



Millennial climatic instability during penultimate glacial period recorded in a south-western France speleothem

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ABSTRACT

This study focuses on an earlier MIS (marine isotopic stage) 6 flowstone deposit (Vil-car-1 core, Villars Cave, SW-France) that recorded millennial scale climatic events. This speleothem provides a high resolution record of the penultimate glacial period on mid-latitude land. The flowstone high growth rate and the episodes with depleted calcite $\delta^{13}\text{C}_c$ values suggest that marine isotopic substage 6e environmental conditions were relatively warm and humid. Its $\delta^{13}\text{C}_c$ profile clearly displays five main climatic oscillations from 177 to 165 ka, correlated with stadial–interstadial events. Similarly to the MIS3 period, the transitions in between those interstadials and stadials are characterised by abrupt changes and intra-interstadial variability is observed.

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1. Introduction

Climatic millennial-scale variability now appears to be a ubiquitous feature over the last 400,000 years (e.g. Dansgaard et al., 1993; Martrat et al., 2007; Wang et al., 2008). The bipolar seesaw mechanism (Broecker, 1998; Stocker and Johnsen, 2003) that explains this climatic variability for the last climatic cycle, still needs additional supporting observations for older periods. The study of millennial scale variability of periods older than the last climatic cycle expressed in terrestrial records is indeed hindered by the scarcity of mid-to-low latitude continuous records with robust chronologies and sufficiently high resolution. For the penultimate glacial (marine isotopic stage (MIS) 6, 185–135 ka), these oscillations are recorded in some Antarctic ice cores and marine cores such as on the Iberian margin and Mediterranean sea (e.g. Petit et al., 1999; de Abreu et al., 2003; Martrat et al., 2004; Jouzel et al., 2007; Martrat et al., 2007; Margari et al., 2010). Over continental mid-to-low latitude areas, three speleothem records of MIS6 variability can be quoted: the Hulu/Sanbao Chinese cave record that yields an impressive resolution back to 224 ka (Cheng et al., 2006;

Wang et al., 2008), the Eastern Mediterranean record based on the Soreq-Peqi'in record (Ayalon et al., 2002; Bar-Matthews et al., 2003), and the Argentarola Italian cave one (Bard et al., 2002b). Despite its coarse resolution, the Gitana cave speleothem record, southeast Spain, also allows to depict a high frequency variability involving more humid interstadials during earlier MIS6 (Hodge et al., 2008). In other caves from central-western, Italy (Drysdales et al., 2004) and in the Israeli Negev Desert caves which are part of the NE Sahara Desert (Vaks et al., 2006), speleothems (re-)started growing towards late MIS6 ~ 142–141 ka. In all these millennial variability records of MIS6, the number of events (from 2 to 17) as well as their characteristics (e.g. frequency, magnitude, duration, age) during MIS6 versus the MIS3 pattern is far from making unanimity. Differences in both time resolution and threshold limit above which an event is considered as being a significant interstadial event, account for most of the observed discrepancies. Furthermore, unlike during MIS3, the lack of Greenland ice core milestone is a source of confusion. Recently a synthetic Greenland temperature proxy profile over the last 800 ka has been modelled, based on a record from Antarctica and the thermal bipolar seesaw concept, and tends to fill this gap (Barker et al., 2011).

The fact that MIS6 is not a “regular” glacial period also makes the study more complex. Symbol of this particular nature, sapropel S6 deposition in the Eastern Mediterranean results from enhanced low latitude hydrological activity through an intensified monsoon system episode in Eastern Africa, associated with an insolation maxima and a strong precession minima that led to a northward migration of

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the inter-tropical convergence zone ITCZ (Melieres et al., 1997; Rossignol-Strick and Paterne, 1999; Laskar et al., 2004; Malaize et al., 2006). The formation of most Mediterranean sapropel occurred during interglacial periods and humid and warm interstadials. The S6 occurrence is dated at 172 ka, earlier MIS6 glacial period (e.g. Lourens, 2004). The low carbon isotopic composition of speleothems from the Eastern Mediterranean (Soreq Cave) typical of the entire MIS6 suggests that even if climate was as cold as much of the last glacial, the conditions were never as dry (Ayalon et al., 2002). Speleothem deposition in the extreme arid region of NE Sahara (discontinuous from 177 to 135 ka) confirms that earlier MIS6 was a major humid interval in eastern Mediterranean (Vaks et al., 2006) and that humid conditions started to develop before the onset of Termination II ~ 142–141 ka. Globally, glacial MIS6 was a time interval of extended ice-sheets with some generally dry and cold conditions as inferred from high benthic $\delta^{18}\text{O}$ composition, low temperature and CO_2 concentration in Antarctica – methane concentration was intermediate (Imbrie et al., 1984; Jouzel et al., 1993; Petit et al., 1999; Lisiecki and Raymo, 2005; Jouzel et al., 2007). A Dole effect anomaly is spotted earlier MIS6 in the atmospheric O_2 of Antarctic ice cores and is mainly attributed to enhanced biospheric productivity, possibly associated with strong monsoon activity (Malaize et al., 1999; Masson et al., 2000; Landais et al., 2010).

In Europe, recent studies suggest that some climate improvements obviously punctuated this glacial MIS6 period, in particular during its earlier part (Marine Isotopic SubStage 6e, MIS6e). Several speleothem growth occurrences suggest humid periods as in Clamouse Cave, Southern France (~169.1–162.3 ka; Plagnes et al., 2002), in Stump Cross (Baker et al., 1996) and Robin Hood Caves, UK (Rowe and Atkinson, 1985). In Argentarola marine cave, Italian coast, speleothem growth is detected at 18.5 m below present day level (~190–145 ka). This implies that climate was humid and warm enough to allow speleothem deposition while sea level was at least 18.5 m below modern one (Bard et al., 2002a). Stump cross caverns' flowstone deposited only during MIS6 (~169.2–166.3 ka), MIS4 to 3 and ~10 ka. This cave seems to be flooded during warm episodes (e.g. MIS5) and the development of a continuous permafrost over the cave prevented speleothem formation during full glacial conditions: the flowstone appears to precipitate only during intermediate conditions (Baker et al., 1996). From a Western Iberian marine core, de Abreu et al. (2003) inferred that despite sea-level was lower than during last glacial, sea surface temperatures (SSTs) were warmer than during MIS4 to 2. Iceberg discharges occurred during

earlier MIS6 but they were significantly different compared to MIS3 regarding the distinct source, the limited amplitude of the related ice rafted debris (IRD) and the fact that they are not characterised by IRD pulses but rather by a group of continuous events mainly marked by the high abundance of *Neogloboquadrina pachyderma* (sinistral) (McManus et al., 1999; de Abreu et al., 2003). These 40°N observations suggested that ice sheet configuration was different and that the globally warmer temperature would have led to iceberg melting closer to the source areas. A recently published pollen study from the Ionnina basin, NW Greece provides a high resolution record of the whole MIS6 vegetation history and highlights the similarity between earlier MIS6 and 3 climatic features, especially regarding the high frequency changes (Roucoux et al., 2011). This study demonstrates that conditions were likely globally wetter in comparison to MIS3, but also cooler which is different compared to the Atlantic records.

