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Speleothems and paleoglaciers

Christoph Spötl [a,*], Augusto Mangini [b]

[a] Institut für Geologie und Paläontologie, Leopold-Franzens-Universität Innsbruck, Innrain 52, 6020 Innsbruck, Austria
[b] Forschungsstelle Radiometrie, Heidelberger Akademie der Wissenschaften, Im Neuenheimer Feld 229, 69120 Heidelberg, Germany

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Abstract

Ice and speleothems are widely regarded as mutually exclusive as the presence of liquid water is a fundamental prerequisite for speleothem deposition. Here we show that speleothems may form in caves overlain by a glacier, as long as the temperature in the cave is above freezing and the conduits are not completely flooded by melt water. Carbonate dissolution is accomplished via sulfide oxidation and the resultant speleothems show high $\delta^{13}C$ values approaching and locally exceeding those of the parent host rock (lack of soil-derived biogenic C). The $\delta^{18}O$ values reflect the isotopic composition of the melt water percolating into the karst fissure network and carry an atmospheric (temperature) signal, which is distinctly lower than those of speleothems formed during periods when soil and vegetation were present above the cave. These 'subglacial' speleothems provide a means of identifying and dating the former presence of warm-based paleoglaciers and allow us to place some constraints on paleotemperature changes.

Keywords: speleotherms; glacier; isotopes; U/Th dating; paleoclimate

1. Introduction

During the past decade speleothems have been firmly established as accurate and precise chronometers of terrestrial climate change on time scales ranging from seasonal to orbital. One advantage of this archive is its wide geographical distribution, from the tropics to the Arctic Circle and from coastal to high-mountain regions. The minimum requirements for (semi)continuous deposition of stalagmites and flowstones – the two types of cave carbonates most commonly utilized in paleoclimate studies – are the presence of water (and hence temperatures above freezing), dissolved calcium and bicarbonate, and the thermodynamic drive of these waters to degas carbon dioxide upon entry into the cave. Recent studies have demonstrated the great potential of speleothems as paleo-wetness indicators in regions influenced by drought and/or monsoonal precipitation, e.g., [1–4]. Speleothem deposition in high-latitude and high-altitude caves likewise responds like an on/off switch, but here the controlling parameter is temperature. Once the interior temperature drops below freezing, speleothem deposition is halted and – depending on the hydrological regime – perennial ice will accumulate. Speleothems such as stalagmites, flowstones and stalactites only form when the cave's interior temperature is permanently above the freezing point of water (volumetrically insignificant, so-called cryogenic carbonates, however, are known to form in some ice caves [5,6]). Speleothem growth identified by

* Corresponding author. Tel.: +43 512 507 5593; fax: +43 512 507 2914.
E-mail address: christoph.spoetl@uibk.ac.at (C. Spötl).

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U-series dating can thus be used as a robust and quantitative paleotemperature indicator in cold climate regions. e.g., [7–9].

A consequence of climate cooling in high-latitude and mountainous regions is the expansion of glaciers and ice sheets (or permafrost in arid regions) due to a lowering of the equilibrium line altitude (ELA). Little is known how caves and their hydrology and interior climate are affected by overriding ice masses. Studies from Norway [10,11] suggest that during the Pleistocene caves overlain by ice were flooded by melt water resulting in phreatic conditions and the local deposition of siliciclastic sediments (as opposed to speleothems). On the other hand, speleothem deposition currently occurs in Castleguard Cave, Alberta. The interior of this cave lies beneath the Columbia Icefield and moderate carbonate (and gypsum) is forming today [12,13].

During the past seven years we have been working on cold-climate cave systems in the Eastern Alps, which show parallels to both the Norwegian sites and Castleguard Cave. Based on U-series dates and stable isotope data from a number of speleothem samples we conclude that speleothems in at least some of these alpine caves were also formed during times when the caves were covered by ice. As a corollary, we propose that speleothems formed in caves beneath glaciers can be used to identify and date the presence of warm-based paleoglaciers.

