussed by Risi et al (2008a), Tuinenburg et al (2015) and Gao et al
336 (2013). Stable isotopes are imprint of convection along air parcel trajectories, and in the case
337 of Douala, precipitation seems to acquire its signature on average 72h before reaching the
338 station. Figure 6 shows the precipitation averaged over the past 72h (3 days) along the
339 backward trajectories, for each month. The precipitation $\delta^{18}\text{O}$ is significantly anti-correlated
340 with this average precipitation ($r^2=0.60$). This supports our hypothesis that convective activity
341 along trajectories significantly controls the seasonality of precipitation $\delta^{18}\text{O}$ in Douala.
342 In summary, the moisture “source” at Douala is most of the year in the GOG. Two seasonal
343 precipitation maxima along the trajectories (Figure 6) are related to the seasonal migration of
344 the ITCZ. Thus, even if precipitation in Douala shows a unimodal regime with a rainfall peak
345 in August, $\delta^{18}\text{O}$ records bimodal cycle as in the “source” area. This suggests that the regional
346 convective activity is the main control of the isotopic composition of precipitation in Douala.

347 In the following section, we use the isotopic information of the water vapor to provide
348 additional insights about the main processes controlling the isotopic composition of
349 precipitation.

350

351 Figure 4.

352

353 Figure 5.

354

355 Figure 6.

356 ***4.2.3. Information from the isotopic composition of water vapor (δv)***

357 The isotopic composition of the water vapor provided by the TES instrument can help to
358 understand the variation of δ in precipitation: it allows separating the relative effect of
359 processes acting along the trajectories of the water vapor (δD_v) and local post-condensational
360 processes ($\delta D_p - \delta D_v$) following the equation below:

$$361 \quad \delta D_p = \delta D_v + (\delta D_p - \delta D_v) \quad (1).$$

362 δD_v and δD_p (Figure 7a) show a similar seasonality ($r^2 = 0.57$). Both δD_v and δD_p present
363 depleted values in April-May and September-October (Figure 7a), corresponding to more
364 active convection in the GOG. In addition, there is always a positive offset between the
365 isotopic composition of water vapor and precipitation. Thus, the isotopic information on the
366 advected water vapor is preserved in the isotopic composition of the rain.

367

368 Figure 7.

369 However, absolute values of δD_p and TES δD_v should not be directly compared. TES was
370 calibrated using in-situ measurements on local sites (Worden et al 2012) but not specifically
371 in West Africa. In addition, TES data at 900hPa represent an average over several vertical
372 levels in the lower troposphere. Therefore, focus is given instead to temporal variability. Yet,
373 assuming that the vertical δD gradient between 1000 and 900hPa is constant, $\delta D_p - \delta D_v$
374 variations can be interpreted as variations in the rain-water vapor interaction processes.

375 The regression between δD_p and δD_v gives a slope of $a_1 = 1$ whereas the values for δD_p and
376 $\delta D_p - \delta D_v$ are $a_2 = -0.68$ and $r^2 = 0.18$. This insignificant r^2 attests that the rain-water vapor
377 interaction processes do not control the δD_p variations. Based on the slope a_1 , δD_v accounts
378 for 100% of the δD_p variability. A similar case has been reported in western tropical Pacific
379 (Conroy et al. 2016). This confirms that the seasonal variability of the isotopic composition of
380 rainfall is predominantly influenced by δD_v , i.e. by the processes that affect water vapor along
381 the trajectories. This means that the variability of $\delta D_p - \delta D_v$ acts only to dampen and to blur the
382 variability of δD_p . For instance, the evolution of $\delta D_p - \delta D_v$ presents a maximum in spring
383 (Figure 7b) and autumn, when δD_v is most depleted.

384 **4.3. Main controls of precipitation isotopic composition at daily scale**

385 ***4.3.1. Temporal evolution of δ and link with upstream convection***

386 Daily $\delta^{18}O$ in Douala varies from -0.3‰ to -7.4‰ , with a mean value of -3.0‰ close to the
387 monthly weighted mean of -2.8‰ calculated for the 2006-2016 period. These daily data
388 appear then to be representative of the 2006-2016 mean seasonal cycles. Moreover, as for the
389 seasonal scale, daily rainwater (Figure 8) is most depleted in spring (April-May) and more
390 enriched in summer (July-August).

391

392

Figure 8.

393 In order to investigate upstream convection effects, we calculated the correlation between the
394 $\delta^{18}\text{O}$ of rainfall and the precipitation recorded at 1 to 3 days earlier, along back trajectories
395 (Table 3). In March, April and August, the correlations are significant ($\geq 95\%$), suggesting
396 that, like for the seasonal timescale, the daily evolution is driven by convection along
397 trajectories. Yet, this mechanism is insufficient to explain completely the variability of δ in
398 daily precipitation, especially in May, June and July.