To sum up, the earlier MIS6, as a glacial period, was a time interval with some extended ice-sheets, generally drier and colder conditions over Europe and Mediterranean compared to interglacial periods. At lower latitudes, hydrological activity was enhanced via both Asian and African monsoon system intensification, consequence of the north hemisphere summer insolation maxima and precession minima at 174 ka. This period turns out to be intermediate between full glacial and interglacial, with conditions rather similar to MIS3, with some millennial climatic oscillations, associated with limited iceberg discharges in the Northern Atlantic. Whether climate was globally warmer or not during earlier MIS6 to MIS3 is still a matter of debate. The discrepancy regarding this element of comparison from on record to the other, might just reflect small temperature difference between MIS3 and early MIS6, and local discrepancies. To improve our knowledge about the mechanisms triggering these high frequency oscillations during glacial periods, the theory of the bipolar see-saw emitted for the last climatic cycle (Broecker, 1998; Barbante et al., 2006; Capron et al., 2010) needs to be studied for MIS6 under a fairly distinct orbital configuration, as Margari et al. (2010) have initiated. Additionally, some more details about the characteristics of this variability (e.g. amplitude, length, frequency) are now needed as well as a description of the expressions of these oscillation's impact on terrestrial areas.

In this study, we focus on the bottom part of the Vil-car-1 flowstone core (Villars Cave, SW France), corresponding to the early penultimate glacial, and whose framework was already published in Wainer et al. (2011).



Fig. 1. Site locations.

2. Regional setting

Villars Cave (45°30'N, 0°50'E, 175 masl) is a 10 km long network lying within a Bathonian–Bajocian oolitic limestone, 10 to 40 m below surface in the Dordogne region (SW France, Fig. 1). The local climate is oceanic temperate with humid and mild winter, largely influenced by the Atlantic Ocean, only distant by ~180 km. According to the nearby Nontron meteorological station, mean annual air temperature above Villars cave is 12.1 ± 0.7 °C from September 1984 to August 2007. For this period, the winter temperature is 6.4 ± 1.3 °C and summer temperature of 18.8 ± 1.1 °C in average (Genty et al., 2006; Wainer et al., 2009). Total annual rainfall varies from year to year but is distributed all over the year (1020 ± 206 mm).

Rain and drip water isotopic studies suggest that both the high density of the fissures network in the host carbonate and the thickness of the soil (≤ 20 cm) allow persistent infiltration all year along, even during periods when theoretical evapotranspiration estimates exceed rainfall amounts (Genty et al., 2006). Stalactite drip rates do not cease but slow down during summer and autumn months. The

C3 type vegetation above the cave consists of a deciduous forest with juniper, oak and hornbeam trees.

Two different levels of galleries separated by 10–20 m can be distinguished: the mean temperature in the upper galleries is 12.4 ± 0.4 °C since 1995, oscillating seasonally. The temperature in the lower level is lower and does not show any seasonal changes (11.6 ± 0.1 °C since 2005 (Genty, 2008)).

3. Materials and methods

3.1. Drilling site and sample description

Vil-car-1 flowstone core was collected in the Balcon Chamber, ~150 m from the nearest entrance, ~25 m below surface. The flowstone is ~5 m² and gently slopes away (~20°) from a large pillar. Its surface was wet upon collection, suggesting that the speleothem was still active when drilled. The entire core is 114.2 cm long and 7.5 cm wide (Fig. 2). Its petrography varies along the core from

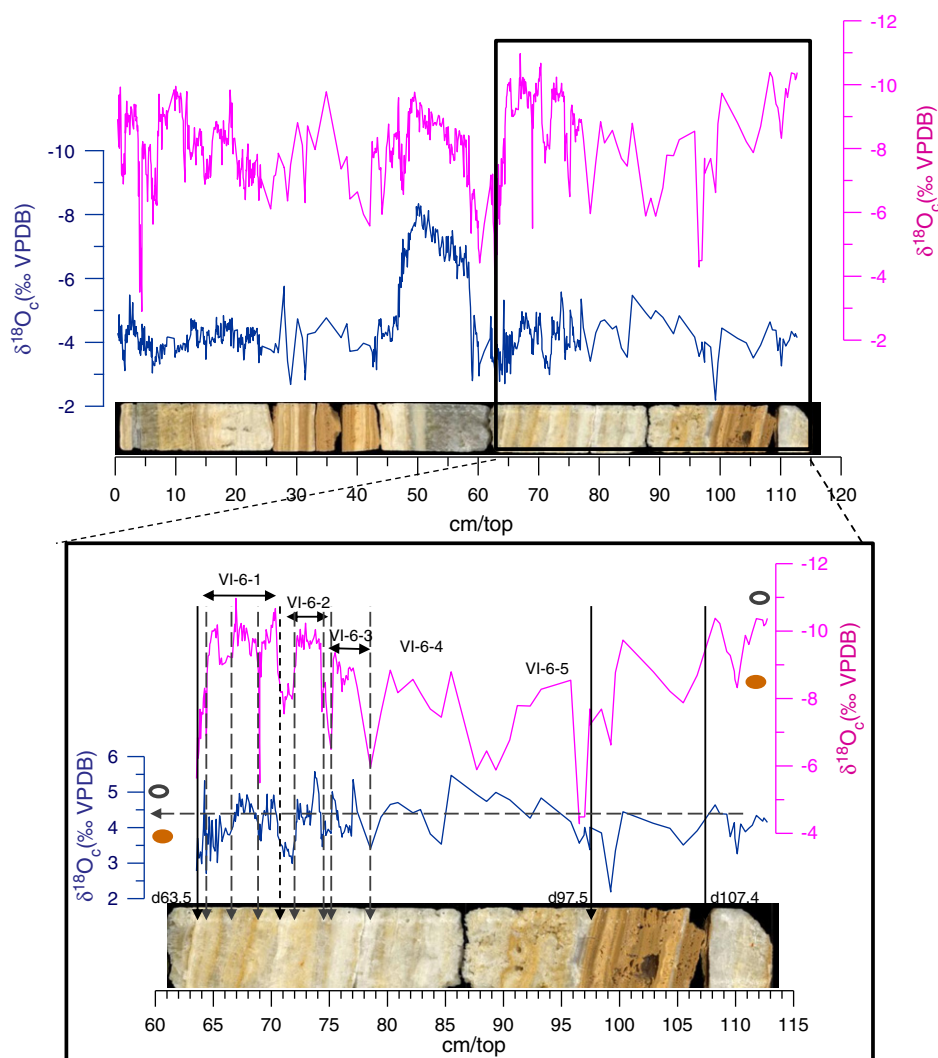


Fig. 2. Polished section of core Vil-car-1 and $\delta^{18}\text{O}_c$ – $\delta^{13}\text{C}_c$ profiles versus depth from the top of the core: whole core at the top, part of interest for this study at the bottom. Black arrows indicate the major discontinuities identified by the following code: d + depth. For a better understanding, we named the $\delta^{13}\text{C}_c$ depleted events as indicated on the bottom figure: VI (for Villars Interstadial)-6 (for MIS6) + number from the youngest. Brown ellipses indicate the composition of detrital carbonate found in modern cave clay, the empty grey represent modern calcite composition. Dotted arrows link isotopic composition main changes to core petrography. Horizontal arrow indicates that the average $\delta^{18}\text{O}_c$ composition of the calcite from 96 to 90 cm/top is distinct from carbonate clay composition.

