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French summer droughts since 1326 CE: A reconstruction based on tree ring cellulose $\delta^{18}O$

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Abstract.

The reconstruction of droughts is essential for the understanding of past drought dynamics and can help evaluate future drought scenarios in a changing climate. This article presents a reconstruction of summer droughts in France based on annually resolved, absolutely dated chronologies of oxygen isotope ratios ($\delta^{18}O$) in tree ring cellulose from *Quercus spp*. Samples were taken from living trees and timber wood from historic buildings at two sites: Fontainebleau (48°23'N, 2°40'E; 1326–2000 CE) and Angoulême (45°44'N, 0°18'E; 1360–2004 CE). Cellulose $\delta^{18}O$ from these sites proved to be a good proxy of summer climate, as the trees were sensitive to temperature and moisture availability. However, offsets in average $\delta^{18}O$ values between tree cohorts necessitated a correction before joining them to the final chronologies.

Using the corrected $\delta^{18}O$ chronologies, we developed models based on linear regression to reconstruct drought, expressed by the standardized precipitation evapotranspiration index (SPEI). The significant correlations between the SPEI and cellulose $\delta^{18}O$ ($r \approx -0.70$), as well as the verification of the models by independent data support the validity of these reconstructions. At both sites, recent decades are characterized by increasing drought. Fontainebleau displays dominantly wetter conditions during earlier centuries, whereas the current drought intensity is not unprecedented in the Angoulême record.

While the $\delta^{18}O$ chronologies at the two studied sites are highly correlated during the 19th and 20th centuries, there is a significant decrease in the correlation coefficient between 1600 and 1800 CE, which indicates either a weaker climate sensitivity of the tree ring proxies during this period, or a more heterogeneous climate in the north and the south of France. Future studies of tree ring isotope networks might reveal if the seasonality and spatial patterns of past droughts can explain this decoupling. A regional drought reconstruction based on a combination of the two sites shows good agreement with historical records of grape harvest dates in France, which provide another proxy of summer climate.

1 Introduction

Droughts can have severe impacts on ecosystems and on human activies (Büntgen et al., 2010; Seneviratne et al., 2012; Diaz and Trouet, 2014), as they influence water availability, groundwater recharge, algal blooms (Paerl et al., 2011), forest productivity (Allen et al., 2010), carbon sink saturation (Nabuurs et al., 2013), crop yields, forest fires, soil degradation, building subsidence and insurance costs (Corti et al., 2011, 2009), as well as human health (Haines et al., 2006; OBrien et al., 2014). More intense and longer meteorological droughts have already been identified in southern Europe (Sousa et al., 2011; Vicente-Serrano et al., 2014) and southern France (Giuntoli et al., 2013) in past decades. In response to increased greenhouse gas concentrations, climate models project a marked increase in heat waves and droughts in France for the 21st century (Vidal et al., 2012; Dai, 2013; Kovats et al., 2014), augmenting the pressure on water resources and challenging the sustainability of the current agricultural system (Itier, 2008; Levrault et al., 2010).

Soil moisture feedbacks have shown to play an important role in recent European hot summers (Zampieri et al., 2009) and their seasonal predictability (Quesada et al., 2012). Vidal et al. (2010) have provided a homogenized reference framework for French droughts since the 1950s based on high resolution meteorological reanalyses. However, prior to the instrumental period, little is known about the past intensity, return period and spatial extent of drought events, mostly based on historical sources (Garnier, 2011). The detection and attribution of recent hydroclimate changes and their link with the ongoing warming trend therefore remain an open issue. A reconstruction of the temporal and spatial extent of drought in the past would constitute a valuable basis for expanding the knowledge on past droughts using natural climate archives.

In semi-arid regions, tree ring width chronologies can provide records of past drought periods (e.g. Cook et al., 2004; Li et al., 2006; Esper et al., 2007; Stahle et al., 2007; Linares et al., 2012; Yadav, 2013). Drought, however, can also occur in high-rainfall areas and is a recurrent feature of the European climate (European Environment Agency, 2001). A great amount of high-resolution proxy data from Europe exists, but many tree-ring based reconstructions focus on high-altitude or high-latitude sites, where ring width and density show a strong sensitivity to temperature. Pan-European signature years in oak ring width series have been related to soil moisture anomalies (Kelly et al., 2002), but they only give access to extreme events.

Stable isotopes in tree ring cellulose have proven to be the more reliable proxy in areas where tree growth is not strongly dependent on climate, typically at low elevation, mid-latitude sites (e.g. Loader et al., 2008; Saurer et al., 2008; Young et al., 2012). They can help extend climate reconstructions into regions which are not yet well covered (Leavitt et al., 2010), and also provide information about precipitation amount (Rinne et al., 2013; Young et al., 2015). Recent studies have revealed the sensitivity of cellulose $\delta^{18}O$ to the combined effects of temperature and humidity, and therefore of drought (Raffalli-Delerce et al., 2004; Masson-Delmotte et al., 2005; Etien et al., 2008, 2009; Labuhn et al., 2014). This influence can be explained by the physical processes and mechanisms, such as the isotopic enrichment of leaf water, which relate the cellulose $\delta^{18}O$ signal to climate (Hill et al., 1995; Barbour and Farquhar, 2000; Barbour et al., 2004; Cernusak et al., 2005; Sternberg, 2009; Gessler et al., 2009, 2013).

A complete understanding of past climate variability must include the whole frequency spectrum from inter-annual to multi-centennial scales. A challenge of tree ring based climate reconstructions is the preservation of low-frequency variability in the proxy records. The construction of ring width chronologies necessitates the standardization of individual ring width series to correct non-climatic growth trends (Cook et al., 1990). However, standardization can also eliminate low-frequency climatic trends in the series. Esper et al. (2004) identified the standardization method applied to ring width and density data as the most important cause of differences in low-frequency trends between different hemispheric scale temperature reconstructions of the past millennium. If appropriate data and methods are used, multi-centennial climate variability can be preserved in ring width chronologies (Briffa et al., 2002; Esper et al., 2002), but large numbers of samples are required.

For isotope proxies, a strong common signal between trees at a site is usually found, and often only a few trees are used to extract climate information (e.g. Leavitt, 2010; Daux et al., 2011; Shi et al., 2011). This is especially beneficial if the number of available samples is limited, e.g. when subfossil wood or ancient building material is used. However, Loader et al. (2013) demonstrated that despite a strong common signal, higher levels of replication than those typically adopted in isotope dendroclimatology would be needed for an accurate estimation of the population mean. Although stable isotope series have a greater potential to retain long-term climatic trends as they are usually not standardized, a problem can arise when combining different trees or tree cohorts into an isotope chronology, if replication is low and site-related factors lead to offsets in average isotope values between individual trees or sub-series. Several recent studies propose methods to deal with offsets (Esper et al., 2010; Gagen et al., 2012; Hangartner et al., 2012; Naulier et al., 2015). An offset correction might induce not only a partial loss of low-frequency climate information, but also makes it difficult to relate differences in absolute isotope values between distant sites to climatic gradients, because the effects of these gradients can be superimposed by the influence of the local environment.