399 Table 3.

400 ***4.3.2. Factors besides the upstream convection intensity: continental recycling and*** 401 ***organization of convective systems***

402 ***Continental recycling?***

403 The goal of this section is to test whether continental recycling has significant influence on
404 $\delta^{18}\text{O}$ at the daily time scale. To that end, we calculated the percentage of time of the air parcel
405 over the continent during the last 3 days (72h) along the trajectory, hereafter F_{land} . This
406 calculation was relatively simple since air back trajectories have been computed at 6h time
407 steps and then for instance, the travel time of 36h of air mass over the continent will
408 correspond to F_{land} equal to 50%. However, for this issue, we calculated air back trajectories
409 at different heights: 800, 850, 900, 950 and 980 hPa since moisture is typically transported at
410 a spectrum of heights. As it is the boundary layer that feeds the convection and it is with the
411 water vapor from low layers that the rain is re-evaporated, investigation was not performed
412 above 800 hPa. Results from this quantitative calculation (not shown here) of moisture
413 transport for different heights show that there is no difference between them, especially since
414 we focus only on air back trajectory at 3 days before Douala. In June, July and August, F_{land} is
415 0% for all rainy days (except on June 11th when $F_{\text{land}} = 25\%$). Table 4 presents the detailed

416 results for spring period and shows an anti-correlation between $\delta^{18}\text{O}$ and F_{land} . Precipitation
417 are thus more depleted, when the transit time of air parcels over the continent is long, during
418 the last 3 days along the back trajectory This is in contradiction with the expected effect of a
419 continental recycling.

420 Table 4.

421
422 Since δ is controlled in spring by a convection upstream the sampling site, the fact that $\delta^{18}\text{O}$ is
423 more depleted with the increase of F_{land} , suggests a significant influence of convection
424 intensity over the continent. Indeed, [Xu and Zipser \(2012\)](#) demonstrated that convection is
425 generally more intense over the land than over the ocean. The more intense the convection
426 along trajectories, the more depleted the water vapor. Therefore, moisture origin has only an
427 indirect effect on isotopic composition of precipitation, depending on whether air mass goes
428 through regions of a strong convection. The positive and significant correlation between F_{land}
429 and GPCP precipitation confirms this mechanism.

430 Therefore, we suggest that, δ variations at a daily timescale are partly controlled by the
431 intensity of the convection along air parcel trajectories and that continental recycling is not
432 involved.

433

434 ***Influence of the organization of convective systems?***

435 Several studies have shown that convective systems deplete the low-level water vapor more
436 efficiently when they are more organized ([Lawrence et al. 2004](#), [Risi et al. 2008](#), [Tremoy et al](#)
437 [2014](#)), probably because moister air in larger systems allows for more efficient rain-vapour
438 diffusive exchanges ([Risi et al. 2008a](#)). In addition, convective systems deplete more vapor as
439 they extend to the upper troposphere ([Lacour et al 2018](#)) and as the extent of their anvils

440 (measured by the fraction of stratiform clouds) is large (Aggarwal et al. 2016). These features
441 are typically associated with higher degrees of organization.

442 Here, we test whether the type of convective organization has a significant impact on the $\delta^{18}\text{O}$
443 observed at Douala. By using satellite image from the SAF classification, we define 3 classes
444 of clouds in the region from March to August according to their altitude: (i) very low to low;
445 (ii) medium to high and (ii) very high clouds. The low cloud class is mostly observed in July-
446 August (Table 5 and Figure 9). The mid-level to high cloud class is dominant in June and the
447 very high clouds are mostly present in spring (Figure 9). The area of cloud systems increases
448 throughout spring, reaching a maximum in May, and decreases again from June. In July-
449 August, the areas are too small to be calculated.

450

451  Figure 9.

452 Spatial organization of convective systems also determines rainfall event duration. The latter
453 increases globally from March to August (Figure 10), ranging from 12 minutes to 1278 min
454 (i.e. more than 21 hours).

455

456  Figure 10.

457 Combining this information on cloud altitude, cloud area and rainfall duration; we can infer
458 that in March-April, most convective systems correspond to deep mesoscale convective
459 systems (MCS; Mapes and Houze 1993; Laing and Fritsch 1997; Mathon et al. 2001; Fink et
460 al. 2006; Tremoy et al. 2014). In May-June, convective systems are MCS that become even
461 larger and organized, but extend less deep in altitude (in June). In contrast, in July-August,

462 convective systems are isolated; characterized by small cumulonimbus in the middle of low-
463 to-medium clouds (Figure 9b) that follow one another throughout the day, explaining the
464 apparent long durations of rain events.

465 The type of convective organization emerges then as an important control on the precipitation
466 isotopic composition. Cloud surface shows a significant negative correlation with
467 precipitation $\delta^{18}\text{O}$ from March to June (Table 5). The larger the cloud area, the more depleted
468 the precipitation. In addition, the more the convective system is organized, the longer is the
469 duration of the event (Laurent and Machado 2002, Fiolleau et al. 2013). Consistently, event-
470 scale $\delta^{18}\text{O}$ is anti-correlated with event duration in March ($r^2 = 0.44$), April ($r^2 = 0.54$), May
471 ($r^2 = 0.64$) and more weakly in June ($r^2 = 0.34$). The longer is the convective system, the more
472 depleted is the precipitation.

473 These different types of convective organization can also explain the isotopic evolution at the
474 seasonal scale. In spring, the organized, deep MCS deplete the water vapor efficiently,
475 leading to the observed precipitation $\delta^{18}\text{O}$ minimum. In contrast, in July-August, the
476 shallower, small isolated cumulonimbus clouds deplete the water vapor less efficiently,
477 leading to the observed precipitation $\delta^{18}\text{O}$ maximum in spite of the local precipitation
478 maximum observed at Douala.