Table 1

U–Th results (Wainer et al., 2011). All quoted uncertainties are at 95% confidence level (2σ). Dates used in the age model are highlighted in grey.

Sample	Location (cm/top)	$\pm$ (cm)	^{238}U (ng/g)	$\pm$	$\delta^{234}\text{U}$ (‰)	$\pm$	$(^{230}\text{Th}/^{238}\text{U})$ (activity ratio)	$\pm$	$(^{230}\text{Th}/^{232}\text{Th})$ (activity ratio)	$\pm$	Uncorr. age (ka)	$\pm$	Corr. age (ka)	$\pm$	Err age (%)	Laboratory
Lsup	63.50	0.40	38.82	0.14	10.88	10.11	0.7748	0.0143	34.0	0.6	158.050	7.786	155.8	7.8	5.0	GEOTOP
U	63.85	0.30	37.67	0.02	5.71	1.57	0.7817	0.0049	59.0	0.4	163.549	2.467	162.2	2.6	1.6	GSI-mai06
V	64.45	0.25	35.82	0.04	7.53	2.36	0.7897	0.0069	69.3	0.6	166.616	3.587	165.5	3.6	2.2	GSI-mai06
AK	70.00	0.30	58.37	0.36	8.42	6.68	0.8008	0.0083	64.8	0.7	171.855	5.457	170.6	5.5	3.2	GEOTOP
M	76.74	1.05	37.63	0.18	3.35	9.16	0.7935	0.0105	106.3	1.5	170.589	7.090	169.8	7.1	4.2	GEOTOP
AL	80.00	0.30	47.88	0.30	3.32	5.01	0.8120	0.0091	72.7	0.8	180.626	5.892	179.5	5.9	3.3	GEOTOP
B	96.95	0.50	32.00	0.02	9.03	2.20	0.8124	0.0052	130.2	1.3	177.692	3.060	177.1	3.1	1.7	LSCE
P	109.95	0.65	45.67	0.16	6.58	9.00	0.8110	0.0070	217.8	2.0	178.247	6.202	177.9	6.2	3.5	GEOTOP

$\delta^{234}\text{U} = ((^{234}\text{U}/^{238}\text{U})_{\text{activity ratio}} - 1) * 1000$. Decay constant used are $9.1577 * 10^{-6} \text{ yr}^{-1}$ for ^{230}Th , $2.826 * 10^{-6} \text{ yr}^{-1}$ for ^{234}U (Cheng et al., 2000) and $1.55125 * 10^{-10} \text{ yr}^{-1}$ for ^{238}U (Begemann et al., 2001). All ages are corrected for detrital ^{230}Th using $(^{238}\text{U}/^{232}\text{Th})_{\text{activity ratio}} = 0.9 \pm 0.45$ measured on clay of Villars cave (Wainer et al., 2009). After applying it, the ages were younger by a few centuries to 2.3 ka. This correction was done thanks to ISOPLAT3, kindly provided by Kenneth Ludwig, Berkley Geochronological Center (Ludwig, 2003).

more to less porous calcite with columnar fabrics (see also Fig. 2 in Wainer et al., 2011 for detailed description).

Here we focus on the part located below 63.5 cm from the top of the core. Three types of calcite petrography can be distinguished:

- from 114 to 108.7 cm/top, calcite is dark and compact DCC (Genty, 1992; Genty, 1993; Quinif et al., 1994);
- from 97.5 to 85 cm/top, calcite is characterised by vacuoles and gour features;
- from 85 to 63.5 cm/top, calcite is mostly white porous WPC (Genty, 1992; Genty, 1993; Quinif et al., 1994) and laminated; some of the laminae contain high concentration of clay particles.

From 108.7 to 97.5 cm/top, the core contains high concentration of clayey particles with only a few “clean” calcite laminae (Fig. 2). The presence of clay particles might result from flooding episodes of the underground river that was flowing through the lower part of this chamber until recent outside works modified its pathway.

The three major optical discontinuities in this section are used as landmarks hereafter and are named relatively to their distance from the top: d63.5, d97.5 and d107.4 (Fig. 2). They all correspond to a major change in the petrography. The two latter bracket the clay rich interval. The d63.5 discontinuity corresponds to the end of the laminated porous calcite deposition and the beginning of the DCC deposition. Microscope observations show that a new generation of calcite crystals started growing after d63.5. This specific feature is not observed for the other clay layers from 71 to 63.5 cm/top (Wainer, 2009). The observation of a new generation of calcite crystals surrounding d63.5 together with the growth gap as inferred from the results of U–Th dating and the peak in Mg concentration, indicates a severe growth hiatus, likely followed by a flush episode (Wainer et al., 2011).

3.2. Methods

Calcite oxygen and carbon isotopic composition ($\delta^{18}\text{O}_\text{c}$ and $\delta^{13}\text{C}_\text{c}$) were determined on 190 sub-samples on the bottom part of the core, 114–63.5 cm/top. They were drilled along the growth axis using a 0.7 mm diameter dental drill. The sampling resolution ranges from 0.15 to 2.6 cm where gour features (rimstone) are numerous enough to impede a higher resolution. Calcite $\delta^{18}\text{O}_\text{c}$ and $\delta^{13}\text{C}_\text{c}$ composition was analysed on a VG OPTIMA mass spectrometer after orthophosphoric acid reaction at 90 °C (LSCE) and are reported to V-PDB standard. The analytical error is 0.08‰ for the $\delta^{18}\text{O}_\text{c}$ and 0.05‰ for the $\delta^{13}\text{C}_\text{c}$ (2σ). Hendy test was performed on two laminae at 97.45 and 63.65 cm/top.

The U–Th dating methods used on the lower part of the core are detailed in Wainer et al. (2011). The eight dating analyses were

performed in three different laboratories: The GEOTOP (Montreal, Canada) using a VG sector TIMS ($n = 5$), see details in Marshall et al. (2009); the Geological survey of Israel (GSI) using a Nu Instruments MC-ICP-MS ($n = 2$), see details in Vaks et al. (2006); the LSCE, using a Finnigan MAT-262 TIMS ($n = 1$), see details in Plagnes et al. (2002).