In this article, we present annually resolved chronologies of oxygen isotope ratios in latewood cellulose ($\delta^{18}O_c$) constructed from *Quercus spp.* living trees and timbers at two sites in France: (1) an extension of the previously published series from Fontainebleau (1596–2000) (Etien et al., 2008, 2009) back to 1326; and (2) an extension of the previously published series from Braconne forest near Angoulême (1860–2004) (Labuhn et al., 2014) back to 1360. The previous studies have demonstrated that $\delta^{18}O_c$ at these sites responds to summer temperature and moisture availability. Here, we present reconstructions of summer droughts for Angoulême and Fontainebleau based on the $\delta^{18}O_c$ chronologies, as well as a regional drought reconstruction combining these sites. Furthermore, we investigate the absolute and relative variability between individual trees and between tree cohorts from the same site, and address methodological issues related to the construction of long isotope chronologies from several cohorts. Then, we compare $\delta^{18}O_c$ between Fontainebleau and Angoulême at different time scales to evaluate the spatial coherence of the records and the temporal stability of their relationship. As meteorological data are only available for the 20th century, this comparison can give indications if the climatic forcing for northern and southwest France droughts which influences $\delta^{18}O_c$ is stable over time.

2 Data and methods

2.1 Study sites characteristics and climate

The two studied sites, Fontainebleau (FON; 48°23'N, 2°40'E) and Angoulême (ANG; 45°44'N, 0°18'E; Fig. 1) are about 300 km away from each other and lie at a similar low altitude: FON 144 $\mathrm{m\,a.s.l.}$, and ANG 110 $\mathrm{m\,a.s.l.}$ The soil at FON is about 1.5 m deep and the texture is dominated by sand mixed with loam and clay. ANG is located in a hilly karstic landscape with a cambisol layer of 0.3 to 0.6 m depth.

Monthly meteorological data (average temperature, maximum temperature, and precipitation, 1901–2009) were obtained from the CRU TS3.10 data set (Harris et al., 2013). Time series were extracted for the grid cells containing the study sites. Both sites are characterized by a temperate oceanic climate. The average annual temperature is 11.5 $^{\circ}\mathrm{C}$ at FON and 11.9 $^{\circ}\mathrm{C}$ at ANG, the average summer (JJA) temperature is 18.8 °C at both sites. Average annual precipitation sums range from 600 mm (FON) to 770 mm (ANG), while average summer precipitation at both sites is about 160 mm. FON has remarkably drier winters than ANG (140 mm and 210 mm respectively).

The Standardized Precipitation Evapotranspiration Index (SPEI; 1901–2001; Beguería et al., 2010; Vicente-Serrano et al., 2010) for the site grid cells was obtained from http://sac.csic.es/spei/database.html at different temporal resolutions. The index calculation is based on CRU data. We selected this drought index for reconstruction because it yielded consistently higher correlations with $\delta^{18}O_c$ compared to other drought indices. The SPEI includes potential evapotranspiration, and it might therefore be more representative of the influence on tree physiological processes than e.g. the Standardized Precipitation Index (SPI; McKee et al., 1993), which is based on precipitation alone. Drought indices which incorporate soil moisture, like the self-calibrating Palmer Drought Severity Index (PDSI; Palmer, 1965; van der Schrier et al., 2006) or the Standardized Soil Wetness Index (SSWI; Vidal et al., 2010) should reflect the hydrological conditions which are relevant for tree growth. However, the soil water holding capacity is derived from coarse gridded data sets of soil properties, which might not be representative of the local conditions. Furthermore, all the drought indices cited here identify coherent patterns of drought intensity and frequency at the study sites.

Both sites, FON and ANG, display very similar trends and interannual variability of temperature, precipitation and SPEI throughout the 20th century, and there is no marked climatic gradient between the sites (Fig. 2). If summer droughts influence the tree ring proxies, we can therefore expect similar variability of the proxy time series from both sites.

2.2 Tree samples

The previously published FON $\delta^{18}O_c$ chronology (1596–2000; Etien et al., 2008, 2009), constructed from living oak trees from Fontainebleau forest and oak timbers from Fontainebleau castle, is extended in this study using 27 oak timber cores from three buildings of Fontainebleau castle, which constitute different construction periods: Porte Dorée (PD), Chapelle (CH), Petites Ecuries 1 (PE1), and Petites Ecuries 2 (PE2). One core was taken per timber beam. The building wood likely originates from the neighboring forest (Dan, 1642; Domet, 1873). Fourteen cores were selected for isotope analysis, resulting in a sample depth of two to nine trees.

The previously published ANG $\delta^{18}O_c$ chronology from Braconne forest (1860–2004; Labuhn et al., 2014) consists of two groups of living trees from different locations in the forest (1 km apart) with an average age difference of 210 years: Braconne "young trees" (ANG-B), and Braconne "old trees" (ANG-GR). The chronology is extended here using the older part of the ANG-GR samples (1626–1859) and samples from oak timbers in historic buildings in the city of Angoulême (ANG-TW), 15 km from the forest. A total of 36 cores was taken from three different buildings (one core per timber beam): Angoulême Maison du Comte (AMDC), Poullignac church (POUL), and La Rochefoucauld castle (LRF). According to the shape and length of the beams, a beam likely corresponds to one tree. The provenance of the timbers is not documented, but dendroprovenancing (Bridge, 2012) supports the assumption of a local origin of the wood, as correlations between the timber samples and different reference chronologies decrease with increasing distance. A subset of cores was selected for isotope analyses to obtain a sample depth of four to six trees for each year with sufficient overlap between samples; only between 1556 and 1591 is the number of trees < 4.

The living trees at FON are *Quercus petraea*. At ANG, the species has not been determined in the field, but *Q. petraea* and *Q. robur* are the dominant species in the forest. For all timber wood, the oak species is unknown. Although methods to determine the species based on wood anatomy exist (for a review see Feuillat et al., 1997), an unambiguous discrimination between *Q. petraea* and *Q. robur* is not possible (Schoch et al., 2004). To our knowledge, the oxygen isotopic composition of cellulose in the two oak species has never been compared, but specific differences in the $\delta^{18}O$, as already shown for $\delta^{13}C$ due to differences in water use efficiency (Ponton et al., 2001), cannot be ruled out. Even if there is no direct species effect, different species have different site preferences (Lévy et al., 1992), and the site hydrology influences the $\delta^{18}O$ (Labuhn et al., 2014). However, site effects can also influence the $\delta^{18}O$ of trees within a species. We therefore consider that the site-related uncertainty is more important that the uncertainty due to the ambiguity in species determination.

Tree ring width was measured under a binocular microscope using a LINTAB measuring table (Rinntech) with a precision of 0.01 mm, and cores were crossdated. The crossdating was verified visually and statistically with the program COFECHA (Holmes, 1983; Grissino-Mayer, 2001).

2.3 $\delta^{18}O$ measurements, pooling strategy and corrections

For isotope analysis, the selected cores were cut ring by ring with a scalpel and earlywood was separated from latewood. Only the latewood was analyzed. Corresponding rings of multiple cores taken from the same tree were pooled. The wood samples were homogenized and cellulose was extracted according to Green (1963) and Leavitt and Danzer (1993). The $\delta^{18}O$ of cellulose was measured using a thermal combustion elemental analyzer (Finnigan Thermo TC-EA) coupled with a Finnigan MAT252 mass spectrometer at LSCE in Gif-sur-Yvette, France. The measurements were corrected using the cellulose reference Whatman CC31. Each sample was measured at least twice. The analytical uncertainty was 0.20‰. $\delta^{18}O$ values are reported with reference to VSMOW (Coplen, 1994). A total of 1371 samples (787 ANG, 584 FON) was measured for this study, excluding replicates.