479 Surprisingly, the r^2 between cloud area and $\delta^{18}\text{O}$ are close and even higher (in May and June)
480 than that observed between $\delta^{18}\text{O}$ and upstream precipitation (Table 3). After verifying that
481 upstream convection (GPCP precipitation recorded at 72h to Douala along the trajectory) and
482 cloud area are uncorrelated (Table 5), we can assert that two independent parameters control
483 $\delta^{18}\text{O}$: upstream convection intensity and the size of the convective system (both at local scale
484 and at upstream, see the Table 5).

485

486

Table 5.

487

488 **5. Conclusion and outlook**

489 This study aims at investigating the processes controlling the year-to-year (inter-annual),
490 month-to-month (seasonality) and day-to-day (intra-seasonal) variability of rainfall isotopic
491 composition in the Gulf of Guinea (GOG) region and especially in Douala (Cameroon). We
492 observed that the annual cycle of δ is quite reproducible. Most years presents a W shape with
493 minima in spring and autumn. $\delta^{18}\text{O}$ and δD_v appear to be mainly controlled by upstream
494 convection and by the size of convective systems. We identified that the continental recycling
495 doesn't impact the rainwater isotopes in the humid tropical area of the GOG. In particular, the
496 $\delta^{18}\text{O}$ minima in spring and autumn are associated with strong convective activity in the GOG
497 and large, long-lived and deep mesoscale convective systems, whereas the $\delta^{18}\text{O}$ maximum in
498 July-August is associated with reduced convective activity in the GOG and isolated shallow
499 convective systems. The importance of upstream convective activity in controlling the $\delta^{18}\text{O}$ of
500 precipitation at various time scales is in line with a large body of recent research in different
501 tropical regions influenced by a monsoon system. The importance of the type of convective
502 system (size, organization, vertical extension) is also consistent with a growing number of
503 recent studies but this is the first time that it is demonstrated through a detailed analysis of
504 such a large number of individual convective systems. The findings of this study (obtained by
505 integrating in situ and satellite measurements) advance our understanding of the temporal
506 variation of precipitation stable isotopes in humid tropical area such as Douala, and shed a

507 new light on the importance of large scale meteorological conditions controls on precipitation
508 stable isotope composition in the GOG.
509 Notwithstanding, numerical climate modelling could be a useful complementary approach to
510 further analyse the factors controlling rainfall isotopic composition in and around Douala
511 Moreover, the climate in Douala being representative of a small area with monomodal rainfall
512 regime surrounded by area with bimodal rainfall, high resolution modelling is necessary to
513 correctly capture such particularity. More robust conclusions could be obtained with larger
514 samples, which could be obtained by extending the length of the record. To better understand
515 the role of convective organization on the isotopic composition, sampling at the intra-event
516 scale is necessary. Finally, measuring the isotopic composition in the water vapor in addition
517 to precipitation would be useful to isolate the post-condensation effects.

518

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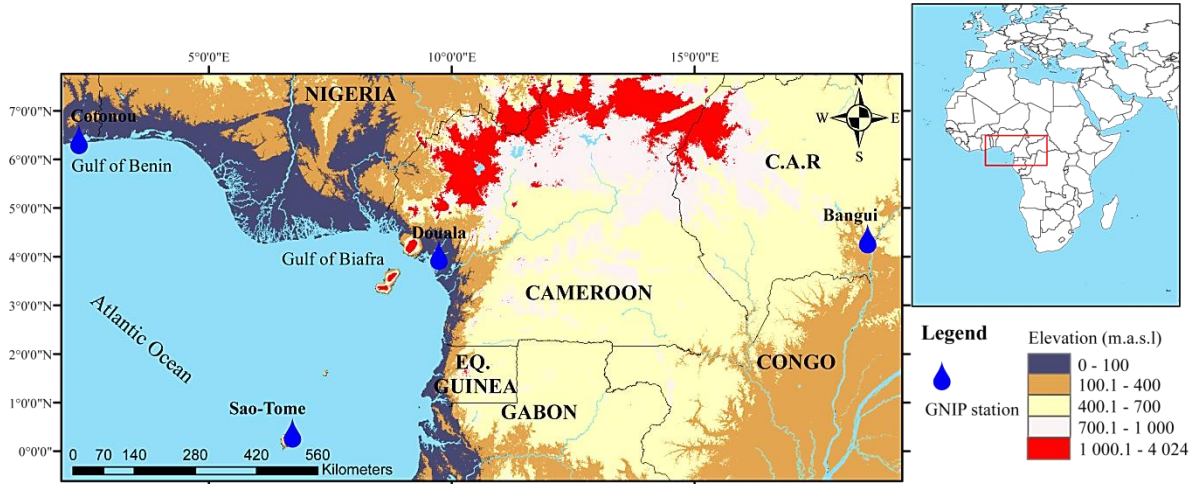
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946 **Figures and captions.**

