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Projected changes in the atmospheric dynamics of climate extremes in France

Pascal Yiou ¹ , Davide Faranda ^{1,2,3}, Soulivanh Thao ¹ and Mathieu Vrac ¹

¹ Laboratoire des Sciences du Climat et de l'Environnement, UMR8212 CEA-CNRS-UVSQ, U Paris-Saclay & IPSL, Gif-sur-Yvette, France

² Affiliation 2; e-mail@e-mail.com

* Correspondence: pascal.yiou@lsce.ipsl.fr

Abstract: Extremes of temperature, precipitation and wind have caused damages in France, in the agriculture, transportation and health sectors. Those types of events are largely driven by the atmospheric circulation. The dependence on the global climate change is not always clear, and is the subject of extreme event attribution (EEA). This study reports an analysis of the atmospheric circulation over France for seven events that struck France in the 21st century, in various seasons. We focus on the atmospheric dynamics that leads to those extremes and examine how the probability of atmospheric patterns and their predictability responds to climate change. We analyse how the features of those events evolve in simulations following an SSP585 scenario for future climate. Using a range of CMIP6 simulations helps determining uncertainties linked to climate models.

Keywords: Extreme events; climate change; France

1. Introduction

Rather than a global perspective, decision makers have emphasized the necessity of projections of regional extremes for adaptation to climate change [1]. Therefore, this paper will focus on a few extreme climate events that recently occurred in continental France. Insurance, health, agriculture and energy sectors (at least in France) are often affected by thermal extremes (cold and hot), extreme precipitation or wind speed. Observations show that the duration of such events rarely exceeds a couple of weeks for a maximum impact. Therefore, the key time scale we will consider in this paper is subseasonal.

A climate or meteorological extreme event is often defined when a key variable (temperature, precipitation, wind speed) exceeds a predefined threshold. This allows computing the probability of the event, and then the change of probability of that event, conditional to a climate change scenario, which is the essence of extreme event attribution (EEA) [2–4]. Many papers of EEA are based on the estimate of probabilities of events, from statistical modeling of exceeding a threshold. This requires tools from extreme value theory (EVT) [5]. The main caveat of some of those studies is that they do not take into account the physical processes leading to the extreme events, like features of the atmospheric circulation. Attempts to connect the atmospheric circulation to variables that define the extremes have been recently devised [6,7]. Shepherd [8] argued that the atmospheric circulation is a key element of the uncertainty in attribution studies. This motivates our focus on features of the atmospheric variability that drive a few key events.

New mathematical tools of extreme event attribution have been recently devised [9]. Those tools focus on *dynamical* properties of the atmospheric circulation. Dynamical properties of physical systems correspond to time derivatives of the variables of the system, which can be determined from well-chosen mathematical and statistical indicators

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38 [10–12]. Faranda *et al.* [10] have argued that such dynamical indicators are related to the
39 predictability of the (atmospheric) system. Hence our paper will deal with estimates of
40 dynamical features linked to the atmospheric predictability. The dynamical indicators
41 we will consider include the local dimension, persistence, and pattern likelihood.

42 In this paper, we study the atmospheric circulation that prevailed during several
43 extreme climate events that occurred in France since 2000. Our goal is to determine how
44 the atmospheric motion properties can be altered by climate change.

45 Without being exhaustive, we decided to have a wide panel of types of events.
46 Hence, the events we consider are based on temperature, precipitation or wind speed,
47 and which occurred in the four seasons. In the midlatitudes, those quantities are linked
48 to the large-scale atmospheric circulation [13]. This justifies a focus on the atmospheric
49 dynamics that is related to the considered extremes. Hence, we shall examine how the
50 synoptic scales of the atmosphere that occur during regional temperature, precipitation
51 or wind events can be affected by climate change.

52 We will determine the values of dynamical indicators of the atmospheric circulation
53 in a reanalysis dataset, for reference extreme events. Then we will assess how those
54 dynamical indicators during those reference events will change, as a response to a future
55 climate scenario (here SSP585). This will be performed by sampling the atmospheric
56 variability due to future climate change from a multi-model ensemble (the Coupled
57 Model Intercomparison Project, phase 6: CMIP6 [14]).

58 Data and methods will be exposed in Section 2. Results appear in Section 4. Discus-
59 sion and conclusion appear in Section 5.

60 2. Materials and Methods

61 2.1. Observations and reanalyses

62 We considered the ERA5 reanalysis [15] for the determination of key extreme events
63 since 2000. The ERA5 period is available from 1979 to now, with a horizontal resolution
64 of 0.25° . We computed daily averages for sea-level pressure (SLP), geopotential height at
65 500hPa (Z500), 2-m temperature (T2m), total precipitation (Pr, which combines rainfall,
66 snowfall and hail), 10-m wind speed and 10-m peak wind speed (W , W_{\max}). SLP and
67 Z500 fields were extracted for a region covering 80W–30E and 30N–65N. Temperature,
68 precipitation and wind speed fields were extracted over a region covering France (4.5W–
69 8.5E; 42N–51.5N). We then selected the 1018 gridpoints that are included in continental
70 France (excluding Corsica).

71 2.2. Climate model simulations

72 We analyze daily output of the Coupled Model Intercomparison Project phase 6
73 (CMIP6) [14] for 11 historical simulations (see Table 1 for references), and one socio-
74 economic pathway (SSP) scenario [16]. This selection was dictated by the availability of
75 Z500, temperature and precipitation fields on daily time scales at the time of analyses:
76 we have only selected models whose data were fully available for the whole period
77 1950–2100. The historical simulations cover the period 1950–2014. The forcings are
78 consistent with observations. They include changes in: atmospheric composition due
79 to anthropogenic and volcanic influences, solar forcing, emissions or concentrations of
80 short-lived species and natural and anthropogenic aerosols or their precursors, as well
81 as land use. The SSP585 scenario corresponds to a representative concentration pathway
82 scenario (RCP) with a radiative forcing increase of 8.5 Wm^{-2} in 2100 due to greenhouse
83 gas emissions, relative to pre-industrial conditions [16].

Table 1: List of CMIP6 simulations used in this study, their approximate horizontal resolution and references.