For the previously published FON $\delta^{18}O_c$ chronology, wood from all living tree and timber cores had been pooled for each year prior to analysis. In this study, the new FON samples were pooled by building, thereby grouping together trees of approximately the same age. Although the inter-tree variability of $\delta^{18}O_c$ at a site was generally low, a systematic offset had been observed between the old (ANG-GR) and the young (ANG-B) trees at Braconne forest, which is likely due to differences in the site hydrology (Labuhn et al., 2014). This observation raises concerns about a possible site-related increased variability in timber samples. To address this question, we performed an inter-tree comparison for the ANG timber wood by analyzing individual rings at a 5-year interval. For the remaining years, the rings of all timber cores were pooled prior to cellulose extraction.

A juvenile effect of increasing $\delta^{18}O_c$ during the first 20 years of a tree's life had been observed in the trees from site ANG-B (Labuhn et al., 2014). In the ANG data and the new FON data (1326–1595), the juvenile effect was taken into account by omitting the first 20 years of each tree core from the analysis. For the previously published series from FON (1596–2000), several trees had been pooled and a correction was not possible anymore. Furthermore, for these samples, only the calendar year of each ring was known but not their cambial age. It was therefore not possible to verify whether the trees had been cored to the pith, and we do not know whether the innermost rings of a core actually correspond to a period of juvenile growth.

Both the FON and ANG chronologies are composed of several sub-series (cohorts), which have been analyzed separately. The cohorts showed differences in the mean isotope values for their periods of overlap of up to 1‰ (see Sect. 3.2). As all trees have been exposed to the same climatic influences, these offsets are likely due to the local conditions (competition, microclimate, rooting depth, soil hydrology). We tested different methods to merge the overlapping series (following Hangartner et al., 2012) in order to investigate the effects of such offsets on the final chronology: (1) no correction was applied; (2) the mean $\delta^{18}O_c$ values of the older cohorts were adjusted to match the mean of the corresponding younger cohorts in their period of overlap; (3) all cohorts were normalized to have the same mean ($\mu = 0$) before merging them. In all cases, the averages of two cohorts were calculated weighting each cohort by the number of trees.

2.4 Statistical analyses

Statistical analyses were performed using the R software (R Development Core Team, 2015). The site chronologies were decomposed to high-pass and low-pass filtered data using a cubic smoothing spline with a 50% variance cutoff at a period of 10 years, which enables a comparison of the inter-annual and decadal variability between series. Running correlations between sites and running standard deviations (SD) for each site were calculated using a window of 51 years. Correlation coefficients were calculated between $\delta^{18}O_c$ and the meteorological variables at monthly and seasonal time scales, in order to investigate the climate response of the trees and to identify the climate parameters which can be reconstructed from the tree ring proxy.

A model for reconstruction of the drought index SPEI was developed based on a linear regression between the index and $\delta^{18}O_c$. The validity of the model was tested by dividing the SPEI data into a calibration (two-thirds of the values) and a verification data set (one-third of the values). This enables a quantitative comparison of the reconstructed drought index with independent data not used in the calibration. The subsets of data were selected randomly and the procedure was repeated 1000 times. To evaluate the skill of the model in estimating the SPEI, we calculated the correlation coefficient (r), the reduction of error statistic (RE), and the coefficient of efficiency (CE) (see Briffa et al., 1988) for each iteration.

After verification of the model performance using the split data set, a new model was calibrated using the entire period of the observed SPEI. Based on the final model, the drought index was reconstructed for the full length of the $\delta^{18}O_c$ chronology. The confidence interval around the reconstruction was determined based on the differences between the measured and the reconstructed SPEI values (±2 SD of the differences).

3 Results

This section presents the comparison of isotopic values between individual trees (3.1) and between cohorts (3.2), the comparison of the $\delta^{18}O_c$ chronologies at the two sites, FON and ANG (3.3), their correlations with meteorological variables (3.4), and lastly the drought reconstructions (3.5).

3.1 Inter-tree comparison for ANG timber samples

The difference in the average $\delta^{18}O_c$ values of ANG timber cores analyzed individually at a 5-year interval is up to 1.8‰ (Fig. 3). For single years, the range of $\delta^{18}O_c$ values is between 0.01‰ and 4.20‰, the average difference between the maximum and minimum values being 1.18‰. Despite the differences in average values, these series display a coherent year-to-year variability at the 5-year time step. The average correlation coefficient between cores is $r = 0.72$, the average Gleichläufigkeit (Schweingruber, 1988) is 76%, and the expressed population signal (EPS; Wigley et al., 1984) is 0.97.

3.2 Cohort offset correction for FON and ANG

At both FON and ANG, separately analyzed tree cohorts show offsets in the average $\delta^{18}O_c$ values of up to 1‰ during their overlap periods (Table 1; Fig. 4). The stacking method applied for the construction of the chronologies significantly influences their long-term trends (Fig. 5). For FON, the offset-corrected values before 1600 are on average $> 1‰$ higher than the uncorrected values. For ANG, the corrected values are lower on average than the raw values before 1640, and higher on average after this year.

The uncorrected chronologies of FON and ANG differ in long-term trends and $\delta^{18}O_c$ values diverge prior to 1620, although the decadal trends are largely synchronous. The corrected chronologies, on the contrary, show a good agreement for the centennial trends (Fig. 5). Correlation coefficients between the chronologies improved from 0.26 (raw) to ~ 0.50 (both corrections) at the inter-annual scale, and from 0.12 to ~ 0.47 for the low-pass filtered data. The subsequent analyses are based on the chronologies which are corrected by subtracting the mean from each cohort before merging them.

3.3 Comparison of $\delta^{18}O$ site chronologies

The $\delta^{18}O_c$ chronologies from FON and ANG are significantly correlated during the 20th century ($r = 0.71$ for high-pass and $r = 0.69$ for low-pass filtered data). The correlation shows a marked decrease between 1600 and 1800. These patterns are similar for the inter-annual variability and the decadally smoothed chronologies (Fig. 6; Fig. 7).

The standard deviation calculated for a 51-year moving window illustrates common patterns of changes in the magnitude of inter-annual variability, e.g. periods of low inter-annual variability around 1470, 1620 and 1800, and increased variability before 1400 and around 1700. In general, the variability is slightly lower at FON (mean SD = 0.52) that at ANG (mean SD = 0.60).

3.4 Correlations with meteorological variables

The strength and direction of the correlations of $\delta^{18}O_c$ with different meteorological variables are similar for FON and ANG (Table 2). $\delta^{18}O_c$ is positively correlated to temperature during the summer months, with stronger correlations for maximum temperature than for mean temperature. Negative correlations are found with precipitation and the drought index SPEI. Generally, correlations are improved when considering the integral of the summer months rather than single months. The correlations between the mean chronology and meteorological variables averaged over the study region yielded comparable results to those for each site. The highest correlation between $\delta^{18}O_c$ and the SPEI are obtained using the SPEI with a 3-month resolution, which represents drought conditions during June–August. This variable was therefore chosen as the reconstruction target.

The relationship between $\delta^{18}O_c$ and the climate variables is relatively stable throughout the 20th century, as correlation coefficients (r) calculated for a 31-year moving window vary only slightly (standard deviation of r between 0.04 and 0.08 for different pairs of variables; not shown).

