



Reconstructing Historical Droughts in Spain (1806–1865): Atmospheric Dynamics Versus Qualitative Impact Records

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Abstract. Significant droughts occurred during the nineteenth century in the Iberian Peninsula causing severe social and economic consequences. Although such events are typically studied through quantitative indices like the Standardized Precipitation Index (SPI); in this case study, instrumental rainfall data only cover the last thirty years of the century. The limited spatial coverage of these early instrumental data justifies the use of historical documentary sources for the study of droughts during this transitional century between the purely pre-instrumental period and the modern instrumental era. This study aims to reconstruct nineteenth-century droughts in the Iberian Peninsula and to assess the extent to which documentary drought records reflect large-scale atmospheric dynamics.

In this work, we developed drought indices based on rogation ceremonies from municipal documentary sources compiled in the AMARNA® Repository and analyzed two atmospheric indicators from the 20CRv3 reanalysis: sea-level pressure and geopotential height at 500 hPa (Z500). The correlation between drought indices and the atmospheric variables reveals a sustained decrease in the sensitivity of historical drought records starting in the mid-19th century, likely due to social, political and other long-term changes. This decline delimits a more reliable period for the documentary-based analysis.

Our findings reveal that years with the most drought records were accompanied by persistent sea-level pressure and Z500 anomalies over the Iberian Peninsula, consistent with known circulation patterns associated with drought. These anomalies occurred predominantly during the preceding winter (DJF), indicating that atmospheric conditions favorable to drought had already developed several months before the late-winter/spring social response captured in the documentary record.

This underscores the value of combining documentary and reanalysis data to reconstruct pre-instrumental climate variability, while highlighting the need to account for both source reliability and temporal lags in future historical drought studies.



1 Introduction

20 Drought is a phenomenon defined as a period of persistent water deficit lasting from weeks to years, usually covering a large area and often leading to social and economic crises (Wilhite and Pulwarty, 2017). It can be classified into different categories depending on its impact on the population and the environment: meteorological drought (characterized by prolonged rainfall deficits covering a large area), agricultural and ecological drought (e.g., soil moisture exhaustion leading to crop failures and reduced primary productivity in forests) and hydrological drought (manifesting as water shortage in reservoirs, rivers and lakes) 25 (Seneviratne et al., 2021). Droughts onset and variability are governed by large-scale atmospheric circulation anomalies. These anomalies are often associated with anticyclonic systems centered over the affected region, which decrease cloudiness and inhibit rainfall (Seneviratne et al., 2021; Woollings et al., 2018).

Droughts, when measured in terms of their impacts and understood as natural hazards, show a high variability of effects. This means that the same rainfall anomaly, as a measure of natural hazard, can lead to widely differing impacts depending on 30 the vulnerability and exposure of societies to this hazard (Gil-Guirado, 2013). For this reason, quantifying the area affected by drought, as well as its start and end dates or its intensity, makes it a complex phenomenon to assess. Collectively, these factors have led to decades of research into indices designed to facilitate the monitoring of droughts. Most studies related to drought analysis have utilized three indices whose construction is based on instrumental records : the Palmer Drought Severity Index (PDSI; Alley 1984), the Standardized Precipitation Index (SPI; McKee et al. 1993) and the Standardized Precipitation 35 Evapotranspiration Index (SPEI; Serrano et al. 2010). While these indices can be calculated for earlier periods, they become much more reliable from the twentieth century onward. Prior to this period, the scarcity and lower quality of instrumental records often introduce increased uncertainty or biased results.

Studying past droughts can also be done using various methods depending on the time scale and area of interest. In some regions, high-quality instrumental series or long-term reconstructions are available. A notable example is the precipitation 40 series of Barcelona, ranging from 1786 to 2014, by Prohom et al. (2015). This series has allowed to calculate some of the indices mentioned above. For broader spatiotemporal analyses, natural proxies such as tree rings can help identify decadal or singular long drought events (Sauchyn and Skinner, 2001). The assessment of droughts in the terms described above is primarily focused on the hazard component. However, for periods predating instrumental records, historical documentation provides an opportunity to examine droughts from the perspective of their socioeconomic impacts. In this regard, documentary evidence 45 from historical archives, such as parish records, chronicles, and logbooks, constitutes a valuable source of climate information for the period prior to the instrumental era. A comprehensive inventory that lists nearly 700 documentary series covering Europe since the fifteenth century, with the highest concentration occurring after 1800, highlights the relevance of such documentary proxies (Burgdorf, 2022). Rogations are a religious liturgical mechanism to confront or alleviate some extraordinary adversity, whether social, environmental, or climatic. Within this institutional proxy, the motivations for rogations allow us to identify the 50 impacts of extraordinary environmental or climatic situations. For the Iberian Peninsula, "pro pluvia" rogations to face drought are the most frequently used type of rogation. Recent research has successfully calibrated these ceremonies against instrumental



data, confirming their reliability as drought indicators (Domínguez-Castro et al., 2008; Tejedor et al., 2019; Barriendos et al., 2024).

During the nineteenth century Spain suffered numerous droughts (Moruno, 2020, 2026) that severely affected the population and its territories. This century generally presents difficulties for drought reconstruction for two reasons. First, there is a decline in the interest of recording rogation ceremonies even though they continue to take place, leading to potential gaps and biases. Secondly, this period also corresponds to the early stage of institutional meteorological records, characterized by a limited number of observation points and still-developing methodologies. For example, before the late nineteenth century the SPI can reliably be applied only to the Barcelona (start: 1786) and Cadiz (start: 1817) precipitation series. Although the index can be calculated for a single location, the spatial coverage of the available instrumental data is insufficient to support a robust regional analysis before the 1860s–1870s, thereby justifying the use of historical documentary sources to study droughts throughout this transitional century.

2 Data and methods

2.1 Atmospheric drivers and variable selection

Among existing atmospheric variables to study droughts, two were used: Geopotential height at 500 hPa (Z_{500}) and sea-level pressure (SLP). A persistent positive Z_{500} anomaly is a robust indicator of an anticyclonic state, which makes it a common variable for assessing atmospheric drivers of drought (Peña et al., 2022; Bakke et al., 2023; Kingston et al., 2015). Additionally, SLP characterizes the lower tropospheric structure and quantifies the equivalent barotropic nature of the system and identify signs of persistent atmospheric blocking. Together, Z_{500} and SLP provide a complementary vertical perspective.

Beyond atmospheric anomalies, thermodynamic processes are also important drivers of drought dynamics. In particular, land-atmosphere feedbacks can amplify and sustain drought conditions independently of the prevailing circulation. High temperatures increase evapotranspiration demand, leading to soil moisture depletion which in turn enhances surface warming and creates a positive feedback loop. Other mechanisms in the literature explore the link between land-atmosphere feedbacks and precipitation deficits (Guillod et al., 2015; Herrera-Estrada et al., 2019).

An alternative approach is to directly analyze precipitation observations which provide complementary insights; however their utility is often constrained by heterogeneous spatial coverage and limited historical availability. These limitations will be explored further in Section 2.2.3 .

2.2 Data

2.2.1 20th Century Reanalysis V3

To obtain values of physical variables such as Z_{500} and SLP from the nineteenth century, reanalysis data can be utilized to compensate for the lack of systematic observational data during this period. The Twentieth Century Reanalysis Version 3 (20CRv3) provides such data for the 1806–2015 period (Slivinski et al., 2019). The reanalysis assimilates surface pressure



observations and prescribes sea-surface temperature and sea-ice concentration as boundary conditions (Compo et al., 2011). It employs an 80-member ensemble Kalman filter, operating at T254 spectral truncation that yields horizontal resolution of $1^\circ \times 1^\circ$, with 64 vertical levels up to 0.3 *hPa*. Temporal coverage includes 3-hourly, daily, and monthly average values. Although 20CRv3 provides atmospheric fields from 1806 onwards, the 1806–1835 period corresponds to an experimental extension of the dataset and should therefore be interpreted with additional caution owing to the substantially smaller number of assimilated surface-pressure observations (Slivinski et al., 2021). This point has also been explicitly noted in more recent studies using the dataset (e.g., Imfeld et al., 2024).

In addition to Z_{500} and sea-level pressure, 20CRv3 provides precipitation amounts; however this variable is less reliable as it is entirely model-derived and unconstrained by direct observations. A preliminary comparison with the instrumental precipitation records in Barcelona (detailed in Section 2.2.3) revealed that precipitation amounts from the reanalysis did not correlate well, particularly they failed to represent precipitation extremes. Consequently, precipitation data from 20CRv3 were not considered as a drought indicator in the present analysis.

2.2.2 Drought database

The AMARNA® Repository (Arxius Multidisciplinaris per a l'Anàlisi del Risc Natural Antròpic / Multidisciplinary Archives for the Analysis of Anthropogenic Natural Risk) provides a comprehensive inventory of extreme meteorological events and their social impacts for the 1035–2025 CE period comprising a total of 43,060 cases grouped into 12,023 episodes and documented through 23,044 pages of thesaurus entries, all geographically distributed across Spain (Tuset et al., 2022; Barriendos et al., 2024). In this work, only drought records covering the nineteenth century and most of Spain were retained (see Section 2.3 for excluded regions of Spain). Thereafter, there are 2 study periods (i) the full diagnostic period, 1806–1899, and (ii) the more reliable calibration and composite period, 1806–1865. The second was taken as the capacity of rogation ceremonies to capture large-scale droughts is most reliable in this window.