2. Physiographic setting

The study area is located in the Eastern Alps 30 km SSE of Innsbruck (Austria), where some 30 caves are cut in calcite marbles above the modern timberline. The largest one, Spannagel Cave, has a total length of ca. 10 km and is located between 2195 and 2524 m a.s.l. (for details and maps see [14–16]). At 2500 m the mean annual air temperature is about 0 °C with temperatures in summer (April–September) averaging ca. 7 °C and in winter (October–March) ca. −4 °C. The air temperature inside the cave is barely above the freezing point (+1.4 to +2.5 °C) allowing water–rock interactions and (slow) formation of speleothems. Today, the area above Spannagel Cave (and all the other caves in this region) ranges from a sparsely vegetated glacier fore field to alpine meadows. As recent as during the "Little Ice Age" about two thirds of the area above this cave were covered by the Hintertux Glacier, whose sharp-crested lateral moraine ridges clearly mark the prominent advance during 1850–1860 (plus a smaller one during the 1920s). As the

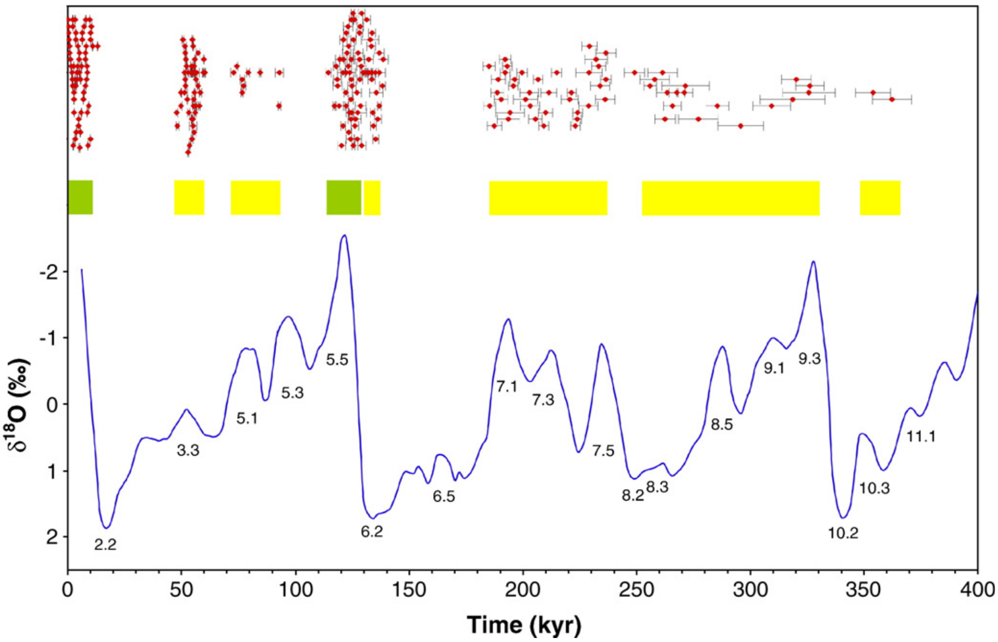

Fig. 1. Compilation of 251 TIMS U–Th age determinations of U-rich speleothems from Spannagel and adjacent caves (shown with 2-sigma error bars; for analytical details see [20,21]. The stacked benthic O isotope record of [22] (only some substages are labelled for sake of clarity) serves as reference curve for global glacial–interglacial climate variability. Note that speleothem deposition was not restricted to peak interglacial conditions at this high-alpine site (green bars) and U–Th data in conjunction with C and O isotope data suggest that deposition beneath a warm-based glacier was common during the past ca. 350 kyr (yellow bars). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

advances during the "Little Ice Age" were the most extensive ones in the entire Holocene [17,18] this observation is central for the interpretation of pre-Holocene glaciers and their concomitant ELA depressions. Mapping of older moraine features indicates that the entire cave system was in a subglacial position during the Younger Dryas and certainly during the Last Glacial Maximum, when the ice above the cave reached a thickness of up to 250 m [19].