4. Results

4.1. Age model: the earlier penultimate glacial

U–Th data are detailed in Wainer et al. (2011) for the entire core, and in Table 1 for the bottom part. From 114 to 63.5 cm/top, all ages are in stratigraphic order within the error bars. The uranium concentration is low (32 to 58 ng/g), and together with the need to correct for the presence of detrital material (Table 1) result in relatively large age uncertainties (~ 5.2 ka in average). The age model is based on linear interpolation between dated points. Two points were excluded due to their slight offset toward older ages, although they fit with the age model within 2σ uncertainty (Fig. 3 and Supplementary material A). According to the proposed chronology, the part of the core from 114 to 63.5 cm/top grew from 178.1 ± 6.2 ka to 155.8 ± 7.8 ka, i.e. during earlier MIS6. The growth rate of the relatively clean calcite parts ranges

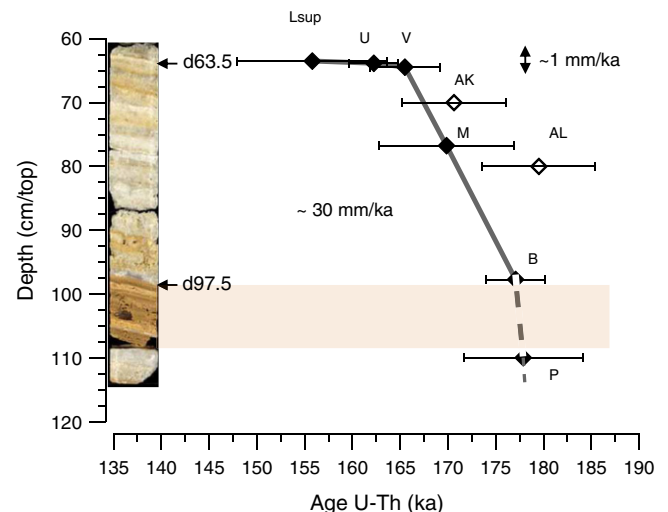


Fig. 3. Growth curve. Empty diamonds are dates excluded from the age model (see Supplementary material A). 2σ error bars are provided.

from 1 mm/ka between 165.5 and 155.8 ka to ~30 mm/ka between 177.1 and 165.5 ka (Figs. 3 and 4).

4.2. Carbon and oxygen isotopic analyses

4.2.1. Isotopic equilibrium

Hendy test and the Replication tests (Hendy, 1971; Schwarcz, 1986; Dorale and Liu, 2009) indicate deposition under equilibrium conditions (Wainer et al., 2011) which is in contrast to modern day

conditions (Genty, 2008). Combined inclusion and clumped isotopes analyses in the section between 65 and 50 cm/top on the same core also show that slight kinetic effects occurred during the 165 to 125 ka time interval (Daëron et al., 2011; Wainer et al., 2011). Based on these observations, it is likely that although the Hendy and Replication tests indicate equilibrium, this part of the core deposited between ~178 and 156 ka (Table 1, Fig. 2) may have deposited also slightly out of equilibrium. Thus, we focussed on qualitative changes rather than quantitative reconstructions.

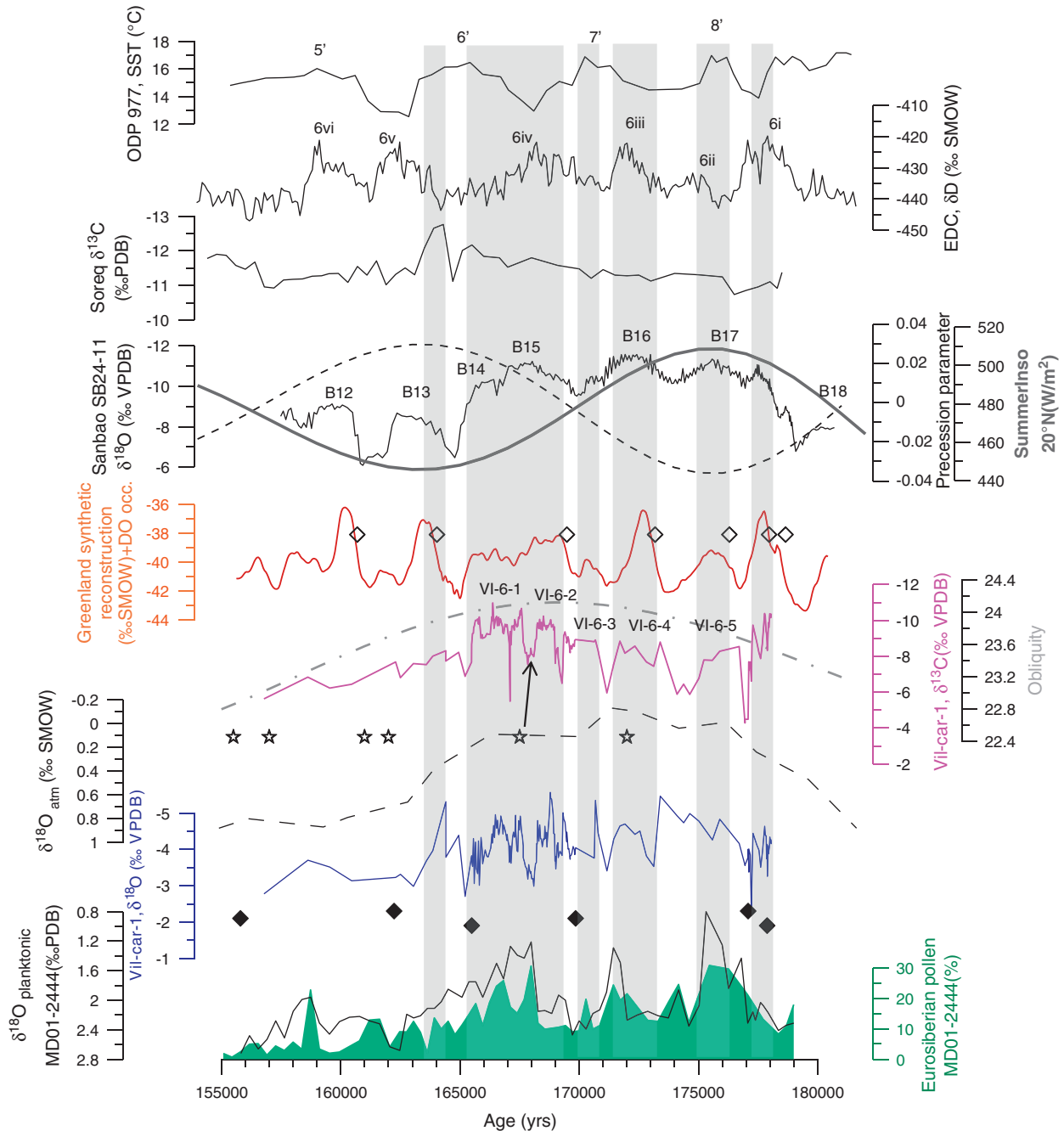


Fig. 4. Comparison between the Vil-car-1 isotopic profiles, $\delta^{18}\text{O}_c$ as a thin blue line, $\delta^{13}\text{C}_c$ as thin pink line, black diamonds stand for Vil-car-1 U–Th dating (this study) to (from bottom to top): Eurosiberian pollen percentages and $\delta^{18}\text{O}$ composition of planktonic foraminifera from MD01-2444 marine core (off Lisbon (Margari et al., 2010)), O_2 atmospheric $\delta^{18}\text{O}$ represented as a thin black dashed line (Jouzel et al., 1993); stars represent IRD deposition on the Iberian Margin (40°N, MD95-2040 (de Abreu et al., 2003)); synthesised Greenland temperature proxies as a brown curve; empty diamonds stand for the predicted Dansgaard-Oeschger (DO) events, both plotted against the sepeleoge chronology (Barker et al., 2011), stalagmite SB24-11 Sanbao cave, China $\delta^{18}\text{O}_c$ profile (Wang et al., 2008), Soreq cave $\delta^{13}\text{C}_c$ record, East Mediterranean (Ayalon et al., 2002), δD record from EDC core, Antarctica (Jouzel et al., 2007), SST from the alkenones U_7^* index from the ODP 977 marine core, Alboran sea (Martrat et al., 2004). Orbital parameters including low latitude 20°N summer insolation data as a thick grey line are from (Laskar et al., 2004), superimposed to Villars $\delta^{13}\text{C}_c$ pink curve, thick dashed line is for the obliquity, thin dashed line is for precession parameter are presented superimposed to the Sanbao record.