Historical data were compiled from various documentary sources, including ecclesiastical institutions, municipal administrative records and private sources. Droughts in the database are mainly estimated from pro-pluvia rogation ceremonies which were organized by civil and ecclesiastical entities, thereby providing an institutional consistency to the record. However, as documented in several studies (Gil-Guirado, 2013; Gil-Guirado et al., 2020), the frequency of rogation ceremonies decreased with the gradual secularization of society; furthermore their occurrence was also sensitive to social crises such as civil wars, introducing potential gaps and biases in the record during the second half of the nineteenth century.

Each record in the database contains several variables relevant to this study. These include the type of meteorological episode (such as drought, fluvial flood or persistent rainfall, among others), the precise date of occurrence (daily resolution if available), the municipality where the event was documented, and the affected river system (stream, sub-basin and general basin). A final variable captures the intensity of the rogation ceremony associated with each record, coded on a scale ranging from 0 (no ceremony) to 5 (pilgrimage to an external sanctuary), thereby reflecting the perceived severity of the event by contemporary observers. It is worth noting that the hierarchy of these five levels of rogation is determined by a specific typology of liturgical



ceremonies, whose formality and homogeneity have been preserved across the centuries down to the present day (Martín-Vide and Barriendos, 1995; Barriendos, 1997).

2.2.3 Barcelona precipitation records

To address the lack of reliable data for precipitation amounts in the 20CRv3, the long-term instrumental precipitation series of Barcelona (Prohom et al., 2015), covering the 1786–2022 period, was incorporated into the analysis. The relationship between this series and drought variability over the Iberian Peninsula was previously established by Barriendos et al. (2024), providing a strong basis for its use here. Nevertheless, as a single-station record, this series does not capture precipitation variability across the entire peninsula. A precipitation deficit in Barcelona does not necessarily reflect drought conditions in southern Spain (Parracho et al., 2016; Rodrigo and Barriendos, 2008). Consequently, the results derived from this series should therefore be interpreted with caution and considered complementary to the spatial analysis

2.3 Study Area

The study area focuses primarily on the Mediterranean region of the Iberian Peninsula, encompassing Catalonia, Valencia, Murcia, and Andalusia regions, alongside their associated river basins. Within this geographical scope, the final boundary selection accounts for the spatial distribution of historical drought records. Although AMARNA® Repository compiles records across Spain, records are unevenly distributed, with higher concentrations in specific areas as represented by the circles' size in Fig 1. The chosen study area retain all cities with the highest number of documented droughts over the 1806–1899 period while ensuring a sufficient number of grid points for subsequent statistical analysis. Considering that the 20CRv3 reanalysis has a spatial resolution of $1^\circ \times 1^\circ$, a more precise mathematical definition of boundaries is needed and is detailed in the following. By defining the western, eastern, southern and northern borders to include the cities with the highest number of droughts (Sevilla, Murcia, Palma de Mallorca and Vic), two conditions were established on longitude and latitude. Let λ denotes longitude and ϕ latitude (in degrees), a first spatial selection defines the domain $\mathcal{D}_0$ as

$$-6 < \lambda < 4 \quad 36 < \phi < 44 \quad (1)$$

This corresponds to the blue and red crosses in Fig. 1. Grid points located in the northwestern and southwestern corners of $\mathcal{D}_0$ were removed because they do not contain cities with significant drought events over the period. Finally, a point was retained in $\mathcal{D}$ if its coordinates satisfied :

$$\lambda + 36 \leq \phi \leq \frac{7}{3} \times \lambda + 54 \quad (2)$$

where the two linear bounds were derived from the coordinates of the grid points to be excluded. This procedure yields the 84 grid points shown as blue crosses in Fig. 1. In conclusion, both drought records and atmospheric variables out of this area will not be considered.

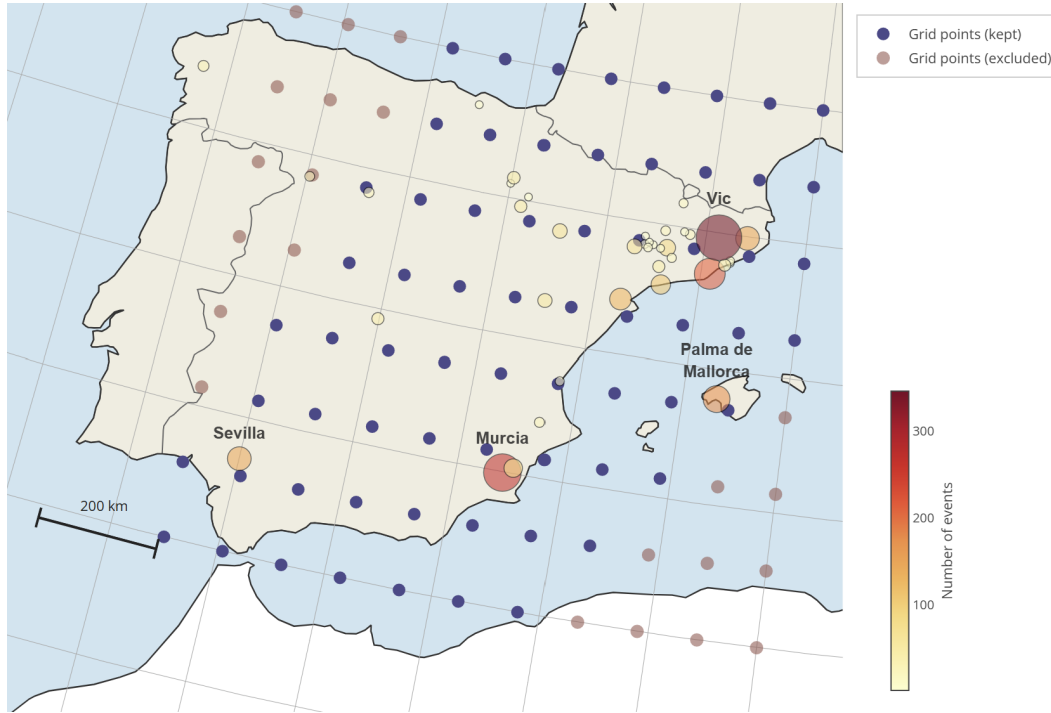


Figure 1. Drought cases documented over the Iberian Peninsula during the 1806–1899 period in the AMARNA® Repository (circles, scaled by event count). Blue crosses indicate the 84 grid points of the final study domain $\mathcal{D}$ corresponding to the $1^\circ \times 1^\circ$ grid of the 20CRv3 reanalysis. Red crosses indicate grid points within the initial bounding box $\mathcal{D}_0$ that were subsequently excluded.

145 2.4 Methods

2.4.1 Data processing

Defining drought intensity from a database is challenging and requires robust methodological choices. Four indices are introduced below, divided into two categories. Their definitions, advantages, limitations and relationships are discussed in turn. The first category is based on the number of drought cases documented per year over our study area, regardless of their spatial distribution. For each year i an anomaly index z_i is computed as

$$z_i = \frac{n_i - \mu}{\sigma} \quad (3)$$

where n_i denotes the number of drought cases documented in year i and μ and σ are respectively the mean and standard deviation of annual case counts over the 1806–1899 period. Although this index provides information on the overall societal impact and the apparent intensity and duration of drought events, it is sensitive to variations in reporting behavior and social context. Moreover, it does not account for the spatial distribution of records: a concentration of 20 records within a single municipality yields the same index value as 20 records spread across 20 different cities throughout Spain, despite reflecting very different spatial extents of drought.



To address this limitation, the second index is based on the number of unique municipalities affected by drought in a given year over the study area. The municipality level was chosen because municipal councils were the authorities responsible for deliberating on the need for a rogation ceremony, deciding upon it, and transmitting the order to the ecclesiastical authority for execution. The anomaly is computed using the same formulation, with μ and σ now derived from the annual count of unique affected municipalities over the study period. However, this index does not distinguish between a year in which 200 cases are documented across 20 municipalities and a year in which only 20 cases are documented across those same municipalities.

The second category applies the same two indices but restricts the analysis to records occurring in the first five months of the year. This choice was greatly motivated by the potential seasonal bias of historical drought records. Seasonal analysis of rogation ceremonies has previously proven effective for capturing drought variability in central Spain (Domínguez-Castro et al., 2008). Documentary evidence reflects the social perception and impacts of drought rather than meteorological conditions directly, and its timing may therefore be influenced by agricultural calendars, water-demand cycles and institutional responses, a prayer is most likely to succeed when people most want rain (Espín-Sánchez and Gil-Guirado, 2022; Espín-Sánchez et al., 2026). Table 1 presents the seasonal distribution of drought records for the full 1806–1899 period in the AMARNA® Repository. Records are strongly concentrated in winter and spring, accounting for 70.2% of the total. Excluding December records, which represent only 4% of the total, does not significantly change this pattern. Restricting the index to the January–May window, therefore, captures the majority of the drought signal while providing a temporally consistent basis for comparison with atmospheric variables at specific dates, as discussed in the following sections. In conclusion, drought anomaly indices only use the municipality affected, year and month of occurrence from the database. An index based on the intensity of rogation ceremonies was also evaluated, but it yielded virtually identical results to the case-count index, likely due to the high collinearity between these two variables. It was therefore excluded from further analysis. Similarly, an index taking into account the duration of the drought episode could be explored but need an aggregation of drought records from a same period.

Table 1. Seasonal distribution of drought records in the AMARNA® Repository, expressed as a percentage of annual totals. Seasons are defined as: Winter (December–February), Spring (March–May), Summer (June–August), Autumn (September–November).

