Holocene interaction of maritime and continental climate in Central Europe: New speleothem evidence from Central Germany

Sebastian F.M. Breitenbach⁎, Birgit Plessenb, Sarah Waltgenbachc, Rik Tjallingiib, Jens Leonhardtd, Klaus Peter Jochume, Hanno Meyerf, Bedartha Goswamig, Norbert Marwang, Denis Scholzc

a Sediment- and Isotope Geology, Institute for Geology, Mineralogy and Geophysics, Ruhr-Universität Bochum, Universitätsstr. 150, 44801 Bochum, Germany
b GFZ-Potsdam, Section Climate Dynamics and Landscape Evolution, Potsdam, Germany
c Institut für Geowissenschaften, Johannes Gutenberg-Universität, Mainz, Germany
d Thüringer Höhlenverein, Erfurt, Germany
e Climate Geochemistry Department, Max Planck Institute for Chemistry, Mainz, Germany
f Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research, Periglacial Research Section, Potsdam, Germany
g Potsdam Institute for Climate Impact Research (PIK), Member of the Leibniz Association, Potsdam, Germany

ARTICLE INFO

Keywords:
Germany
Speleothem
Holocene
Oxygen isotopes
Carbon isotopes
S/Ca
Sr/Ca
Palaeoclimate
Continental climate
Maritime climate
Climate boundary

ABSTRACT

Central European climate is strongly influenced by North Atlantic (Westerlies) and Siberian High circulation patterns, which govern precipitation and temperature dynamics and induce heterogeneous climatic conditions, with distinct boundaries between climate zones. These climate boundaries are not stationary and shift geographically, depending on long-term atmospheric conditions. So far, little is known about past shifts of these climate boundaries and the local to regional environmental response prior to the instrumental era.

High resolution multi-proxy data (stable oxygen and carbon isotope ratios, S/Ca and Sr/Ca) from two Holocene stalagmites from Bleßberg Cave (Thuringia) are used here to differentiate local and pan-regional environmental and climatic conditions Central Germany through the Holocene. Carbon isotope and S/Ca and Sr/Ca ratios inform us on local Holocene environmental changes in and around the cave, while δ18O (when combined with independent records) serves as proxy for (pan-)regional atmospheric conditions.

The stable carbon isotope record suggests repeated changes in vegetation density (open vs. dense forest), and increasing forest cover in the late Holocene. Concurrently, decreasing S/Ca values indicate more effective sulfur retention in better developed soils, with a stabilization in the mid-Holocene. This goes in hand with changes in effective summer infiltration, reflected in the Sr/Ca profile. Highest Sr/Ca values between 4 ka and 1 ka BP indicate intensified prior calcite precipitation resulting from reduced effective moisture supply.

The region of Bleßberg Cave is sensitive to shifts of the boundary between maritime (Cfb) and continental (Dfb) climate and ideally suited to reconstruct past meridional shifts of this divide. We combined the Bleßberg Cave δ18O time series with δ18O data from Bunker Cave (western Germany) and a North Atlantic Oscillation (NAO) record from lake SS1220 (SW Greenland) to reconstruct the mean position of the Cfb-Dfb climate boundary. We further estimate the dynamic interplay of the North Atlantic Oscillation and the Siberian High and their influence on Central European climate. Repeated shifts of the Cfb-Dfb boundary over the last 4000 years might explain previously observed discrepancies between proxy records from Europe. Detailed correlation analyses reveal multi-centennial scale alternations of maritime and continental climate and, concurrently, waning and waxing influences of Siberian High and NAO on Central Europe.

1. Introduction

European climate is characterized by heterogeneous climate conditions, with distinct boundaries that demarcate large-scale geographical regions with a coherent climatic pattern (Kottke et al., 2006). The coherence between the climatic characteristics within each such climate zone depends on large scale atmospheric systems and can be detected using climate network techniques (Rheinwalt et al., 2016).
Central European climate is strongly influenced by intricately linked North Atlantic Oscillation (NAO) and Siberian High (SH), which govern (winter) precipitation and temperature over Europe. Both, NAO and SH, are expressions of (mainly wintery) semi-permanent, quasi-stationary surface pressure features that reside over the North Atlantic and northern Eurasia that closely interact with each other (Cohen et al., 2001). We will illuminate the non-stationary interaction between NAO and SH and its relation to the boundary that separates maritime and continental climates in the discussion below.

The boundary between the modern maritime Cfb and continental Dfb climatic zones according to the Köppen-Geiger classification (Peel et al., 2007), is not stationary, but shift in space and time (Kottek et al., 2006). Future shifts of the Cfb-Dfb climatic boundary mirror circulation changes that result from global warming and might lead to more frequent extreme weather patterns like heat waves and droughts, with significant repercussions for society (Cohen et al., 2014; Mann et al., 2018). To delineate the mechanisms related to Central European climate, it is imperative to understand the interaction of maritime and continental climates and the positioning of the border between the two. Climatic changes from multi-annual to centennial timescales can affect human society (Büntgen et al., 2011, 2016; Kennett et al., 2012; Ludlow et al., 2013; Tan et al., 2015). Pre-industrial communities experienced spatio-temporal variations in regional precipitation and temperature (droughts, floods, unusual cold spells) more severely due to the direct impact on agricultural yield, in combination with high production and re-distribution costs. While multiple underlying causes for temperature or precipitation changes have been invoked, including solar and volcanic forcing, coupled with changes in atmospheric circulation patterns (Hurrell, 1995; Brönnimann, 2007; Ludlow et al., 2013; Ridley et al., 2015; Thébement et al., 2015; Büntgen et al., 2016), little is known about the associated changes in seasonal climate hydrological and thermal conditions in Central Europe for the Holocene (Simonis et al., 2012).

Temporally highly resolved (i.e. seasonal to decadal) multi-proxy reconstructions are the most promising tool to put the local and regional environmental response into perspective to global changes, including rising temperatures. Most importantly, there is an urgent need for palaeoclimate reconstructions, which are sensitive to seasonal aspects of regional and local climate as well as environmental conditions (Wong and Breeker, 2015; James et al., 2015; Rehfeld et al., 2016). Only with sufficiently high resolved and spatially distributed reconstructions can we begin to quantify shifts in climate boundaries (Seager et al., 2018). Holocene climatic and environmental variability in Central Europe (and the responsible forcings) remains insufficiently understood, due to seasonal biases in some palaeoclimatic time series, and a lack of chronological control in others. Although tree ring records give very detailed insights into changes in summer temperatures and precipitation, they cover only the last ~2.5 ka (Büntgen et al., 2011, Büntgen et al., 2016). Available high-resolution lacustrine reconstructions, based on robust varve counting and/or radiocarbon chronologies, often cover only parts of the Holocene (e.g. von Grafenstein et al., 1999; Martin-Puertas et al., 2012; Czycmik et al., 2016; Malkiewicz et al., 2016).

Speleothems (secondary cave carbonates) constitute powerful paleo-environmental archives that can be dated very accurately radio metrically and provide a large number of environmental and climate proxies (Dorale et al., 2007; Fairchild and Baker, 2012). However, so far, only a few speleothem-based Holocene records are available from Europe, mainly from southern and western Europe (Fohlmeister et al., 2012; Frisia et al., 2005a; McDermott et al., 2011; Mischel et al., 2017; Niggemann et al., 2003; Smith et al., 2016; Verheyden et al., 2000). One exception is the well dated high-resolution isotope record from Spannagel Cave, but due to its position in the Alps, it remains difficult to separate the influencing factors involved (Mangini et al., 2005; Fohlmeister et al., 2013). Highly resolved, well-dated time series are available, but limited to the last few thousand years (e.g. Proctor et al., 2000; Boch et al., 2009; Mangini et al., 2005). While these time series have added important details to our understanding of the factors that regulate European climate variability, additional spatial coverage is required to fully comprehend local to regional variability.

Here we present a multi-decadally to decadal resolution Holocene multi-proxy reconstruction from two stalagmites from Bleßberg Cave, Thuringsia, Germany, and investigate shifts of the mean position of the maritime-continent climatic boundary through time. The location of Bleßberg Cave at the western limit of continental climate region makes this study site particularly well suited to investigate spatio-temporal changes of the Cfb-Dfb boundary, δ13C and δ18O are combined with Sr/Ca and Sr/Ca records to gain insights into past local and (pan-)regional environmental dynamics in Central Europe over the last 11,000 years. We attempt to single out environmental responses to climatic changes at a more continental setting and to reconcile apparently divergent palaeoclimatic reconstructions from different European regions. We argue that east-west shifting of the maritime and continental climate boundary might explain observed differences between palaeoclimatic reconstructions from western and eastern sites. Our results suggest that reconstructions from western sites might not be representative of conditions further east and vice versa and that time series networks are needed to better understand local responses to global climatic changes.

2. Geographic and geologic setting

2.1. Geology and geography of Bleßberg Cave

The NW-SE oriented Bleßberg Cave is located at 50°25′28″ N and 11°01′13″ E at ca. 500 m a.s.l. at the southern fringe of the Thuringian schist mountains ca. 7 km from the town of Eisleben, Germany (Fig. 1). Bleßberg Cave developed in Triassic marly limestones (Lower Muschelkalk/Anisium) and is orientated parallel to the Franconian Line, a Hercynian (NW-SE) directed, deep-reaching Cretaceous reverse fault (Reichert et al., 2008). The cave was discovered in 2008 during construction of the new high-speed rail line between Berlin and Munich. Blasting work opened the cave ~240 m from the south portal of the ~8 km long Bleßberg tunnel.

The cave had no naturally accessible entrance prior to its discovery, but is drained by a small stream. The lack of a natural entrance preserved the cave from human or animal disturbance before 2008. The cave’s atmosphere is relatively stable, but the stream running through the main passage and airflow through breakdown could influence cave ventilation.