4.2.2. Earlier penultimate glacial carbon isotopic profile

Carbon isotopic values ranges from -4.3 to -11% V-PDB (Fig. 2) with frequent oscillations. Increasing trend is observed from the bottom of the core (-10.4%) to 96.55 cm/top (-4.5%) and decrease to -11% at 66.95 cm/top with rapid increase to -5.6% before the d63.5 discontinuity. Superimposed is a higher frequency trend: large negative peaks with steep flanks ranging from 2.3 to 4‰ in magnitude are visible from d98.3 to d63.5. Second-order oscillations (amplitude $\sim 1\%$) within these broad negative peaks are also observed.

4.2.3. Earlier penultimate glacial oxygen isotopic profile

Oxygen isotopic composition varies from -2.1 to -5.6% , oscillating mainly around -4% (Fig. 2). Most of the $\delta^{18}\text{O}_c$ oscillations follow the $\delta^{13}\text{C}_c$ high frequency with a magnitude ~ 1 – 2% . This is less prominent in the section between 95 and 85 cm/top, and may be attributed to the lower sampling resolution in this section.

5. Discussion

U–Th ages obtained on this part of the core suggest that the time resolution is among the highest obtained for a continental record during this period: 34 years in average from 169.9 to 165.5 ka and before 177.6 ka. It is coarser (250 years in average) elsewhere due to spatial sampling resolution (see Section 3.2). It is thus one of the rare records that display high frequency climate variability on land during MIS6 over Western Europe.

5.1. Growth period and growth rate during earlier MIS6

The age model shows that the speleothem grew from ~ 178.1 ka to ~ 155.8 ka. The growth ceased between 155.8 and 148.3 ka, most probably due to colder and drier climatic conditions associated with the MIS6 glacial maximum (Wainer et al., 2011).

The relatively high growth rates of this Villars flowstone (up to 30 mm/ka, Fig. 3) from 177.8 to 165 ka (Fig. 3) are suggestive of favourable conditions for speleothem growth, i.e., water supply, bicarbonate and calcium ions concentration. This latter implies that biogenic CO_2 production in the soil was relatively high due to relatively dense vegetation and bio-pedological activity above the cave. Thus, mean annual temperature was probably higher during this part of MIS6 glacial period in the area of Villars cave. This is supported by speleothem deposition in other parts of Europe, some of which also yielding increased growth rates at the same time, such as Argentarola Stalagmite I, Italy between ~ 172 and 165 ka (Bard et al., 2002a).

5.2. Characterisation of climate instabilities during earlier MIS6 as recorded in Vil-car-1

5.2.1. Interpretation of the $\delta^{13}\text{C}_c$ signal

Previous studies have demonstrated that Villars speleothem $\delta^{13}\text{C}_c$ composition mainly reflects changes in vegetation and bio-pedological activity density above the cave, through the biogenic CO_2 production in the soil up to its saturation, and thus is a robust proxy for millennial to orbital climate variations in the region (e.g. Genty et al., 2001; Genty et al., 2003). The denser the vegetation, the higher $p\text{CO}_2$ in the soil, the more depleted the soil and the speleothem $\delta^{13}\text{C}_c$ values; whereas during climate deteriorations, reduced bio-pedological activity associated with an enhanced proportion of more enriched atmospheric carbon in the soil, results in higher speleothems $\delta^{13}\text{C}_c$ values. In addition to the soil biogenic CO_2 production, there are two main additional effects that can modify the $\delta^{13}\text{C}_c$ signal: (1) kinetic effects which occur more likely during dry (usually associated with cold in W Europe) periods, through an enhanced evaporation, and tend to increase the $\delta^{13}\text{C}_c$ composition; and (2) external contamination by clayey detrital particles (containing detrital carbonate with distinct $\delta^{18}\text{O}_c$ and $\delta^{13}\text{C}_c$ compare to calcite) input. As evoked in paragraph 4.2.1, kinetic effects are minor here. To verify if the presence of detrital component could

have influenced the $\delta^{13}\text{C}_c$ values, we analysed the isotopic composition of the carbonated fraction in Villars cave modern clay, and investigated the potential link between the isotopic curve morphology and the calcite clay-rich layers (Fig. 2). The carbonate-clay shows an isotopic composition close to the one of the Vil-car-1 calcite: $\delta^{18}\text{O}_{\text{clay}}$ is -3.9% and $\delta^{13}\text{C}_{\text{clay}}$ is -8.5% V-PDB (Fig. 2). On the portions of the core (i.e. 74–65 cm/top and below 108 cm/top) where the isotopic signal is the lowest, $\delta^{13}\text{C}_c$ reaches -9 to -10% and -4 to -5% for the $\delta^{18}\text{O}_c$; it is thus clear that this signal reflects solely the isotopic composition of the speleothems calcite.

On the basis of the visible clay content on the polished section of the Vil-car-1 core, clay contamination may be possible at two single locations: 111 and 69 cm/top (Fig. 2). Trace elements were analysed between 63 and 73 cm/top and yield an increase in Al/Ca ratio at 69 cm (Wainer et al., 2011). Given the significant contribution of Al in clay chemical composition, this is supporting the hypothesis of a possible clay contamination at that point. The negative $\delta^{13}\text{C}_c$ event from 96 to 91 cm/top coincides with an increase in clay content too, but the $\delta^{18}\text{O}_c$ values of this part of the core suggest that the $\delta^{13}\text{C}_c$ signal is mainly controlled by the speleothem calcite; thus we suggest that the $\delta^{13}\text{C}_c$ oscillations are climate controlled.

Interestingly, similar $\delta^{13}\text{C}_c$ values are observed for MISS6e, the last and the present interglacials and the MIS3 interval. This translates that the biogenic CO_2 production was such that the $p\text{CO}_2$ has reached a threshold of saturation of the soil voids/pores. Any extra biogenic CO_2 produced above this threshold probably escapes to the atmosphere. Thus, the CO_2 production might have been more important but this was not recorded.

The time interval between 177 and 165 ka (97.5–63.5 cm/top) contains five low $\delta^{13}\text{C}_c$ events (large negative peaks with steep flanks (Figs. 2 and 4)), interpreted as episodes of warmer and more humid climate conditions. By analogy to rapid climatic events during the last glacial period, the Vil-car-1 major low $\delta^{13}\text{C}_c$ events were labelled from the youngest to the oldest as “VI-6-X” for “Villars Interstadial within MIS6 number X”. According to the age model these events lasted ~ 2.2 , 0.8, 1.5, 2.6 and 1.7 ka respectively, although the accurate durations are not certain due to the errors in the age model. Given the extreme steepness of these peak flanks, we assume that changes between stadials and interstadials were abrupt.