Period	Seasonal Distribution (%)			
	Winter (DJF)	Spring (MAM)	Summer (JJA)	Autumn (SON)
All years (1806–1899)	20.2	50.0	11.0	18.8

To compare droughts with atmospheric variables during each event, standardized anomaly indices were constructed for each atmospheric field. Let $P_{i,m,y}$ denote the sea-level pressure at grid point i , month m and year y , where each grid point i is located at latitude λ_i and longitude ϕ_i . Let $\mu_{i,m}$ and $\sigma_{i,m}$ be the mean and standard deviation of P at grid point i for month m , computed over the 1806–1899 reference period. It is important to note that the mean and standard deviation of none of the variables changed significantly over the entire century; consequently, anomalies can also be calculated using different reference periods and yield similar results. The standardized anomaly is defined as



$$185 \quad z_{i,m,y} = \frac{P_{i,m,y} - \mu_{i,m}}{\sigma_{i,m}} \quad (4)$$

The regional index $Z_{m,y}$ is then obtained by computing the cosine-latitude-weighted spatial mean over the study domain $\mathcal{D}$:

$$Z_{m,y} = \frac{\sum_{i \in \mathcal{D}} z_{i,m,y} \cos(\phi_i)}{\sum_{i \in \mathcal{D}} \cos(\phi_i)} \quad (5)$$

The same procedure is applied to the 500 hPa geopotential height field Z_{500} . For the instrumental precipitation series in Barcelona, the same standardization was used, but no spatial average was required.

190 Since monthly anomalies are available for all 12 months of each year, seasonal and five-month anomalies were computed and are used hereafter. Computing such averages is further justified by the physical conditions underlying drought formation; drought events typically result from persistent atmospheric conditions lasting several months rather than from short-duration anomalies, making seasonal or multi-month averages more physically meaningful predictors than individual monthly values.

2.4.2 Atmospheric drivers and statistical analysis

195 To assess the relationship between SLP and Z_{500} and drought anomalies, both Spearman and Pearson correlations were computed as they both carry their own limitations on those types of data (detailed in Section 3.3). In addition, there might be a temporal offset between the appearance of atmospheric anomalies and drought anomalies. The database likely compiles records of different drought types, which respond to atmospheric forcing on different timescales. Meteorological droughts, defined primarily by precipitation deficits, tend to manifest shortly after the atmospheric anomaly, whereas agricultural droughts, characterized by soil moisture depletion, may lag by several months (Fuentes et al., 2022). Since no uniform offset can therefore be assumed, aggregating anomalies over seasonal or multi-month windows is more appropriate than relying on individual monthly values. Hydrological droughts which can occur several months or years later, are less likely to be recorded because they depend on the water level of natural reservoirs or infrastructures which were less common at that time.

To further assess the robustness of these correlations, two bootstrap resampling procedures were applied. In both methods, atmospheric and drought anomalies were resampled jointly at each iteration to preserve the covariance structure between the two series. The first method consists of drawing, at each iteration, a random year from the full record. Although this approach assumes that observations are independent and identically distributed, which may be violated in the presence of autocorrelation, it enables the subsequent identification of individual years that exert a disproportionate influence on the overall correlation. The second method draws blocks of three consecutive years at each iteration, which is more physically meaningful, as it accounts for potential inter annual persistence in atmospheric and drought conditions. For both methods, the bootstrap estimate of the correlation is defined as:

$$210 \quad \hat{\rho}^* = \frac{1}{B} \sum_{b=1}^B \hat{\rho}^{(b)} \quad (6)$$



where B is the number of iterations and $\hat{\rho}^{(b)}$ is the correlation coefficient computed on the b -th resample. Given the relatively small sample size ($n = 94$) and the presence of years with high-magnitude anomalies, the bias-corrected and accelerated (BCa) method is used to compute bootstrap confidence intervals. Unlike standard percentile-based intervals, the BCa method corrects for both bias and skewness in the bootstrap distribution, making it more reliable when the sampling distribution of the estimator is asymmetric (Efron, 1987). The interval bounds are defined by the quantiles of order

$$\alpha_k = \Phi \left(\hat{z}_0 + \frac{\hat{z}_0 + z_{\alpha_k}}{1 - a(\hat{z}_0 + z_{\alpha_k})} \right) \quad (7)$$

where Φ denotes the standard normal cumulative distribution function, z_{α_k} is the corresponding normal quantile, $\hat{z}_0$ is the bias-correction term, and a represents the acceleration coefficient.

Autocorrelation is calculated for lags of 1 to 3 years and reported in Table 2, for drought anomalies (computed from the number of unique municipalities affected) and both Z_{500} and SLP anomalies; results for other drought and atmospheric anomalies aligns with these findings. Following the standard approximation for white noise processes, autocorrelation coefficients smaller than $\frac{1.96}{\sqrt{n}}$ are considered insignificant, corresponding to a threshold of 0.210 for $n = 94$. Only sea-level pressure anomalies present an absolute value of autocorrelation coefficient greater than 0.21 for a lag of 1 year. Drought anomalies and Z_{500} data support the assumption of independence underlying the single-year resampling method.

To assess the temporal offset discussed previously, the bootstrap procedure was applied considering different temporal lags.

Table 2. Autocorrelation coefficients of drought anomalies (number of unique municipalities affected), Z_{500} and sea-level pressure (SLP) anomalies at lags of 1 to 3 years. Absolute values above 0.21 (dashed threshold, $1.96/\sqrt{n}$, $n = 94$) are significant at the 95 % confidence level and are shown in bold. For the 1806–1865 period ($n = 60$), the threshold is 0.253, yielding consistent results.

Lag	Drought anomalies (municipalities) ^a	Z_{500} anomalies	SLP anomalies
1	0.15	−0.08	−0.25
2	−0.01	−0.17	−0.13
3	0.05	0.04	−0.04

^a Results for other drought anomalies are matches with those reported here (all < 0.21).

Finally, composite maps are computed for the Z_{500} and SLP anomaly fields over Europe for dry and non-dry years (see Section 3.5). This analysis aims to identify the dominant spatial patterns of atmospheric variability associated with drought conditions. For each atmospheric variable and each grid point, the seasonal mean was computed separately over dry years ($\mathcal{D}$) and non-dry years ($\mathcal{ND}$). The composite anomaly is defined as their difference:

$$\Delta \bar{X}_s = \frac{1}{|\mathcal{D}|} \sum_{i \in \mathcal{D}} X_{s,i} - \frac{1}{|\mathcal{ND}|} \sum_{j \in \mathcal{ND}} X_{s,j} \quad (8)$$

where $X_{s,i}$ denotes the seasonal mean of a given atmospheric variable during season s of year i , and $|\mathcal{D}|$ and $|\mathcal{ND}|$ are the respective sample sizes. Dry and non-dry years are defined in Section 3.5.



235 3 Results and Discussion

3.1 Records sensitivity over the 19th century

The AMARNA® Repository covers the entire nineteenth century; however, as previously discussed, the sensitivity of the historical records indicating drought periods decreased over time. Identifying the precise onset of this decline is challenging, as it reflects a gradual process driven by evolving social conditions rather than an abrupt change. If the reanalysis data are assumed, within their intrinsic limitations, to accurately represent atmospheric conditions throughout the study period, the correlation between SLP and Z_{500} anomalies obtained from the 20CRv3 and drought anomalies from the AMARNA® Repository should remain approximately stationary over time.

Figure 2 presents the temporal evolution of the pearson correlation coefficient between all drought anomaly indices and both SLP and Z_{500} anomalies. The correlation was computed over a 30-year moving window centered on each year y . This approach was preferred over a cumulative window, in which early years would carry disproportionate weight. Spearman correlations show the same overall temporal decline (within ± 5 years) but exhibit systematically lower values in the early part of the record. This may partly reflect the reduced reliability of 20CRv3 prior to 1835, a period corresponding to the dataset's experimental extension with fewer assimilated observations, rather than a genuine weakening of the documentary-atmospheric relationship.

To further reduce interannual noise, a 5-year moving average was applied to all series, though the overall trend remains the same. All four drought anomaly indices were considered to assess whether differences in reporting behavior are reflected in their respective correlations with atmospheric variables. It should be noted that, given the centered window, a degradation in reporting quality beginning around year y_0 will first affect the correlation estimated at approximately $y_0 - 15$, introducing a systematic 15-year lead in the apparent onset of the decline visible in the figure.

Correlations with SLP (Figure 2a) show a sustained decline from around 1855, implying that from 1870 onward, years exhibit a poorer correlation. This downward trend is not driven by isolated years of poor correlation; indeed, the correlation coefficient declines almost monotonically with each successive rolling window yielding a lower value than the preceding one, suggesting a systematic rather than an episodic deterioration. Overall, the correlation coefficient for the four indices declines by an average of 62% between its peak around 1855 and 1884.

Results for Z_{500} (Figure 2b) are broadly consistent. Correlations begin to decline around 1851, which, after correcting for the centered-window offset, corresponds to a degradation in reports' ability to describe drought situations from approximately 1866 onward, in agreement with the SLP results. A marked and sustained decline is then observed through the end of the century, with the correlation coefficient declining by 61% to 83% across the four indices. Additionally there is a significant decrease from 1831 to 1840 which drives the poor correlation observed between 1846 and 1855.