Bleßberg Cave is overlain by 12 to 50 m (35–40 m above the sampling site) of marly limestone (Lower and Middle Muschelkalk), which carries a relatively thin soil. The soil can be classified as agriculturally altered Leptosol (Rendzina, Ap-Ah-(T)-Cv-Cn, Boden, 2006), possibly with clay enrichment below the Ah horizon. The vegetation directly above the cave is anthropogenically altered farmland, consisting mainly of weeds and crops, whereas up hill, mixed deciduous and pine forests take over.

2.2. Climatic characterisation of the study site

The cave site is located near the watersheds between Weser, Elbe and Rhine rivers. Modern climate can be characterized as Cfb climate in the Köppen classification, with a mean annual air temperature of ca. 6°C at Neuhau am Rennweg, ca. 12 km NE from the cave (Kottek et al., 2006). Rather high precipitation (862 mm/year, DWD Climate Data Center, www.dwd.de/EN/ourservices/cdctfp/cdctfp.html) results from orographic precipitation on the southwest-facing slope of the Thuringian mountains. Meteorological data from the station Neuhau am Rennweg shows no significant seasonality in precipitation, whereas air temperature is highest in July (average TJuly = 14.6°C) and coldest in January (average TJan = −3.2°C) (Fig. 2).

The seasonality in temperature leads to significant variation in...
potential and real evapotranspiration, which in turn leads to maximum effective infiltration between September and March (Fig. 2). Thus, the isotopic composition of dripwater entering Bleßberg Cave should be slightly biased towards the winter season, with implications for the interpretation of the $\delta^{18}$O_speleothem signal, similar to other caves in Germany (e.g. Wackerbarth et al., 2010; Mischel et al., 2015).

Observational rainfall and temperature data reveals that winter temperature at Neuhaus am Rennweg correlates positively with the NAO index (data from ftp://ftp.cpc.ncep.noaa.gov), which is in agreement with the results of Baldini et al. (2008). The modern location of the Cfb-Dfb boundary east of Bleßberg Cave mirrors the oceanic influence in Central Europe (Kottek et al., 2006). In line with earlier results (Baldini et al., 2008; Riechelmann et al., 2017), we are confident that $\delta^{18}$O in dripwater and speleothems can be used to study winter climate when the site is influenced by the Siberian High, which brings cold and dry air masses to Bleßberg Cave. The position of the cave near the western limit of the continental Dfb climate regime makes it highly sensitive to shifts of the boundary between the Cfb and the Dfb climate further west.

This position however also means that any link between Bleßberg and the NAO system might be lost if the climate boundary is located west of the cave and continental climate prevails (see discussion).

### 2.3. Samples

The stalagmite samples BB-1 and BB-3 were recovered following the discovery of the cave. Sample BB-1 is 17.7 cm long and ca. 7 cm wide, while BB-3 is ca. 15 cm long with a diameter of ca. 9 cm. BB-1 was broken before collection, and the original drip site feeding this sample is thus unknown. Sample BB-3 was collected from the main passage and showed a fresh and wet surface (Fig. 1c). The samples were cut along their growth axes and polished and show a very clear, pale-ochre colour, with rarely visible lamination and very little detrital material.
3. Methodology

3.1. Cave monitoring

In order to gain insight into cave ventilation and infiltration dynamics, which influence the geochemical proxies in Bleßberg Cave, microclimatic parameters are monitored. Between 2009 and 2012, monitoring was severely limited; since January 2013, data collection improved with renewed access to the cave. A CORA device (Luetscher and Ziegler, 2012) is used to log temperature, humidity, air pressure and pCO2 at 4-h intervals. Dripwater was collected in airtight 5–12 mL vials during our visits and stored in cool, dark conditions until analysis.

Stable isotope values ($\delta^{18}$O and $\delta^D$) of dripwater were determined at the Alfred Wegener Institute (AWI) in Potsdam. 18 samples were collected to cover the different seasons (Table 1, Fig. 3). A Finnigan MAT Delta-S mass spectrometer equipped with two equilibration units was used for the online determination of hydrogen and oxygen isotopic composition. The external errors for standard measurements of hydrogen and oxygen are better than 0.8‰ and 0.1‰, respectively (Meyer et al., 2000).

3.2. Mineralogy and geochemistry

Optical microscopy and fluorescence microscopy were performed...
on a Zeiss Jenalumar, SEM imaging and X-ray diffraction (XRD) at the Helmholtz Centre Potsdam, Deutsches GeoForschungsZentrum, Potsdam, Germany, to verify the mineral structure and lamination of the stalagmite samples. BB-1 consists of calcite and has been tested with cathodoluminescence (CL) microscopy to test for any diagenetic alterations. BB-3 grew on a flowstone in the middle of a ca. 8 m high passage and consists of calcite, too. Most of the stalagmite shows very clear parallel calcite crystals and horizontal growth layers, making the stalagmite ideally suited for palaeoclimate studies. The lowermost centimeters of the sample are already part of the underlying flowstone and show brownish clay layers and radial fan-structured crystals. Examination of thin sections under blue light reveals growth intervals with bright green laminae, suggesting the presence of organic material (Fig. 4b).

3.3. 230Th/U dating

For 230Th/U-dating, sub-samples with masses between 50 and 200 mg were cut from the growth axes of the individual stalagmites using a precision diamond wire saw as well as a micro band saw. In total, 41 230Th/U-ages were determined (N BB-1 = 18, NBB-3 = 23, Table 2). Chemical preparation and mass spectrometric analysis were performed at the Max Planck Institute for Chemistry, Mainz. For separation of Th and U, the samples were dissolved in 7 N HNO3, and a mixed 229Th-233U-236U spike was added (see Gibert et al., 2016, for a detailed description of spike calibration). Potential organic material was removed by addition of a mixture of concentrated HNO3, HCl and H2O2. After evaporation and re-dissolution in 6 N HCl, U and Th were chemically separated using ion exchange column chemistry (Yang et al., 2015). For mass spectrometric measurements, the separated fractions of U and Th were dissolved in 2 ml of 0.8 N HNO3 and analysed with a Nu Plasma multi-collector inductively coupled plasma mass spectrometer (MC-ICPMS). Analytical details are described in Obert et al. (2016). Ages were calculated using the half-lives of Cheng et al. (2000), and the correction for potential detrital contamination assumes an average 229Th/238U weight ratio of the upper continental crust (3.8 ± 1.9) and 230Th, 234U and 238U in secular equilibrium.

### Table 1

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Sampling date</th>
<th>Location</th>
<th>δ18O [‰ VSMOW]</th>
<th>δD [‰ VSMOW]</th>
<th>d excess</th>
</tr>
</thead>
<tbody>
<tr>
<td>BBH-NG-23.01.09</td>
<td>23.01.2009</td>
<td>N. Passage</td>
<td>−9.61</td>
<td>No data</td>
<td>No data</td>
</tr>
<tr>
<td>BBH-NG-23.01.09</td>
<td>23.01.2009</td>
<td>N. Passage</td>
<td>−9.62</td>
<td>−65.2</td>
<td>11.7</td>
</tr>
<tr>
<td>BBH-NG-23.01.09</td>
<td>23.01.2009</td>
<td>N. Passage</td>
<td>−9.59</td>
<td>−66.1</td>
<td>10.6</td>
</tr>
<tr>
<td>BBH-SG-24.01.09</td>
<td>24.01.2009</td>
<td>S. Passage</td>
<td>−9.11</td>
<td>−61.6</td>
<td>11.3</td>
</tr>
<tr>
<td>BBH-SG-24.01.09</td>
<td>24.01.2009</td>
<td>S. Passage</td>
<td>−9.13</td>
<td>N.A.</td>
<td>N.A.</td>
</tr>
<tr>
<td>BBH-SG-24.01.09</td>
<td>24.01.2009</td>
<td>S. Passage</td>
<td>−9.11</td>
<td>−61.6</td>
<td>11.3</td>
</tr>
<tr>
<td>BBH-MF1 24.11.2012</td>
<td>24.11.2012</td>
<td>Station 1</td>
<td>−9.72</td>
<td>−66.9</td>
<td>10.9</td>
</tr>
<tr>
<td>BBH-MF2 24.11.2012</td>
<td>24.11.2012</td>
<td>Station 1</td>
<td>−9.72</td>
<td>−66.6</td>
<td>11.2</td>
</tr>
<tr>
<td>dripwater 24.08.2013 BB-1 24.08.2013</td>
<td>S. Passage</td>
<td>−9.71</td>
<td>−67.3</td>
<td>10.4</td>
<td></td>
</tr>
<tr>
<td>dripwater 24.08.2013 BB-1 24.08.2013</td>
<td>S. Passage</td>
<td>−9.64</td>
<td>−66.7</td>
<td>10.4</td>
<td></td>
</tr>
<tr>
<td>dripwater 24.08.2013 BB-1 24.08.2013</td>
<td>S. Passage</td>
<td>−9.67</td>
<td>−66.7</td>
<td>10.7</td>
<td></td>
</tr>
<tr>
<td>dripwater 24.08.2013 2 m 86</td>
<td>24.08.2013</td>
<td>S. Passage</td>
<td>−9.20</td>
<td>−64.9</td>
<td>8.7</td>
</tr>
<tr>
<td>dripwater 24.08.2013 103</td>
<td>24.08.2013</td>
<td>S. Passage</td>
<td>−9.56</td>
<td>−66.1</td>
<td>10.4</td>
</tr>
<tr>
<td>dripwater 24.08.2013 2 m 103</td>
<td>24.08.2013</td>
<td>S. Passage</td>
<td>−9.52</td>
<td>−66.0</td>
<td>10.2</td>
</tr>
<tr>
<td>drip water 27.05.2017, MP90</td>
<td>27.05.2017</td>
<td>Baseball Arena</td>
<td>−9.63</td>
<td>−64.9</td>
<td>12.2</td>
</tr>
<tr>
<td>drip water 27.05.2017, MP90</td>
<td>27.05.2017</td>
<td>Baseball Arena</td>
<td>−9.42</td>
<td>−64.6</td>
<td>10.8</td>
</tr>
<tr>
<td>drip water 27.05.2017, MP90</td>
<td>27.05.2017</td>
<td>Baseball Arena</td>
<td>−9.58</td>
<td>−64.6</td>
<td>12.1</td>
</tr>
<tr>
<td>drip water 27.05.2017, MP90</td>
<td>27.05.2017</td>
<td>Baseball Arena</td>
<td>−9.67</td>
<td>−64.6</td>
<td>12.8</td>
</tr>
</tbody>
</table>