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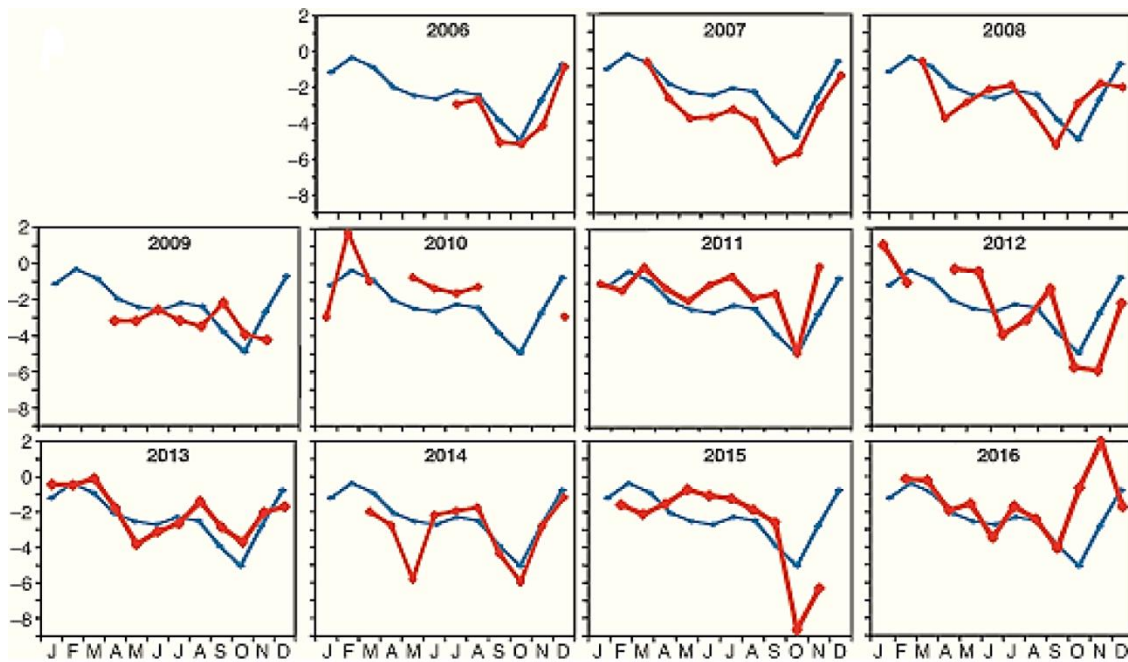
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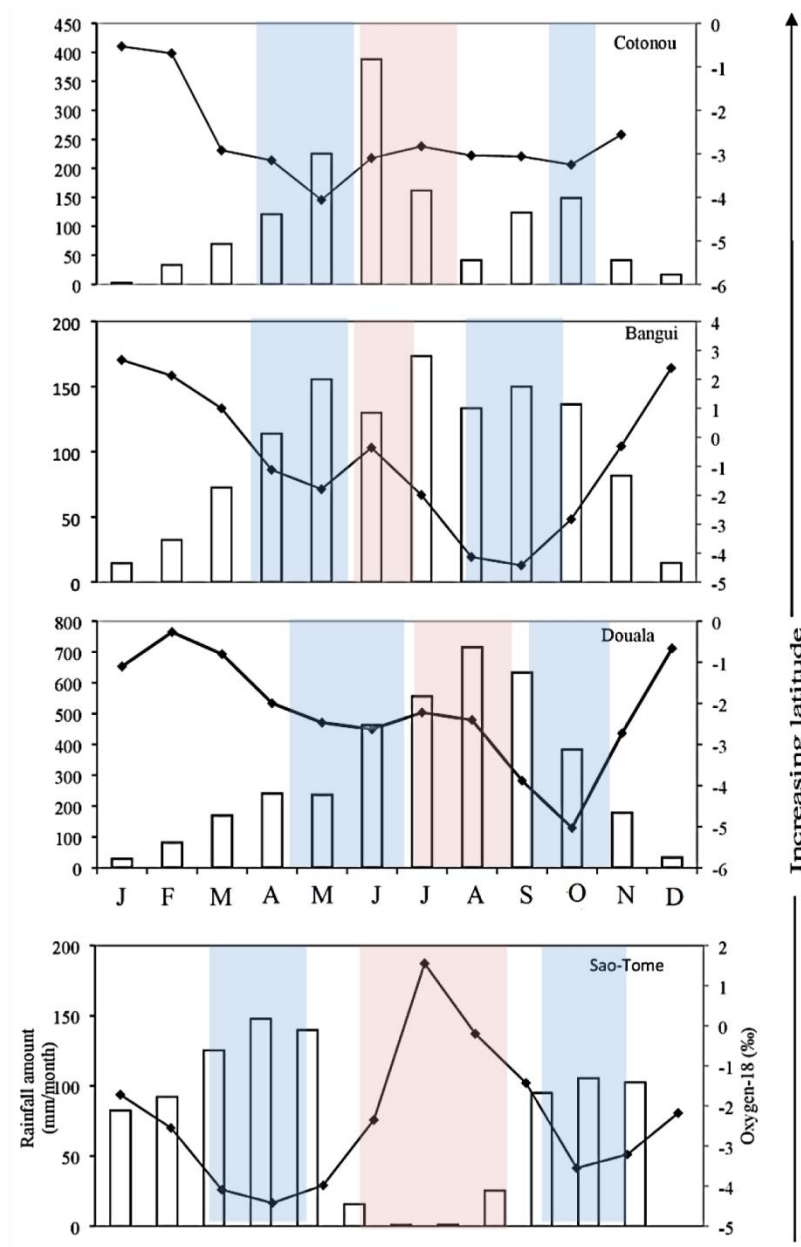
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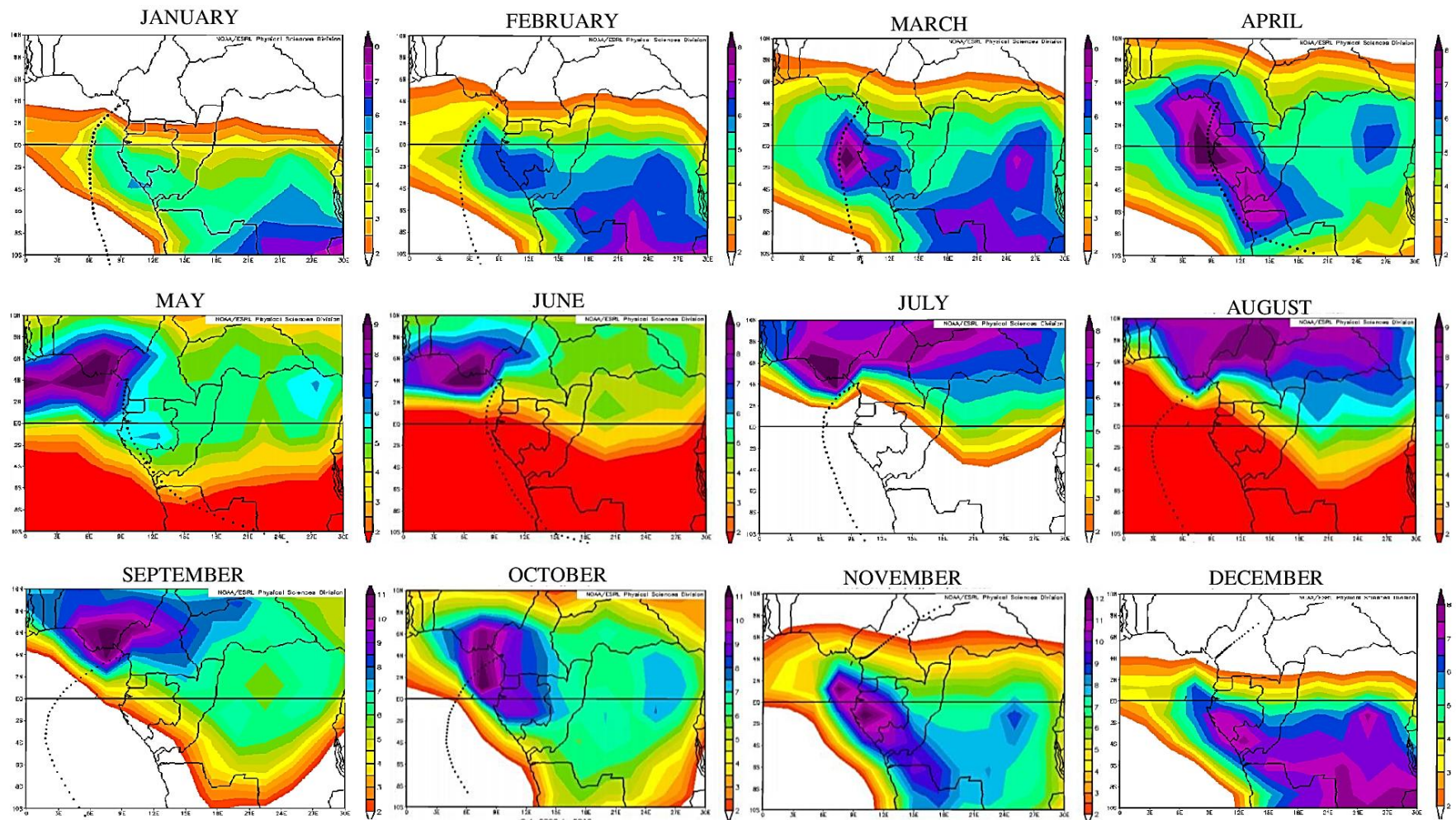
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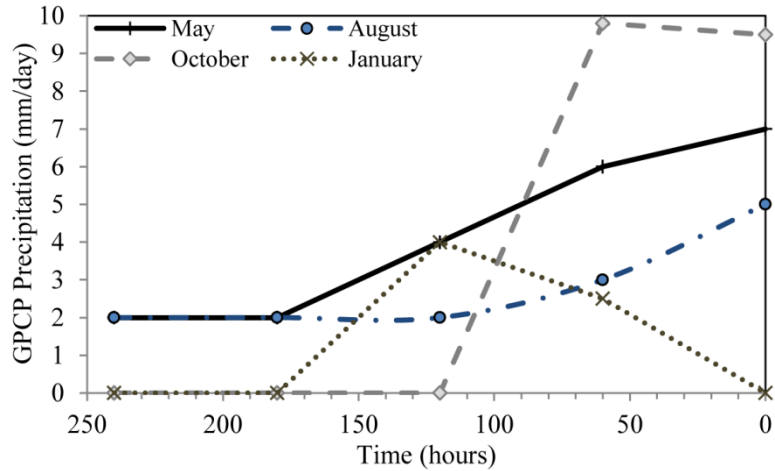