To identify changes in a historical drought-related series a Pettitt test (Pettitt, 1979) can be used (Espín-Sánchez and Gil-Guirado, 2022). This test was applied to each series (both restricted to windows of ± 10 and ± 15 years and to the full record) consistently identifies a significant breakpoint ($p < 0.01$) between 1860 and 1868 across all correlation series and both atmospheric fields, confirming that the onset of the decline is robust to the precise choice of breakpoint year (1865, after the 15



years correction). The correlation with drought anomalies based on the number of municipalities affected in a year shows a slightly earlier shift (1854–1860), consistent with its earlier visual departure from the other series in Fig. 2.

270 We suggest that this loss of sensitivity does not affect the proxy itself, but rather the documentary record that transmits it. Rogation ceremonies continued to reliably capture specific and verifiable social responses to rainfall deficits; it is the written record documenting them that progressively lost sensitivity over the course of the nineteenth century, unevenly across the cities studied, as a function of shifting social, political, and institutional contexts. A complementary hypothesis can be proposed : as documentary records reflect the social impacts of drought rather than meteorological conditions alone (Espín-Sánchez and Gil-
275 Guirado, 2022; Espín-Sánchez et al., 2026), any long-term changes in societal vulnerability, exposure, sensitivity, or adaptive capacity could also have contributed to the weakening relationship between documentary records and atmospheric drivers. It is, however, not possible to determine the relative significance of each mechanisms in the overall decrease.

3.2 Temporal evolution of drought and atmospheric anomalies

The four indices of drought anomalies defined previously reflect different physical and social realities, but they present similarities in their overall trends. Spearman correlations between the different drought indices range from 0.87 to 0.93 (all significant at
280 $p < 0.01$), indicating that neither the seasonal resolution nor the aggregation method significantly affects the resulting anomaly series. Furthermore, the indices are found to co-vary consistently across the full record, identifying the same drought years regardless of the method used. Anomalies derived from the number of municipalities affected tend to be of smaller magnitude and exhibit lower variance than those based on raw case counts, as the number of affected municipalities grows more slowly
285 during severe events.

For this reason, and in addition to the justification provided in the previous section, regarding the historical reliability of this metric, Figure 3 displays drought anomalies derived from the number of municipalities affected between January and May of each year. Four categories of atmospheric anomalies are plotted to compare their evolution with these drought anomalies: SLP and Z_{500} anomalies calculated over winter (December to February) and anomalies calculated over a five-month span (December
290 to April). The latter was chosen based on the monthly distribution of drought records in the AMARNA® Repository (See Table 1).

Four prominent drought anomalies are visible at the beginning of the century (1817, 1822, 1824 and 1825); these events have been documented in previous studies (Domínguez-Castro et al., 2012; Barriendos et al., 2024; Moruno, 2020). They are divided into two distinct episodes: the year 1817, which is described as the driest year of the 1786–2022 period in the Barcelona series
295 (Moruno, 2020) and multi year episode 1822–1825. These two episodes occurred during a major drought period that lasted from 1812 to 1825, coinciding with the Dalton Minimum and intense volcanic activity (notably the eruption of Mount Tambora in 1815).

The two types of analyzed windows (winter and five-month average) do not yield the same results, as the five-month average tend to be of lower magnitude due to the aggregation method. Additionally, for the 1806–1868 period, elevated SLP and Z_{500}
300 anomalies tend to co-occur with elevated drought anomalies. As a first approximation, we can consider instances where both the Z_{500} and drought anomalies exceed the fourth quartile, a threshold chosen to isolate the most extreme events while retaining a

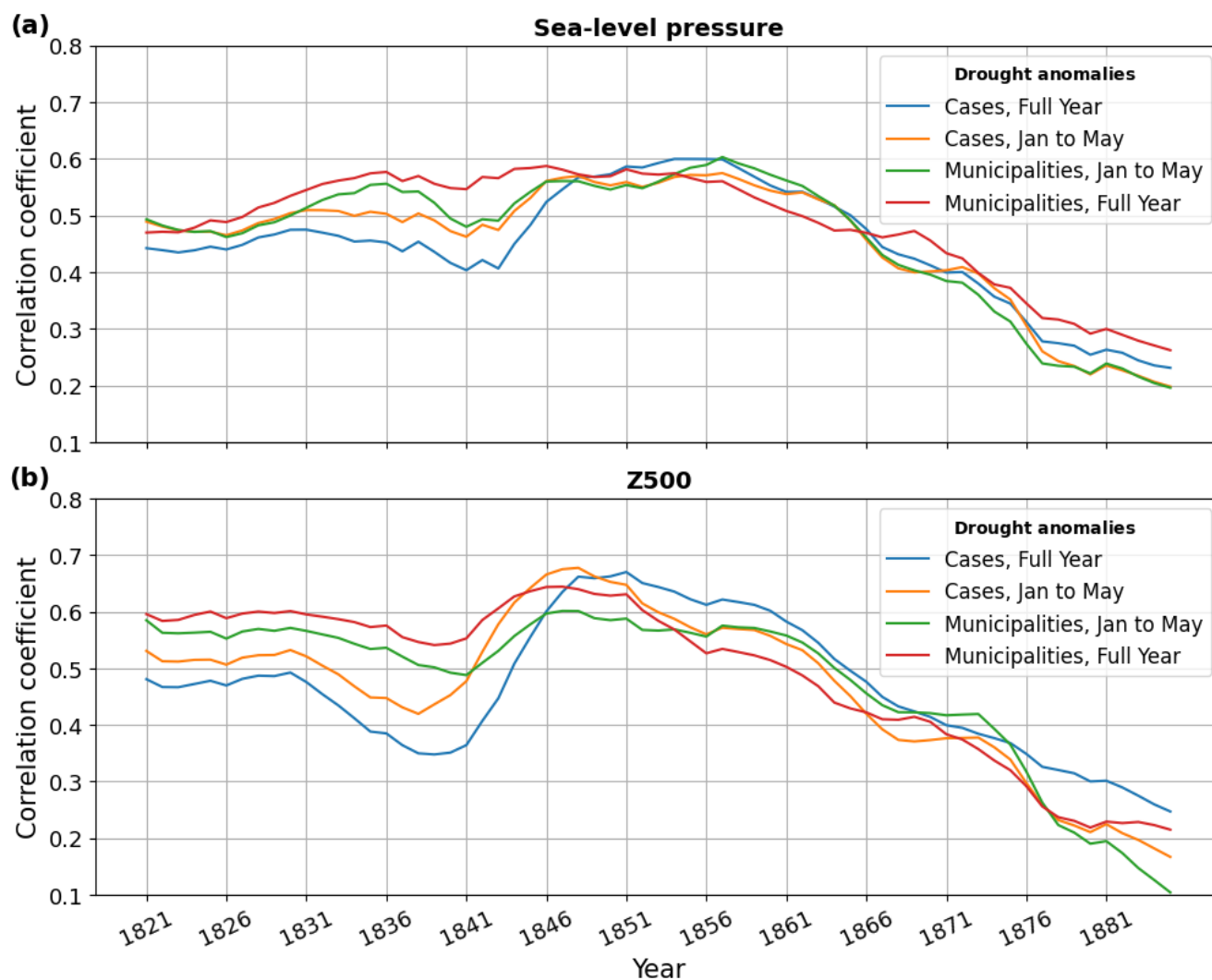


Figure 2. Temporal evolution of the Pearson correlation between standardized atmospheric anomalies and drought anomaly indices, computed over a 30-year sliding window centered on each year and smoothed with a 5-year moving average. (a) Sea-level pressure (SLP) standardized anomalies. (b) Geopotential height at 500 hPa (Z_{500}) standardized anomalies. The apparent onset of the decline is shifted approximately 15 years earlier than the true degradation in reporting quality due to the centered window (see text).



sufficient sample size, yielding cut-off values of 0.348 and 0.333. This period encompasses 16 such years. Among the years in which Z_{500} anomalies exceed 0.348, 11 out of 16 are also associated with drought anomalies exceeding 0.333 (69%). Applying the same approach to the subsequent period yields a lower proportion of 45%, which aligns with the hypothesis, formulated in the preceding section, suggesting that records from the latter part of the century are less reliable to describe drought conditions.

These figures should nonetheless be interpreted with caution. First, the sample of the 16 years with high Z_{500} anomalies is too small to draw statistically robust conclusions about the strength of the relationship. Second, while atmospheric dynamics provide a physical basis for expecting a link between positive pressure anomalies and precipitation deficits, standardized anomalies do not reflect absolute atmospheric states. Third, the drought index itself is derived from qualitative historical records, which capture societal responses to water stress rather than meteorological conditions directly; consequently, these records cannot be straightforwardly mapped onto physical atmospheric conditions. This analysis should therefore be understood as a preliminary exploration of the relationship between atmospheric and drought anomalies.