3.5 Drought reconstructions

The correlation between SPEI and $\delta^{18}O_c$ is $r = -0.69$ ($p < 0.001$, $N = 100$) for FON, $r = -0.68$ ($p < 0.001$, $N = 104$) for ANG, and $r = -0.62$ ($p < 0.001$, $N = 104$) for the mean chronology. The equations describing the linear regression line between the two variables are used to estimate the SPEI from $\delta^{18}O_c$ for each site (Fig. 8). The r, RE and CE statistics confirm the validity of the models (high r, positive RE and CE; see Briffa et al., 1988, Table 3). The squared correlations are $R^2 = 0.48$ for FON and $R^2 = 0.46$ for ANG, which means that almost 50% of the variability in the drought index is accounted for by the respective models. For the mean chronology $R^2 = 0.38$. The standard deviation of the differences between modeled and observed SPEI (0.72 for FON; 0.70 for ANG; 0.71 for the mean chronology) defines the confidence interval for the reconstructions (Fig. 9).

For ANG, the oldest part of the reconstruction is characterized by a prolonged period of relatively wet conditions (1360–1450), followed by a period of drier conditions until 1600 (Fig. 10). Another wet period until 1760 follows, with a short increase to relatively drier conditions around 1720. The most marked trend in the reconstruction is the shift to dry conditions from 1760 to 1850. The late 20th century is characterized by increasingly dry summers, but these SPEI values are not unprecedented in the record.

The reconstructed SPEI at FON indicates relatively wet conditions for large parts of the record prior to 1800, except two drier intervals in the 14th century and in the earlier half of the 16th century. Like for ANG, there is a marked step to drier conditions around 1800, and the late 20th century is characterized by increasing drought.

4 Discussion

Dry summer conditions influence both the source water isotopic composition and the physiological processes in a tree, which together determine the oxygen isotope ratios in tree ring cellulose (Roden et al., 2000; McCarroll and Loader, 2004). This leads to a correlation between $\delta^{18}O_c$ and the SPEI that can be interpreted as the trees' climate response, providing the basis for the drought reconstruction. The $\delta^{18}O$ of precipitation has numerous influences such as the water vapor source, air mass trajectories and the rainout history of the air mass, the condensation temperature and the precipitation amount (Darling et al., 2006). The isotopic composition of the source water is principally controlled by the $\delta^{18}O$ of local precipitation, although it can be modified by evaporation of water from the soil and the mixing of water from different seasons, which depends on water residence times in the soil (Tang and Feng, 2001; Brooks et al., 2010). During warm and dry summers, a higher condensation temperature of the rain water, as well as stronger evaporation from falling raindrops and from soil water increase the $\delta^{18}O$ of the source water (e.g. Gat, 1996). The water taken up by the roots is transported to the leaves where it is enriched in the heavy isotope due to evaporation. The enrichment depends on the difference in vapor pressure of leaf air and ambient air, and is therefore controlled by relative humidity (e.g. Barbour et al., 2004; Sternberg, 2009; Gessler et al., 2013). Drought conditions therefore lead to an enrichment of ^{18}O in both the source water and the leaf water, yielding higher $\delta^{18}O_c$ values.

The relationships between $\delta^{18}O_c$ and different meteorological variables observed at our sites match the results of previous studies in temperate regions regarding the strength and direction of the correlations (e.g. Szczepanek et al., 2006; Reynolds-Henne et al., 2007; Treydte et al., 2007; Loader et al., 2008; Saurer et al., 2008; Hilasvuori and Berninger, 2010; Haupt et al., 2011). The validity of the local drought reconstructions from Fontainebleau and Angoulême is supported by the high correlations between the drought index and $\delta^{18}O_c$, as well as by the verification of the models with independent data. However, the climatic influences on $\delta^{18}O_c$ are complex and although they appear to be constant during the 20th century (Sect. 3.4), they might have changed over time. In order to discuss the reliability of our reconstructions beyond the period of modern instrumental data, and to asses possible sources of uncertainty, we have to look at the variability between trees and between sites, as well as compare our reconstructions to other paleoclimate records.

4.1 Variability between trees and between cohorts

Local site conditions and influences on individual trees (e.g. soil hydrology, competition, microclimate) affect the oxygen isotopic ratios in cellulose. Despite the relative coherence in the inter-annual variability, the absolute differences in average $\delta^{18}O_c$ between trees and between tree cohorts at the studied sites can be more than 1‰. According to our model (Sect. 3.5, Fig. 8), changes of this magnitude in $\delta^{18}O_c$ would translate into a change in the reconstructed SPEI of 0.8, which is slightly larger than our defined confidence interval. With such site-related offsets, it is not appropriate to compare absolute isotope values at distant sites, e.g. to infer climatic gradients. When averaging uncorrected series, there is a risk of introducing artificial shifts or trends in the final chronology that are not climatic. Applying an offset correction seems necessary, and our approach is supported by the better agreement between sites after correction (Sect. 3.2). However, the correction also implies a partial loss of the low-frequency variability, as the variability which exceeds the length of the cohorts may not be preserved. As a consequence, the longer-term drought variability could be underestimated, and periods of prolonged drought would be more difficult to identify than extreme events relative to an overall mean.

Further uncertainty arises from the sometimes low sample depth, and it was impossible to keep it constant due to limited number of available samples from historic buildings. The confidence interval around reconstructions should therefore not be constant over time, but the influence of reduced sample depth cannot be quantified without individual tree measurements of $\delta^{18}O_c$ during the calibration period. A solution to these problems would be to increase the sample depth and to measure trees individually, possibly at the expense of the temporal resolution, which would allow identifying outliers and calculating a confidence interval around the mean $\delta^{18}O_c$ value. However, it could be problematic to find appropriate samples for periods where no living tree material is available. This investigation demonstrates the importance of assessing $\delta^{18}O_c$ variations between individual trees, and of correcting offsets when combining different sets of trees, in order to carry out meaningful inter-site comparisons and reliable identifications of multi-centennial trends.

4.2 The coherence between site chronologies and its temporal stability

The $\delta^{18}O$ chronologies from FON and ANG reveal common patterns of variability on different time scales, but their relationship is not stable over time. High correlations between the two isotope chronologies, e.g. during the 19th and 20th century, indicate common forcing factors on $\delta^{18}O_c$ variability. The only possible forcing acting on this spatial scale is climate. The changes in correlation strength are likely not linked to the construction of the chronologies, as high correlations occur both with small and large numbers of sampled trees (e.g. during the 15th and 20th century). The series variability (expressed by the running standard deviation, SD) changes simultaneously at both sites, and does not seem to be systematically linked to sample depth and/or switches from one cohort to another. We propose two hypotheses to explain the periods of decreased correlation between sites: (1) a changing climate response of the trees; and (2) changing climate patterns over France.

Proxies often show a non-linear response to climate (Schleser et al., 1999). The climate sensitivity of trees is known to change with time (e.g. Reynolds-Henne et al., 2007; Friedrichs et al., 2009; Dorado Liñán et al., 2011; Mérian et al., 2011; Voelker, 2011; Linares et al., 2012; Rozas and Olano, 2013; Candela-Galván et al., 2015), due to tree age, changes in the timing and length of the growing season or in the atmospheric CO_2 concentration. In addition, the response of trees to climate changes may depend on their local environment, e.g. soil conditions. The periods of low correlation between sites could be explained by a weaker sensitivity of the proxy to climate when some climate factors are below a certain threshold. For example, during relatively humid periods, $\delta^{18}O_c$ would not strongly depend on humidity and it would not exhibit large inter-annual variations. However, the amplitude of inter-annual variability at both sites (illustrated by the running SD; Fig. 7d) does not support this hypothesis, as SD and the site-to-site correlation do not show common patterns. The SD is certainly influenced by the number of trees, but even when the number of trees is constant (e.g. in ANG 1680–1880), large changes in the SD occur. If the climate response of the trees changes, this has implications for our climate reconstruction, which is based on the climate-proxy relationship during the 20th century, where instrumental measurements of meteorological variables are available.