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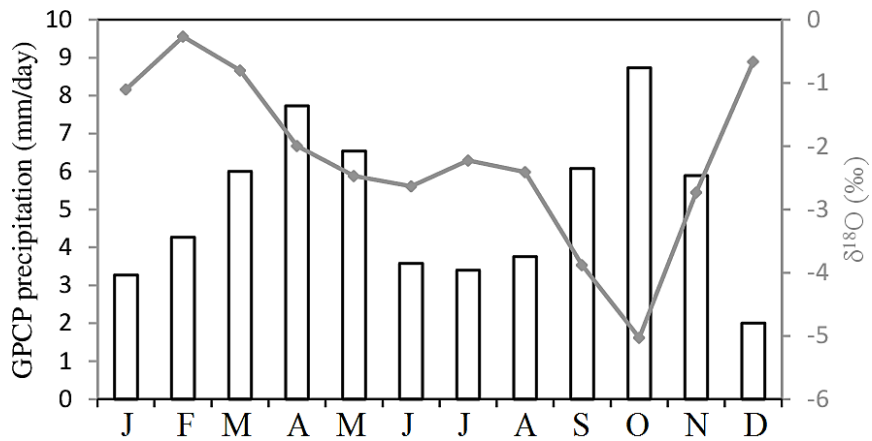