Fig. 3. Stable isotope data from dripwater and precipitation. A) Dripwater δ18O data from Bleßberg Cave compared with mean monthly δ18O of precipitation at Wasserkuppe/Rhön (squares), Leipzig (circles) and Hof (triangles) (IAEA/WMO, 2017). B) Cross-plot of dripwater and monthly mean precipitation δ18O and δD values from the same stations. Large empty symbols denote mean values for the winter season (September to March).
established using COPRA (Breitenbach et al., 2012). In order to calculate 2.5% and 97.5% quantile confidence limits (ca. 2σ), 5000 Monte Carlo (MC) simulations were run using a piecewise cubic interpolating. COPRA transposes the chronological uncertainty to the proxy domain, so that the x-axis (time) is error-free (Breitenbach et al., 2012). This procedure allows statistical reanalysis with the entire ensemble of age models.

3.4. Stable carbon and oxygen isotope analysis

Fig. 4. Images of stalagmites BB-1 (A) and BB-3 (B), together with fluorescence images, stable isotope profiles (δ¹³C and δ¹⁸O) and μXRF Sr/Ca, and S/Ca ratios. The lower brownish section of BB-3 belongs genetically to the flowstone onto which BB-3 grew and the geochemical data from this interval is not discussed in the main text.

Stable carbon and oxygen isotope ratios of the stalagmites (δ¹³C and δ¹⁸O) were determined at the Helmholtz-Centre Potsdam (GFZ). Stable isotope samples have been obtained using a “Sherline 5410” vertical milling machine with an inhouse-built sample stage and a micrometer-drive to allow precise movement of the speleothem block, for stepwise milling at a resolution of 0.1 mm. Carbonate aliquots were analysed for δ¹³C and δ¹⁸O values using a Finnigan MAT253 isotope ratio mass spectrometer (ThermoFisher Scientific) connected to an automated carbonate-reaction device (KIEL IV). Samples of around 20–60 μg were automatically dissolved with 103% H₃PO₄ at 72 °C, and the isotopic composition was measured on the released and cryogenically purified CO₂. Results are reported relative to VPDB, and replicate analysis of international reference material (NBS19) yielded external standard deviations of 0.06‰ (1σ) for both δ¹³C and δ¹⁸O.

3.5. Micro-XRF analysis

The distribution of elements detectable with micro X-ray
fluorescence (μXRF) has been determined in both samples, BB-1 and BB-3 (Fig. 4a, b). Both samples were scanned along the stable isotope track using an EAGLE III XL μXRF spectrometer (Röntgenanalytik, Germany) at the GFZ. The μXRF is equipped with a 50 W Rh X-ray source that was operated at 40 kV and 300 μA. Measurements were performed under vacuum every 40 μm with a spot size of 50 μm and a counting time of 60 s. The resulting intensities are given in counts per second (cps) and the reproducibility is proven with repeated measurements. Besides Ca, significant count variations were found for Sr and S. Element intensities have been normalized over Ca to minimize sample density effects and scaled by multiplying the intensity ratios by 1000.

4. Results

4.1. Cave microclimate monitoring

Bleßberg Cave has a rather stable thermal regime, with a mean temperature of 8.7 ± 0.1 °C and a relative humidity of 99.8 ± 0.2%. Cave air pCO2 is slightly elevated relative to outside air and varies from 780 to 824 ppm, with a maximum in winter, which probably results from enhanced effective infiltration and introduction of soil CO2 in winter. Although the level of the cave stream changes significantly in the eastern cave passages, it seems not to strongly alter the cave atmospheric conditions.

Stable oxygen and deuterium isotope values (Table 1) from 18 Bleßberg Cave dripperwater samples are relatively invariable, with δ18O = −65.5 ± 1.7‰ and δ2H = −9.5 ± 0.2‰ (Fig. 3a) and agree well with the winter season (Sep-Mar) averages of precipitation from GNP stations Wasserkuppe/Rhône (δ18O-Sep-Mar = −69.6‰, δ2H-Sep-Mar = −10.1 ± 1‰, IAEA/WMO, 2017), Hof (δ18O-Sep-Mar = −71.3‰, δ2H-Sep-Mar = −9.8 ± 1.3‰) and Leipzig (δ18O-Sep-Mar = −67.1‰, δ2H-Sep-Mar = −9.6 ± 1.4‰) (Fig. 3a, b, IAEA/WMO, 2017; meteorological data cover 1983–2003 for Hof and Wasserkuppe/Rhône, and 1986–2003 for Leipzig). The slightly lower mean winter value for Wasserkuppe/Rhône is due to the altitude effect (the station’s altitude is 921 m a.s.l., compared to 567 m in Hof and 125 m in Leipzig). All values fall on the Global Meteoric Water Line (GMWL, Fig. 3b), which indicates that no secondary evaporation effects influence our samples. The d excess is in the range of global average of precipitation. Temporal and spatial dripper δ18O variability is rather small, with only about 0.6‰ variation between different sampling days and sampling spots in different cave passages, but denser sampling is needed to evaluate this aspect.

4.2. 230Th/U dating and age modeling

The results of 230Th/U-dating are presented in Table 2. The 238U concentration of the stalagmites varies between 0.614 ± 0.009 (BB-3)
and 3.74 ± 0.02 μg g⁻¹ (BB-1). BB-1 has a substantially higher ²³⁸U-concentration compared to BB-3. The ²³²Th-concentration is generally low (< 1 ng g⁻¹; Table 2) and even below the detection limit for some samples.

Only four samples show an elevated input of detrital material with a ²³³Th-content between 1.09 ± 0.01 ng g⁻¹ (NR-31 of BB-1) and 55.87 ± 0.53 ng g⁻¹ (BB3-5.2 of BB-3), respectively. All four samples were taken at the base of the two stalagmites. The relatively high concentrations of uranium (up to 3.4 μg g⁻¹) and lack of detrital thorium (Table 2) result in mean dating errors of ± 30 years in BB-1 and ± 86 years in BB-3 (2σ).

Stalagmite BB-1 grew between 5.6 and 0.6 ka BP (kilo years before present, i.e. 1950 CE), while BB-3 grew between 11.9 and 5.4 ka BP (Fig. 5). Unfortunately, the last ca. 600 years are missing and the reconstruction cannot be directly linked with meteorological data.

The age-depth slopes indicate that both stalagmites grew at different rates; slow growth prevailed between ca. 11 and 7 ka BP and 3.7 and 0.2 ka BP, whereas faster growth occurred in the mid-Holocene between 6 and 4 ka BP (Fig. 5). BB-3 shows relatively steady growth in the early Holocene and a slight increase after ca. 7 ka BP. A hiatus has been identified during ²³⁸Th/U-dating of BB-3, located at 100.2 mm from top, which is characterized by yellow-brownish coloring of the stalagmite.

This hiatus covers a ca. 400-year-long interval between 8.26 and 7.85 ka BP. BB-1 shows a drastic reduction in growth rate after ca. 4.7 ka BP, approaching values more similar to those in the early Holocene. While growth rates differ, the convergence in both stalagmites during the overlapping period suggests common forcing(s) that adjust the growth pattern. The interval where both stalagmites overlap (ca. 5.6 to 5.2 ka BP) reveals that BB-3 (the older of the two) has lower ¹³C and S/Ca, and higher Sr/Ca and ¹⁸O values compared to BB-3. BB-1 also shows higher high-frequency variability.

Before an interpretation in terms of environmental changes can be attempted the observed offsets must be explained as both stalagmites are apparently precipitated from different feeding systems. BB-1 was likely fed from fracture flow, whereas BB-3 received seepage flow water from a low-permeability host rock matrix (Fairchild and McMillan, 2007).

Seepage flow from the epikarst matrix goes in hand with prolonged residence time of infiltrating water and mixing of waters of different age. This ultimately leads to smoothing of the proxy signals, which explains the reduced high-frequency variability in BB-3. Seepage flow from a less permeable matrix forces longer interaction of the infiltrating water with soil and host rock, which can well explain the observed lower ¹³C and S/Ca ratios, and higher Sr/Ca levels in BB-3. Lower ¹³C in BB-3 compared to BB-1 can be explained by less CO₂-degassing in a matrix with lower permeability, i.e. seepage from a host rock matrix is less prone to open conditions and PCP, compared to fracture flow where open conditions might occur faster and more frequently (Fairchild and McMillan, 2007). Lower S/Ca ratios in BB-3 might result from better sulfur retention in the host rock, compared to larger fractures. Prolonged dissolution of host rock in a seepage flow system can also increase the Sr/Ca baseline in BB-3 (which is subsequently modulated by PCP dynamics). On the other hand, a fracture-flow fed BB-1 would reflect a faster response to changes in infiltration, higher S/Ca caused by lower sulfur retention in the soil and epikarst, and a lower Sr/Ca baseline due to shorter water-rock interaction. Finally, the lower ¹⁸O values observed in BB-1 might result from a more direct response to cold-season infiltration and/or snowmelt through the fractures feeding BB-1, and less mixing with summer-season water in the epikarst. These factors must be kept in mind when interpreting the proxy variability in both stalagmite records. We note however that millennial-scale trends and centennial dynamics in both records are still interpretable in terms of environmental conditions at the cave site."