The second explanation for the temporal instability of the correlations between the FON and ANG oxygen isotope chronologies could be a change in the regional homogeneity of the climate. The 20th century is characterized by coherent patterns in meteorological variables at FON and ANG (Fig. 2), and during this period we observe high correlation between the $\delta^{18}O_c$ chronologies. However, the climate in France might have been more heterogeneous in the past. Changes in seasonal aspects of climate or in the relationship between temperature and humidity can alter the way these variables act on $\delta^{18}O_c$ (Barbour et al., 2002; Masson-Delmotte et al., 2005; Reynolds-Henne et al., 2007). While temperature variations are likely to be coherent between the sites at a time scale which is relevant for the proxy, precipitation patterns are generally more variable in space. The season of the trees water supply (e.g. winter or growing season precipitation), depends on soil properties, root depth, and precipitation seasonality, and can vary over time and between sites (Bréda et al., 1995; Schulze et al., 1996; Hruska et al., 1999; Hanson et al., 2001).

Correlations with ring width show that FON trees are more sensitive to precipitation, and the site receives less winter precipitation than ANG. The periods of decoupling between the sites could be characterized by changes in winter precipitation, which would influence the drought signal at FON, whereas at ANG, where summer evaporative enrichment dominates the $\delta^{18}O_c$ signal (Labuhn et al., 2014), winter precipitation would not have a strong influence. According to our reconstruction, the period of low correlation coincides with a prolonged wet period at FON. Changes in the spatial and seasonal distribution of precipitation could therefore be responsible for the temporal instability of the relationship between the FON and ANG $\delta^{18}O_c$ chronologies.

It cannot be ruled out, however, that the variable coherence between sites is at least partly linked to methodological issues which influence the quality of the reconstructions: the correction of cohort offsets, and the variable sample depth which is low during certain periods (Sect. 4.1). These possible uncertainties lead us to combine the two site chronologies to produce a regional drought reconstruction, as the distance between FON and ANG is not far and the sites show very coherent drought patterns in the 20th century (Fig. 2). The mean chronology yielded correlations with averaged climate comparable to those at each site (Table 2).

4.3 Comparison of the drought reconstructions with other records

Historical records of grape harvest dates (GHD) have been interpreted as proxies of summer temperature, and GHD records are available from the Burgundy (near FON) and Bordeaux (near ANG) wine regions (Daux et al., 2012). Tree ring $\delta^{18}O_c$ at ANG is significantly correlated with GHD from Bordeaux at an inter-annual scale ($r = -0.50$, $p < 0.001$, $N = 272$). Warm and dry summers therefore seem to lead to high $\delta^{18}O_c$ values and early grape harvest. At FON, on the contrary, $\delta^{18}O_c$ and GHD from Burgundy are not well correlated ($r = -0.22$, $p < 0.001$, $N = 621$). It is possible that FON $\delta^{18}O_c$ is less dependent on temperature (weaker correlations than ANG) but more on precipitation (slightly higher correlations than ANG for summer and growing season averages). This is in line with our previous hypothesis (Sect. 4.2): trees at FON respond to droughts caused primarily by a precipitation deficit. For the 20th century, both high temperatures and low precipitation play a role in the causes of drought. During this period, the ANG and FON $\delta^{18}O_c$ are therefore correlated, and FON is also correlated with the Burgundy GHD ($r = -0.60$, $p < 0.001$, $N = 93$). Yiou et al. (2012) have identified a shift in the North–South temperature gradient in France during the Little Ice Age (1500–1850) based on GHD records, which they relate to changes in the prevailing atmospheric circulation over the North Atlantic. Although this shift does not coincide with the decline in correlations between FON and ANG, their study illustrates that atmospheric circulation changes can have different impacts on the local climate on spatial scales which correspond to the distance between our sites.

The regional SPEI reconstruction based on the mean chronology from both sites might be the more reliable drought record due to the increased number of samples. The correlation of the regional reconstruction with the French composite GHD record is good at an interannual scale ($r = -0.37$, $p < 0.001$, $N = 623$). Other available drought records, an SPEI calculated using early instrumental records of temperature and precipitation from Paris (Slonosky, 2002; Rousseau, 2009) and a reconstruction of the Palmer Drought Severity Index (PDSI) extracted for the study region from the "Old World Drought Atlas" Cook et al. (2015), show a similar high frequency variability as our reconstruction. The correlation coefficients at the interannual scale are $r = 0.48$ ($p < 0.001$, $N = 303$) for the Paris SPEI and $r = 0.34$ ($p < 0.001$, $N = 679$) for the PDSI. However, these drought records differ considerably in long-term trends. Nevertheless, the low frequency variability is very coherent between our reconstruction and the GHD record except for the earliest part prior to ca. 1450, displaying common trends, e.g. an increase from 1520 to 1750 CE, followed by a decrease in the mid-18th and 19th century (Fig. 11).

Further comparisons with precipitation reconstructions from different locations in Europe (Cooper et al., 2012; Rinne et al., 2013; Wilson et al., 2005, 2013) indicate a heterogeneity between sites. Different precipitation patterns can be expected at this spatial scale, and some reconstructions differ in seasonality (see Rinne et al., 2013, for a comparison of the reconstructions). Of these reconstructions, only the Southern England May–August precipitation reconstruction by Rinne et al. (2013)

showed some clear common characteristics with the SPEI reconstruction presented here, e.g. drying trends from ca. 1770–1820 and in the late 20th century. A more detailed investigation of the differences between European drought and precipitation reconstructions and how they compare to the present-day spatial patterns in precipitation is worth investigating in future studies.

5 Conclusions and perspectives

This study has demonstrated that oxygen isotope ratios in tree ring cellulose can provide records of past summer droughts. The observed offsets in absolute $\delta^{18}O$ values between trees and between cohorts highlight the importance of isotope measurements on individual trees, which should be considered in future work in order to detect such characteristic in the isotope series. Using a higher and constant number of samples is recommended to avoid issues with offsets. If corrections need to be applied, a part of the low-frequency variability in the chronology might be lost. Moreover, the effect of non-climatic factors on the average cellulose $\delta^{18}O$ values makes it challenging to use tree ring isotope networks to reconstruct spatial gradients in climate.

The chronologies from Fontainebleau and Angoulême presented here constitute the longest continuous cellulose $\delta^{18}O$ time series in France. Combined, they provide a regional reconstruction of summer droughts covering more than six centuries, which is coherent with other proxies of summer climate. The trees display a very good agreement in their inter-annual variability, and highly significant correlations with meteorological variables during the 20th century. The changes in the coherence between the two sites during earlier centuries indicate that response of the proxy to climate might be non-linear, or that the spatial patterns of climate in France have changed. A detailed comparison of other proxy records from France and Europe will be necessary to confirm this hypothesis. Together with model simulations, such comparisons will contribute to our understanding of spatial drought patterns in the past.