Simulation name	Atmospheric resolution	Data reference
BCC-CSM2-MR	100 km	[17]
CanESM5	500 km	[18]
CNRM-CM6-1-HR	100 km	[19]
CNRM-CM6-1	250 km	[20]
CNRM-ESM2-1	250 km	[21]
INM-CM4-8	100 km	[22]
INM-CM5-0	100 km	[22]
IPSL-CM6A-LR	250 km	[23]
MIROC6	250 km	[24]
MRI-ESM2-0	100 km	[25]
UKESM1-0-LL	250 km	[26]

84 The simulations of each climate model came in ensembles of several members. For
 85 simplicity and because not all ensemble sizes were equal, we picked one member of each
 86 model ensemble.

87 2.3. Bias correction and trend removal

88 Given that the models have biased representation of the Z500, we apply a statistical
 89 bias correction [27] on the Z500 fields allowing to account for climate change [28].

90 The statistical bias correction method applied is the Cumulative Distribution Func-
 91 tion - transform (CDF-t) method, developed by Michelangeli *et al.* [27]. The bias correc-
 92 tions were made through the "CDF-t" R package (available at [https://cran.r-project.org/
 93 web/packages/CDFt](https://cran.r-project.org/web/packages/CDFt)). More theoretical and technical details, as well as first validations
 94 and comparisons can be found in [9,28].

95 This approach links the cumulative distribution function (CDF) of a climate variable
 96 (here Z500) from GCM simulations to be corrected, to the CDF of this variable from a
 97 reference dataset, here the ERA5 reanalysis dataset [15]. CDF-t can be considered as
 98 a variant of the empirical quantile-mapping method but within the appropriate target
 99 (here future) time period and therefore accounts for changes of CDF from the calibration
 100 period to the projection one.

101 This bias correction method is applied for each grid-point separately in two different
 102 ways. First, CDF-t is applied on a monthly basis to the "raw" ERA5 reanalyses and 11
 103 CMIP6 GCM simulations. The results are called the "non-detrended bias corrections".

104 Secondly, CDF-t is applied (also on a monthly basis) to ERA5 reanalyses and GCM
 105 simulations from which a spatial and seasonal trend is removed. To do so, for each day
 106 (in ERA5 and CMIP6), the Z500 spatial average is calculated. Next, for each calendar
 107 day (e.g., each January, 1) over the periods of interest (1979–2019 or 2061–2100), a linear
 108 fit of the daily Z500 spatial average as a function of time is estimated. This spatial trend
 109 is then removed from each Z500 grid-cell value for the specific calendar day. Then
 110 the spatial average value estimated for the model during the year 2000 is added to the
 111 calendar day. This ensures that a seasonality (estimated for 2000) is preserved, with no
 112 trend in the resulting Z500 data. Those seasonally and spatially detrended data are the
 113 inputs of CDF-t, providing adjusted values. Hereafter, we refer to those adjusted values
 114 as "detrended bias corrections". The removed Z500 spatial average trend corresponds
 115 to the spatially uniform shift of Z500, mainly caused by the warming over the region.
 116 Hence, by removing this trend, we also removed the first-order thermodynamic effect
 117 of warming on Z500 fields. Therefore, the resulting anomalies, which are further bias
 118 corrected, indicate changes that are mostly due to dynamical changes of the state of the
 119 atmosphere.

120 CDF-t bias correction can be applied on detrended and non-stationary data [28–32].
121 Indeed, unlike other bias correction methods that require stationary distributions, the
122 CDF-t approach is explicitly designed to account for changes in the distributions — i.e.,
123 to account for non-stationarity — and is thus suited when trends are present in data.
124 This does not imply that CDF-t corrects the trends of the model, but rather that CDF-t
125 mostly preserves the trends from the model data to be corrected.

126 2.4. Variations of dynamical indicators due to climate change

127 We extract the daily Z500 fields corresponding to the selected extreme event in the
128 ERA5 reanalysis. We then embed these observed trajectories into historical simulations
129 (1979–2019), and projections (2061–2100) under a high (SSP585) emissions scenarios of
130 the CMIP6 models [14]. This embedding of observed trajectories in climate simulations
131 is performed by looking at best analogues. Using the extreme event attribution vocabu-
132 lary, we consider the historical simulations (1979–2019) as the *factual* world, i.e. the
133 actual world with the current level of anthropogenic emissions. The SSP585 scenario
134 corresponds to a *counterfactual* world. Contrarily to the traditional counterfactual world
135 representing the world that could have been without climate change, our counterfactual
136 world explores the world that could be under projected trajectories of anthropogenic
137 emissions [2]. This *forward* EEA approach has been used by Van Oldenborgh *et al.* [33],
138 who calculated trends up to 2100 from model outputs, Sweet *et al.* [34] who evaluated
139 the annual maximum storm tide level for four different scenarios of sea level rise (see
140 also Sweet *et al.* [35]) and by Kay *et al.* [36], Yoon *et al.* [37] to project the evolution of
141 fire risks in the future. When we embed the portion of Z500 trajectories corresponding
142 to extreme events, we assume that the circulation patterns associated with the extreme
143 event could be observed in the climate model simulations. This assumption is justified by
144 previous studies where it has been verified that the average analogue distances between
145 observed atmospheric patterns (for Z500 and sea-level pressure) and the Historical
146 model simulations are within the analogue distance of the ERA5 reanalysis [38]. For each
147 extreme event, we follow the approach of Faranda *et al.* [9] by computing the analogs of
148 the observed synoptic patterns in each set of model simulations, and determine their
149 properties. For each daily Z500 field observed during extreme events, we compute the
150 Euclidean distance from all the other daily Z500 fields by a spatial average of grid-point
151 distances and we then select the closest 2% daily Z500 fields. This defines our analogs
152 ensemble. Note that the results do not crucially depend on this percentage provided that
153 it is in the range of 0.5 to 3%. The values of the Euclidean distance allow to determine
154 how well the circulation patterns associated with extreme events fit in the simulations.
155 This metric is what we call "analog quality", defined as the average of the Euclidean
156 distance of the 2% closest fields. In addition, we compute the local dimension d and
157 a persistence θ metrics, which are linked to the predictability of the circulation (see
158 [10,39,40]). We have also checked in a previous study that the dynamical systems metrics
159 found for historical simulations are within one standard deviation of those of the 20CR
160 reanalysis datasets [41]. The local dimension and persistence characterize the recurrences
161 of a system around a state in phase space. In our case, the state is the Z500 map for
162 a given extreme event. Values of d and θ are obtained for every day in the dataset of
163 interest. d provides information on the number of pathways the system can take to reach
164 and leave a state [42,43]. It acts as a proxy for the system's active number of degrees
165 of freedom around the state of interest. $0 < \theta \leq 1$ is a metric of persistence [44] of an
166 atmospheric circulation state in time, i.e. how long the system typically stays around the
167 state of interest. A very persistent state (i.e., with θ^{-1} close to zero) is highly stable (and
168 therefore also highly predictable), while a very unstable state yields $\theta = 1$ and therefore
169 low persistence. The information provided by persistence and predictability are differ-
170 ent: persistence is related to very short term predictability, namely the possibility of
171 observing tomorrow a pattern which resembles the one observed today. The the metric

