Late quaternary speleogenesis and landscape evolution in the northern Apennine evaporite areas

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Abstract

Gypsum beds host the majority of the caves in the north-eastern flank of the Apennines, in the Emilia Romagna region (Italy). More than six hundred of these caves have been surveyed, including the longest known epigenic gypsum cave systems in the world (Spipola-Acquafredda, ~11 km). Although this area has been intensively studied from a

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geological point of view, the age of the caves has never been investigated in detail. The rapid dissolution of gypsum and uplift history of the area have led to the long-held view that speleogenesis commenced only during the last 130,000 years.

Epigenic caves only form when the surface drainage system efficiently conveys water into the underground. In the study area, this was achieved after the dismantling of most of the impervious sediments covering the gypsum and the development of protovalleys and sinkholes. The time necessary for these processes can by constrained by understanding when caves were first formed.

The minimum age of karst voids can be indirectly estimated by dating of the infilling sediments. U-Th dating of carbonate speleothems growing in gypsum caves has been applied to 20 samples from 14 different caves from the Spipola-Acquafredda, Monte Tondo-Re Tiberio, Stella-Rio Basino, Monte Mauro, and Castelnuovo systems. The results show that: i) caves were forming since at least ~600 kyrs ago; ii) the peak of speleogenesis was reached during relatively cold climate stages, when rivers formed terraces at the surface and aggradation caused paragenesis in the stable cave levels; iii) ~200,000 years were necessary for dismantling of most of the sediments covering the karstifiable gypsum and the development of a surface mature drainage network.

Besides providing a significant contribution to the understanding of evaporite karst evolution in the Apennines, this study refines our knowledge on the timescale of geomorphological processes in a region affected by rapid uplifting.

Keywords: carbonate speleothems; evaporite karst; landscape evolution; cave formation; palaeoclimate

1. Introduction

Caves can form under two main hydrochemical conditions: epigenic and hypogenic (Klimchouk, 2007). Surface waters that derive their aggressiveness from the land surface and the atmosphere carve epigenic caves, while fluids attaining their acidity from deep sources (i.e. rising H_2S or CO_2, or mixing of two different solutions such as in coastal settings) are responsible for the formation of hypogenic caves. Hypogenic speleogenesis, often happening deep below the Earth surface, can act over very long time spans (Polyak, 1998). Epigenic speleogenesis, on the other hand, occurs closer to the land surface, and erosional processes tend to intercept karst voids in a few millions of years.

Karst dynamics are intimately associated with environmental variations at the surface (Ford and Williams, 2007), and the chronological refinement of dissolution processes and karst history may help us to better understand base-level (i.e. local water table) variation, which is related to geological events such as uplift and other tectonic movements, as well as valley incision. In this regard, the study of epigenic caves carved in gypsum provides some advantages compared to those hosted in limestone/dolostone. Gypsum is highly soluble, and speleogenesis occurs up to hundred times faster than in carbonates (Klimchouk, 2000). The explorable caves at present, if not confined in deeper levels of the geological sequence (i.e. hypogenic Ukrainian systems, Klimchouk, 2012), must be relatively young because the rapid underground erosion combined with surface weathering lead inevitably to the fast dismantling of the bedrock, demolishing the underlying caves. On the other hand, high gypsum solubility provides an almost contemporaneous response of speleogenesis to surface climatic and environmental changes, even at intra-Milankovitch timescales (Columbu *et al.*, 2015). It follows that dating epigenic gypsum caves offers the double opportunity of constraining the timing of cave formation and allowing a better understanding of the link between external and underground geological processes. Consequently, because of the link between epigenic caves and surface realms, understanding the age of the first proto-caves provides important insights on the timing of geomorphological evolution at the surface. Specifically, it constrains the time necessary for: 1) the exhumation of the karst terrain (i.e. when karstifiable bedrock is no longer confined and starts to undergo surface weathering and erosion), and ii) the development of surface drainage capable to efficiently convey water into the subterraneous system, thus forming epigenic caves (i.e. formation of protovalleys, sinkholes, and dolines).