3. Modes of speleothem deposition

Several passages and chambers of Spannagel Cave contain speleothems, and some are active today (based on U–Th dates and hydrochemical studies of drip waters showing supersaturation with respect to calcite). Two fundamentally different processes give rise to speleothem deposition in this cave system which can be discriminated using the stable C and O isotopic composition: (a) partitioning of soil-derived carbon dioxide into the vadose waters giving rise to speleothems with low $\delta^{13}C$ values, and (b), dissolution of marble via sulfide oxidation, resulting in high $\delta^{13}C$ values in speleothems very similar to those of the marble. The first mechanism requires a moderately well developed soil and alpine vegetation above the cave (as prevails today), whereas the latter process only depends on the presence of liquid water in the aquifer and may operate even when the cave is buried beneath a (warm-based) glacier. Oxidation of disseminated sulfides in the gneiss that caps the marble bed is evident also today by elevated sulfate concentrations in seepage waters, and precipitation of gypsum in several passages and its low S isotopic composition [16].

4. Speleothem age distribution

While the topography above Spannagel Cave shows a strong imprint of repeated glacial activity up to the end of the "Little Ice Age" leaving little traces of older sediments, speleothem samples provide a unique record of carbonate deposition in the subsurface going back a few hundred thousands of years. Results from some 250 U–Th measurements by thermal ionization mass spectrometry from ca. 30 samples from various locations in Spannagel and neighboring caves show several interesting features (Fig. 1): (a) a bias toward periods characterized by a high global sea level suggesting warm, interglacial-type climate in the Alps, (b) a suite of samples that formed at times of intermediate sea level, e.g. during Marine Isotope Stage (MIS) 3 and MIS 7.4, and (c) a lack of samples that formed during glacial maxima (with the possible exception of MIS 8).

5. 'Subglacial' speleothems

There are no clear macroscopic differences between stalagmites or flowstones that formed during the

Table 1
Th/U ages of stalagmite SPA 119 and flowstone SPA 4

Lab #	Distance from top (cm)	$\delta^{234}U$		Conc. ^{238}U		Conc. ^{232}Th		Conc. ^{230}Th		Age	
		(‰)	± (‰)	(μg/g)	± (μg/g)	(ng/g)	± (ng/g)	(pg/g)	± (pg/g)	(kyr)	± (kyr)
SPA 119											
2931	0.5	118.5	3.6	7.088	0.0099	20.0176	0.0440	115.57	0.44	220.5	3.7
4113	2.8	53.0	1.5	31.522	0.032	7.736	0.015	477.69	0.96	221.3	2.0
4114	5.5	43.7	1.5	33.319	0.033	11.160	0.022	500.6	1.0	223.0	2.0
4115	10.2	35.0	1.5	35.135	0.035	11.306	0.023	523.2	1.0	224.0	2.1
4116	14.2	34.6	1.3	46.463	0.046	33.030	0.066	691.3	1.4	223.8	2.0
4117	19.4	29.1	1.5	45.870	0.046	6.434	0.021	682.5	2.0	228.8	2.7
2872	21.3	28.8	2.3	49.025	0.0735	75.0148	0.1725	729.29	1.68	228.8	4.2
4118	22.4	30.0	1.6	43.645	0.044	25.556	0.077	648.5	2.1	226.9	2.9
SPA 4											
3948	1.2	10.8	1.6	251.01	0.25	0.3693	0.0021	3799.5	10.3	265.7	3.9
4121	1.9	3.0	1.4	324.33	0.32	<0.05		4883.3	9.8	271.0	3.5
4122	2.4	4.0	1.4	399.22	0.40	<0.05		6001.1	13.2	267.6	3.4
4120	3.0	4.7	1.7	103.90	0.10	<0.05		1557.4	5.8	263.4	4.8
3949	3.8	49.7	2.1	223.51	0.25	<0.05		3691.4	8.1	320.2	6.5
4123	4.7	7.8	1.4	147.79	0.15	<0.05		2320.5	4.9	326.2	6.2
3950	4.9	2.1	1.4	156.96	0.16	<0.05		2482.9	5.0	362.3	8.5
4124	5.3	0.7	1.4	179.59	0.18	<0.05		2827.0	5.7	353.9	7.7

Errors are quoted as 2-sigma standard deviations.