Intra-event variability is well visible during the youngest and the longest event: three second order peaks $\sim 1\%$ in amplitude and lasting ~ 400 years occur within the plateau of the VI-6-1 event. This centennial scale $\delta^{13}\text{C}_c$ variability within VI-6-1 is also recorded in the $\delta^{18}\text{O}_c$ profile. Events VI-6-3 and VI-6-4 also display such oscillations but they are fewer, one during each event.

5.2.2. Interpretation of the $\delta^{18}\text{O}_c$ signal

Numerous factors influence speleothem calcite oxygen composition, i.e., source, air mass trajectory, cave air temperature, rainfall amount, season of infiltration and therefore drip water oxygen isotopic composition ($\delta^{18}\text{O}_{\text{dw}}$).

In most of previously studied Villars speleothems, $\delta^{18}\text{O}_c$ variations range within 1% and rarely display long term trends, making this signal difficult to interpret. The largest $\delta^{18}\text{O}_c$ shift studied in Villars Cave has been documented in Vil-car-1 core: a depletion of -5% from late MIS6 to the last interglacial optimum – Termination II (TII) (Wainer et al., 2011). A study associating clumped isotope, fluid inclusions and calcite isotopic composition suggests that this shift is mainly due to temperature increase and, to a lesser extent to variations of kinetic effects amount, $\delta^{18}\text{O}_{\text{dw}}$ (estimated from fluid inclusions measurements) remaining surprisingly constant through TII (Daëron et al., 2011; Wainer et al., 2011). During MIS6 the $\delta^{18}\text{O}_c$ range is of $\sim 3\%$. The reason for such large $\delta^{18}\text{O}_c$ variations during this glaciation is not clear yet and this needs to be further studied.

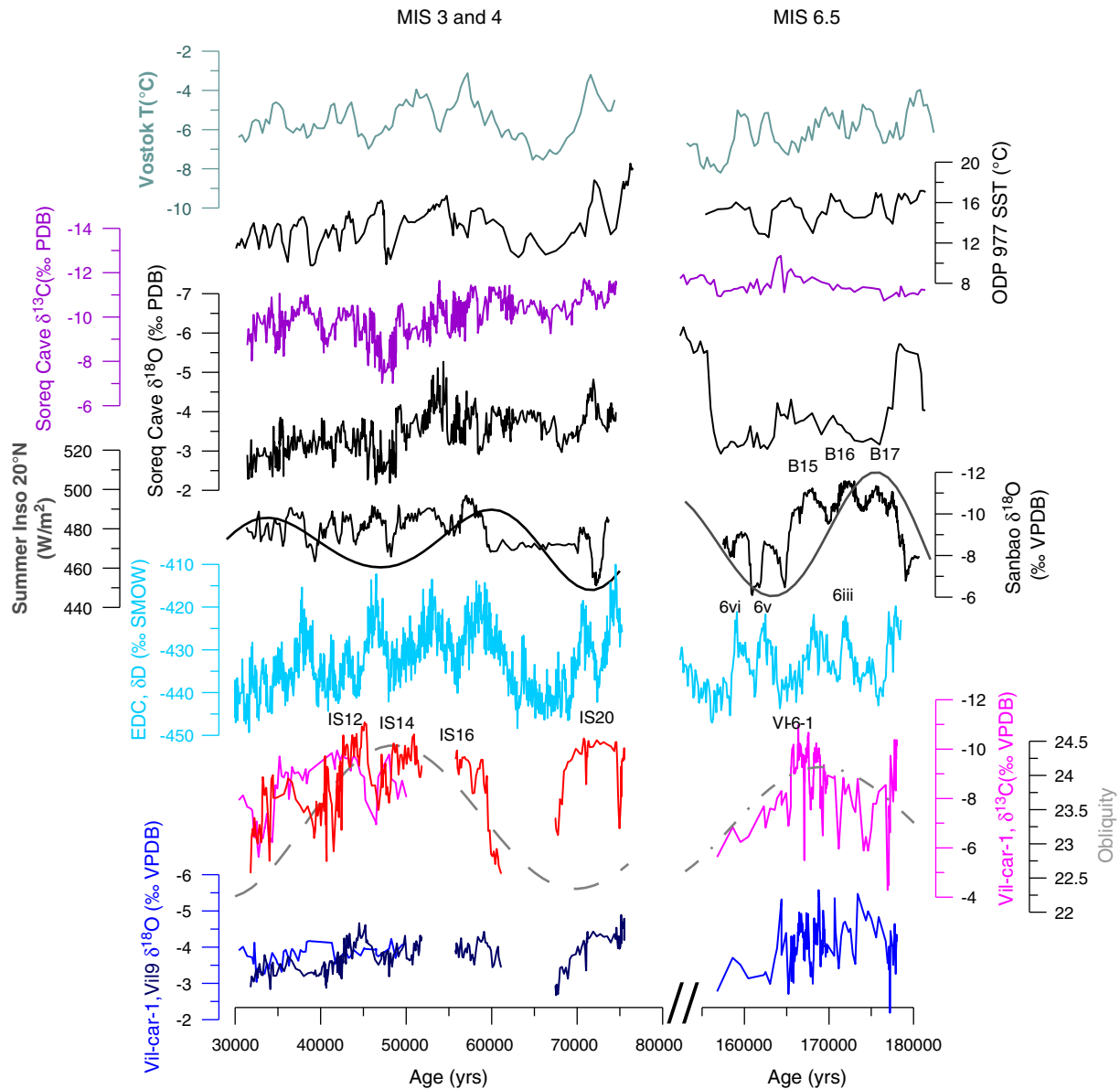


Fig. 5. Comparison of the climatic variability of early MIS6 and MIS3–4 in Villars cave (blue and pink curves for Vil-Car-1 profile, thin red and dark blue for Vil 9 profiles (Genty et al., 2003)) and in other climate archives: δD record from EDC core Antarctica in thin blue (Jouzel et al., 2007), stalagmite SB24-11 Sanbao cave, China $\delta^{18}O_c$ profile (Wang et al., 2008), Soreq cave $\delta^{18}O_c$ (black) and $\delta^{13}C_c$ (violet) record, East Mediterranean (Ayalon et al., 2002), SST from the alkenones U_{37}^K index from the ODP 977 marine core, Alboran sea (Martrat et al., 2004), Antarctic temperature record at Vostok (Cuffey and Vimeux, 2001) as a thick grey-blue curve. Orbital parameters are from Laskar et al. (2004): low latitude $20^\circ N$ summer insolation data as a thin black line superimposed to the Sanbao record, obliquity as a thick grey dashed line superimposed to the $\delta^{13}C_c$ Villars curve.