4.3. Stable isotopes

In total, about 2000 samples have been analysed for stable isotopes in both stalagmites (1451 in BB-1, and 538 in BB-3). The total amplitude in δ¹⁸O across the entire Holocene is −2‰, from −5‰ around 9 ka BP to −7‰ in the last millennium (Figs. 4a, 6). In the younger stalagmite BB-1, δ¹³C values range from −7‰ to −5.7‰, with an increasing trend from ca. −6.5‰ in the lower part towards a maximum value of −5‰ at 1.3 ka BP. At the top, the BB-1 profile shows a large shift to values around −6.8‰. Superimposed on the millennial to centennial changes, multi-decadal variability is detected with amplitudes of 0.3 to 0.5‰. In the older stalagmite BB-3, δ¹³C values vary from −8‰ to −5‰. A trend to lower values is observed in the upper half of this stalagmite (since −9 ka BP), which is only interrupted by two prominent excursions before and after the hiatus around 8 ka BP. A first notable shift of −0.8‰ to lower values occurred just before the hiatus. A second shift is found after the hiatus, with a rapid change from values around −5.4‰ to −6‰ (Figs 4b, 6). No clear trend is found between 7.2 and 5.3 ka BP, but multi-centennial scale variability of −0.2–0.4‰ characterizes this section. Where BB-3 overlaps with BB-1, it is offset and less negative by −0.5‰.

Across both stalagmites, δ¹³C values vary −8.5‰, from −9.8‰ to −1‰ (Figs. 4, 6), with a general trend towards lower values through the Holocene (Fig. 6). A plateau, with values around −7‰, is found between 11 ka BP and 9 ka BP. This stable phase is interrupted at ~11 ka BP, where δ¹³C increased around 1‰ for 400 years. Around 8.7 ka BP, a rapid 2‰ shift to more negative values occurred, followed by the hiatus (~8.3–7.8 ka BP). This hiatus lasted 400 years, after which growth commenced again. The δ¹³C profile restarted with values around −6‰, decreasing steadily to −8‰ around 7 ka BP (Fig. 6). Variable δ¹⁸O values are found during the mid-Holocene. After 3 ka BP, a further decrease occurred, with the lowest values of the entire profile at around 2 ka BP, and a last shift to higher values since 1.5 ka BP.

In addition to the samples obtained for proxy records, ~250 samples were drilled for Hendy tests (Hendy, 1971). In BB-1, 5 Hendy test transects were performed, with 10 to 15 samples each. Neither a significant positive correlation between δ¹³C and δ¹⁸O, nor an increase of δ¹⁸O values with distance from the apex is observed. From BB-3, no Hendy tests have been obtained.

4.4. µXRF data

In the older stalagmite BB-3, the Sr/Ca ratio varies from 65 and 160, while in BB-1, it ranges from 100 to 195 (Fig. 4). In BB-3, Sr/Ca increases slowly until ca. 6.8 ka BP, and reaches then a relatively stable level at ca. 150, before the values fall back to ca. 120 at the top of the stalagmite. The general trend varies only little. In BB-1, the Sr/Ca
profile is characterized by a higher degree of variability. Starting at ca. 120, it increases to ca. 160 until ca. 3.3 ka BP, when the values drop to ca. 130. Afterwards, Sr/Ca values increase again until ca. 2 ka BP, before the trend reverses, and Sr/Ca values are lowest in BB-1 (ca. 100). High-frequency changes are found superimposed on the long-term trends. Sharp multi-decadal increases are found at 4.8, 3.5, 3.2, and 1.5 ka BP, with Sr values reaching 195.

S/Ca ratios show a long term trend opposite to Sr/Ca over the Holocene. In BB-3, a multi-millennial trend to lower S/Ca values is found (Fig. 6c), from 80 in the early Holocene to 30 at around 6 ka BP. Few significant deviations are observed, although centennial-scale changes are present. In BB-1, S/Ca values further decline between 5.6 and 4.5 ka BP, but then stabilize at a level of 30 to 20 (Fig. 4a). Only minor fluctuations are observed at centennial scale, with one stronger shift to higher S/Ca values at 3.3 ka BP and a second possibly at 1.7 ka BP.

4.5. Statistical analyses of spatio-temporal relationships between proxy records

To statistically test for periods of east- and westward Cfb/Dfb boundary migration, we performed a rigorous correlation analysis on the δ¹⁸O time series from BB-1 and BU-4 and the DSM record from lake SS1220. As the sampling resolution of the proxy records is different and irregular, and the time axes contain a certain amount of uncertainty, we use a special correlation analysis that considers the chronological uncertainties of each record when calculating correlations. For all palaeoclimate records, we consider a moving window of 500 years, moved in steps of 50 years over the time span 4 to 0.5 ka BP. At each window position the COPRA proxy record ensembles (with 5000 members each) for the two caves are taken and in a first step 5000 pairs of proxy records are randomly selected, one from each cave. For each pair, we extract the part of the proxy time series that falls within the time window, and estimate the cross-correlation at lag zero between these extracted sub-time series using a Gaussian kernel-based correlation approach (NESTool, normalized kernel width $h = 0.75$) (Rehfeld et al., 2011). This results in an empirical test distribution of 5000 correlation estimates for that particular window location. Repeating this for all other timings of the window, we obtain an ensemble of 5000 time series of correlation estimates. The statistical significance of this ‘observed’ correlation distribution is obtained on the basis of a 2-sample Kolmogorov-Smirnov (KS) test, which tests the null hypothesis that the probability distribution of the correlation samples (from the observed data) equals the probability distribution of correlation values of uncorrelated times series of the same statistical properties than the observed data). The random correlation estimates are obtained by repeating the correlation calculation as before, but with the difference that the randomly chosen proxy time series are further uniform-randomly shuffled before estimating the kernel-based cross correlation. The KS test statistic is based on the maximum difference between the
two empirical cumulative distribution functions obtained from the two correlation samples and is not necessarily related to an overlap of their interquartile ranges. This test provides a p-value, the statistical significance of which is obtained at a confidence level of 1%, after taking into account multiple comparisons by using Holm's method combined with the Dunn–Šidák correction factor.

5. Discussion

Stable oxygen and carbon isotope data from speleothem carbonates have been studied for decades, but ambiguities remain with regard to identification of the processes underlying changes in these isotope proxy records (e.g. Gascoyne, 1992; McDermott, 2004; Mangini et al., 2005; Lachniet, 2009; Fairchild and Baker, 2012). To fully understand individual controls on environmentally sensitive proxies, monitoring is required at the local level. Here, the combination of stable isotope and μXRF elemental data allows detailed insights into seasonally-biased controls on the proxies in speleothems from Bleßberg Cave and the reconstruction of local and (par-)regional environmental changes.

Since the hydrological system in the epikarst acts as a low pass filter, the climatic signal does not reach the cave immediately, thus imposing a lag of unclear but likely annual to multiannual length to speleothem records from Bleßberg Cave.

5.1. Interpretation of geochemical proxies

5.1.1. δ13C, S/Ca and Sr/Ca records as proxies for vegetation, soil development and infiltration

Both stalagmite δ13C values and S/Ca ratios (Fig. 6a, b) can serve as proxy for changes in vegetation composition and density, soil microbial activity as well as changes in infiltration (Lechleitner et al., 2017). In temperate locations, these factors contribute most to the carbon budget in the dripwater and can be reflected in speleothem δ13C values (Genty and Massaut, 1999; Genty et al., 2001; Fairchild and Baker, 2012). Speleothem Sr/Ca constitutes a sensitive proxy for changes in infiltration, with higher values being recorded during times of reduced moisture availability (Fairchild and Treble, 2009). Since Bleßberg Cave had no natural entrance before its discovery during tunnel construction, ventilation-induced in-cave CO2 degassing can be regarded as subordinate process affecting the δ13C values of the dissolved inorganic carbon (DIC) and in stalagmites. CO2 dynamics could potentially be mediated by changing stream volume and/or flow velocity (Troester and White, 1984). Unfortunately, in Bleßberg Cave, this process cannot be tested as a consequence of the massive concrete injection during the tunneling process (ILEK, 2011).

The Bleßberg site is characterized by higher soil microbial and vegetation activity during spring and summer, which, together with higher winterly effective infiltration induces elevated (relative to surface atmosphere) CO2 levels with low soil δ13C signature. In the epikarst, this δ13C signal is adjusted to more positive values, depending on the residence time of the percolating water, and ultimately transferred to the cave (Lechleitner et al., 2017). The stalagmites then record a complex signal of epikarst and in-cave degassing, vegetation composition and density as well as soil microbial activity above the cave (McDermott, 2004). This mechanistic explanation is corroborated by the observed low δ13C values between −7‰ and −10‰ in the time series from Bleßberg Cave. If δ13C would solely reflect the composition of the host rock limestone much higher values (+1 to −2‰) would be expected. Thus, we infer that Holocene vegetation and soil development left a recognizable imprint on the δ13C signal. Denser C3 vegetation (temperate forest) cover or soil development would lead to lower speleothem δ13C, while more open conditions or thin soils would shift δ13C towards higher values (e.g. Scholz et al., 2012). A more open forest and/or thinner soil with reduced microbial activity and reduced production of soil CO2 on the other hand would result in less negative δ13C values (Genty et al., 2001, 2003).