Holocene (or MIS 5.5) and speleothems that formed during MIS 3 or MIS 7.4, i.e. at times when the ELA was most likely lower than during the "Little Ice Age". Speleothems in these high-alpine caves are characterized by inclusion-poor, coarsely crystalline, columnar, low-Mg calcite fabrics. Occasionally, layers are stained gray due to the presence of fine detritus disseminated in the calcite. Micritic or dendritic fabrics, common in many cave sites (e.g., [23]) are absent. In addition, all studied samples exhibit slow growth rates and – particularly in the case of flowstones – common growth interruptions (hiatus).

The most important difference between fully interglacial speleothems and presumably 'subglacial' ones is their stable C isotopic composition: the former show low $\delta^{13}C$ values ranging from ca. −3 to ca. −9‰, while

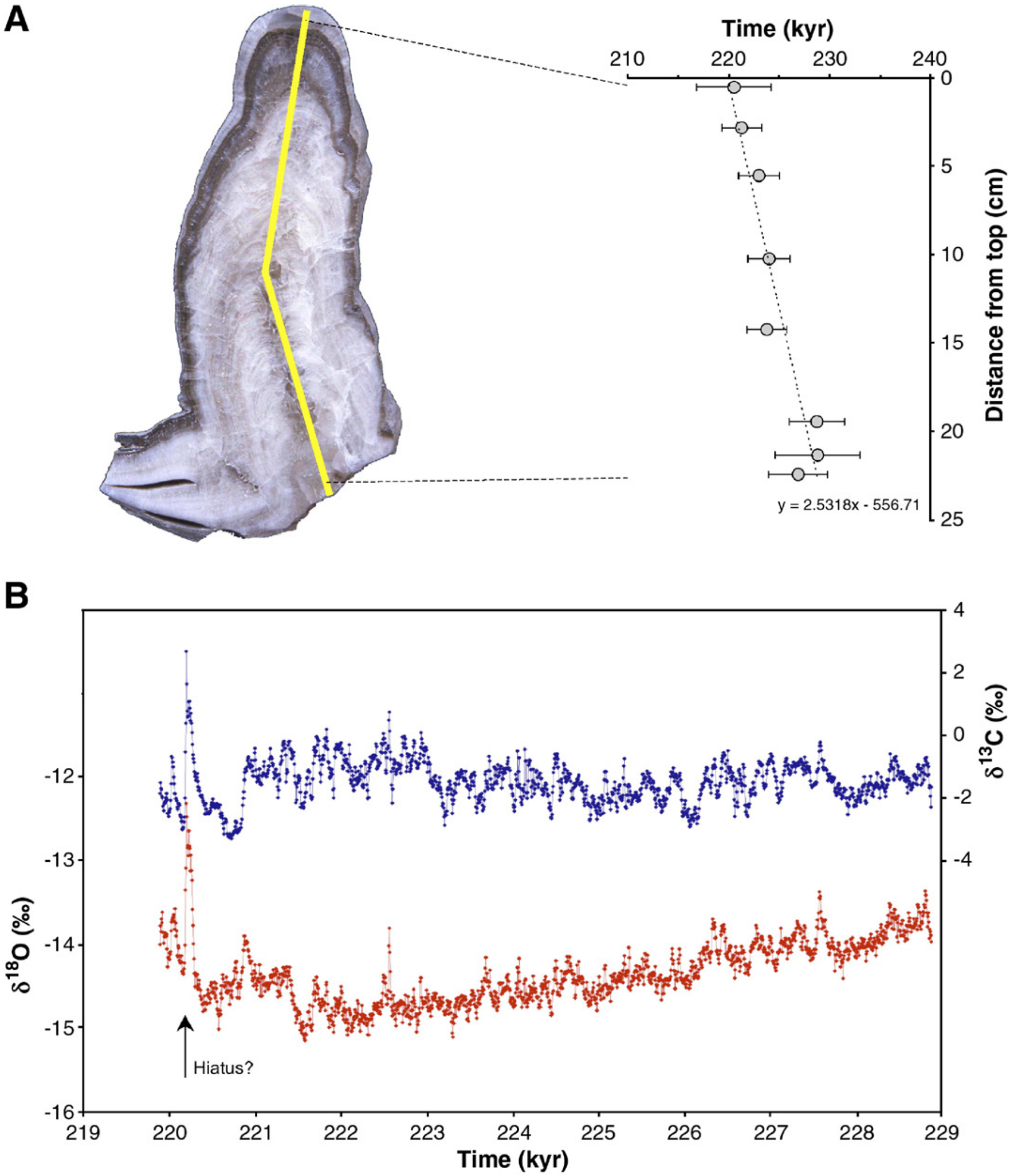

Fig. 2. Stalagmite SPA 119 from Spannagel Cave formed during MIS 7.4, a period of low global sea level (cf. Fig. 1) when this alpine karst was most likely covered by ice. (A) Longitudinal cross section and U–Th age-depth relationship used to convert depth into age, (B) high-resolution stable isotope traverse (yellow line in (A)). The $\delta^{18}O$ values are on average 6‰ lower than fully interglacial calcite and the $\delta^{13}C$ values show no impact of biogenic C from the soil zone. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

the latter group shows $\delta^{13}C$ values typically between ca. −2 and +2‰ and hence approaches the C isotopic composition of the host marble (ca. +2‰). We attribute these compositions to the lack of pedogenic C input into the subsurface. Locally, kinetic fractionation also played a significant role during deposition of some flowstones, as shown by positive $\delta^{13}C$ values as high as +21‰, covarying with $\delta^{18}O$ values (see below).

The O isotopic compositions of presumably 'subglacial' speleothems is distinctly lower than those of