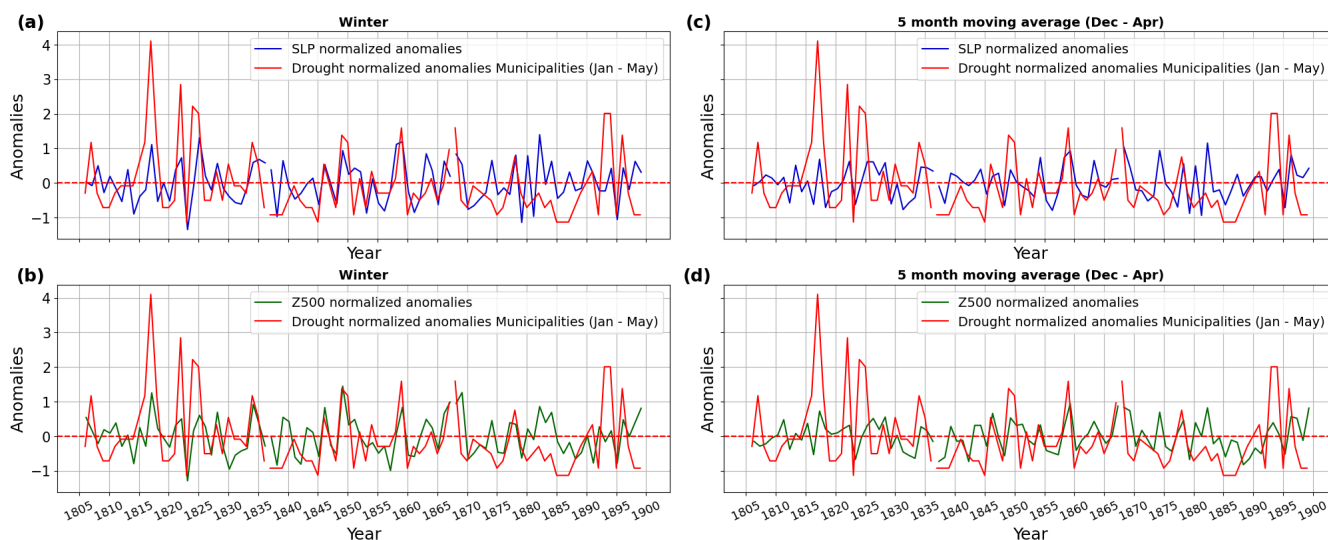


Figure 3. Temporal evolution of normalized drought anomalies based on the number of unique municipalities affected each year between January and May, compared with standardized atmospheric anomalies. Left panels (a) and (b) show anomalies computed over winter (December to February); right panels (c) and (d) show anomalies computed over a five-month window (December to April). The top panels, (a) and (c) display sea-Level pressure (SLP) anomalies, while bottom panels (b) and (d) display 500 hPa geopotential height (Z_{500}) anomalies. The horizontal dashed red line indicates the zero baseline.

3.3 Statistical correlation and limitations in the analysis

To assess the robustness of the results, Figure 4 presents a Spearman correlation matrix between the four drought indices and the anomalies of three atmospheric variables, SLP, geopotential height at 500 hPa (Z_{500}), and precipitation amounts recorded in Barcelona. The left panel shows correlations computed over the 1806–1865 period (the upper bound of which was selected as



the year beyond which drought records sensitivity begins to decline), while the right panel shows the results for the 1866–1899 period.

The choice of Spearman’s ρ over Pearson’s r reflects the nature of the drought anomaly indices. Although these indices are expressed as continuous values, they are constructed from qualitative historical records, meaning that a twofold increase in the index does not necessarily correspond to a twofold increase in drought severity. The cardinal interpretation required by Pearson’s r is therefore not warranted.

Spearman’s ρ , which depends only on the rank ordering of observations, is more appropriate in this context. It should be noted, however, that rank-based measures also carry limitations: a small to moderate difference in the anomaly index between two years does not necessarily imply a difference in their actual intensity. Pearson correlations were computed as a robustness check and yielded broadly consistent results.

All correlations from the first period are statistically significant at the 0.01 level, whereas none of those from the second period reach the 0.05 threshold. Correlation coefficients are also systematically lower in the second period, consistent with the apparent decline in proxy sensitivity discussed above. Within the first period, correlations are stronger with Z_{500} anomalies than with sea-level pressure. Correlations with precipitation, although limited by the local nature of the measurements relative to the entire country, are nonetheless significant and negative, in accordance with the physical definition of drought.

Several factors may account for the overall moderate strength of these correlations. First, the drought anomaly indices are constructed from qualitative historical records translated into quantitative values. The subjectivity inherent in each records introduces non-systematic noise into the index; unlike a physical measurement error, this noise has no systematic relationship with the analyzed atmospheric variables, and therefore reduces the observed correlation without reflecting a genuine absence of a physical signal. The limitations of indices derived from qualitative sources have been discussed in the literature by Brázdil et al. (2010).

Second, the relationship between atmospheric anomalies and drought indices is subject to temporal displacement. An anomaly leading to precipitation deficit can have an influence on soils, which play a determining role in harvest yields, months later. Moreover, soil moisture feedback and hydrological memory mean that a drought may itself reinforce and prolong the conditions that generated it (Fuentes et al., 2022). Both mechanisms introduce a lag between the atmospheric signal and the reported drought that seasonal averaging only partially captures.

Lastly, geopotential height and sea-level pressure are not the sole atmospheric drivers of drought in the Iberian Peninsula. Blocking anticyclones, sea-surface temperature anomalies in the Atlantic and Mediterranean, and large-scale teleconnections such as the North Atlantic Oscillation are known to modulate precipitation variability independently (Bladé et al., 2012).

3.4 Quantifying correlation uncertainty and synoptic drivers of outlier droughts

Bootstrap method described in section 2.4.2 was applied to the Spearman correlation between winter Z_{500} anomalies and drought anomalies derived from the number of municipalities affected over the 1806–1865 period ($n = 60$). The BCa confidence intervals are [0.317, 0.720] with a 0.54 mean for the 1-year block and [0.348, 0.706] with a 0.55 mean for the 3-year block. The 1-year scheme yields a broader interval, this is expected because when years are resampled independently, a single

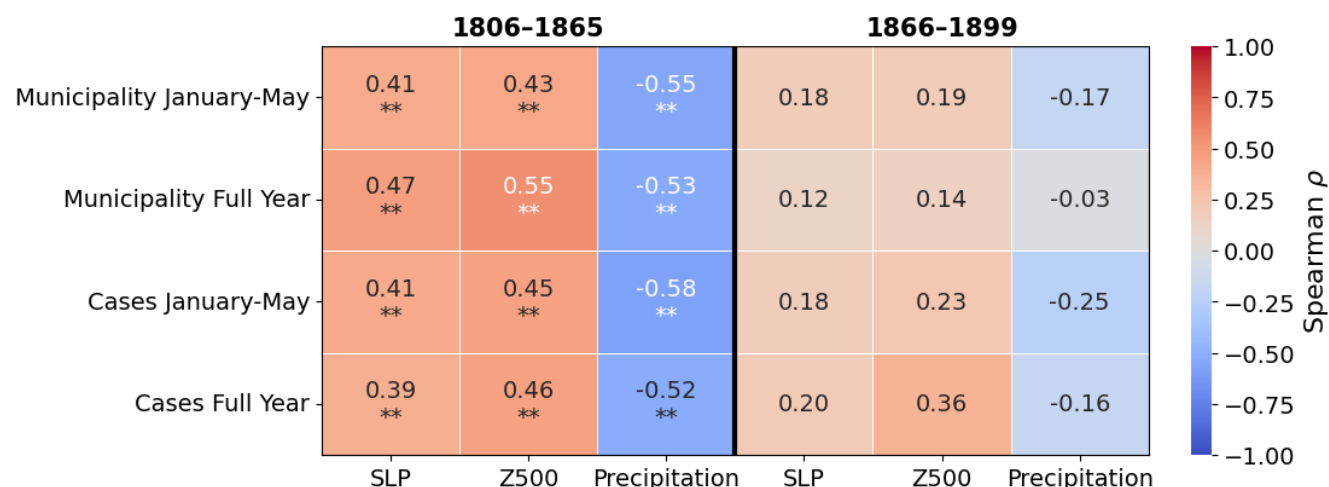


Figure 4. Spearman correlation coefficients (ρ) between drought anomaly indices computed from the AMARNA® Repository and three atmospheric variables: geopotential height at 500 hPa (Z_{500}) and sea-level pressure (SLP) from the 20CRv3 reanalysis, and the precipitation amount recorded in Barcelona. Left panel: correlations computed over the 1806–1865 period. Right panel: correlations computed over the 1866–1899 period. A single asterisk (*) indicates significance at the 0.05 level; a double asterisk (**) indicates significance at the 0.01 level.

influential year can substantially shift the correlation in a given resample, thereby increasing the variability across resamples and widening the interval. Nevertheless, the intervals remain wide in both cases, and a 0.4 difference between two different ρ is physically substantial, underscoring the uncertainty in the estimate.

To further characterize this uncertainty, the 1-year resampling results were examined in detail. To investigate the physical drivers of low-correlation years, the following analysis focuses on two years : 1830, which appear the most in bootstrap simulations with a ρ lower than 0.32 (lower bound of the 95% interval), and 1824, a year with high drought anomalies (ranking as the fourth largest of the century according to the drought index based on the number of municipalities affected), but with winter Z_{500} anomalies ranked only 19th. It is important that these years have positive anomalies, indeed, a year with no documented droughts and a high atmospheric anomaly may reflect data gaps or social underreporting, which would require a dedicated historical analysis. By contrast, a year with a strong drought anomaly and a weak atmospheric anomaly is more likely to reflect a physically interpretable decoupling between surface conditions and large-scale circulation or simply a bias in the seasonal calculation of drought anomalies.

Figures 5 and 6 present the mean atmospheric configuration for each month of 1824 and 1830 respectively. Normalized sea-level pressure anomalies (color contours) and Z_{500} (contour lines) are both represented. Red shading indicates positive pressure anomalies (high-pressure conditions), while blue shading denotes negative anomalies (low-pressure conditions).