Further information on soil and vegetation dynamics can be gained from the S/Ca signal (Fig. 6b). Generally, multiple sulfur sources have been identified, including sea spray, dust and volcanic aerosols as well as anthropogenic pollution (Frisia et al., 2005b; Wynn et al., 2010; Wolff et al., 2017). However, the observed long-term trend in S/Ca is most likely not caused by changes in volcanic aerosol loadings, which vary randomly. Similarly, changes in sea spray are unlikely to cause for the observed long-term trend because Bleßberg Cave is a continental site. Anthropogenic influence is similarly unlikely, as a major impact would be expected with industrialization, which is not covered by the youngest section of BB-1.

Inorganic and organic S is also derived from weathering and organic matter decomposition (Edwards, 1998). Sulfur reduction and cycling also depends on redox potential Eh and pH conditions in the soil, although this is generally more significant in waterlogged soils (Connell and Patrick Jr., 1968; Hussin, 2012). The Rendzina above Bleßberg Cave is characterized by high pH, low water holding capacity and strong ventilation. The predominant inorganic form of sulfur in aerobic soils is easy-leachable sulfate in the aqueous solution and attached to minerals. This also means that plant uptake of sulfate ions competes with direct leaching to the cave, with better developed soils resulting in improved retention and reduced S-loss through leaching. Sulfur can be retained in the soil by two main mechanisms, i.e. immobilization and adsorption, the relative importance of which can vary (Anderson, 1988). The retention times of S can be long, depending on adsorptive capacity and the input of S (Edwards, 1998). Importantly, increasing contact time between S and soil results in reduced SO42− leaching due to adsorption and/or transformation of S to less mobile components (Edwards, 1998). The negative multi-millennial trend in S/Ca observed in BB-3, which is anticorrelated with Sr/Ca (Fig. 6b, c), suggests that increasing prior calcite precipitation (PCP) (as evidenced by higher Sr/Ca) results in enhanced S retention (lower S in speleothems), due to reduced infiltration and prolonged contact between S components and soil. The decrease of S in BB-3 through the early to mid-Holocene, thus, likely results from soil development, in tandem with diminishing (summertime) infiltration.

Interestingly, visual inspection of the S/Ca and Sr/Ca profiles reveals a positive correlation on multi-decadal to centennial timescales independent from the long-term trend (Fig. 4), indicating that other factors partake in forcing sulfur mobility on these timescales. Changes in weathering rates, aerosol input or soil activity are all potential candidates for this high-frequency variability. Still, the responsible mechanism must affect both S/Ca and Sr/Ca to cause the observed pattern. Enhanced S immobilization and mineralization and low PCP under colder and wetter conditions, reflected in lower S and Sr content in the stalagmites (Edwards, 1998) is a viable explanation. In summary, both proxies are interpreted as indicators of local conditions, which of course are embedded and moderated by the larger-scale climatic milieu.

The climate in the study region shows no clear seasonality in precipitation (Fig. 2b), but evapotranspiration, and thus effective infiltration, varies throughout the year, with a maximum during the winter months (Fig. 2d). Infiltration-sensitive proxies should therefore be able to record infiltration changes, with a bias towards the warm season and reduction in spring to autumn infiltration. Dry conditions can lead to (more) open conditions in soil and epikarst, and associated CO2-degassing from infiltrating water into epikarst air pockets and the cave atmosphere, which in turn allows PCP (Fairchild and McMillan, 2007, Breitenbach et al., 2015). PCP can alter the elemental composition of dripwater feeding the stalagmites and increases the Sr concentration in the aqueous solution (Fairchild and Treble, 2009).

The Sr/Ca profile across both stalagmites shows a positive trend throughout the early to mid-Holocene (11 ka BP to 6 ka BP, Fig. 6c). This trend is interpreted as resulting from increasing summertime PCP with increasing insolation and a more negative moisture balance in summer (Kalis et al., 2003; Russo and Cubasch, 2016). Subsequent to the Holocene Thermal Maximum (Wanner et al. 2011), Sr/Ca ratios
remain on a relatively constant level (even if the offset between both stalagmites is taken into account). The observed higher centennial-scale variability after ca. 5.5 ka BP is likely due to the fact that this stalagmite received fracture flow with a faster response to surface conditions. Several positive Sr peaks at 3.5 ka, 3.2 ka and 1.5 ka BP suggest intense, but short-lived periods of moisture reduction and intensified PCP. Since about 2 ka BP, Sr/Ca values decrease again, possibly related to weakening summer insolation (Russo and Cubasch, 2016) and improved hydroclimatic conditions with higher effective infiltration. Strong PCP also results in increased degassing of CO₂ and precipitation of calcite (Fairchild and McMillan, 2007) and can, thus, also be reflected in less negative δ¹³C values. However, as outlined above, the δ¹³C signal is largely unrelated to Sr and governed by vegetation dynamics.

5.1.2. Controls on δ¹⁸O in precipitation and dripwater

5.1.2.1. Regional factors controlling δ¹⁸O in precipitation. Precipitation is delivered to Bleßberg Cave throughout the year and mainly from the North Atlantic, without clear seasonal maximum (Fig. 2) but higher effective infiltration in the winter season (Fig. 3). The oxygen isotope composition in precipitation (δ¹⁸Oprecip) is strongly linked to air temperature and the NAO conditions (Baldini et al., 2008). These relationships are exemplified with δ¹⁸O data from meteorological stations Hof, Leipzig and Wasserkuppe/Rhön in Fig. 3a; at all three stations δ¹⁸O is highest in summer, with a total annual amplitude of ca. 4 to 5‰. Thus, under higher temperatures and/or a positive NAO, δ¹⁸O is expected to increase. The temperature control on δ¹⁸O in the region (using Hof, Leipzig and Wasserkuppe/Rhön as reference) is ca. +0.23‰°C⁻¹, thus counter-balancing the negative temperature effect on carbonate precipitation. This is different to Bunker Cave, for which Wackerbarth et al. (2010) observed an additional effect of the amount of winter precipitation on precipitation δ¹⁸O, which we do not observe for the reference stations used here.

5.1.2.2. Local factors controlling δ¹⁸O in infiltrating water. Local factors that influence δ¹⁸O in soil, epikarst and cave include i) evaporation of soil water, ii) evaporation of dripwater, iii) CO₂-degassing of dripwater in the cave, iv) temperature, and v) mixing of older and younger water in the epikarst. Evaporation of soil water is more pronounced in the summer season (Fig. 2), but is unlikely to contribute to changes in Bleßberg Cave dripwater on short time scales because May and August samples are not significantly enriched relative to winter month samples (Fig. 3a) and because no secondary evaporation trend, with values following a slope lower than 8, is observed (Fig. 3b). Evaporation of dripwater can be excluded based on the same observations and the fact that in cave relative humidity is stable > 98%. CO₂ degassing from dripwater into the cave air during times of lower drip rates might affect δ¹⁸O (Mühlinghaus et al., 2009). However, this effect is very small (i.e. ∆δ¹⁸O ~0.1‰/1000 s. at T = 10 °C, Deininger et al., 2012) and would not explain δ¹⁸O changes in the range of 0.5 to 2 permil over multi-decadal to centennial time scales. Temperature effects on the dripwater are unlikely because the cave air temperature does not vary appreciably (Tcave air = 8.7 ± 0.1 °C). Finally, mixing of infiltration waters of different age is likely the main reason for the relatively stable dripwater δ¹⁸O (Fig. 3a) and a very narrow range along the GMWL (Fig. 3b). The small offset between the two individual stalagmites, and the dampened variability found in BB-3 are evidence for mixing of waters in the epikarst, which differ for the two stalagmite sites.

5.1.3. δ¹⁸O as proxy of atmospheric circulation

The Bleßberg Cave δ¹⁸O (δ¹⁸Ospel) record (Fig. 6d) integrates precipitation history and the corresponding δ¹⁸O signal in precipitation (δ¹⁸Oprecip), seasonal changes in infiltration, and temperature in a complex fashion and thus records longer-term pan-regional environmental dynamics. Given the observed seasonal infiltration pattern (Fig. 2), it can be argued that both dripwater and speleothem δ¹⁸O values represent a mean signal that is slightly biased towards the winter season. This is similar to other German caves (e.g. Mischel et al., 2015; Wackerbarth et al., 2010). Thermal conditions in the cave are stable and cave air temperature changes over the course of the Holocene are unlikely to have caused the observed 2‰ amplitude in δ¹⁸Ospel values. Assuming a (cave) temperature dependence of the isotopic composition during carbonate precipitation of ca. −0.24‰°C⁻¹ (McDermott, 2004), counter-balanced by a positive δ¹³Cspel/T relation of ca. +0.23‰°C⁻¹ would imply ΔT Holocene ~8 °C to explain the observed 2‰, which seems unrealistically high for the last 9–10 ka BP. Several lines of evidence suggest an important influence of atmospheric circulation, moisture source dynamics and δ¹³Cspel on δ¹⁸Ospel in Bleßberg Cave. A significant positive correlation has been found between the NAO and δ¹⁸O over Central Europe (Hurrell, 1995; McDermott et al., 2011; Baldini et al., 2008; Comas-Bru et al., 2016). The strong positive impact of the winter NAO on δ¹⁸O is caused by more frequent inflow of cold, isotopically depleted precipitation during negative NAO phases (Baldini et al., 2008). A tentative link between NAO and Bleßberg δ¹⁸Ospel is found in the visual similarity between BB-1 δ¹⁸O and the updated detrital silicate mineral (DSM) record from lake SS1220 (DSMSS1220 = δ¹⁸O, Zn, Tl; Olsen et al., 2012, Fig. 6d). For this study we updated the SS1220 chronology by using the Intcal13 calibration curve (Reimer et al., 2013) to re-calibrate the ¹³C measurements originally reported in Olsen et al. (2012). We then sampled 5000 stratigraphically ordered age models from the multimodal re-calibrated ages, which resulted in a 5000 member DSMSS1220 proxy record ensemble. Warmer conditions with increased meridional flow of warm maritime air over SW Greenland are normally associated with negative NAO, while positive NAO conditions are characterized by cold, dry air masses over southern Greenland (Olsen et al., 2012). In Central Europe, negative (positive) NAO conditions are linked to colder (warmer) and drier (wetter) air under increased (decreased) Siberian High influence, reflected in more negative (positive) δ¹⁸Ospel values (Baldini et al., 2008; Trouet et al., 2009).