172 d for Z500 is meant to extend towards subseasonal to seasonal scales, as this metric is
 173 linked with the underlying local Lyapunov exponents of the systems [11].

174 The $d - \theta$ parameters are computed for each selected event, in all models and
 175 scenario runs (historical and SSP585), so that we can detect changes in the atmospheric
 176 circulation observed during extreme events. A change in the analogs quality tells
 177 whether the atmospheric configuration is more or less likely in the historical than in
 178 the scenario experiments. A change in the dynamical indices informs on the change
 179 of predictability and persistence of the circulation pattern associated with the extreme
 180 event.

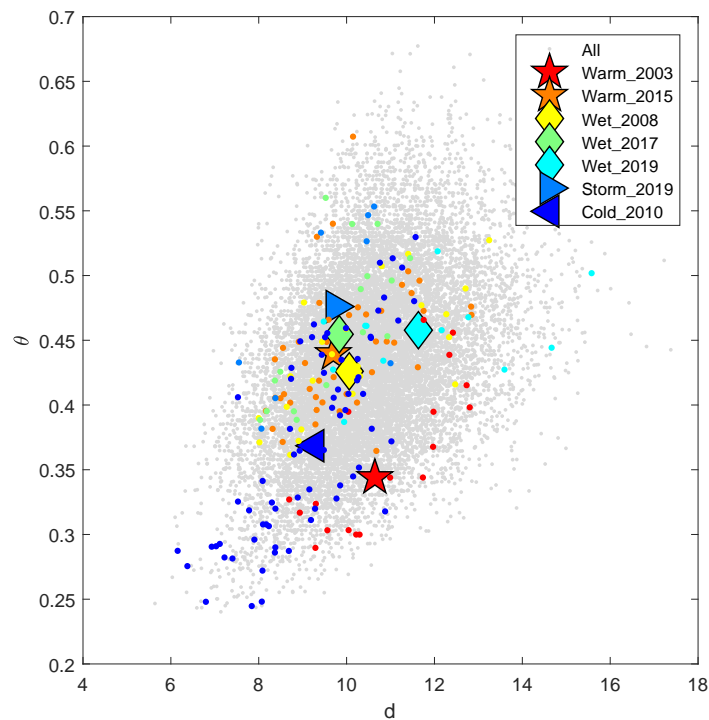


Figure 1. Scatter plot of the daily values of local dimension d and local persistence metric θ determined from the Z500 of the ERA5 reanalysis data. The symbols indicate the median d and θ values for the seven events selected in Sec. 3, stars for heatwaves, diamonds for precipitation events, right triangle for the 2019 storm and left triangle for the 2010 cold spell. The colored dots correspond to daily values of $d-\theta$ for each identified event.

181 3. Selection of events

182 We decided to chose extreme events that hit France since the beginning of the 21st
 183 century. This paper will evaluate how the probability of such events change according to
 184 scenarios of climate change, following the approach of Faranda *et al.* [9] that we recalled
 185 before.

We determined key years for extremes of temperature, precipitation and wind that occurred in the 21st century. We used a simplified version of the approach of Cattiaux and Ribes [45]. We first considered each season separately (Winter: December-January-February, DJF; Spring: March-April-May, MAM; Summer: June-July-August, JJA; Fall: September-October-November, SON), as the features of extremes depend on the season. For each season and each gridpoint of the ERA5 reanalysis, we determined the 5th (q_{05}) and 95th (q_{95}) quantiles of near-surface temperature (T2m), total precipitation (Pr), wind speed (W) and maximum wind speed (Wmax), between 1979 and 2020. Then for each day t or year y and each gridpoint s , we determine whether the value of climate variable $X \in \{T2m, Pr, W, Wmax\}$ exceeds a 95th quantile threshold. The reason for this

procedure is the high dependence of precipitation and wind speed on altitude: wind speeds exceeding 100 km/h or precipitations rates over 100mm/day are not particularly extreme in mountainous areas, while they are rather rare in plains. Therefore, it is necessary to consider location dependent thresholds (i.e. local quantiles) to detect extremes.

$$\pi(s, t, y) = \begin{cases} 1, & \text{if } X \geq q_{95} \\ 0, & \text{if } X < q_{95}. \end{cases} \quad (1)$$

For each year y , we compute the empirical probability that X exceeds a high threshold (q_{95}) (or is below a low threshold (q_{05})):

$$p(y) = \frac{1}{N_{FR}N_{season}} \sum_s \sum_{t \in y} \pi(s, t, y), \quad (2)$$

186 where N_{FR} is the number of grid point in France ($N_{FR} = 1018$ in ERA5) and N_{season} is
 187 the number of days in a season ($N_{seas} \approx 92$). This quantity accounts for the intensity of
 188 events (the variable X has to exceed a high threshold), spatial extent (fraction of grid cells
 189 for which X exceeds a threshold) and duration (number of days for which X exceeds a
 190 threshold).