One of the most intriguing - yet complicated – conundrums when studying the genesis and evolution of cave systems is an understanding of the exact timing of void formation, i.e., the problem of assigning a numeric age to “what is not there” (Sasowsky, 1998); a range of strategies have been proposed in order to overcome this problem. The ages obtained by the dating of alunite in Carlsbad and Lechuguilla caves are considered contemporaneous with the formation of the cave passages (Polyak *et al.*, 1998), because this mineral forms when the acidic cave-forming waters encounter clays. In the same area, dolomite has also been reported as a speleogenetic byproduct (Polyak *et al.*, 2016) Similarly, replacement gypsum in sulphuric-acid caves can provide reliable ages of cave formation, as long as gypsum did not undergo alteration - and did not loose uranium - over time (Plan *et al.*, 2012; Piccini *et al.*, 2015). These can be considered the nearest direct approaches to dating a void. However, the survival of alunite and/or gypsum is restricted to those parts of the cave that have never experienced flooding, dripping or seepage since their formation. In addition, formation of speleogenetic dolomite is possible in restricted circumstances. Also, these minerals are only found in sulphuric-acid caves, strongly limiting the utilisation of this direct-dating approach. As a result, karst voids are mostly indirectly dated (Audra *et al.*, 2006) using a range of infilling materials that logically are formed subsequent to the carving of the cave. These materials include sediments, archaeological remains, fossils and speleothems.

The speleogenesis of epigenic limestone and dolostone karst systems is often driven by changes of the local base-level at the surface (Williams, 1982; Ford and Williams 2007). This means that bedrock exhumation, allowing the start of speleogenesis, may have happened many millions of years ago. Additionally, with the exception of tropical environments (Farrant *et al.*, 1995; Audra *et al.*, 2011), cave development in carbonate rocks is a slow process (White, 1988). It follows that the formation of those karst networks currently showing complex multi-kilometric extensions may have begun several millions of years before present (De Waele *et al.*, 2012a; Tassy *et al.*, 2013; Calvet *et al.*, 2015; Häuselmann *et al.*, 2015). In such cases the datable materials traceable to the first stages of cave evolution may have been irretrievably lost, and even the primordial asset of the karst passages is likely overwritten by more recent speleogenetic phenomena. For example, dated speleothems from the middle levels (~900 m a.s.l.) of the Corchia system (central Italy) approach one million years in age (Woodhead *et al.*, 2006; Bajo *et al.*, 2012), while the exhumation of the area occurred not earlier than 5.0-4.5 Ma ago (Balestrieri *et al.*, 2003). The formation of the earliest upper levels (~1400 m a.s.l.) is assigned to the Late Pliocene (Piccini, 2011), and further U/Pb dating of speleothems from the highest cave levels would be needed to refine this chronological constraint.

We studied several still-active karst systems carved in the gypsum sequence outcropping in the Northern Apennine foothills, Emilia Romagna region, Italy (De Waele *et al.,* 2011). Earlier studies have also reported Late Messinian karst systems in this area, now entirely filled with sediments. These 5 million year old cave segments have formed before the gypsum sequence was covered with the thick less permeable Argille Azzurre Fm. and tilted in its actual monocline position (De Waele and Pasini, 2013). Epigenic caves in gypsum could not have formed since the Late Messinian submersion, and the most recent karst cycle started only after the Apennines emerged from the sea and most of the covering sediments were removed from the gypsum sequence. In this portion of the Apennine piedmont, karst networks are mostly composed of superimposed, more or less horizontal tunnels that reflect the position of past local base levels (Columbu *et al.*, 2015). We integrated the U-Th dating of twenty carbonate speleothems, which provide minimum ages of the passages in which they grew, with in-cave and external morphological features that are indicators of the palaeo base levels. These data have then been combined with regional geological relationships to document the Quaternary evolution of the underground and surface drainage of the area of study. The main aims and motivations of the paper therefore are:

1) To establish the timing of the inception of speleogenesis. For the aforementioned reasons the ages of these caves have been largely underestimated in the past. Previous research has linked the speleogenesis to the last 130,000 years (Demaria, 2002; Forti, 2003; Pasini, 2012); limited radiometric ages (Forti and Chiesi, 2001; Forti, 2003), and archaeological (Miari, 2007; Negrini, 2007) and paleontological (Pasini, 1969) findings have corroborated this idea.

2) To understand the periods during which the excavation of the different cave levels constituting the systems was most effective.

3) To evaluate the time necessary for the dismantling of impervious rocks originally covering the karstifiable gypsum unit and for the excavation of the first valleys, dolines and sinkholes, which allowed the penetration of water into the epigenic karst network.

Considering the above-said link between the subterranean and surface environments, the outcomes may reinforce our knowledge of the connection between the rapid uplift that is characteristic of the study area (Cyr and Granger, 2008; Picotti and Pazzaglia, 2008) and the landscape evolution of the middle to upper Pleistocene.

2. Study area

We explored five karst systems located in the northern Apennine piedmont (Emilia-Romagna region), carved in the *Vena del Gesso* (= Gypsum vein) formation (Vai and Martini, 2001; Lugli *et al.*, 2010): Spipola-Acquafredda, Monte Tondo-Re Tiberio, Stella-

Rio Basino, Monte Mauro and Castelnuovo (Figure 1). The first lies at the south-eastern periphery of Bologna (Figure 2A), while the remaining caves are situated ~40 km south-east of the city (south of the city of Imola), where the gypsum vein outcrops with its classical cuesta-like morphology (Figure 2B). The gypsum vein is mostly composed by macrocrystallyne selenitic gypsum. In macrocrystalline gypsum, pervasive primary porosity is scarce. Superficial water infiltrates into the underground once the bedrock is exposed by vadose fractures and joints originated by tectonic and/or diagenesis. Superficial karst landforms are also well developed in the study area (i.e. dolines and blind valleys, Figure 2), having a primary role in the current and past development of the underground drainage network. Gypsum in the Italian peninsula, excluding the few Permian and Triassic outcrops, was deposited during the Messinian Salinity crisis (Krijgsman *et al.*, 1999; Roveri *et al.*, 2014). In the Northern Apennines, it is exposed along a NW-SE elongated belt connecting the peaks of the mountain chain in the south with the Po plain foredeep in the north (Figure 1). The structure of the caves comprising the explored karst systems is intimately related to the geological and environmental factors that have characterised the area since the uplift of the Apennines. First, the majority of caves follow NW-SE tectonic features (De Waele and Piccini, 2008); less frequently WNW-ESE and W-E lineaments, or the SW-NE anti-Apennine directions, are exploited. Second, all caverns are epigenic (Klimchouk, 2000), developed in an unconfined aquifer in the first few hundred metres of the exposed bedrock. Considering that the karst conduits convey the infiltrating waters to the local base level, the altitude and elongation of the caves is related to base-level oscillations over time, associated with Quaternary climate changes and uplift (Columbu *et* *al.*, 2015). Several superimposed levels constitute the Spipola-Acquafredda and Monte Tondo systems, of which only the lowermost is currently active. Two levels form the Castelnuovo system and the Stella-Basino system, possibly reflecting the younger age of these two networks (Chiarini *et al.*, 2015). At Monte Mauro, on the other hand, the Grotta dei Banditi (Bandit cave) is the only branch left of a much bigger cave system, now completely eroded. Remains of carbonate flowstones that can still be found at the surface close to Grotta dei Banditi are all that is left of the other branches of this system. In fact, any review of the gypsum caves in the Emilia-Romagna territories should also consider those caves that are no longer visible today, destroyed by the continuous surface and subterranean erosion of the soft gypsum beds.