The drought signal for 1824 is strong across all indices: regardless of the method used, records consistently indicate a drought year, with the majority of records concentrated in the first five months (accounting for 73% of annual records). Spatially, these records cluster in the north-eastern part of Spain.

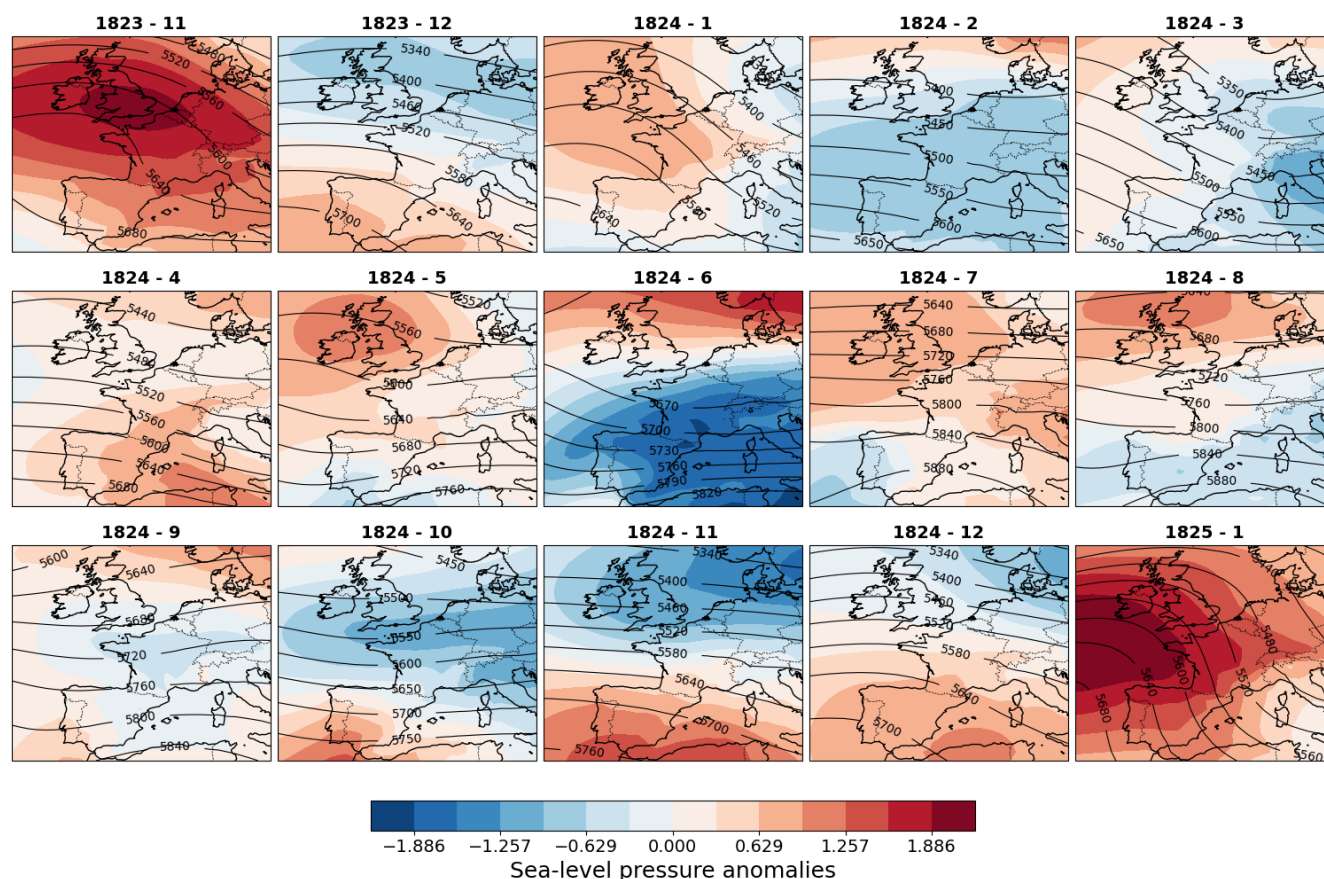


Figure 5. Sea-level pressure normalized anomalies (color contours), geopotential height at 500 hPa (black contours) on each month of 1824. Obtained from Source : NOAA CIRES Twentieth Century Global Reanalysis V3

In November 1823 (see Figure 5), strong positive SLP anomalies are visible over north-eastern Spain, and the Z_{500} contour lines suggest a blocking configuration inhibiting Atlantic moisture from reaching northern Spain.

However, the mean December–February atmospheric anomalies calculated for 1824 appear weak, resulting in a structural divergence between the reconstructed winter atmospheric circulation and the documented drought severity. Rather than an analytical artifact, this discrepancy highlights two structural constraints of the analytical framework. First, restricting the atmospheric aggregation window to December–February doesn't fully capture the multi-month accumulation of deficits, particularly during early spring, for example here in April 1824 where there was a situation likely unfavorable to rain. Second, qualitative historical archives reflect socio-economic impacts and reporting sensitivity rather than strictly physical metrics. Following the extreme drought of 1822 (Domínguez-Castro et al., 2012; Barriendos et al., 2024), heightened social memory and public awareness may have amplified the historical recording of subsequent, even moderate, rainfall deficits.



Consequently, the case of 1824 serves as a boundary case that highlights the limits of relying solely on early-winter atmospheric averages to capture complex, multi-seasonal drought events recorded in qualitative historical sources.

The drought signal for 1830 differs across the four indices. When the full year is considered, it reaches an anomaly value of approximately 1.5, ranking 1830 as the seventh year with the most records in the century. However, when only the first five months are considered, the anomaly decreases to 0.5, placing it close to the climatological mean. This suggests that droughts were mainly documented in the second half of the year; yet, since the atmospheric anomalies computed in this study are restricted to the December–April window, any late-year signal falls outside the analysis period and cannot be captured. This interpretation is supported by the seasonal distribution of drought records in the AMARNA® Repository: 45% of all records are concentrated in April and May, 32% in October and November, and only 2% span January through March, confirming that the late-year signal is both real and systematically excluded from the December–April anomaly window.

As shown in Figure 6, October 1830 exhibits atmospheric conditions highly conducive to drought, with strong positive SLP anomalies along the western Mediterranean coast. The corresponding Z_{500} field reveals a powerful anticyclonic ridge centered over western Europe, indicative of a blocking configuration that suppresses convective ascent and deflects Atlantic fronts poleward, thereby inhibiting precipitation over the Iberian Peninsula and southern France (Woollings et al., 2018). This pattern persists into November at reduced intensity, likely sustaining a cumulative rainfall deficit over the region. In contrast, atmospheric conditions during the winter show no clear atmospheric patterns conducive to drought, which is consistent with the near-absence of droughts records for these months in 1830.

In conclusion, the case of 1830 clearly illustrates the limitations of the analysis. The three-month average largely misses any drought signal from late spring onward, while the five-month index, although capturing late-spring conditions, remains blind to the drought-favourable atmospheric configurations during autumn. Although autumn records are not common in the AMARNA® Repository, this season can play a non-negligible role in modulating harvest outcomes through soil moisture recharge and the conditioning of winter-sown crops (Peña Gallardo et al., 2019).

It should be noted that SLP and Z_{500} primarily serve as indicators of meteorological drought by capturing large-scale atmospheric configurations that inhibit precipitation. However, they do not account for hydrological drought, which develops with a temporal offset as precipitation deficits propagate through soil moisture, streamflow, and groundwater systems (Gaona et al., 2022; Fuentes et al., 2022). Consequently, the absence of significant atmospheric anomalies does not preclude the occurrence of drought. Taken together, the cases examined here illustrate both the potential and the limitations of atmospheric reanalysis as a framework for investigating historical droughts: while SLP and Z_{500} anomalies provide valuable insights into the synoptic drivers of meteorological drought, they offer only a partial view of a complex phenomenon.

3.5 Atmospheric patterns of years with documented droughts

Composite maps were restricted to the 1806–1865 period, after which the quality of drought records degrade substantially (see Section 3.1). A year is classified as dry if its standardized anomaly in the number of municipalities affected exceeds 0.59, which corresponds to at least 12 affected municipalities in the raw record prior to normalization. This threshold yields a dry-year sample of $|\mathcal{D}| = 21$ years and a non-dry-year sample of $|\mathcal{N}\mathcal{D}| = 39$ years; raising this threshold would reduce the dry-year