191 Figure 2 shows the time series of $p(y)$ for temperature, precipitation and wind
 192 speed. The higher values for temperature extremes (Fig. 2a, vertical axis) reflect that
 193 thermal events have a larger geographical extent and a longer duration (2–3 weeks).
 194 Storms travel across France in a couple of days. Precipitation events generally cover
 195 very limited areas and last a couple of days.

196 In this paper, in order to avoid a tedious exhaustive list of events, we focused on
 197 warm winters and summers (which have impacts on human health and ecosystems),
 198 cold winters (which have impacts on energy demand and human health), wet springs
 199 (that have impacts on agriculture and river management) and falls, and windy winters
 200 (due to storms).

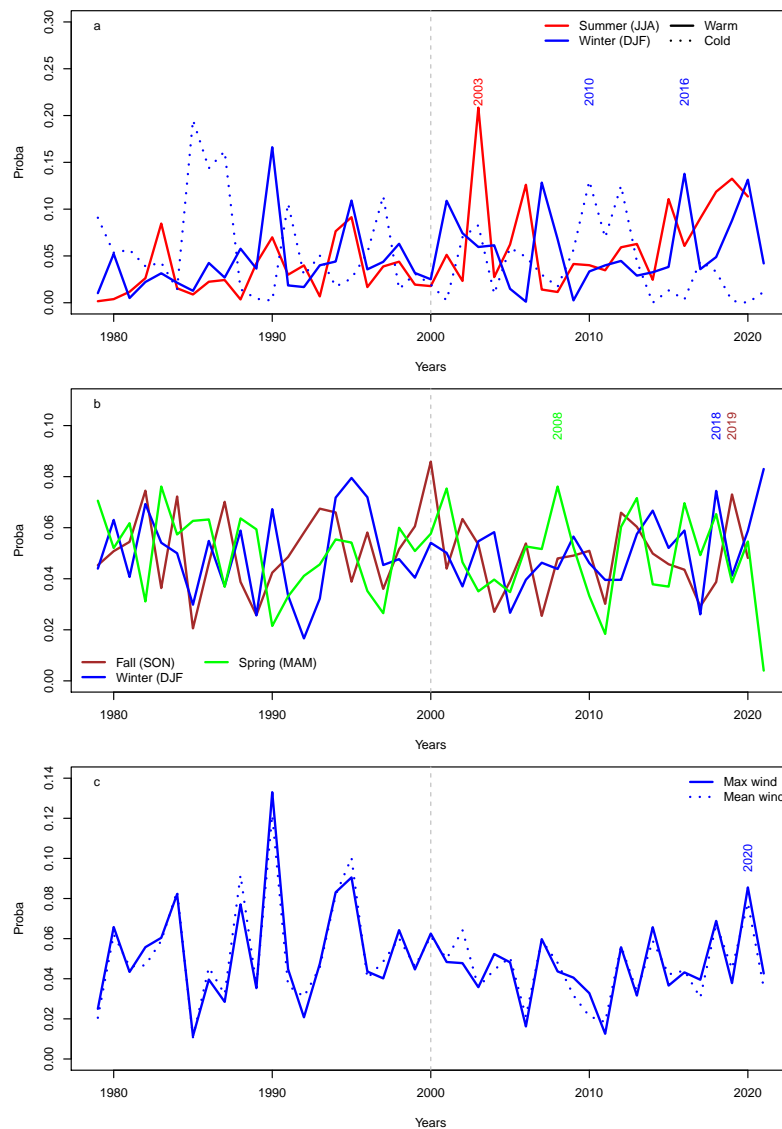


Figure 2. Variations of $p(y)$ for temperature (a), precipitation (b) and wind speed (c). The vertical dashed line is for the year 2000 after which the events are considered. The colors indicate the seasons: red for summer, brown for fall, blue for winter and green for spring. The dotted line in panel (a) indicates the value of p when temperature is below the 5th quantile. The years indicate the records during the 21st century, but excluding 2021.

201 With this approach, we outline seven remarkable recent years (since 2000), with
 202 examples of warm summer (2003) and winter (2016), cold winter (2010), wet spring
 203 (2008), wet winter (2018), wet fall (2019), and windy winter (2020). Then we identified
 204 event dates from (daily) time series of climate variables. We determined clusters of days
 205 when the considered climate variable X exceeds the seasonal average of the 95th quantile
 206 of 1018 geographical points (in continental France), with possible excursions below this
 207 threshold of no more than 2 days. Shorter "excursion" times to less extreme states (i.e. 1
 208 day or a fully continuous cluster) could have been considered, depending on the impacts
 209 of the event. For example, the insurance sector considers that events must be separated
 210 by at least 24h to be treated separately. But this only marginally affects the analyses
 211 presented in this paper.

212 The features of those seven identified events are summarized in Table 2.

Table 2: Table of key extreme events in France since 2000. The median local dimension d and persistence metric θ are indicated for each event.

Date of event	Cold	Warm	Wet	Storm	d	θ
31/07/2003 – 17/08/2003		X			0.64	0.34
09/03/2008 – 01/04/2008			X		10.06	0.43
11/12/2009 – 18/02/2010	X				9.27	0.37
01/12/2015 – 13/01/2016		X			9.70	0.44
25/12/2017 – 09/01/2018			X		9.83	0.45
13/10/2019 – 25/10/2019			X		11.63	0.46
06/12/2019 – 15/12/2019			X	X	9.75	0.48

213 3.1. Summer Heatwaves

214 The summer 2003 was a heatwave epitome, in term of amplitude and duration
 215 [46,47]. This events had huge impacts on the biosphere [48], public health [49] and the
 216 economy [REF]. As of 2021, the mean summer temperature of 2003 in France is still the
 217 all time record since the beginning of meteorological records.

218 The mean temperature reached its climax between July 31st and August 17th 2003.
 219 The mean seasonal temperature anomalies are shown in Fig. 3 (left). The event was
 220 preceded by a drought that started in early May 2003. This prolonged drought hindered
 221 latent heat fluxes and hence exacerbated high temperatures. The heat anomaly was also
 222 characterized by tropical night time temperatures.