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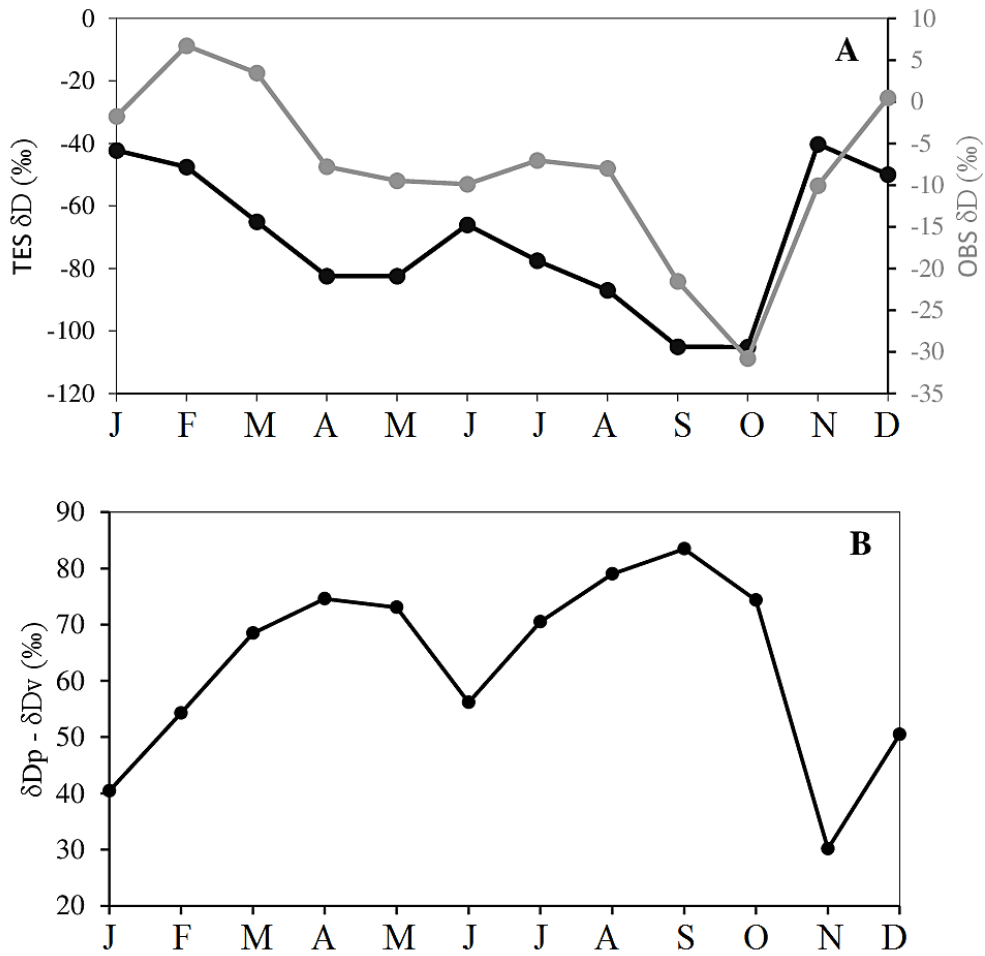


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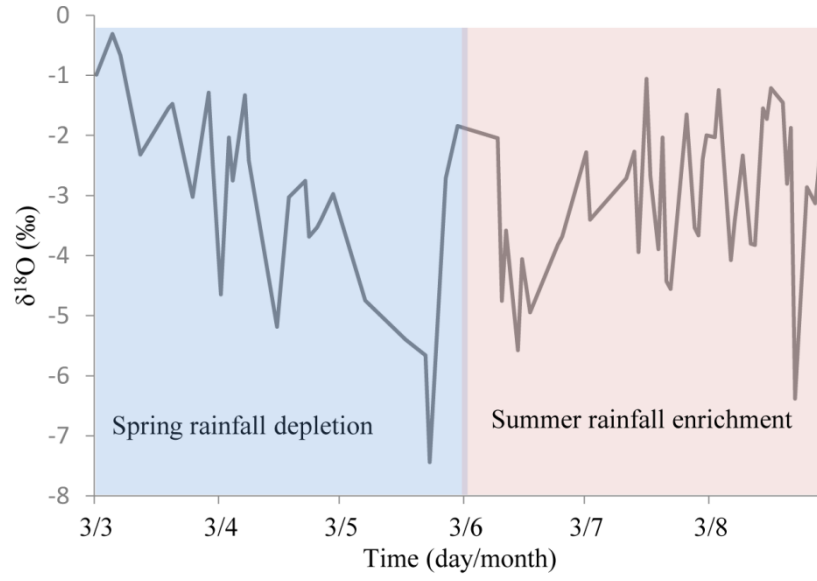


Figure 8. $\delta^{18}\text{O}$ in daily rainfall at Douala from March 2017 to August 2017.