6 Data availability

$\delta^{18}O$ data and SPEI reconstructions can be accessed at https://www.ncdc.noaa.gov/paleo/study/20002.

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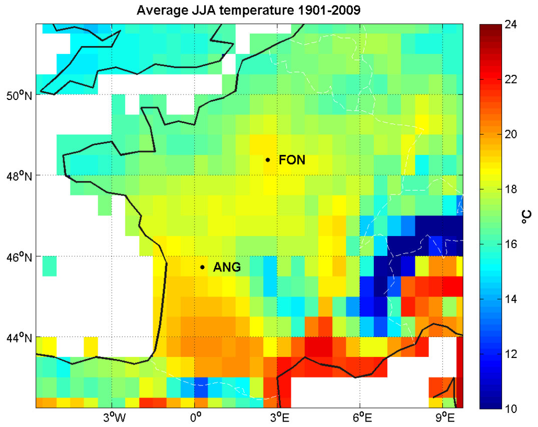

Figure 1. Map of France showing the locations of the tree ring chronologies from Fontainebleau (FON) and Angoulême (ANG). The color scale indicates the average summer (June–August) surface air temperature 1901–2009.

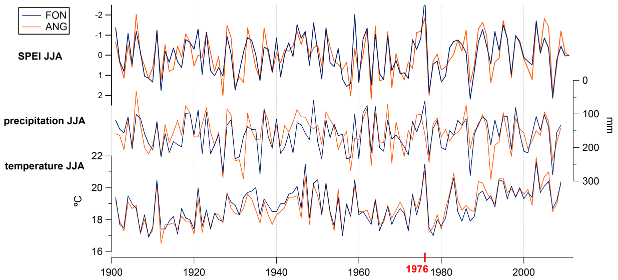

Figure 2. Time series of the drought index SPEI (negative values indicate dry conditions), precipitation and temperature, June–August (JJA) averages for Fontainebleau (FON) and Angoulême (ANG). Note that the axes for SPEI and precipitation are reversed, so that upward values reflect warm and dry years. The year of the exceptional drought 1976 is marked in red. See text for data sources.

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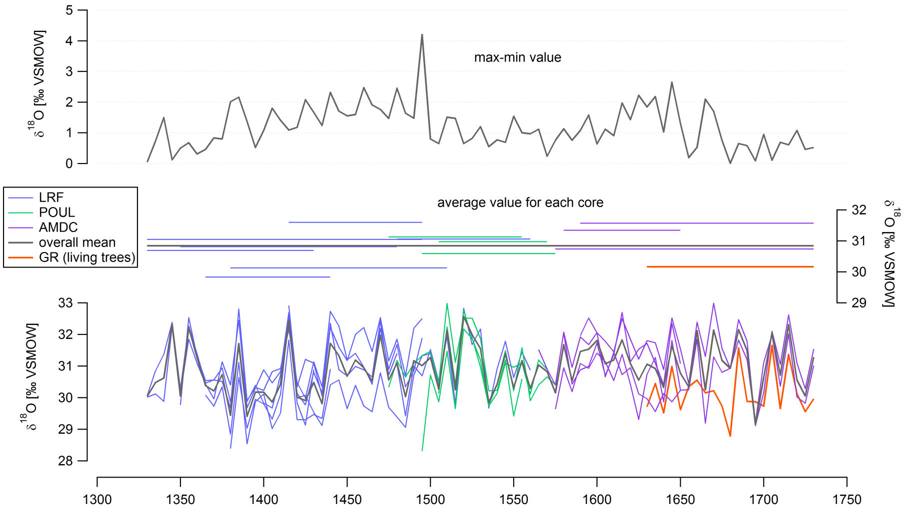

Figure 3. Inter-tree comparison. Bottom: $\delta^{18}O_c$ measured every 5th year for individual timber cores from different buildings in Angoulême (LRF, POUL, and AMDC). The orange line shows the $\delta^{18}O$ of living trees (GR). Middle: Time spans and average $\delta^{18}O$ values for the respective cores. The thick gray line is the mean $\delta^{18}O$ of all timber cores over the whole period. Top: Range of measured values for each year (maximum minus minimum value).

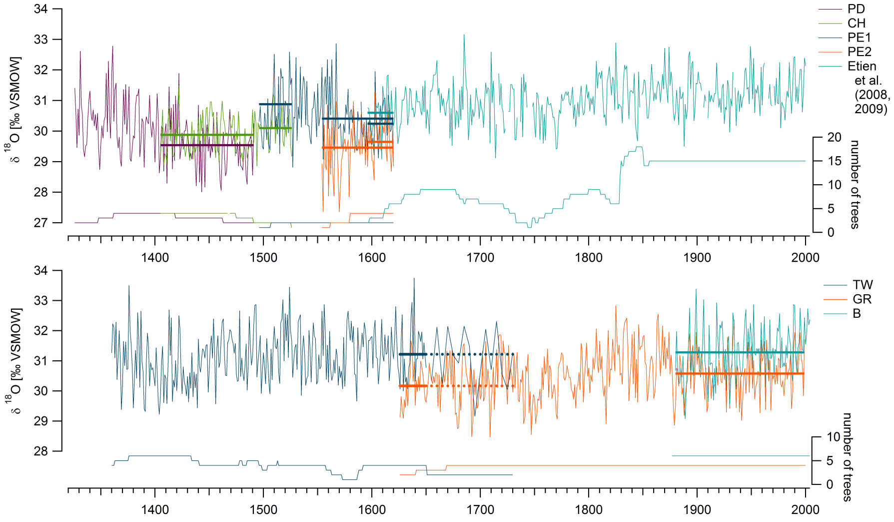

Figure 4. Cohort offsets. Top: The $\delta^{18}O_c$ series for different tree cohorts from Fontainebleau. Bottom: The $\delta^{18}O_c$ series for living tree cohorts (B and GR), and timber wood (TW) from Angoulême. Horizontal lines indicate the mean of each series for the period of overlap. TW samples have been measured every 5th year only in the period covered by the dotted line.

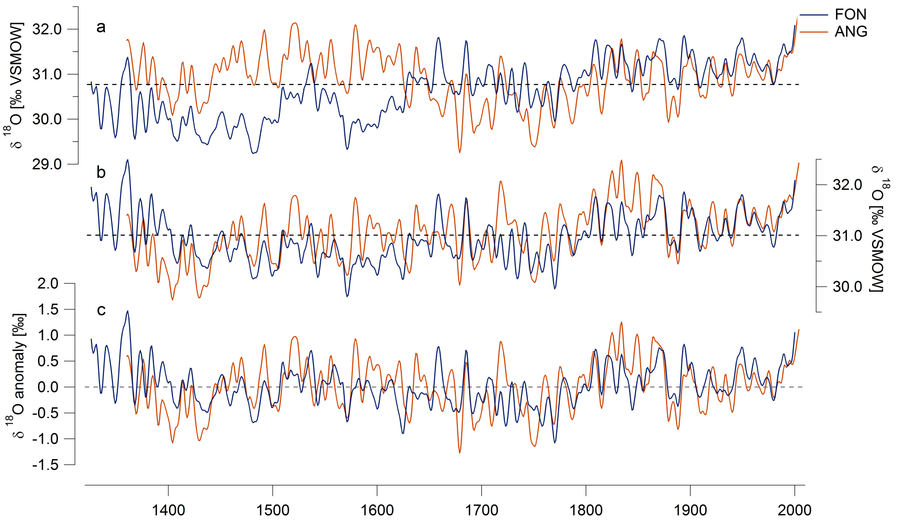

Figure 5. Cohort offset correction for cellulose $\delta^{18}O$ chronologies from Fontainebleau (FON) and Angoulême (ANG) (low-pass filtered data). Different methods were used in order to combine overlapping tree cohorts to a site chronology: (a) raw values; (b) correction of the offset between cohorts (see Fig. 4) by adjusting the mean of the older cohorts to the mean of the respective younger cohorts; (c) all cohorts were normalized by subtracting the mean of each cohort from the respective $\delta^{18}O$ values before combining them. In each case, an average weighted by the number of trees was calculated for overlap periods between cohorts.