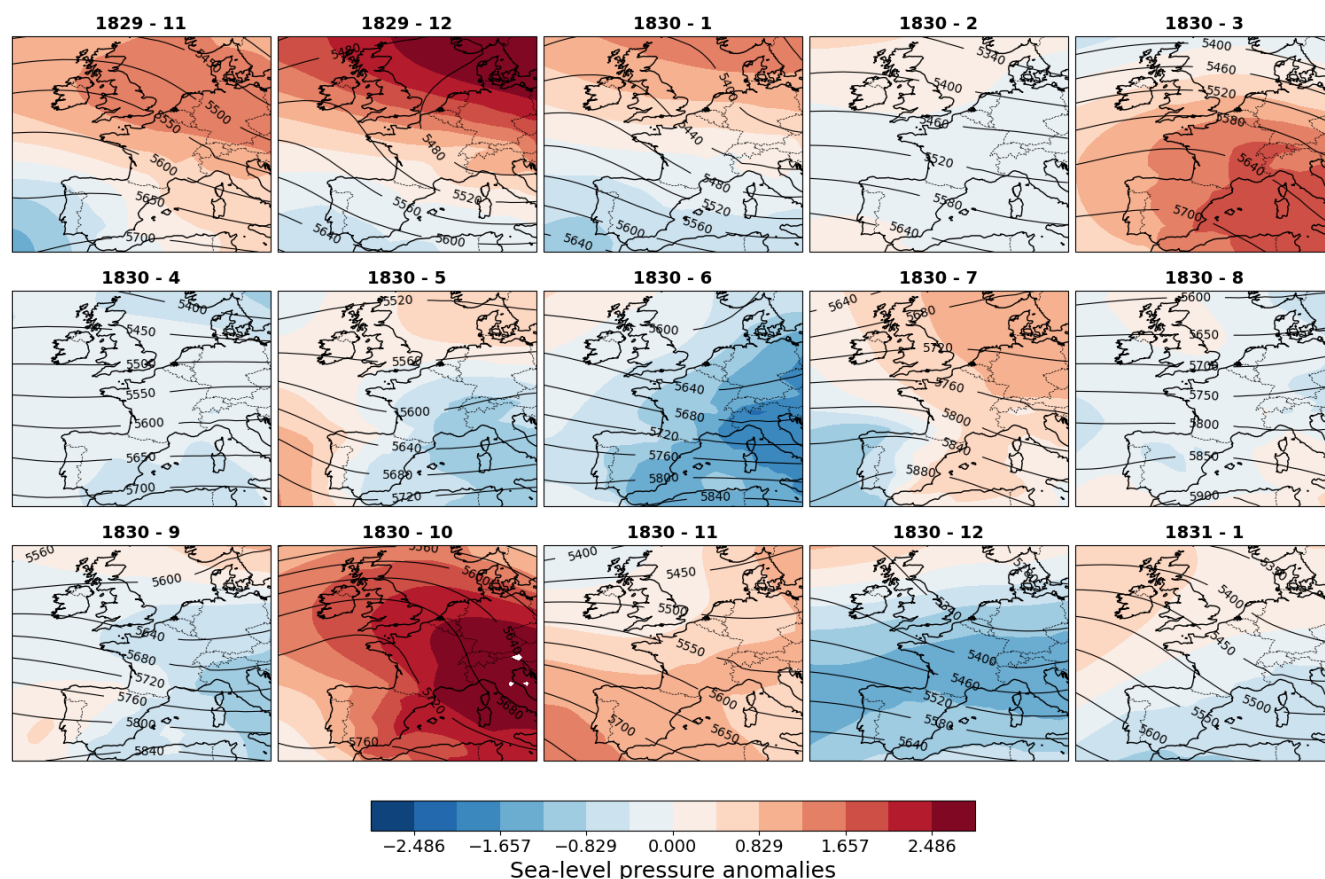


Figure 6. Sea-level pressure normalized anomalies (color contours), geopotential height at 500 hPa (black contours) on each month of 1830. Obtained from Source : NOAA CIRES Twentieth Century Global Reanalysis V3

sample to fewer than 17 years, leaving several seasons with insufficient cases for robust composites. The composite patterns are robust to the choice of drought index: 18 of the 21 dry years are common to all four indices, and only two years are absent
415 from the three remaining indices, confirming that differences between indices are concentrated near zero anomalies. Because the two sub-samples are unequal in size and their variances cannot be assumed to be equal, the statistical significance of the composite differences was assessed using a two-sample Welch's *t*-test.

Figure 7 presents the composite anomaly maps computed for the four seasons of a dry year and for the autumn of the preceding year. The latter was included to assess whether the antecedent atmospheric configuration in autumn provides insight
420 into drought occurrence during the following year. Black stippling indicates grid points where the difference in seasonal means is statistically significant (using Welch's *t*-test, $p < 0.05$). The colorbar scale is held constant across all seasons to facilitate a direct comparison of the relative magnitude of the seasonal anomalies.

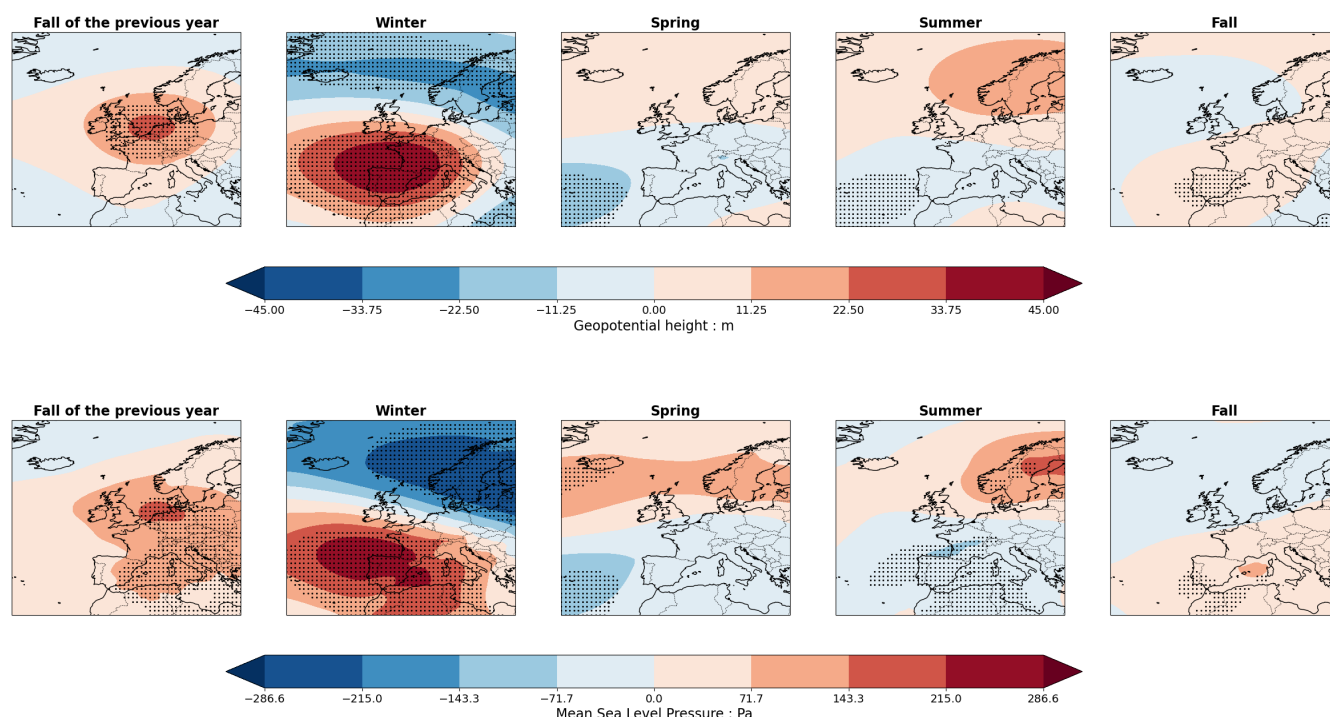


Figure 7. Seasonal composite differences in atmospheric mean state between dry years (standardized drought anomaly > 0.59 based on the AMARNA® Repository) and non-dry years. Black points represent area where the welch $p_{value} < 0.05$. Top panels (a) represent Z_{500} differences and bottom panels (b) represent SLP differences. Obtained from Source : NOAA CIRES Twentieth Century Global Reanalysis V3

The most prominent signal is observed in winter, when both Z_{500} and SLP exhibit large positive anomalies over the Iberian Peninsula, parts of southwestern France, and the adjacent Atlantic margin. Positive anomalies of Z_{500} and SLP are associated with anomalously strong anticyclonic circulation, which promotes subsidence and suppresses precipitation over these regions. In contrast, spring and summer show negligible differences between dry and non-dry years, with few significant grid points. The autumn of the year preceding a drought displays a signal with both a lower amplitude and a reduced spatial extent of significant anomalies compared to winter, suggesting that antecedent atmospheric conditions may contribute to drought development but to a lesser extent than those in winter

Similarly, the autumn of the dry year itself shows modest positive anomalies. This aligns with the presence, in the AMARNA® Repository, of drought records extending through the end of the calendar year (18% in autumn), reflecting the role of autumn precipitation in replenishing soil moisture and sustaining agricultural yields (Peña Gallardo et al., 2019).

In summary, while Table 1 shows that drought events in the AMARNA® Repository were predominantly documented during spring, the accompanying atmospheric conditions unfavorable to precipitation occurred mainly during winter. A plausible explanation for this discrepancy lies in the documented lag between precipitation deficits and their consequences for soil: a



winter precipitation deficit first depletes soil moisture and water reserves, and its effects on the main cultivated crops become apparent only in spring (Fuentes et al., 2022). Since spring soil moisture has been identified as a primary driver of crop yield reductions (Gaona et al., 2022), a dry winter likely translates into agricultural stress at the time of the spring harvest, which is when shortages would have been most acutely perceived and recorded.

440 These findings are also consistent with the seasonal functions of rogation ceremonies identified by Espín-Sánchez and Gil-Guirado (2022). Autumn rogations were primarily evaluative, as they were typically initiated only after a sustained rainfall deficit had become evident following the dry months of September–October, with ceremonies usually taking place in November–December. In contrast, spring rogations appear to have had a more preventive character, often being initiated in response to signs of insufficient rainfall during the preceding winter. This is illustrated in Fig 7: a year is considered dry when
445 rogation ceremonies were held in at least 12 distinct municipalities, and since more than half of these ceremonies typically occur in spring, a dry year is generally characterized by a high number of spring rogation ceremonies. These ceremonies are shown to be accompanied by positive atmospheric anomalies during the preceding winter, potentially indicative of drought conditions. Rogation ceremonies in spring were thus preceded by a probable accumulation of precipitation deficit.