Thus, climatic conditions in Central Europe are not solely determined by maritime (NAO) influence from the Atlantic, but also by the SH. The wintery SH is another major player that influences atmospheric circulation, precipitation, and temperature (Berry and Chorley, 2010, Tubi & Dayan 2013). The interaction between the NAO and the SH results from complex tropospheric and stratospheric linkages (Cohen et al., 2001; Cohen et al., 2014; Ambaum and Hoskins, 2002). Both systems are intimately linked with the Arctic Oscillation (AO), which has been suggested to lag NAO dynamics by a few days (Ambaum and Hoskins, 2002). A positive NAO leads to a stronger stratospheric vortex and a deepened low pressure cell over the North Pole, and effectively a positive AO. This in turn results in stronger Westerlies and fewer and smaller meanders of the zonal jet stream (Cohen et al., 2014). Under such conditions, the strong Westerlies transport more and warmer moisture from the N Atlantic towards Europe and into Eurasia and the SH has rarely the chance to reach Central Europe. In a negative NAO/AO mode on the other hand the Arctic low pressure system is weaker and the zonal jet is characterized by increased meandering which allows frequent atmospheric blocking and intrusion of cold air from the north/northeast. The Westerlies are weaker and frequently redirected southward, thus delivering less moisture to Germany, while the intruding Arctic air is very cold. Thus, the influence of the SH increases under negative NAO/AO conditions. Inflow of north-easterly continental air under strong SH conditions replaces warm maritime air from the west. The increasing SH influence with distance from the Atlantic coast is directly expressed in the change from a maritime Cfb climate to a continental Dfb climate at ca. 10°E (Peel et al., 2007). Thus, δ¹³Cspel from Bleßberg Cave is sensitive to shifts of the boundary between the two climate zones.

5.2. Local and pan-regional environmental changes in Central Europe

The multi-proxy datasets from BB-1 and BB-3 allow detailed insights
into both local and pan-regional environmental conditions throughout the Holocene. Below, we first discuss how the local environment evolved after the last deglaciation. Then, we compare the climatic conditions in Thuringia to those in Europe and the North Atlantic realm and discuss the role of a shifting Cfb-Dfb climate boundary and jet stream as explanation for the differences between sites.

5.2.1. Local environmental changes

We use the δ¹³C, S/Ca, and Sr/Ca time series to reconstruct local vegetation, soil and infiltration dynamics through the Holocene. High δ¹³C values between the Late Glacial and the onset of the Holocene (Fig. 6a) point to an open, and possibly grassland vegetation, typical for periglacial conditions (Hahne, 1991; Bebermeier et al., 2018). After a short-lived excursion to drier and/or colder conditions shortly before 11 ka BP, the δ¹³C record trended towards higher values for the next ca. 2400 years, potentially indicative of an early coniferous forest (Hahne, 1991; Bebermeier et al., 2018). Decreasing S/Ca and low, slowly increasing Sr/Ca values between 11 and 9 ka BP indicate cumulative soil development and improved sulfur retention with sufficient effective infiltration to minimize summerly PCP (Fig. 6b, c). The Preboreal to early Holocene development is comparable to that in the southern Harz Mountains, ca. 100 km to the north of our study site (Bebermeier et al., 2018).

Elevated δ¹³C values after ca. 9 ka BP (the onset of the Boreal, sensu Hahne, 1991) point to a return to a more open forest, or a change in vegetation composition. Pollen profiles from the Rhön (Hahne, 1991), ca. 40 km west of Bleßberg Cave, and Eichsfeld (Bebermeier and others, 2018) suggest replacement of pine forest by Corylus communities which prefer warmer conditions. A change to more open forest is not mirrored in the S/Ca and Sr/Ca ratios, suggesting that it was not accompanied by increased (summer) drought and related soil or infiltration changes. Less negative δ¹⁸O values at ca. 9 ka BP also suggest relatively warm/wet conditions (Fig. 6d). A drastic decrease in both δ¹³C and δ¹⁸O between ca. 8.7 ka BP and 8.4 ka BP, concurrent with a slight increase in Sr/Ca might be explained by a post-glacial afforestation under drier summer conditions and intensified PCP. A scenario where the vegetation composition changed from deciduous to coniferous forest is less likely because one would expect a shift to more positive δ¹³C values (Amiotte-suchet and others, 2007), opposite to the observed trend. A concurrent trend towards more negative δ¹⁸O values suggests cooling and drying and/or increased winter precipitation prior to the 8.2 ka event. The hiatus found in BB-3 covers ca. 400 years, starting 8.26 and ending 7.85 ka BP, with a minimum and maximum length of 64 and 710 years, respectively, when considering the chronological uncertainties in the BB-3 record. The hiatus corresponds closely with the 8.2 ka event (Mayewski and others, 2004; McDermott, 2004) and can only be explained by complete absence of dripwater, caused either by flooding or infilling of the cave passage, re-routing of the water, prolonged drought, or permafrost conditions (Vaks and others, 2010, 2013; Lechleitner and others, 2017). A return of permafrost during the 8.2 ka event seems the most likely scenario because we have no evidence in support of infilling of the cave with sediment or water, or drought with zero effective infiltration, which should be reflected in increased Sr/Ca (increased PCP). The development of permafrost (i.e. at least two consecutive years with ground temperature < 0 °C, Harris and others, 2009) during the 8.2 ka event would require a decrease in mean annual air temperature of > 6 °C with respect to modern conditions, which at first glance seems very high. A drastic, although not reaching 6 °C, decrease in temperature during the 8.2 ka event has been found in western and northern Europe (Seppä and others, 2009; Vincent and others, 2011). However, given the complex interaction between geomorphology, geology, precipitation, air and ground temperatures (Harris and others, 2009) local permafrost development in hilly Thuringia seems plausible. More detailed work is needed to reconstruct the climatic conditions surrounding the 8.2 ka event.

After ca. 7.8 ka BP, conditions improved and stalagmite growth commenced again, with rapid soil development and denser forest vegetation, as seen in lower S/Ca ratios and δ¹³C values. This climatic amelioration was accompanied by slightly increased PCP, likely due to warmer summers (Fig. 6c). After ca. 6.7 ka BP, soil development and/or sulfur retention level out, with only two excursions around 5.5 and 3 ka BP, when conditions seem to have been comparatively drier. Vegetation, however, was likely more dynamic, as reflected in noticeable variations in δ¹³C between 5.5 and 3.5 ka BP (Fig. 6a), sometimes together with increased PCP. A trend to more negative δ¹³C values after 3 ka BP suggests increased forest cover, which was reversed only in the last ca. 1500 years, possibly related to deforestation activities of Bronze Age settlers (Bebermeier and others, 2018).

5.2.2. Pan-regional climatic changes

Today, Central Europe is governed by maritime Cfb climate, while eastern Europe is characterized by continental Dfb climate (Kotte and others, 2006; Peel and others, 2007). The meteorological border between these two climatic provinces is located over Germany (Fig. 1) and has also been identified as a significant shift in the orientation of precipitation isochron pattern (Rheinwalt and others, 2016). Importantly, over time this climatic boundary is not stationary and shifts, depending on climatic conditions (Kotte and others, 2006). It should, in principle, be possible to detect past changes of the geographical position of this climatic divide if sufficient well dated and resolved proxy time series were available from locations distributed across Europe. Links between reconstructions that are located on each side of the boundary and influenced by factors that govern climate in the respective zone can give insights how this boundary shifted in space and time. Using only one additional record from the opposing climate zone would not suffice however, because one would find either positive, negative or zero correlation. While the latter two scenarios could be interpreted to indicate a climate boundary between the two sites, the first scenario could only suggest a common forcing factor, but would not clarify whether the climate boundary is located east or west of both sites. Thus, at least one third record is needed to identify the position of the boundary relative to the sites, and the controls relevant at each location.

To estimate the geographical position of the Cfb-Dfb boundary, we use two time series from the North Atlantic realm in addition to the Bleßberg Cave BB-1 record (see Figs. 1 and 7). The first of these records is the δ¹⁸O time series from stalagmite BU-4 (δ¹⁸OBU4) from Bunker Cave in western Germany (Fohlmeister and others, 2012). The δ¹⁸OBU4 record has been interpreted as reflecting a complex signal of temperature and moisture supply. While the δ¹⁸O signal in Bunker Cave reflects multiple processes, less negative δ¹⁸O values indicate colder and/or drier conditions, while lower values suggest warmer and/or wetter climate. Multi-annual variations in surface temperature and precipitation amount have been linked to NAO dynamics (Wackerbarth and others, 2010; Riechelmann and others, 2017).

The second reconstruction used to test the concept of a shifting climatic boundary is a multi-element proxy record of DSM input from lake SS1220 (DSMSS1220) in southwestern Greenland (Olsen and others, 2012). The DSM record is the sum of Sr, Zn and Ti counts observed in the lake sediment profile, in a similar fashion as proposed by Saarni and others (2016), and is used here to infer NAO conditions. Higher DSM values indicate increased sediment transport associated with snowmelt and runoff into the lake. In southwestern Greenland increased runoff and sediment supply are expected in mild and/or short winters with longer thaw periods and prolonged runoff; conditions that have been associated with negative NAO phases (Olsen and others, 2012). High DSM values are thus interpreted as representative for negative NAO phases (see Fig. 6d).