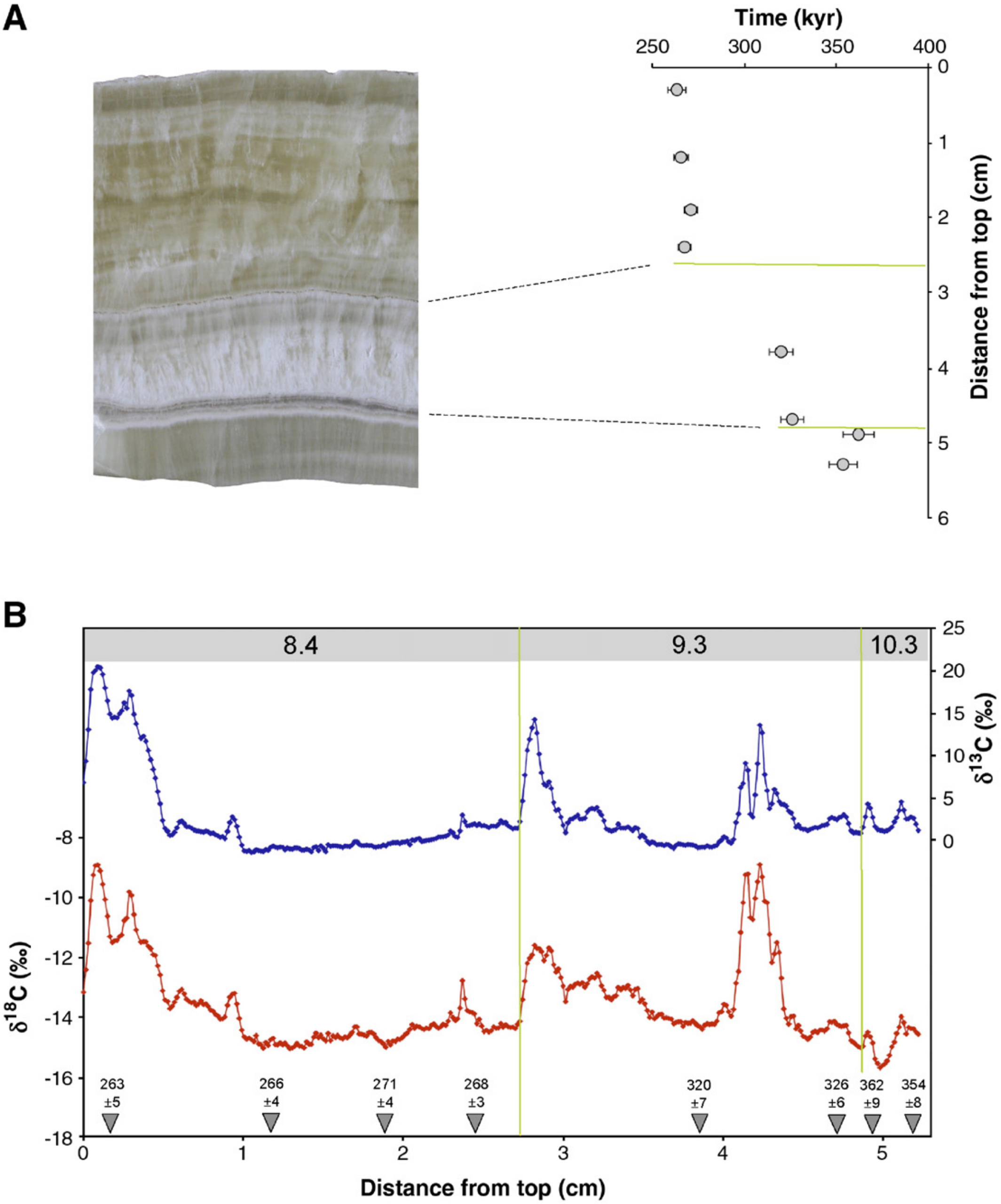

Fig. 3. Flowstone SPA 4 from Spannagel Cave shows at least three discrete growth pulses between MIS 10.3 and 8.4. The location of the hiati (green lines) is based on petrographic observations. Individual U–Th determinations (in kyr, including 2-sigma uncertainties) and the high-resolution stable isotope record are given on a depth scale as the growth phases were too short (relative to the dating uncertainties) to allow for a reliable age model. Note strong positive covariance between O and C isotope values. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Holocene [16,24,25] and MIS 5.5 [26] samples by several per mil. This is consistent with the present-day relationship between air temperature and O isotopic composition of precipitation in temperate continental settings [27], i.e. lower cloud condensation temperatures give rise to lower $\delta^{18}O$ values of precipitation. Quantifying the degree of atmospheric cooling would require accurate constraints on the relationship between air temperature and $\delta^{18}O$, which are currently not available for the late Pleistocene.

6. Two examples

Stalagmite SPA 119 was found detached from its substrate in a paleophreatic gallery of the northern branch of Spannagel Cave. Its internal structure shows evidence of a gradual change in the orientation of the extension axis. For the most part the sample is composed of inclusion-poor, coarsely crystalline low-Mg calcite. The upper part is distinctly gray due to the presence of minor impurities and the top layer, which shows a sharp lower boundary (possibly a hiatus), is again translucent calcite. Like all samples from Spannagel Cave, SPA 119 is very U-rich (Table 1) and essentially free of detrital Th resulting in a robust age model suggesting near-constant deposition between ca. 230 and 220 kyr (Fig. 2). According to the orbitally tuned deep-sea chronology (e.g., [22]) this sample grew during MIS 7.4, a period of markedly low sea-level. The stable isotope data corroborate the interpretation that this stalagmite did not form during conditions comparable to today. The $\delta^{13}C$ values are uniformly high and show no influence of soil-derived biogenic C. On the contrary, the $\delta^{18}O$ values are significantly lower than Holocene or Eemian speleothems from this cave [16,25,26] and decrease gradually from ca. −13.5 to −14.5‰ (Fig. 2). Some kinetic enrichment of both isotope systems is evident, in particular near the top of the sample.