5.3. Characteristics of the millennial variability during earlier MIS6

5.3.1. Comparison of earlier MIS6 to MIS3 millennial variability and climatic background

We use Vil-car-1 flowstone core, Vil 9 and Vil 14 stalagmites records (Genty et al., 2003; Wainer et al., 2009; Genty et al., 2010) to compare MIS6 high frequency variability to the MIS3. The frequencies of the main instabilities (~ 2.4 ka for MIS6 and ~ 2.2 ka for MIS3), the duration of the interstadials and the $\delta^{13}C_c$ amplitude range of intra-interstadial events seem similar (Fig. 5, e.g. IS12, Vil9, Genty et al., 2003). However, whereas the first order $\delta^{13}C_c$ magnitude is similar during both glaciations, $\delta^{18}O_c$ magnitude is much larger during MIS6: the maximum range in $\delta^{18}O_c$ magnitude during MIS3 is $\sim 0.9\%$ to Greenland Interstadial GIS12 event (which is one of the most pronounced MIS3 interstadial events), whereas the $\delta^{18}O_c$ amplitude is ~ 1.5 – 2% during MIS6 (Fig. 5). Moreover, $\delta^{18}O_c$ values during MIS6

are, on average, more depleted than during MIS3 (to -5.6%). Several hypotheses can be raised in order to explain these differences:

- 1) If paleo-drip water oxygen isotopic composition remained constant as was postulated for Termination II (Wainer et al., 2011), then a larger air temperature increase from stadial to interstadial (IS) periods during MIS6 would have been the main factor explaining the MIS3/MIS6 $\delta^{18}O_c$ discrepancy. This is not consistent with the SST records of Iberian marine cores (MD95-2040 and ODP977, Fig. 1 (de Abreu et al., 2003; Martrat et al., 2004)).
- 2) A change in storm track is rather unlikely since the cave is very close to the Atlantic Ocean ~ 180 km away, to produce a significant continental effect on the rainfall $\delta^{18}O$ composition.
- 3) An increase in winter precipitation or a decrease in summer precipitation during the interstadials would result in more depleted $\delta^{18}O_c$ values. Assuming that the observation by Margari et al.

(2010) that summers were rather more humid during MIS6, is also valuable for SW France, a decrease in summer precipitation can be rejected. An increase in winter precipitation in the region of Villars is unfortunately difficult to check in the area.

- 4) About an increased contrast in annual precipitation between stadial and interstadial phases: modelling experiments suggest that the particularly low $\delta^{18}\text{O}_c$ values from 180 to 170 ka in an Argentarola Cave speleothem reflects enhanced precipitation via the amount effect (Bard et al., 2002b). Pollen data from the Ioannina basin also support that MIS6 was wetter than MIS3 (Roucoux et al., 2011). Finally, in the Hulu/Sanbao Cave speleothems (China), the particularly low $\delta^{18}\text{O}_c$ values associated with earlier MIS6 are attributed to the East Asian monsoon intensification (Wang et al., 2008). However, rainwater $\delta^{18}\text{O}$ composition currently does not show any significant relationship with precipitation in Villars, probably due to the coast proximity (Wainer et al., 2011). Consequently, the low $\delta^{18}\text{O}_c$ values cannot be interpreted in terms of increasing precipitation alone. However, the association with low $\delta^{13}\text{C}_c$ values and relatively high growth rates do suggest that rainfall increase is a major factor on the low $\delta^{18}\text{O}_c$. The $\delta^{18}\text{O}$ values of atmospheric O_2 were anomalously low at global scale during earlier MIS6 (Malaize et al., 1999; Landais et al., 2010). This might have impacted the global rainwater $\delta^{18}\text{O}$ composition but the quantification of this impact would need additional work.

About the general climatic conditions, globally warmer SST during MIS6 compare to MIS3 are observed around Iberia (Figs. 4 and 5) and less ice rafted detritus (IRD) reaching 40°N on the eastern Atlantic suggest that extreme SST were generally not as low as during MIS4 to 2 (icebergs melt closer to their source (de Abreu et al., 2003; Martrat et al., 2004)). This involves a reduced southward influence of polar water masses off Portugal (de Abreu et al., 2003). The $\delta^{13}\text{C}_c$ values of Soreq Cave speleothems during MIS6 are more depleted (~2‰) than during MIS3, implying a globally more humid climate (Ayalon et al., 2002). Despite a much lower magnitude of variation, Soreq record displays similar trends to Vil-car-1 during MIS6 (Figs. 4 and 5). The depleted values of the Soreq Cave $\delta^{13}\text{C}_c$ profile together with the higher proportion of Eurosiberian taxa in Iberia are consistent with depleted $\delta^{13}\text{C}_c$ values in Vil-car-1 core which suggests a dense vegetation cover and soil activity, implying relatively high humidity and mild temperature. The high Vil-car-1 flowstone growth rate also supports this assumption (Fig. 3). Finally, the Vil-car-1 decrease in $\delta^{18}\text{O}_c$ signal during this period can be interpreted similarly to reflect higher temperature and/or higher humidity. Given that Soreq Cave and Argentarola cave records (Ayalon et al., 2002; Bard et al., 2002b) also suggest coeval increase in humidity, it seems likely that SW France was also more humid during MIS6 compared with MIS3. Ioannina pollen sequence suggests a MIS6 globally cooler MIS3 (Roucoux et al., 2011). It is delicate to deconvolute the respective contribution of the humidity increase and the temperature increase, in the $\delta^{13}\text{C}_c$ decrease in the Villars area. These two parameters usually change accordingly. Today in Villars cave, the $\delta^{13}\text{C}_c$ in the atmosphere of the cave varies by ~3‰, being more depleted during summer and autumn, and relatively enriched during the winter, following the development of the vegetation above the cave (Genty, 2008). Precipitation being rather constant throughout the year, day light duration and temperature should be the triggering factors for biogenic CO_2 production. The Villars MIS6 interstadial events are thus likely intervals with both increased temperature and humidity. But we cannot answer the question: was MIS6 warmer and/or wetter than MIS3 based on this Vil-car-1 record solely. Fluid inclusion composition and clumped isotope analyses will contribute to answer this question in the future.

5.3.2. Number of cycles and characteristics of climate deteriorations associated with northern Atlantic cold pulses

From ~170.6 ka to ~165.5 ka, the Vil-car-1 record is among the best time-resolved records together with the Chinese Sanbao cave

for mid-to-low latitude area (Fig. 4). Thus, the occurrence of three stadial–interstadial cycles within this time interval can be robustly established. The youngest interstadial of this period, VI-6-1 event (167.5–165.6 ± 3.6 ka), is the most pronounced, the longest and with most internal variability. Abrupt $\delta^{13}\text{C}_c$ increases are recorded at 167.3 and 166.3 ka suggesting climate deteriorations (Fig. 4). Although resolution is lower from ~177–167.5 ka, it is still possible to observe at least two additional rapid climatic cycles with a lower magnitude. The 178–165 ka period covered by our entire record is believed to be the most humid period during MIS6 in China (Wang et al., 2008); it is also the time of sapropel S6 formation due to the intensification of the African monsoon (~172 ka (Lourens, 2004)). From 178 to 155 ka, five oscillations are recorded in the Alboran sea and the SW Iberian margin cores (MD01-2444) yields three main interstadials (Martrat et al., 2004; Margari et al., 2010), whereas five of them are detected in the Vil-car-1 flowstone core (Fig. 4). The difference in resolution likely accounts for most of this discrepancy. Note that during MIS3, few records reproduce the exact number of the stadial/interstadial oscillations recorded in Greenland and their respective magnitude. The Greenland synthetic record predicts seven main Dansgaard–Oeschger-like events, most of which can be correlated with our record (Fig. 4 (Barker et al., 2011)). Interestingly the longest event of this synthetic record (169.4–165.3 ka) can be divided in two steps which are observed in Villars as events VI-6-1 and VI-6-2 respectively. Similarly, correlations of the VI-6 events can be made with the EDC ice core (Fig. 4 (Jouzel et al., 2007)) and Chinese speleothem records (Fig. 4 (Wang et al., 2008)). Thus, VI-6-1 and VI-6-2 events can be correlated with B14–B15 from the Chinese speleothem record and considered as being the counterpart of 6iv EDC ice event; the event, VI-6-4 is correlative to B16 and counterpart of 6iii, and VI-6-5 to B17 and 6ii. The VI-6-3 event is probably correlative to the little IS shown in both the EDC and Greenland synthetic records. The ice event 6i may be correlative to the partly truncated event which shows depleted isotopic values at the base of this Vil-car-1 record. These synchronicities imply that global millennial MIS6e variability strongly impacted the mid-latitude continental environment.