These time series and the Bleßberg δ¹⁸O record from BB-1 (δ¹⁸OBB1) reflect relevant circulation features with sufficient resolution and chronological control for the last ca. 4000 years to allow a first estimation of past boundary dynamics. To establish the Cfb-Dfb boundary, we estimate correlations between the three records, which we then link with the information on the atmospheric circulation patterns (NAO...
index and SH strength) assigned to the proxy records to estimate the position of the climate boundary. It must be kept in mind that the interpretation of $\delta^{18}O$ from Bleßberg and Bunker caves is opposite in sign, with lower $\delta^{18}O$ values in Bleßberg Cave indicating colder and drier conditions under a more pronounced SH and negative NAO influence, while reflecting wetter and warmer (maritime) conditions above Bunker Cave. This relationship could change however due to changes in seasonal precipitation pattern and resulting slopes of the $\delta^{18}O$-T relationship. The obtained correlation patterns allow the identification of the following Cfb-Dfb boundary position scenarios (Fig. 7):

- **Scenario 1 Cfb-Dfb boundary east of Bleßberg Cave**

The first scenario is similar to modern conditions (Kottek et al., 2006), but with the Cfb-Dfb boundary located shifted further east of Bleßberg Cave. The maritime influence reaches towards eastern Central Europe while the Siberian High influence on western Europe is relatively weak. Under such conditions, Bleßberg and Bunker caves would record the same maritime climate and a common forcing with strongly positive NAO influence. The interpretation of $\delta^{18}O_{\text{in}}$ would be reversed, because continentality, and with it the slope of the $\delta^{18}O$-T relationship would be lowered (Bowen, 2008). Consequently, Bleßberg Cave would react much like Bunker Cave, with a positive correlation between the two $\delta^{18}O$ records, while DSM$_{SS1220}$ values should be lowered (Fig. 7). A boundary east of Bleßberg Cave thus would result in a positive correlation between all three records. Modern observational data show a stronger Atlantic influence on Central Europe and a significant positive link between NAO index and precipitation $\delta^{18}O$ at Wasserkuppe/Rhön near Bleßberg Cave (Baldini et al., 2008). Such conditions occur when the zonal jet is stronger and less meandering (Cohen et al., 2014). Unfortunately, since the Bleßberg Cave record does not cover the last few hundred years, we cannot test the inferred links with meteorological records.

- **Scenario 2 Cfb-Dfb boundary between Bleßberg Cave and Bunker Cave**

A second scenario would see the Cfb-Dfb boundary located between the two German caves, with Bunker Cave being governed by maritime climate and Bleßberg Cave by continental climate conditions. Under these conditions, $\delta^{18}O_{\text{in}}$ should decrease (due to increased continentality and resultant colder winters), while the $\delta^{18}O_{\text{BU4}}$ and DSM$_{SS1220}$ records might remain unchanged or increase compared to scenario 1, with the cave records reflecting atmospheric dynamics related to the climatic regime on either side of the Cfb-Dfb boundary. As a result, negative correlation between $\delta^{18}O_{\text{BB1}}$ and $\delta^{18}O_{\text{BU4}}$ should be found, while the Greenland DSM$_{SS1220}$ record would be expected to show negative to nil correlation to $\delta^{18}O_{\text{BB1}}$, but remain positively linked to $\delta^{18}O_{\text{BU4}}$. As long as the Westerlies exert some influence on Bunker Cave (Fig. 7). Scenario 2 would be characterized by frequent SH
Fig. 8. Proxy records from Bleßberg ($\delta^{18}$OBB1, A), Bunker Cave ($\delta^{18}$OBU4, B), and Greenland (DSMSS1220, C) covering the last 4000 years are correlated to highlight non-stationary relationships. Solid lines in A-C denote the median proxy value and shades represent the inter-quartile ranges as obtained from COPRA proxy record ensembles composed of 5000 age-depth simulations each. The DSMSS1220 profile is based on an updated chronology in which the IntCal13 calibration curve was used to calibrate the radiocarbon ages. Randomly chosen pairs proxy records from the COPRA ensembles are used to estimated the correlations (orange bands, D-F) between each pair of records using the NESTool methodology that allows estimation of correlation for irregularly sampled time series. The relationship between the records is compared to that obtained from an ensemble of pairs of randomized time series (blue bands), and this is quantified with the p-value of a 2-sample Kolmogorov-Smirnov test (light blue shaded backgrounds, see Section 4.5 for details). (G) Correlation summary and inferred Siberian High strength and NAO index, as well as estimates of the position of the Cfb-Dfb climate boundary. Solid triangles indicate statistically significant correlations, while empty triangles refer to statistically not significant values that are deduced from the (significant) correlations between the median values of the other pairs. Insufficient data points in the DSMSS1220 record prevent computing the correlations with $\delta^{18}$OBB1 and $\delta^{18}$OBU4 between 3.7 and 3.3 ka BP. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
influence on Central Europe, a consistently changing NAO and stronger meandering of the jet stream, but relatively mild conditions.

- **Scenario 3 Cfb-Dfb boundary east of Bleßberg Cave**

A third scenario might occur where the climate boundary is positioned east of Bleßberg Cave, similar to scenario 1, but with more stable climate conditions. In scenario 3, both speleothem δ¹⁸O records would be expected to be influenced by maritime climate, with rare incursions of cold Arctic air during winter and strongly positive NAO (Fig. 7b). The interpretation of δ¹⁸OBU4 would be reversed (as in scenario 1), because of strongly diminished continentality. Bleßberg and Bunker cave would be more responsive to oceanic conditions, reflected in a positive correlation between both δ¹⁸O records. A negative correlation between δ¹⁸Ospel and DSMSS1220 would result, with lower δ¹⁸Ospel and increased DSM values reflecting negative NAO dynamics and vice versa.

This scenario differs from scenario 1 in that it includes a poleward displacement of the jet stream in a generally warmer atmosphere (Archer and Caldeira, 2008; Woolings and Blackburn, 2012), so that the maximum moisture transport would be redirected towards Scandinavia. Concurrently, the Siberian High influence on Central Europe would be diminished, which in turn would lead to more stable conditions, albeit with decreased winter moisture and higher PCP at Bleßberg Cave (Fig. 6). An intensified and poleward jet and positive NAO result from a large pressure difference between a cold Arctic and very warm mid-latitudes (Archer and Caldeira, 2008). This interpretation would explain higher δ¹⁸OBU4 values (Fig. 7b) as response to drier winter conditions in Thuringia, and lower δ¹⁸OBU4 values, due to warmer and/or wetter conditions at Bunker Cave. Higher winter precipitation would be expected in Northern Europe.

- **Scenario 4 Cfb-Dfb boundary between both caves**

A fourth scenario has been found in the correlation analyses whereby the δ¹⁸OBU4 time series shows an increasing trend, while the δ¹⁸OBU1 time series suggest no clear trend (Fig. 7b) and vice versa, i.e. both cave records are anticorrelated. The DSM record is positively correlated with the δ¹⁸OBU1 time series. The δ¹⁸OBU1 values are slightly more negative compared to scenario 1 (Fig. 7b) while the δ¹⁸OBU1 and DSMSS1220 values are higher. These observations suggest a moderate to strong influence of the Siberian High on Bleßberg Cave and a pronounced negative NAO. The lower δ¹⁸OBU1 values and the positive correlation of δ¹⁸OBU1 with DSMSS1220 suggests winter conditions with frequent SH influence on Bleßberg Cave and negative NAO (i.e. cold winters bringing ¹⁸O-depleted precipitation sensu Baldini et al., 2008). We hypothesize that the climatic boundary is located between both caves.

- **hypothetical Scenario 5 Cfb-Dfb boundary west of Bunker Cave**

Theoretically, a fifth scenario could occur if very cold conditions would lead to a westward retreat of the Cfb-Dfb boundary. Such conditions may have prevailed in periods with southward displaced jet stream, e.g. the late glacial before northern hemisphere warming would allow the maritime climate to reach Central Europe. This scenario is not observed in the discussed late Holocene records. At Bunker Cave increased continentality would result in a positive slope of the δ¹⁸O-T relationship and a reversed interpretation of δ¹⁸OBU4 (higher δ¹⁸OBU4 indicating warmer conditions). Since continentional conditions would prevail, a positive correlation between the two cave records is expected. If the jet stream would be located far west of Bunker Cave, the latter would loose sensitivity to the maritime climate. This would be reflected in a reduced or lacking correlation (in either direction) between the cave proxies and the DSMSS1220 record, which would be expected to still be governed by North Atlantic climate variability and indicative of the NAO state. Western Europe would be expected to be more frequently affected by the wintery Siberian High.

The correlations among the three sites Bleßberg Cave, Bunker Cave and Greenland lake SS1220 (Fig. 8d-f) allow us to estimate changes in the longitudinal position of the Cfb-Dfb boundary and the jet stream over the last ca. 4000 years. NAO and SH pattern are inferred from the DSMSS1220 profile and the δ¹⁸OBU1 time series, respectively. The correlations, the NAO index, relative strength of the Siberian High, and the inferred position of the climate boundary are summarized in Fig. 8g.

Prior to about 3.9 ka BP, concurrent positive correlation between δ¹⁸OBU1 and δ¹⁸OBU4 and negative correlation between the cave records and DSMSS1220 suggest that the Cfb-Dfb boundary was located east of Bleßberg Cave (scenario 3) and that Central Europe was under influence of maritime climate, likely with a moderate wintery SH influence at Bleßberg Cave. With a climate boundary shifted east the influence of the NAO on Central Europe was likely increased. Relatively high DSM values suggest negative NAO conditions, which would go in hand with stronger meandering of the zonal jet and cold air inflow during winter.