223 The atmospheric circulation was characterized by a strong anticyclonic pattern
 224 centered over France (Fig. 3 (right)), during which almost no wind blew over France,
 225 which also enhanced ozone air pollution (and the associated death toll).

226 The atmospheric circulation of this episode was characterized by a medium local
 227 dimension and high persistence (red star in Figure 1).

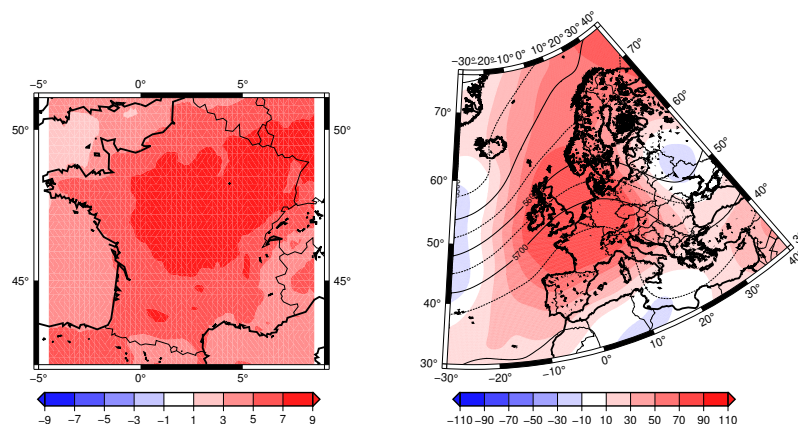


Figure 3. Left panel: mean anomalies of T2m over France between 31/07/2003 and 17/08/2003. Right panel: mean anomalies of Z500 (in m) over the East North Atlantic region (colors). Contour lines indicate the mean Z500 (in m).

228 3.2. Winter cold spells

229 The winter of 2009–2010 was exceptionally cold and snowy in Europe and Eastern
 230 US [50] (Figure 4 (left)). Although the temperatures were not as cold as historical events
 231 such as the winters 1954 or 1963, this cold event created havoc in transportation systems,
 232 and was deemed to be the coldest since the beginning of the 21st century.

233 The atmospheric circulation was cyclonic, with a persisting negative phase of the
 234 North Atlantic Oscillation (Figure 4 (right)) [50,51].

235 The atmospheric circulation of this episode was characterized by a low local di-
 236 mension and high persistence — low values for θ — (deep blue triangle in Figure
 237 1).

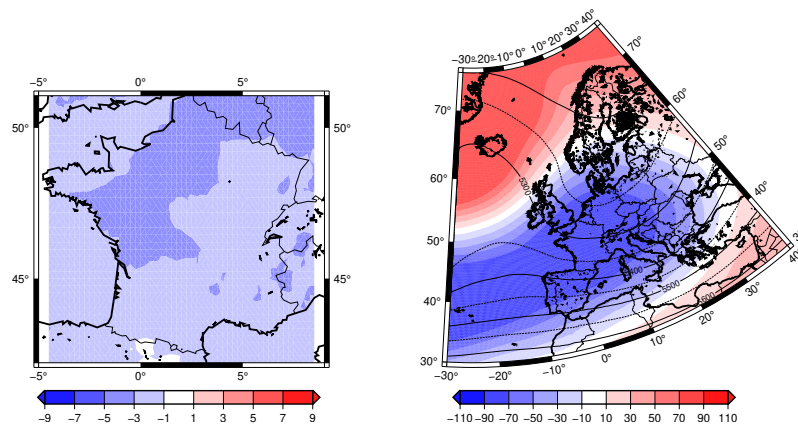


Figure 4. Left panel: mean anomalies of T2m over France between 11/12/2009 and 18/02/2010. Right panel: mean anomalies of (in m) over the East North Atlantic region (colors). Contour lines indicate the mean Z500 (in m).

238 3.3. Winter warm spells

239 The winter of 2015–2016 was particularly warm in France (Figure 5 (left)). The
 240 previous record of winter temperature occurred in 2006–2007 [52]. This warm spell
 241 had a large geographical extent in the northern hemisphere [REF]. In December 2015,
 242 no negative temperatures (in Celsius) occurred in France. This had consequences on
 243 phenological cycles of plants that require freezing temperatures in order to build defenses
 244 against pests [53].

245 The atmospheric circulation had an anticyclonic pattern over France that extended
 246 into North Africa (Figure 5 (right)) [REF]. This warm episode was responsible for air
 247 pollution due to atmospheric stagnation [54] in December 2015.

248 The atmospheric circulation of this episode was characterized by a medium local
 249 dimension and low persistence –large values of θ – (orange star in Figure 1).

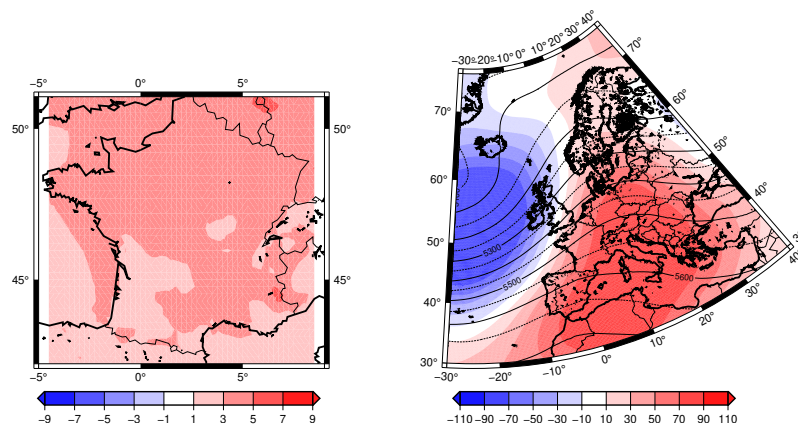


Figure 5. Left panel: mean anomalies of T2m (in K) over France between 01/12/2015 and 13/01/2016. Right panel: mean anomalies of Z500 (in m) over the East North Atlantic region (colors). Contour lines indicate the mean Z500 (in m).

250 3.4. Wet spring events

251 Most of March 2008 was rainy. The daily maxima were not exceptional, but the
 252 cumulated precipitation flooded northern and central France.