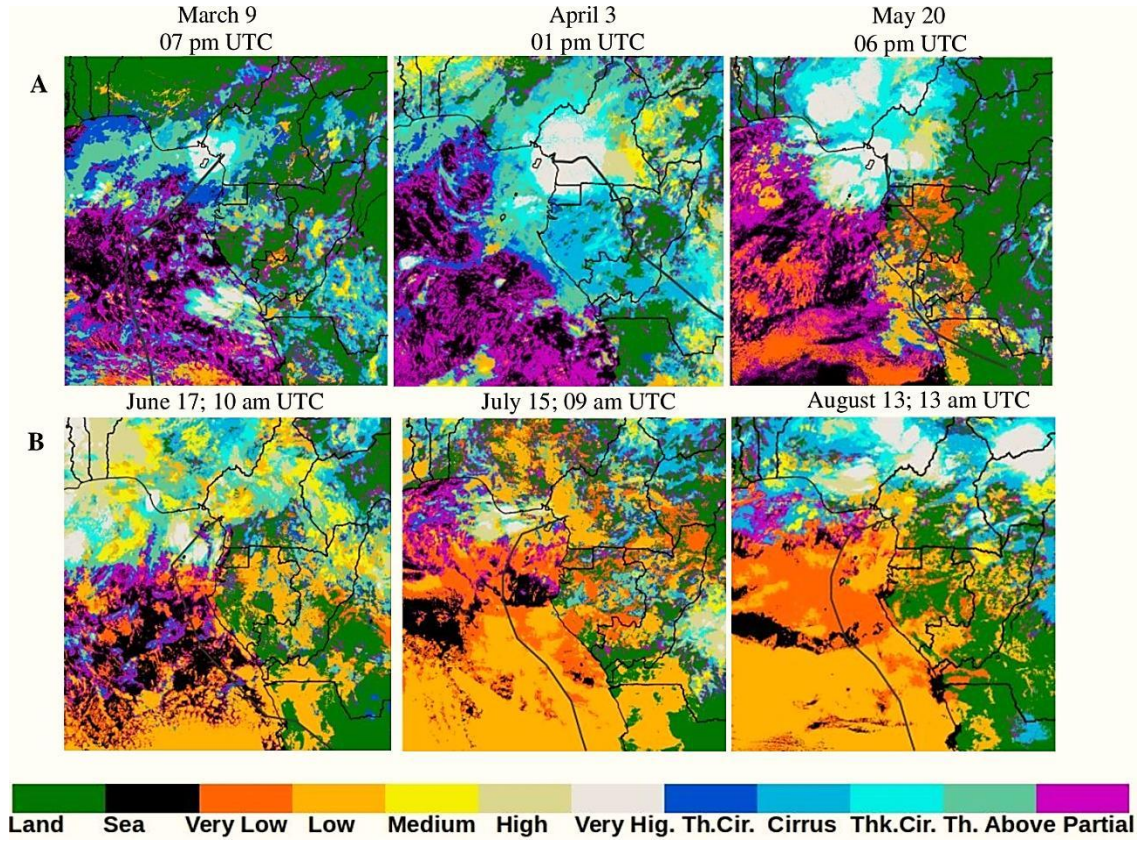


Figure 9. Example of meteorological situations in the study region in 2017 (based on the SAFNWC cloud classification) during (A) spring and (B) summer periods. The backward trajectories (grey line) are plotted for each event.

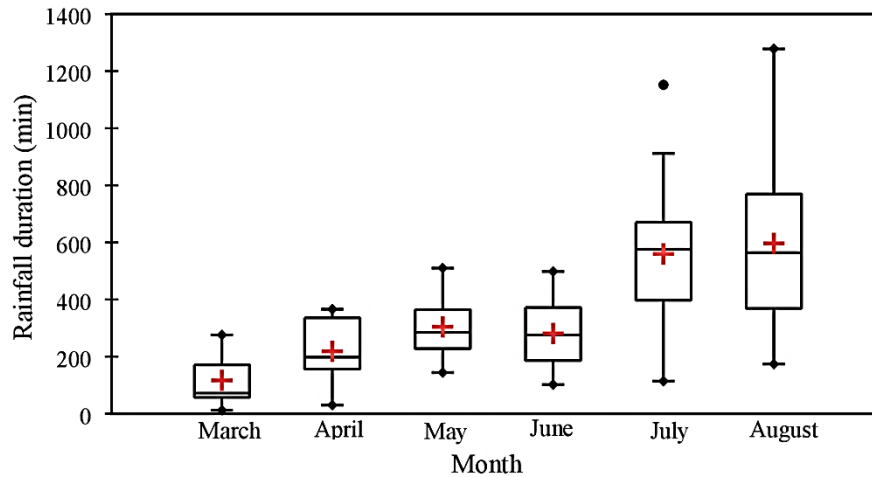


Figure 10. Statistical distribution of daily rainfall duration at Douala as shown by box-and-whisker representation. The boxes have lines at the lower, median and upper quartile values. The whiskers are lines extending from each end of the box to 1.5 interquartile range. Red cross represents the mean and outliers are represented by black points above the maximum

Tables

Table 1. Characteristics of GNIP sampling sites in GOG and Central African regions. Data from Cotonou, Bangui and Sao-Tome have been obtained online from the GNIP database (AIEA/OMM 2018).