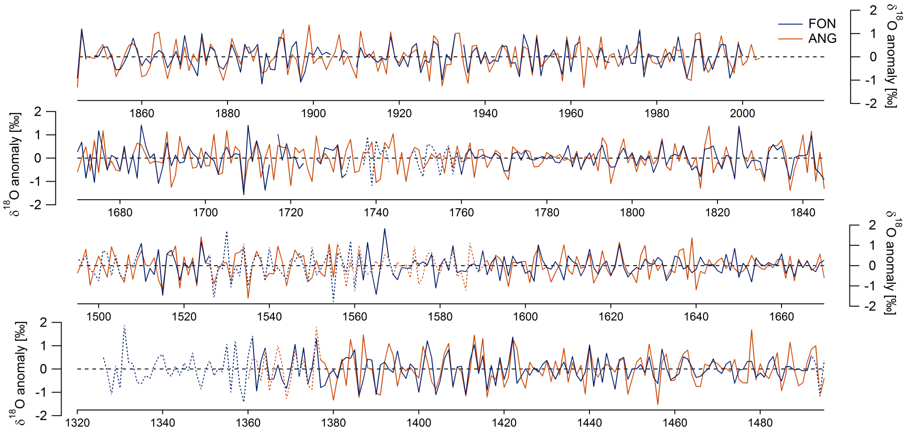

Figure 6. Inter-annual variability in cellulose $\delta^{18}O$ chronologies from Fontainebleau (FON) and Angoulême (ANG), high-pass filtered data. The dotted lines mark the parts of the chronologies where the number of trees is < 4.

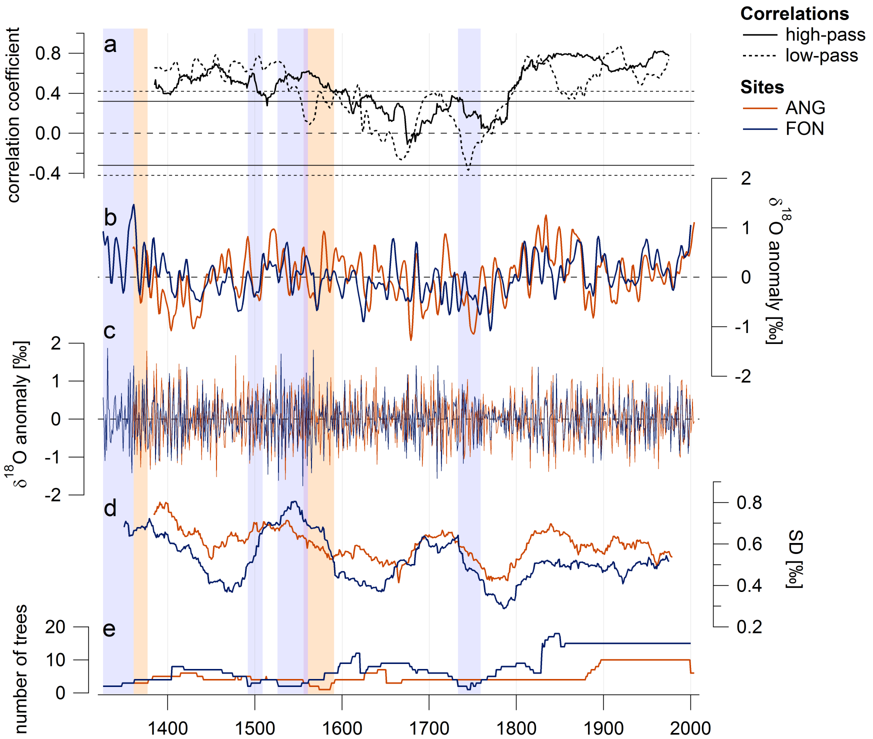

Figure 7. Comparison of cellulose $\delta^{18}O$ chronologies from Fontainebleau (FON) and Angoulême (ANG). (a) 51-year running correlations between FON and ANG, the horizontal lines mark the 0.01 significance level for the high-pass (solid line) and low-pass (dashed line) filtered data; (b) low-pass filtered data; (c) high-pass filtered data; (d) standard deviation (SD) of the high-pass filtered data calculated for 51-year running windows; (e) number of trees. The blue and red shaded areas mark the parts of the FON and ANG chronologies respectively, where the number of trees is < 4.

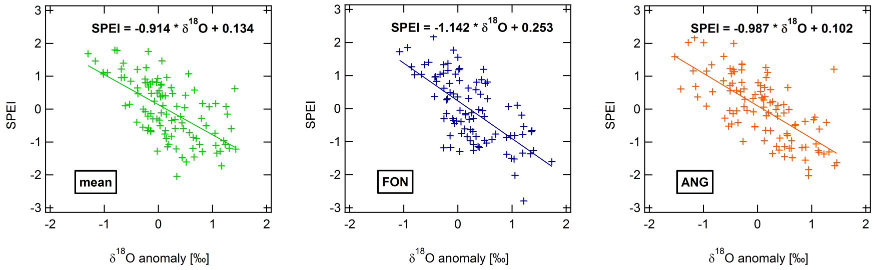

Figure 8. Linear regression between the drought index SPEI (June–August) and cellulose $\delta^{18}O$ for the mean chronology, FON and ANG. The equations describing the regression line provide the models for drought reconstruction.

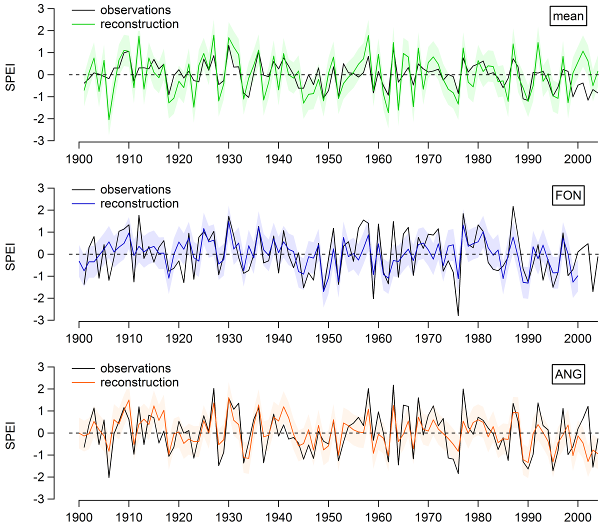

Figure 9. Observed and reconstructed drought indices SPEI (June–August). The shaded areas mark the confidence intervals of the reconstructions.