253 The atmospheric circulation yielded a persisting cyclonic pattern (Fig. 6 (right))
 254 that conveyed moist air into France.

255 The atmospheric circulation of this episode was characterized by a medium local
 256 dimension and medium persistence (yellow diamond in Figure 1).

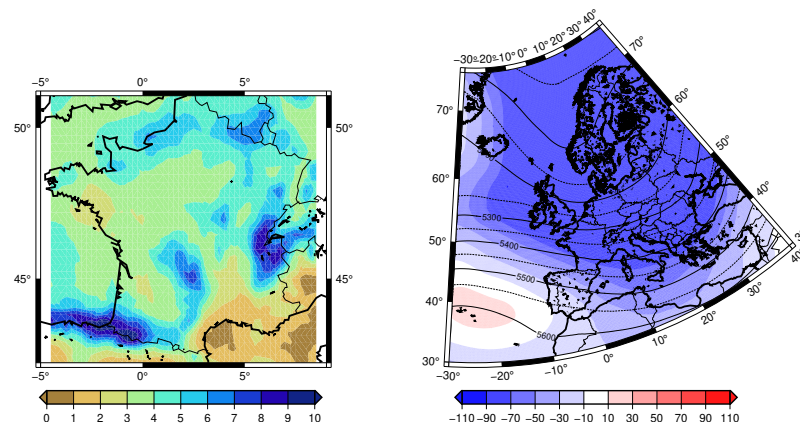


Figure 6. Left panel: mean precipitation rate (in mm/day) over France between 09/03/2008 and 01/04/2008. Right panel: mean anomalies of Z500 (in m) over the East North Atlantic region (colors). Contour lines indicate the mean Z500 (in m).

257 3.5. Wet winter events

258 South eastern France witnessed short and intense precipitation at the beginning of
 259 December 2017 (storm Ana). The end of December 2017 had a longer spell of precipi-
 260 tation, which hit most of France (Fig. 7 (left)), due to a spate of storms (named Bruno,
 261 Dylan, Carmen and Eleanor).

262 The atmospheric circulation yielded a zonal pattern (Fig. 7 (right)) that brought
 263 intense and prolonged precipitations in the southern half of France, along with the
 264 storms.

265 The atmospheric circulation of this episode was characterized by a medium local
 266 dimension and medium persistence (green diamond in Figure 1).

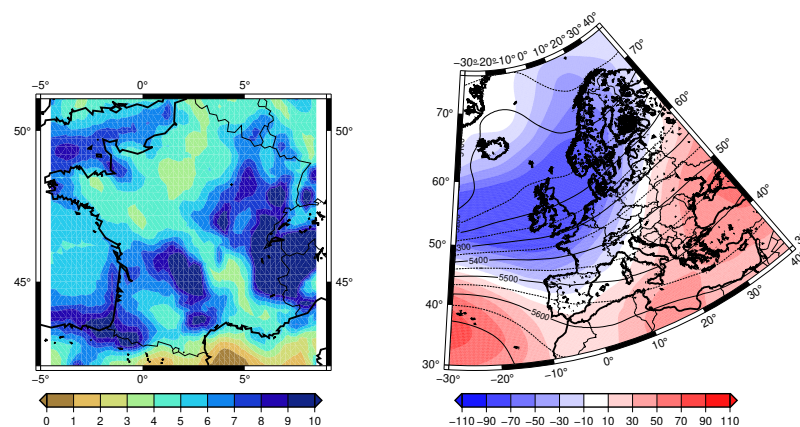


Figure 7. Left panel: mean precipitation rate (in mm/day) over France between 25/12/2017 and 09/01/2018. Right panel: mean anomalies of Z500 (in m) over the East North Atlantic region (colors). Contour lines indicate the mean Z500 (in m).

267 3.6. Fall Mediterranean events

268 Mediterranean precipitation events occur in the Fall season over the mountainous
 269 regions of the Mediterranean arc, when the Mediterranean sea is still warm and high
 270 altitude air has cooled down. Such conditions create strong convective events with
 271 devastating effects.

272 The Fall of 2019 witnessed a large number of Mediterranean events, with a cli-
 273 max in October 2019, during which the Aube region (south east of France) witnessed
 274 catastrophic floods that lead to many casualties.

275 During the October events, the atmospheric circulation had the conjunction of a
 276 cyclonic pattern in the eastern North Atlantic and an anticyclonic pattern over central

277 Europe (Fig. 8 (right)). This meridional circulation pumped moisture into southern
 278 France, with a large amount of precipitation during almost two weeks.

279 The atmospheric circulation of this episode was characterized by a high local
 280 dimension and low persistence –high values of θ – (cyan diamond in Figure 1).

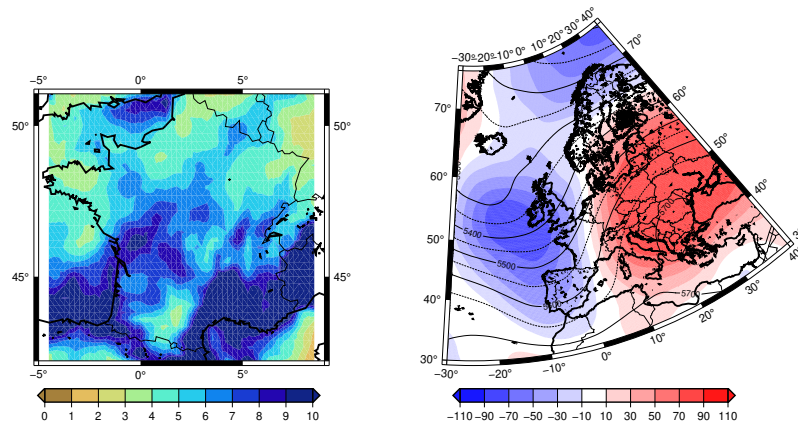


Figure 8. Left panel: mean precipitation rate (in mm/day) over France between 13/10/2019 and 25/10/2019. Right panel: mean anomalies of Z500 (in m) over the East North Atlantic region (colors). Contour lines indicate the mean Z500 (in m).