Site	Country	Sampling period	Elevation (m.a.s.l)	Mean annual precip*(mm)	Vapor Pressure (hPa)	Mean annual T* (°C)
Cotonou	Benin	2005 - 2012	14	1395	28.3	27.1
Bangui	CAR*	2009 - 2015	363	1208	25.7	26.3
Douala	Cameroon	2006 - 2016	18	3720	30.5	27
Sao-Tome	Sao-Tome and Principe	1962- 1976	8	933.7	26.2	25.2

* Precip. = Precipitation, T = temperatures; CAR= Central African Republic; dec. deg = decimal degrees

Table 2. Coefficient of determination (r^2) between $\delta^{18}\text{O}$ and local meteorological settings (precipitation (P) and temperature (T)) for monthly GNIP stations of Cotonou, Bangui, Douala and Sao-Tome, at different sampling periods.

Stations	Sampling period	$r^2 \delta^{18}\text{O}/\text{P}$	$r^2 \delta^{18}\text{O}/\text{T}$
Cotonou (n=95)	2005 - 2012	0.05	0.13
Bangui (n=81)	2009 - 2016	0.24	0.04
Douala (n=106)	2006 - 2016	0.12	0.09
Sao-Tome (n=123)	1962 - 1976	0.19	0.20

Table3. Coefficient of determination (r^2) between $\delta^{18}\text{O}$ and average GPCP precipitation in average over the first day, two first days and three first days along the air parcel trajectory before rainfall in Douala. Correlation coefficients $\geq 95\%$ are highlighted in bold and underlined.

		March (n=08)	April (n=12)	May (n=07)	June (n=09)	July (n=14)	August (n=20)
$\delta^{18}\text{O}$ Vs GPCP Precipitation	One day before	<u>0.53</u>	<u>0.43</u>	0.40	0.03	0.14	0.15
	Two days before	<u>0.55</u>	0.23	0.28	0.14	0.03	<u>0.23</u>
	Three days before	<u>0.64</u>	0.16	0.44	0.00	0.11	<u>0.23</u>

Table 4. F_{land} statistics during the last 3 days along the air parcels trajectories for the months of March, April and May. Relationship (r^2) between F_{land} , $\delta^{18}\text{O}$, and upstream GPCP precipitation. For the summer period, F_{land} corresponds to 0% for almost all rainy days. Correlation coefficients $\geq 95\%$ are in bold and underlined.

	March	April	May
F_{land} (%)	0	0	25
<i>Minimum</i>			
F_{land} (%)	100	100	100
<i>Maximum</i>			
F_{land} (%)	25	69.45	70.83
<i>Mean</i>			
F_{land} (%)	0	83.34	79.17
<i>Median</i>			
r^2 (F_{land} vs $\delta^{18}\text{O}$)	0.23	<u>0.33</u>	0.28
r^2 (F_{land} vs GPCP precip.)	0.51	0.42	0.85

Table 5. Information on the organization of convection (cloud types, cloud area and local/upstream) and its influence on the isotopic composition of daily rainfall over the period from March 2017 to August 2017. Cloud areas were only calculated for very high classes. Correlation coefficients significant $\geq 95\%$ are in bold and underlined.

	March	April	May	June	July	August
Very low to low clouds (%)	0	0	0	0	71.43	54
Medium to high clouds (%)	0	0	18.18	72.73	21.43	40.5
Very high clouds (%)	100	100	81.82	27.27	7.14	5.5
Local without upstream convection (%)	75	27.27	0	22.29	7.14	0
Upstream and local convection (%)	25	72.73	100	77.71	92.86	100
Minimum cloud area (km ²)	5300 ± 100	4500 ± 100	63400 ± 100	2400 ± 100		-
Maximum cloud area (km ²)	35300 ± 100	158100 ± 100	285600 ± 100	280000 ± 100	3077 ± 100	615 ± 100
Mean cloud area (km ²)	17800 ± 100	73700 ± 100	225100 ± 100	138500 ± 100	-	-
r ² Clouds area Vs rainfall duration	<u>0.54</u>	<u>0.63</u>	<u>0.51</u>	0.34	-	-
r ² Clouds area Vs δ ¹⁸ O	<u>0.44</u>	<u>0.55</u>	<u>0.97</u>	<u>0.90</u>	-	-
r ² Clouds area Vs GPCP precipitation	0.30**	0.10**	0.01**	0.46**		
r ² Rainfall duration Vs δ ¹⁸ O	<u>0.44</u>	<u>0.54</u>	<u>0.64</u>	0.24	0.00	0.07

**Anti-correlation

Author contribution statements

B. Nlend , H. Celle-Jeanton , C. Risi , B. Pohl , F. Huneau , S. Ngo Boum-Nkot , G. Seze , P. Roucou c, P. Camberlin , J. Etame f and B. Ketchemen-Tandia conceived the presented paper,. developed the theory and performed the computations.

F. Huneau was particulary in charge of analytical methods.

G. Seze , C. Risi and H. Celle-Jeanton encouraged B. Nlend to investigate cloud type products.

B. Pohl, P. Camberlin and P. Roucou contribute to the understanding and redaction of the climate context.

B. Pohl provided a significant help to improve the English.

All authors discussed the results and contributed to the final manuscript.