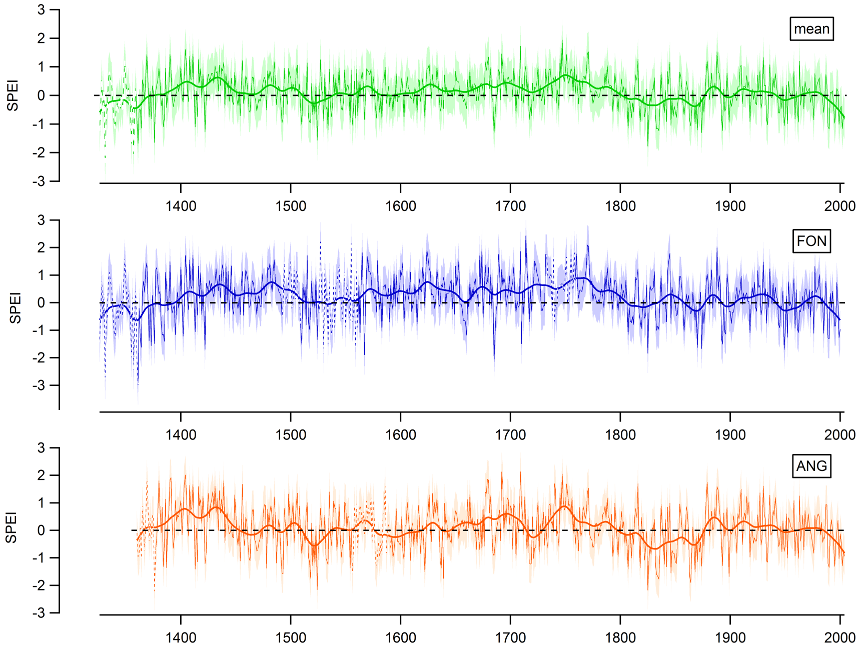

Figure 10. Reconstructions of the drought index SPEI (June–August) based on tree ring cellulose $\delta^{18}O$ for Fontainebleau (FON), Angoulême (ANG), and a composite reconstruction for the two sites (mean). The thick lines are the fitted 30-year cubic smoothing splines. The shaded areas mark the confidence intervals of the reconstructions, and the dotted lines mark the parts of the chronologies where the number of trees is < 4.

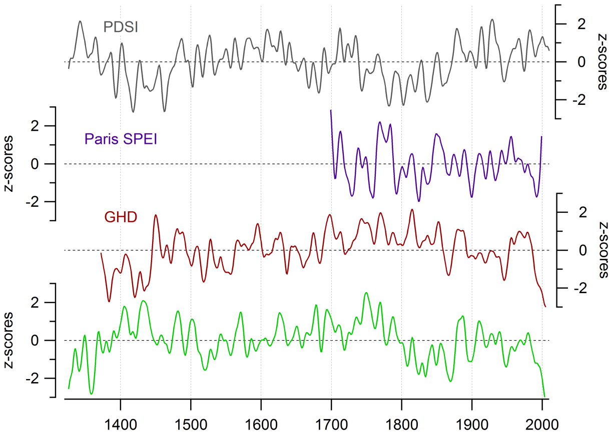

Figure 11. Comparison of the regional drought reconstruction (bottom panel) and other records: A Palmer Drought Severity Index (PDSI) extracted for the study region from the "Old World Drought Altas" (Cook et al., 2015), a Standardized Precipitation Evapotranspiration Index (SPEI) calculated using the historical instrumental records from Paris (Slonosky, 2002; Rousseau, 2009), and a composite record of French grape harvest dates (GHD; Daux et al., 2012).

Table 1. Offsets in average $\delta^{18}O_c$ values and correlations between tree cohorts for their periods of overlap. Italics indicate significant correlations ($p < 0.01$).

Site	**Cohorts**	**Offset [‰]**	**Correlation**	**Years of overlap**
FON	PD–CH	0.34	*0.38*	87
	CH–PE1	0.78	*0.50*	31
	PE1–PE2	0.96	*0.53*	66
	PE1–Etien et al. (2008)	0.36	0.19	24
	PE2–Etien et al. (2008)	0.95	0.22	24
ANG	B–GR	0.76	*0.60*	120
	GR–TW	1.05	*0.63*	41

Table 2. Pearson correlation coefficients (r) between $\delta^{18}O_c$ and monthly meteorological variables, as well as with summer (JJA) and growing season (AMJJAS) averages; 'mean' refers to the combined chronology of the FON and ANG sites. Correlations in italics/bold are significant at the 0.01/0.001 level. See text for data sources.

	Average temperature			Maximum temperature			Precipitation			SPEI		
Month	FON	ANG	mean	FON	ANG	mean	FON	ANG	mean	FON	ANG	mean
J	-0.04	0.12	0.08	-0.03	0.15	0.10	-0.19	-0.09	-0.15	-0.22	-0.09	-0.14
F	0.03	0.15	0.14	0.06	0.17	0.16	-0.18	-0.11	-0.14	-0.17	-0.11	-0.14
M	0.05	*0.23*	0.16	0.08	*0.27*	0.20	-0.13	*-0.24*	-0.21	-0.13	-0.23	-0.19
A	0.17	0.15	0.19	0.15	0.17	0.19	0.03	0.06	0.00	0.01	0.03	-0.02
M	*0.26*	**0.32**	**0.31**	*0.28*	**0.34**	**0.32**	*-0.25*	*-0.25*	*-0.25*	-0.22	*-0.25*	-0.21
J	0.18	**0.41**	**0.35**	*0.25*	**0.48**	**0.42**	**-0.48**	**-0.47**	**-0.57**	**-0.45**	**-0.50**	**-0.57**
J	**0.53**	**0.55**	**0.58**	**0.57**	**0.60**	**0.61**	**-0.39**	**-0.40**	**-0.45**	**-0.42**	**-0.44**	**-0.41**
A	**0.34**	**0.45**	**0.43**	**0.36**	**0.48**	**0.45**	*-0.30*	-0.14	-0.23	*-0.29*	-0.13	-0.23
S	0.19	0.17	0.22	0.20	0.19	0.22	0.15	0.12	0.05	0.15	0.11	0.06
O	0.19	*0.25*	*0.23*	0.20	0.22	0.21	-0.08	0.09	-0.01	-0.12	0.09	-0.01
N	-0.03	0.04	0.00	-0.03	0.03	-0.01	-0.03	0.04	0.08	-0.05	0.08	0.05
D	-0.02	0.03	0.03	-0.03	0.03	0.03	0.08	0.12	0.13	0.07	0.15	0.15
JJA	**0.49**	**0.62**	**0.61**	**0.54**	**0.70**	**0.67**	**-0.60**	**-0.53**	**-0.63**	**-0.69**	**-0.68**	**-0.62**
AMJJAS	**0.48**	**0.57**	**0.59**	**0.51**	**0.65**	**0.63**	**-0.49**	**-0.40**	**-0.54**	**-0.62**	**-0.57**	**-0.52**

Table 3. Comparison of reconstructed and observed drought index SPEI: correlation coefficient (r), reduction of error statistic (RE) and coefficient of efficiency (CE). To calculate these statistics, the data were divided into randomly selected calibration (two-thirds) and verification (one-third) data sets. The given values are the averages and standard deviations of 1000 iterations of this validation.

Site	**Statistic**	**Average**	**Standard deviation**
FON	r	-0.69	0.07
	RE	0.39	0.10
	CE	0.36	0.11
ANG	r	-0.68	0.09
	RE	0.45	0.14
	CE	0.42	0.15
mean	r	-0.62	0.08
	RE	0.37	0.11
	CE	0.34	0.13