281 3.7. Winter storms

282 Winter North Atlantic storms occur every winter. They start near the Labrador Sea
 283 and move across the North Atlantic, following the storm track [55,56]. A few of them
 284 can hit France or Europe, especially when the atmospheric circulation is zonal.

285 The winter of 2019/2020 was particularly stormy, with 10 named storms that hit
 286 France between December 2019 and February 2020. Three named storms (Atiyah, Daniel
 287 and Elsa) hit France between Dec. 6th and Dec 12th 2019, with wind speeds exceeding
 288 150km/h. During that week, the atmospheric circulation was zonal (Fig. 9 (right)), with
 289 cyclonic conditions over Iceland, and anticyclonic conditions near the Azores.

290 The atmospheric circulation of this episode was characterized by a medium local
 291 dimension and low persistence –high values of θ – (light blue triangle in Figure 1).

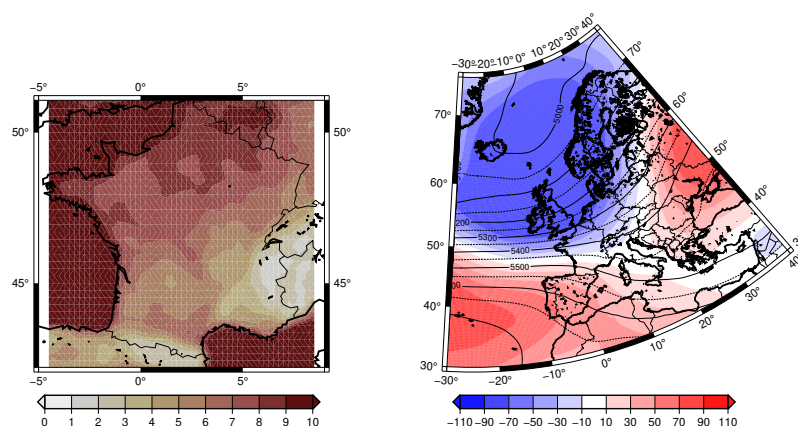


Figure 9. Left panel: mean maximum daily wind speed (in m/s) over France between 06/12/2019 and 15/12/2019. Right panel: mean anomalies of Z500 with respect to the seasonal cycle (in m) over the East North Atlantic region (colors). Contour lines indicate the mean Z500 (in m).

292 This spate of storms also brought a lot of precipitation over France, and caused
 293 local floods.

294 4. Results

295 We show how the dynamical features (dimension d – persistence metric θ) of the
296 atmosphere for the seven emblematic events change in scenario simulations.

297 The present-day values of those dynamical parameters are shown in Figure 1,
298 obtained with the ERA5 reanalysis (1979-2019). We observe that the different events
299 occupy preferential regions of the $d - \theta$ diagram. In particular, the majority of the
300 points for the cold spell of 2010 have high persistence (low θ) and low dimension, which
301 corresponds to the negative phase of the North Atlantic Oscillation [10]. The 2019 wet
302 event occupies a region characterized by blocking patterns. For the other events the
303 identification of the phase space region is not straightforward but one has to keep in
304 mind that in summer the variance of the cloud of points is reduced around the median
305 values. The results discuss the changes of those parameters d – θ for a climate scenario, in
306 CMIP6 simulations.

307 As discussed in Section 2.3, we determine d – θ for the seven events of the beginning
308 of the 21st century, assuming they occur toward the end of the 21st century. We examine
309 the influence of warming on the dynamical indicators by subtracting the trend in Z500.
310 Hence, the "raw" changes of d – θ indicate, to a first order, a thermodynamical contribution
311 of climate change to the dynamics of the event. The d – θ changes with the "detrended"
312 data corresponds to a change in circulation dynamical features related to predictability.

313 The results are summarized in Figure 10, which shows how the dynamical features
314 d – θ change (relatively) for each event.

315 we emphasize that the y -axes are different for panels a) and b). This implies that
316 keeping the trend produces stronger changes in the metrics, with up to 20% of variation
317 (with respect to the historical period) for d, θ in the period 2061-2100 for the circulation
318 associated with the 2003 heatwave. When removing the trends we observe smaller
319 variations of order 2 or 3% for all events. The impact of circulation changes is therefore
320 smaller than the thermodynamical change due to the increase of temperature. In general
321 changes are larger for the end of the century. However we can observe some changes of
322 sign between the two periods for the same metrics.

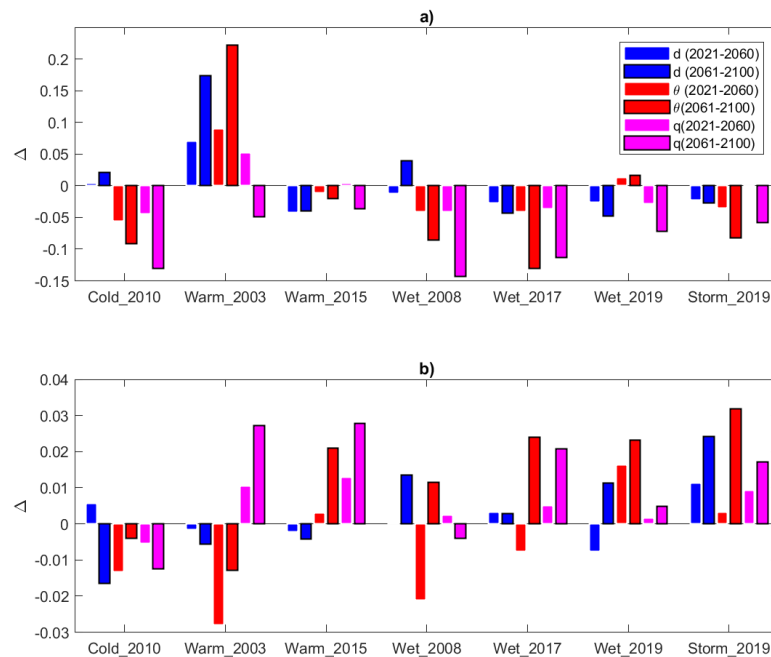


Figure 10. Upper panel (a): d , θ and analog quality (q) changes in the "raw" simulations (Δ values). Lower panel (b): d , θ and analog quality (q) changes in the "detrended" simulations. Each of the seven selected events is represented by six bars. The blue bars are for changes of local dimension d between a reference period 1979–2019 and two SSP585 scenario subperiods (2021–2060 and 2061–2100). The red bars are for changes in persistence θ . The purple bars are for changes in the distances of Z500 analogs.