Sample SPA 4 is a ca. 5 cm thick, inactive flowstone, which formed in the central part of Spannagel Cave. Detailed dating of this exceptionally U-rich sample (up to 399 ppm U, Table 1) revealed a complex growth history starting ca. 360 kyr ago and terminating at ca. 263 kyr. Constrained by two consistent dates the first growth phase occurred during MIS 10 (probably during MIS 10.3 — Fig. 3). Deposition apparently came to a halt during the coldest period of MIS 10 and recommenced during MIS 9 (9.3). The last generation of calcite was deposited again during a rather cold climate period, MIS 8.4. Although the analytical uncertainties of these old ages allow for some adjustments of the growth history it is clear that the latter was not biased toward climatically warm periods. Similar to the previous sample this flowstone also shows strikingly high and invariant $\delta^{13}C$ values coupled with very low $\delta^{18}O$ values. Superimposed on these baselines are several marked positive excursions which could erroneously be attributed to warming events. These high $\delta^{18}O$ values are mirrored by high $\delta^{13}C$ values (opposite of what would be expected for a warm, interglacial climate supporting vegetation), strongly suggesting that both isotopes increased as a result of kinetically controlled fractionation. Note that these anomalous $\delta^{13}C$ values are much higher than those typically measured in speleothems (e.g., [28,29]).

7. Discussion

Spannagel and nearby caves (including Kleegruben Cave [30]) present a natural laboratory of speleothem deposition close to the lower temperature limit even during the present interglacial. Given this sensitive setting, a drop in temperature equivalent to an ELA depression during the “Little Ice Age” (−180 m relative to the mean of 1930–1960 [31]) will result in a glacier expansion covering about two thirds to three quarters of the area above the cave by ice. Only the easternmost segment of the cave was ice-free during the 1850 ice advance allowing stalagmites to grow interruptedly during the Holocene [24,25]. Cooling the atmosphere in the Alps by an amount exceeding the Holocene temperature variability will give rise to an even lower ELA (e.g.,−300 m during the Younger Dryas or −700 m during the Late Glacial Gschnitz stage [32]). This cooling would inevitably propagate into the cave [33,34], freeze the cave and its karst aquifer and stop carbonate precipitation. Previous work at Spannagel and neighboring caves has shown that contrary to these expectations (a) speleothems also formed during MIS 3 [14,21], an interval certainly colder in the Alps than any time during the Holocene [35–37], (b) growth rate did not change significantly between interstadial and stadials of MIS 3 and stalagmites grew even during the Heinrich 5 event [14,30], (c) speleothems also formed during MIS 7 cool periods [21], and (d) flowstones and stalagmites recorded a period of carbonate deposition early during Termination II when $\delta^{18}O$ values were still low [15,21,38]. Rather than postulating ice-free conditions during all times when speleothem deposition formed in these caves, this apparent paradox – speleothem growth during cold conditions at high elevations – can be resolved by a non-pedogenic origin of these speleothems at times when the cave was buried beneath a warm-based alpine glacier. A temperate glacier is characterized by an average air temperature exceeding 0 °C during the summer. The primary energy input (by solar radiation) results in melt-water production which is transported to the glacier bed feeding a complex network of melt-water channels [39,40]. During the winter season the melt-water supply to conduits is greatly reduced and conduits largely collapse [41]. The summer melt water eliminates the winter cold wave from firn by latent heat of refreezing melt-water [42]. The temperature at a depth of 10 to 15 m in a temperate glacier therefore is significantly higher than the temperature above the ice. [42] lists several examples in alpine glaciers where the measured temperature at 30 m depth is 0 °C, contrasting with annual air temperatures of −7 to −8 °C. A temperate alpine glacier therefore acts like a thermostat keeping the top of the rock overburden close to 0 °C and preventing the cave from freezing even when the atmospheric air temperature is well below 0 °C.

In the absence of a carbon dioxide flux from the soil zone the oxidation of sulfides in the karst fissure aquifer provides the impetus for carbonate dissolution [12,43] and the resultant C isotopic composition of the drip water and the speleothems reflects this process by inheriting the C isotopic composition of the host marble (locally modified by strong kinetic fractionation as in the case of sample SPA 4).