After 3.9 ka BP we find a reversed δ¹⁸OBU1 and δ¹⁸OBU4 correlation (scenario 4), which we tentatively interpret as a ca. 200-year long transition phase from scenario 3 to scenario 2. Insufficient data points in the DSMSS1220 record prevent computing the correlations with δ¹⁸OBU1 and δ¹⁸OBU4 between 3.7 and 3.3 ka BP (Fig. 8e, f) and statements regarding the NAO impact on European climate must remain vague. Intermediate DSM values suggests variable NAO conditions. The negative correlation between δ¹⁸OBU1 and δ¹⁸OBU4 can be used however to infer a likely position for the Cfb-Dfb boundary between the two caves (scenario 2). This interval was followed by a shift back to scenario 4. This period with a negative NAO and strong SH lasted for only about a century (~3.3–3.2 ka BP) and was then succeeded by a 500-year long period of the similar scenario 2, with negative correlations between δ¹⁸OBU1 and δ¹⁸OBU4 and δ¹⁸OBU1 and DSMSS1220. At this time the Cfb-Dfb boundary was located between the two caves; both the SH and the NAO exerted their influence on their respective climate zones. A strong SH at that time has also been identified by Mayewski et al. (2004) in the GISP2 ice core potassium record and frequent negative NAO conditions have been noted between ca. 3.1 ka and 2.5 ka BP by Olsen et al. (2012).

Around 2.7 ka BP another transition occurred, found as a ca. 300-year interval with unclear but likely eastward boundary shift. Considering the available evidence, it is likely that this time was characterized by higher climate variability, with continued boundary shifts and variable SH and NAO strength. The low DSMSS1220 values suggests that the NAO shifted to a positive phase at the time. At 2.4 ka BP a prominent, 500-year long interval of near-modern climate set in, with the Cfb-Dfb boundary located between both caves. The SH had influenced winter conditions concurrent with a positive NAO, as indicated by the DSM values.

This maritime interplay was rapidly (within ca. 100 years) replaced by a more maritime climate that lasted from ~1.9 ka BP to ~1.1 ka BP (scenario 3). This period was characterized by a concurrent poleward displacement of the jet stream and more stable conditions in Central Europe. Winters at Bleßberg Cave were most likely drier due to this northward shift of the Westerlies which would bring increased winterly moisture to Scandinavia and Bunker Cave, while the Thuringian summers were seemingly warm and dry (as indicated by increased PCP, Fig. 6). The DSMSS1220 record suggests a positive NAO. Saarni et al. (2016) noted increased snow accumulation during positive NAO, consistent with our interpretation.

A last change is recorded in our time series at around 1.1 ka BP, when the climate boundary shifted westward, to remain between the two caves until at least 500 years ago when our Bleßberg record ends. This last period was characterized by a strong SH and concurrently variable and frequently negative NAO, which together left wintery Germany colder. The inferred strong SH is supported by the GISP2 K record (Mayewski et al., 2004). The inferred negative NAO at the time (covering the Medieval Climate Anomaly) is also supported by a tree
ring-based reconstructions of a multi-centennial mega-drought in Scandinavia (Helama et al., 2009) and humid and mild summers in Central Europe (Büntgen et al., 2011).

Above Bleßberg Cave summertime PCP diminished slowly, indicating increasing summer infiltration. In western Germany (at Bunker Cave) higher climate variability can be inferred from the δ18O record (Fig. 8b), which can be understood if we assume less severe influence of the SH on local conditions at its far western limit.

Because the last 500 years are not covered by the Bleßberg stalagmite, we cannot connect our reconstruction with historical and meteorological data. The Cfb-Dfb climate boundary continued to migrate eastward, is currently located near the eastern border of Poland, and expected to shift even further east under global warming conditions (Kotte et al., 2006).

In summary, the comparison of multiple sites allows the identification of shifts in the mean position of the Cfb-Dfb climate boundary over the last ca. 4000 years, depending on the relative strength and dynamics of the Siberian High and the Westerlies and the position of the jet stream. To refine this reconstruction, a denser network of records would be required, possibly by exploiting the new Speleothem Isotope Synthesis and Analysis database (Atsawawanunt et al., 2018). With improved spatio-temporal coverage it might also be possible to deduce directionality of forcings and determination of the speed of such climate transitions.

6. Conclusions

Two U-series dated stalagmites (BB-1 and BB-3) from Bleßberg Cave, Thuringia, are used to reconstruct local and (pan-)regional environmental conditions. The time series from stalagmites BB-1 and BB-3 cover the entire Holocene with the exception of a ca. 400-year long hiatus at 8.2 ka. All proxies discussed above respond to local processes, including speleothem, tree-ring and pollen data, will help reconstructing past regional climate variability.

In summary, the comparison of multiple sites allows the identification of shifts in the mean position of the Cfb-Dfb climate boundary over the last ca. 4000 years, depending on the relative strength and dynamics of the Siberian High and the Westerlies and the position of the jet stream. To refine this reconstruction, a denser network of records would be required, possibly by exploiting the new Speleothem Isotope Synthesis and Analysis database (Atsawawanunt et al., 2018). With improved spatio-temporal coverage it might also be possible to deduce directionality of forcings and determination of the speed of such climate transitions.

Data availability

The datasets presented here will be available on the website of the corresponding author (SFMB) and on public repository, including https://www.pangaea.de. Data can also be requested by contacting SFMB directly.

Acknowledgements

We thank Brigitte Richert (GFZ Potsdam) for μXRF analysis, and Rudolph Naumann (GFZ Potsdam) for support with XRD and XRF analyses, Heike Rothe (GFZ Potsdam) for ICP-MS analysis and Michael Gömmel for sample preparation. Andreas Hendrich (GFZ Potsdam) helped preparing Fig. 1. Frank Haubrich, Saxony, helped with detailed information on soil composition. The VdHK e.V. kindly supports this study with monitoring equipment and the Thüringer Höhlenverein, Erfurt, is thanked for logistical support. This study received funding from the European Union’s Horizon 2020 Research and Innovation programme under the Marie Skłodowska-Curie grant agreement No 691037. Denis Scholz acknowledges funding by the German Research Foundation (DFG) through grants SCHO 1274/9-1 and SCHO 1274/10-1; Norbert Marwan and Bedartha Goswami acknowledge funding through DFG MA4759/8-1.

Appendix A. Supplementary data

Supplementary data for this article can be found online at https://doi.org/10.1016/j.gloplacha.2019.03.007.

References


Cave, Germany and a detrital silicate material record from lake SS1220 (SW


Cave) higher climate variability can be inferred from the

δ18O record (Fig. 8b), which can be understood if we assume less severe influence of the SH on local conditions at its far western limit.

Because the last 500 years are not covered by the Bleßberg stalagmite, we cannot connect our reconstruction with historical and meteorological data. The Cfb-Dfb climate boundary continued to migrate eastward, is currently located near the eastern border of Poland, and expected to shift even further east under global warming conditions (Kotte et al., 2006).

In summary, the comparison of multiple sites allows the identification of shifts in the mean position of the Cfb-Dfb climate boundary over the last ca. 4000 years, depending on the relative strength and dynamics of the Siberian High and the Westerlies and the position of the jet stream. To refine this reconstruction, a denser network of records would be required, possibly by exploiting the new Speleothem Isotope Synthesis and Analysis database (Atsawawanunt et al., 2018). With improved spatio-temporal coverage it might also be possible to deduce directionality of forcings and determination of the speed of such climate transitions.

6. Conclusions

Two U-series dated stalagmites (BB-1 and BB-3) from Bleßberg Cave, Thuringia, are used to reconstruct local and (pan-)regional environmental conditions. The time series from stalagmites BB-1 and BB-3 cover the entire Holocene with the exception of a ca. 400-year long hiatus at 8.2 ka. All proxies discussed above respond to local processes, which themselves are governed by regional atmospheric dynamics. In this way, they all inform on individual aspects of environmental conditions active at the time of speleothem deposition. δ13C, δ18O and Sr/Ca inform us about local environmental changes, including vegetation and soil changes above the cave, as well as prior calcite precipitation linked to effective infiltration through the course of the Holocene. Stalagmite δ18O is interpreted as recorder of changes in moisture source dynamics and temperature and thus as (pan-regional signal. We combine the Bleßberg Cave δ18O record with distal proxy reconstructions to infer changes in the relative importance of marine and continental conditions on Central European environment.

For the last ca. 4000 years we compared the BB-1 δ18O record with the δ18O profile from stalagmitic BU-4 from Bunker Cave in western Germany and a detrital silicate material record from lake SS1220 (SW Greenland) and estimate the changes in the Cfb-Dfb climate boundary, as well as the strength of the Siberian High and dynamics of the North Atlantic Oscillation. We find a complex temporal pattern with repeated multi-centennial scale E-W shifts of the Cfb-Dfb boundary. The local environment at Bleßberg Cave reacts to an intensified Siberian High mainly with reduced soil development and increased prior calcite precipitation during the warm season, due to reduced wintery precipitation.

The interpretative value of the presented estimate of the dynamics of the climate boundary is currently limited by the low number of well-dated and highly resolved reconstructions from Europe. Using this approach to integrate available datasets from additional archives, including speleothem, tree-ring and pollen data, will help refine and possibly quantify our insights in Holocene climate and vegetation dynamics. We are confident that with improved spatial coverage the history of climate boundaries will become recognizable at much finer scale. Discrepancies between spatially distributed reconstructions might indicate competing environmental forcings and can be exploited to resolve climatic regimes in different climate zones.