323 4.1. Circulation for temperature extremes

- 324 • Cold Spell 20210. From Fig. 10 we see a modest change in the predictability, the
325 persistence of this event increase in the future (negative variation of θ). The analog
326 quality decreases, meaning that the event is less probable in future climate scenarios,
327 in terms of atmospheric circulation (not just temperature).
- 328 • Heatwave 2003. With the trend (Fig. 10a) we have increase of dimension (decrease
329 of predictability) and decrease of persistence and almost zero trends for the analogs
330 quality. We see the impact of removing the trend (Figure 10b). When removing the
331 Z500 trend, the dimension does not change while the persistence increases and the
332 event becomes more probable
- 333 • Warm winter 2015. With the Z500 trend this event has almost zero signal in the
334 three metrics (Fig. 10a). When removing the Z500 trend, this event tends to be less
335 persistent and more probable (better analogs quality) in the future (Fig. 10b).

336 4.2. Circulation for precipitation extremes

- 337 • Wet spring 2008 and wet winter 2017. We see no trend in the dimension d . The per-
338 sistence increases, and the quality of analogs decreases (Fig. 10a). When removing
339 the Z500 trend, we get contrasting signal in the metrics, in particular with better
340 analogs and hence a higher probability of occurrence of the spatial patterns.
- 341 • Wet fall 2019. The parameter changes with the "raw" data shows small negative
342 trends in dimension and analogs quality (Fig. 10a). The event become less persistent
343 when removing the Z500 trend, and also more probable with an increased analog
344 quality (Fig. 10b).

345 4.3. Circulation for a wind extreme

346 Similarly to the wet (and stormy) winter of 2017, the dynamical metrics changes
347 yield opposite signs for the stormy winter of 2019: the values of dimension and per-
348 sistence slightly (but consistently) decrease with the thermodynamical effect (due to
349 warming) (Fig. 10b), but the residual signal in dimension and persistence (due to po-
350 tential pattern changes) increases with a similar amplitude (Fig. 10b). This emphasizes
351 the importance of the dynamical signal (Fig. 10b), with respect to the thermodynamical
352 signal for the genesis of storms. The trends are enhanced between the first subpe-
353 riod (2021–2060) and the second subperiod (2061–2100) of the CMIP6 SSP585 scenario
354 simulations (blue and red bars in Figure 10).

355 5. Discussion and conclusion

356 We illustrate a statistical methodology to assess how the atmospheric circulation
357 properties can change for emblematic recent events that hit France since the beginning
358 of the 21st century, with a multi-model ensemble of climate simulations. This study
359 provides an update of the analyses of Yiou *et al.* [57] and Faranda *et al.* [9] with CMIP6
360 [14] and ERA5 [15], who essentially used the CMIP5 database [58] and NCEP reanalysis
361 [59] for attribution.

362 The results were obtained from the physical hypothesis that extreme values of
363 temperature, precipitation or wind speed in France are linked to synoptic atmospheric
364 circulation. This hypothesis is supported by quite a few studies [13,60,61].

365 The "raw" trends in the dynamical metrics reflect a general increase of the Z500
366 values (which depends on temperature), so that the quality of analogs seem to increase
367 for the summer, and decrease for the other seasons, as reflected by the theoretical study of
368 Robin *et al.* [62] or the analyses of Vrac *et al.* [63] on seasonal changes of the atmospheric
369 circulation.

370 Most of the signals conveying dynamical changes are rather small. This is due
371 to the fact that the variability of the atmospheric circulation might be underestimated
372 by climate models [64]. The trends of the dynamics in model simulations can even be
373 opposite to the trend of reanalyses for stormy events [65]. This is the main caveat of
374 our study, which relies on CMIP6 model simulations, whose spatial resolution is on
375 average ten times coarser than the resolution of ERA5. On the other hand, taking a
376 multi-model ensemble allows taking into account implicitly interdecadal variability due
377 to oceans. We find that the changes are almost always amplified with the amplitude of
378 climate change (from 2021–2060 to 2061–2100) when keeping the thermodynamic trend.
379 This enhancement suggests the robustness of the results (which are not obtained "at
380 random"). However, we cannot exclude that interdecadal variability affects the results
381 when removing the trend.

382 Our results should be interpreted as changes in the likelihood of synoptic circula-
383 tion (quality of analogs) and predictability of events ($d-\theta$) as a response to climate
384 change, conditional to an emission scenario (SSP585) and a set of climate simulations
385 (from CMIP6). The contrasting signals between the "raw" and "detrended" Z500 in
386 CMIP6 illustrate the relative contributions of thermodynamical and dynamical effects
387 in EEA. The orders of magnitudes are similar to what was found for a specific event
388 (winter 2013/2014) [7,66,67], albeit with other datasets. We observe that the changes of
389 dynamical indicators are rather weak for wet events.

390 This (admittedly) technical approach to the analysis of extreme events and their
391 relation with the large scale atmospheric circulation is a necessary step for *conditional*
392 *attribution* (to the atmospheric circulation) of events [4]. We emphasize that this approach
393 is very generic, and could be applied to other events, in other places of the world [e.g.
394 9,57].

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 402 on the esgf platform at IPSL (<https://esgf-node.ipsl.upmc.fr/>).

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 405 design of the study; in the collection, analyses, or interpretation of data; in the writing of the
 406 manuscript, or in the decision to publish the results.

407 Abbreviations

408 The following abbreviations are used in this manuscript:

409	CMIP6	Coupled Model Intercomparison Project phase 6
	SSP585	Socio-economic Pathway No. 5 with 8.5 W.m^{-2} forcing
	Z500	Geopotential height at 500 hPa
410	SLP	Sea-level pressure
	d	Local dimension
	θ	Persistence

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