Speleothem deposition in Spannagel and neighboring caves therefore may show different modes which allow us to make inferences about the presence of warm-based glaciers and the magnitude of (annual, summer) temperature changes:

Pleniglacials. The cave is covered by a cold-based glacier and the temperature in the cave adjusts to a value close to the external temperature. As the present-day average summer temperature outside Spannagel Cave is ca. 7 °C a cold-based glacier cover therefore corresponds to a summer temperature at least 7 °C lower than today. Growth of stalagmites is not possible under these conditions and this is also shown by the U–Th data (Fig. 1). Despite a high density of individual dates none fall into the glacial maxima of MIS 2, 6 or 10 (with the possible exception of MIS 8.2, but this is based on a single data point — Fig. 1).

Terminations. Deposition of stalagmites at the end of a glacial period will begin as soon as the glacier above the cave becomes temperate. The threshold for this conversion requires average summer temperatures >0 °C, corresponding to average summer temperatures about 7 °C lower than today. Mean annual air temperatures at 2500 m elevation should be significantly lower than the present-day mean annual air temperature of 0 °C. Assuming the summer temperatures to be close to 0 °C, the average temperatures were at least 5 °C lower than today. This occurred ca. 136 kyr ago, when speleothem formation recommenced, while fully interglacial conditions (as deduced from stable isotope data) were only reached by ca. 129 kyr [21].

Interglacials. Speleothem deposition is possible if the mean annual air temperature in the cave exceeds 0 °C. In the absence of a temperate glacier above the cave, the growth of speleothems becomes more sensitive to cold spells. Taking the present-day conditions as representative for interglacials, a lowering of the mean annual air temperature by about 2 °C will stop growth of stalagmites. Consequently, the formation of stalagmites during interglacials becomes sensitive to a rather small deterioration of the annual temperature by a couple of degrees. The rock cover above the cave, however, delays the response to the cold spell. Applying a thickness (x) of 30 m for the rock cover, the time scale for the response ($t = x2/D$) amounts to 14 yr, assuming a diffusion coefficient (D) for the heat in the rock of 2×10^{-2} cm^2/s [44]. Thus, only cold spells with duration longer than 14 yr will affect the growth of stalagmites during interglacials.

Our stable isotope data clearly show that the interglacials during the past ca. 350 kyr differed substantially, in that only MIS 5.5 and the Holocene resulted in largely ice-free conditions above Spannagel and adjacent caves giving rise to soil development (as recorded by low δ^{13}C values in the speleothems). None of the MIS 7 warm periods apparently reached this level of ice retreat (and/or were long enough to sustain soil formation), which is consistent with the generally lower global sea level as compared to MIS 5.5 [45]. Interestingly, not even MIS 9 appears to have reached Holocene conditions (cf. sample SPA 4). We therefore argue that these samples record periods when the Hintertux Glacier was larger than today and larger than during the "Little Ice Age" and apparently warm-based.

Interstadials and stadials. Our earlier results demonstrated that at Spannagel and Kleegruben Cave growth of speleothems occurred during MIS 8 and MIS 3, respectively, when the ambient temperature was certainly lower than during interglacials [21,30]. As we also find growth periods during at least part of the stadials (with more negative δ^{18}O values than in the interstadials [30]), we conclude that the average temperatures during interstadials were higher than −5 °C if the location was covered by a temperate glacier, enabling formation of speleothems during at least part of the stadials.

8. Conclusions

While speleothems in most caves stop growing during cold periods because of lack of CO_2 input from the soil zone (e.g. [46,47]), deposition may continue to form in high-alpine (and possibly also high-latitude) caves irrespective of the presence of soil and vegetation above the cave as long as the cave's interior temperature is above freezing and sulfides are present. Castleguard Cave (Alberta) and Spannagel and surrounding caves in the Austrian Alps are two examples. These 'subglacial' speleothems therefore provide a new and potentially unique means to identify and reliably date the presence of warm-based paleoglaciers in previously glaciated regions.

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