To assess the robustness of the temporal lag, the bootstrap procedure applied previously was repeated; however, instead of
450 calculating the correlation between winter atmospheric anomalies and drought anomalies, correlations were computed between drought anomalies and atmospheric anomalies averaged over distinct three-month periods ($n = 10000$ resamples). A lag of zero was assigned to the spring period (MAM), as most documentary records fall within this season. Table 3 presents the Spearman correlation coefficients between Z_{500} anomalies, averaged over each of these periods, and drought anomalies based on the number of distinct municipalities affected in a given year, together with the corresponding 95% bootstrap confidence intervals
455 obtained using a 3-year block resampling scheme. The strongest correlations are found for winter (DJF) and for the January–March (JFM) period, suggesting that January and February conditions play a substantial role, in agreement with the findings of the previous section. When only March, April, and May are considered, the correlation coefficient falls close to zero and its confidence interval includes negative values. This result points to the preventive character of spring rogation ceremonies, and suggests that these records reflect the social response to, and societal impact of, drought conditions rather than concurrent
460 meteorological configurations.

It can be noted that, in addition to this temporal analysis, a regional analysis was also attempted, with the aim of assessing whether the large-scale averaging performed here might be masking stronger local relationships. Averaging both the documentary and atmospheric information over a relatively large domain may indeed smooth out regional differences in drought development and atmospheric circulation, potentially weakening the observed correlations. However, this regional breakdown
465 proved difficult to interpret, as the amount of available data is often insufficient to draw robust conclusions. This is particularly true for the documentary sources: for a given city, the number of records available for a specific year can vary considerably from one year to another. In Barcelona, for instance, the average number of reports per year is 4.48, with a standard deviation of 5.27, a disparity large enough to make the statistical significance of any correlation highly variable from one year to the next. Furthermore, conducting this type of analysis on a $1^\circ \times 1^\circ$ grid would reduce the number of available spatial points from
470 84 (over the full study domain) to a single point, which is clearly too restrictive to support meaningful conclusions.



A more promising alternative for future work could be to compute correlations for a selection of specific regions (e.g., Catalonia, Valencia, or Murcia) rather than for the full $1^\circ \times 1^\circ$ grid, which would allow more data points to be retained (on the order of twenty). This approach would, however, be more complex to implement, as it would require systematically checking the statistical significance of each regional correlation individually, given the uneven and often limited number of records per region and per year.

Table 3. Spearman correlation coefficients between Z_{500} anomalies averaged over successive 3-month periods and drought anomalies (number of distinct municipalities affected per year), together with the 95 % confidence intervals derived from a 3-year block bootstrap ($n = 10\,000$ resamples). Lag 0 corresponds to the spring season (MAM), in which most documentary drought records are concentrated. Increasing lag values correspond to atmospheric periods shifted progressively earlier relative to this reference season.

Lag	Season	r	95 % Confidence Interval
0	MAM (Spring)	0.015	[−0.249, 0.274]
1	FMA	0.315	[0.048, 0.538]
2	JFM	0.496	[0.248, 0.692]
3	DJF (Winter)	0.551	[0.322, 0.722]
4	NDJ	0.314	[0.042, 0.530]

4 Conclusion

The AMARNA® Repository, which compiles historical documentary sources on droughts, enabled us to systematically develop a quantitative drought index for nineteenth-century Spain based on qualitative data, thereby addressing the lack of instrumental data for that period. The ability of the indices to capture the major droughts documented in the literature was demonstrated, at least for the early part of the century.

By using sea-level pressure (SLP) and geopotential height at 500 hPa (Z_{500}) from the NOAA Twentieth Century Reanalysis we constructed normalized atmospheric indices to compare with the documentary-based drought indices. The results show a sustained decline in the correlation between the atmospheric and drought indices from the 1860s onward, with correlation coefficients dropping from 0.68 to as low as 0.18 over 35 years. This indicates that the AMARNA® Repository, taken as a whole, became progressively less effective at tracking drought episodes over this period. This is likely linked to the socio-political changes affecting Spanish society during the second half of the nineteenth century, which affected religious practices such as rogation ceremonies unevenly across cities and time. While, taken individually, a reported ceremony remains reliable, the loss of reliability appears mainly at the scale of the Repository as a whole.

In the database, a clear seasonality in droughts is observed, with half of the reported events occurring in spring and just above 20% in winter. To characterize the synoptic conditions of dry years, we applied a rigorous methodology to classify years as either dry or non-dry and computed the difference in their mean configurations. The results show that dry years are characterized by strong, positive SLP and Z_{500} anomalies over the Iberian Peninsula and its adjacent Atlantic margin during



winter and, to a lesser extent, during autumn of the preceeding year. This observed temporal offset is likely the consequence of the lag between the anticyclonic configuration, which likely inhibited rainfall(meteorological drought) and its consequence for soil moisture and water reserves occurring months later (agricultural and hydrological droughts).

The identification of a temporal lag between the winter/autumn anticyclonic anomalies and the onset of documented drought episodes supports the view that the AMARNA® Repository is able to capture a physically coherent climatic signal, rather than a purely qualitative record of perceived scarcity. This helps to improve the understanding of historical records as drought indicators, together with their limitations, and suggests that the approach developed here could be extended to other regions of the world or to records documenting different types of climatic episodes. These findings also highlight the dual nature of historical documentary records. Although their consistency as climatic proxies decreases over time, this does not diminish their value. Instead, they may become increasingly informative for investigating the evolution of drought impacts, societal vulnerability, institutional responses, and adaptation strategies.

Consequently, while continued efforts to improve the calibration of historical proxy records with instrumental observations under climatically coherent conditions remain essential for climate reconstruction, these sources also provide unique opportunities to investigate how societies have perceived, responded to, and adapted to drought through time.

Some limitations must, however, be considered. Data availability may create gaps in the database, as a substantial amount of potentially relevant material preserved in Spanish municipal, ecclesiastical and other historical archives has yet to be systematically examined. In addition, the 1806–1835 period of 20CRv3 corresponds to the experimental extension of the dataset, and several of the key drought episodes analysed fall within this interval, including 1817 and the multi-year drought episode of 1822–1825. Beyond data coverage, this study relies solely on SLP and Z500 as indicators of atmospheric drivers of drought. Other mechanisms, such as land–atmosphere feedbacks, are known to play a significant role and were discussed qualitatively, but were not quantitatively incorporated into the indices or systematically analysed. This is a reminder that, given the restricted set of drivers considered, the numerical values of the correlations reported here are best interpreted as indicative of the overall tendency rather than as precise quantitative estimates.

Future work could also make the drought indices derived from the Repository more complex by incorporating episode duration, when available, or severity (approximated through the intensity of rogation ceremonies) through weighted indices. This would allow a more accurate definition of dry and non-dry years, and potentially the introduction of intermediate categories. This work demonstrates the robustness of the proxy and highlights its limitations, allowing us to identify the issues affecting it in order to improve the methodologies for collecting and processing information, and also to design future data collection campaigns. Finally, Spain’s documentary heritage also offers considerable potential for further research through administrative records, both supra-municipal and ecclesiastical, increasing the number of drought records for the nineteenth century.

Code and data availability. The 20th Century Reanalysis Version 3 (20CRv3) data are freely available from the NOAA Physical Sciences Laboratory at <https://psl.noaa.gov> (Slivinski et al., 2021). The Barcelona precipitation series is available from (Prohom et al., 2015),



525 <https://doi.org/10.1002/joc.4537>. <https://www.meteo.cat/wpweb/climatologia/dades-i-productes-climatics/serie-climatica-de-barcelona-des-de-1780/>

The underlying raw data is the exclusive property of the AMARNA collective and remains restricted due to applicable intellectual property rights. However, these data may be requested directly from the repository managers: Mariano Barriendos (mariano.barriendos@gmail.com) and Josep Barriendos (josep.barriendos@gmail.com).

530 The python code used in this work is available at <https://doi.org/10.5281/zenodo.22765770>

Author contributions. E.R and D.P conceptualized the study; E.R conducted the data analysis and wrote the original manuscript draft, with methodological development support from D.P, M.B, J.B and S.G. M.B and J.B provided and curated historical documentary resources. D.P and M.B provided supervision throughout the project. All authors contributed to the interpretation of the results and the critical review and editing of the final manuscript.

535 *Competing interests.* The authors declare that they have no conflict of interest.

Acknowledgements. Support for the Twentieth Century Reanalysis Project version 3 dataset is provided by the U.S. Department of Energy, Office of Science Biological and Environmental Research (BER), by the National Oceanic and Atmospheric Administration Climate Program Office, and by the NOAA Earth System Research Laboratory Physical Sciences Laboratory. NOAA/CIRES/DOE 20th Century Reanalysis (V3) data were provided by the NOAA PSL, Boulder, Colorado, USA, from their website at <https://psl.noaa.gov>. These data were produced
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The research has been supported by the grants from from DRYMED (PID2024-157662OB-C21) and FLOODMED (PID2024-157662OB-C22), funded by the Spanish Ministry of Science, Innovation and Universities (MICIU/AEI/10.13039/501100011033) and by ERDF/EU.

Ajuts Joan Oró, AGAUR, Generalitat de Catalunya, 2023 FI-1 00518.



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