Both planktonic $\delta^{18}\text{O}$ profile and Eurosiberian pollen abundances in the Atlantic MD01-2444 record a long interstadial beginning shortly before 168 ka, which is consistent with the features of the VI-6-1 interstadial (Fig. 4 (Margari et al., 2010)). Based on time-correlation, the marine interstadial event centred at 172 ka would correspond to VI-6-4, and the event started at 177 ka to VI-6-5. Based on this marine record, European climate was likely the most humid and the warmest during the oldest interstadial (i.e. 177–175 ka), which is not the case on the Vil-car-1 record. Regional discrepancies might be the cause as it was already shown between South Iberia and Bay of Biscay records during MIS3 (Sánchez Goñi et al., 2008).

Margari et al. (2010) suggested that stadial phases are longer during MIS6 than during MIS3 on the Iberian Margin. It is also the case for Soreq Cave speleothem record (Ayalon et al., 2002). Given the uncertainties associated to our chronology, we cannot confirm that this is the case in Villars. The MIS6 larger magnitude of variation, especially in the $\delta^{18}\text{O}_c$ composition of Vil-car-1 makes them easier to distinguish from the climatic background compare to MIS3. This is consistent with the recent study on the MD01-2444 core (Margari et al., 2010) but contradictory with the observation by De Abreu et al. (2003). These latter evoked poorly differentiated cooling phases in MD95-2040 core record (40°N, Fig. 1), due to the slightly lower sea level of MIS6 (which is still a matter of debate (e.g. Margari et al., 2010)) contributing to the higher stability of ice sheets. However, the ‘poorly differentiated’ configuration holds true particularly for late MIS6, after 164 ka. The 167.5 ka IRD pulse event observed in this marine core seems to have been well imprinted in the Villars record by the sharp increase in both $\delta^{13}\text{C}_c$ and $\delta^{18}\text{O}_c$ (> 1‰ Fig. 4). From 167 to 163 ka in this MD95-2040 marine core, several reduced

cooling event are detected (summer SST and cold fauna concentration). We propose that they are related to the intra-interstadial variability observed in VI-6-1 interstadial in the Vil-car-1 record. From 164 ka, the frequency and the magnitude of cold pulses in the Northern Atlantic increase, leading to global climate degradation in eastern Atlantic borderland towards full glacial conditions. This also appears in the pollen record in the MD01-24444 core record further South along the Iberian margin, in the decreasing growth rate and progressive increase of $\delta^{13}\text{C}_c$ composition in the Vil-Car-1 sample (Figs. 3 and 4). The Vil-car-1's glacial growth stop is coeval to a densification of *Neoglobobulimina pachyderma* (sinistral) concentration peak occurrences found in core MD95-2040 (de Abreu et al., 2003).

Given these observations together with the similar amplitude of $\delta^{13}\text{C}_c$ changes displayed in the Villars cave speleothems during MIS5e and MIS3, both humidity and air temperature were likely widely affected during the stadial phases insofar as the vegetation and soil activity were significantly impacted in the SW France.

6. Conclusion

Vil-car-1 flowstone core provides a discontinuous record of the last 178 ka, including exceptionally high resolution (up to 34 years) terrestrial stable isotope profiles of the earlier penultimate glacial climatic instabilities from ~177 to 156 ka. Despite large age uncertainties (average 2σ is 5.2 ka, for comparison EDC3 ice core chronology's uncertainty ~6 ka over this time interval (Parrenin et al., 2007)), correlation of events can be found between Villars speleothem and the Greenland synthetic record on the Chinese speleothem time scale (Barker et al., 2011). The main features of the earlier MIS6 Vil-car-1 record show that:

- climatic conditions were rather humid and mild as inferred from depleted $\delta^{13}\text{C}_c$ values and high growth rate; this is consistent with the finding from the Iberian core pollen studies as well as Mediterranean cave speleothem; climate was rather similar to MIS3
- climatic instabilities, Dansgaard-Oeschger-cycle like, appear in the record: five clear climate improvement associated to $\delta^{13}\text{C}_c$ and $\delta^{18}\text{O}_c$ depletion attest for soil activity and vegetation cover densification above the cave.

The best resolved records over this time period i.e. the Vil-car-1 (this study) and the Sanbao cave speleothem's, the MD01-2444 marine, and the Greenland synthesised record tend to converge to a number of 4–6 cycles between 180 and 162 ka. A common remarkable features emerging from the comparison of all these 4 records, is that the interstadial centred on 167 ka is particularly well imprinted, lasted several thousand years and contains an internal variability. Correlation between records is hampered by low resolution in some records and by the ambiguity associated with intra-interstadial variability. The “less dramatic” nature of Atlantic cooling events might also contribute to make it more difficult to distinguish each event. Thus, additional high quality and high resolution records are still needed to robustly establish a consensus regarding the number of climate events during earlier penultimate glacial.

The five cycles found in the Vil-car-1 record, are superimposed to a climate improving trend, consistent with the obliquity curve as already observed for Villars records (Genty et al., 2010), the youngest interstadial (~167.5 to ~165.6 ka) being the longest and the most pronounced. From this optimum at ~166 ka, $\delta^{13}\text{C}_c$ composition increased and growth rate decreased until the complete growth cessation, coevally to iceberg discharges intensification within the climate deterioration trend towards full glacial conditions settlement. This climate degradation pattern through these two speleothem-related thresholds is identical to the ‘MIS3 to MIS2’ pattern observed in Villars Cave, possibly linked with the obliquity signal as explained in (Wainer, 2009; Genty et al., 2010).

Despite the more subdued signature in the Atlantic sediments of the Northern Atlantic cooling phases during MIS6 compared to MIS3 (McManus et al., 1999; de Abreu et al., 2003), the MIS6 ones are associated to a similar or slightly lower $\delta^{13}\text{C}_c$ increase in the flowstone record, suggesting that earlier MIS6 climate degradations were sufficiently drastic to impact continental humidity and temperature conditions. This study provides new insights on the number and magnitude of rapid climatic variability during earlier MIS6 already highlighted at global scale, which impacted SW European continental area.

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Appendix A. Supplementary data

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