3. Materials and methods

In order to promote cave conservation (Fairchild and Baker 2012; Scroxton *et al.*, 2016), no in situ carbonate speleothems were removed; samples were principally found as broken specimens inside the caves. In a few cases, speleothems were subjected to core drilling (Spötl and Mattey 2012; Figure 2C) or found at the surface, in the vicinity of the cave entrance. Six flowstones were collected at the Monte Tondo-Re Tiberio system (RTf, A50, 3A, RTy, PP and GO) (Figure 2E). The first was found in the quarry at ~340 metres a.s.l., while the remaining samples were collected in cave galleries respectively at 270, 220, 180, 160 and 130 metres a.s.l. Two flowstones and one stalagmite were sampled in both the Spipola-Acquafredda and Castelnuovo karst systems. In the first, the SpD flowstone was found at the Croara quarry (250 metres a.s.l.) close to the Spipola doline

(Figure 2E) (Forti and Sauro, 1996), the Sp1 flowstone cored in a cave level at 120 m a.s.l. (Figure 2D) and SpS found in a cave level at 125 m a.s.l. In Castelnuovo, the Mor2 stalagmite was found in the Mornig cave at 190 m a.s.l, the P2 flowstone was recovered at the surface at 180 m a.s.l and the P3 flowstone was cored in a cave level at 185 m a.s.l. The exploration of the Monte Mauro cave system allowed us to recover a large stalagmite (BA_Big) from Banditi Cave at 450 m a.s.l., and two flowstones (Ba1 and Ba2) near the cave but at the surface (unroofed remains of a larger cave system). Two other flowstones (MM4 and MM2) were taken at the surface close to the highest elevation of Monte Mauro, around 480-490 m a.s.l. Three flowstones also come from the Stella-Rio Basino system: RBT was found at the surface at 160 m a.s.l, RB1 and RB3 collected underground at 170 m a.s.l. The predominance of flowstones over stalagmites is not a sampling bias: stalagmites are generally extremely rare in the Northern Apennine gypsum karst. Refer to Figure 3 for the size and macroscopic fabric of the samples.

The speleothems were sliced in two with a diamond saw. The analyses were performed on the fresh surface of one half after polishing; the other half was archived. The sub-samples (calcite prisms) used for the U-Th dating weighed between 10 and 120 mg, and were drilled along the growth layers using a dental hand-drill. The dating aimed to recover the base and top age of all the speleothems. When the stratigraphic bottom/top was considered unsuitable for dating (i.e. due to the visible presence of detrital material in the carbonate layers, evidence of carbonate dissolution, etc.), samples were taken of the closest clean and unaltered calcite. Intermediate ages were also determined in order to improve the chronological control of the longest samples.

In total, 70 U-Th ages were produced. The sub-samples, the drilling locations of which are reported in Figure 3, were first dissolved in HNO_3 following the procedure of Hellstrom (2003). A spike of known $^{236}U/^{233}U/^{229}Th$ ratio was added to the solution and the U-Th fraction was separated from the carbonate matrix using Eichrom TRU-Spec resin in columns. The U and Th fractions were collected together and evaporated overnight at 80°C before being taken up in 5% HNO3 / 0.5% HF ready for isotopic analysis. The majority of measurements were performed on a Nu Plasma multi collector – inductively coupled plasma - mass spectrometer (MC-ICP-MS) at the School of Earth Sciences, The University of Melbourne, following the methodology established in Hellstrom (2003) and refined in Drysdale *et al.* (2012). Three samples were instead analysed at t the Laboratoire des Sciences du Climat et de l'Environnement (LSCE) at Gif-sur-Yvette (France) according with Pons-Branchu *et al*. (2014) (Table 1). Th-corrected U-Th ages were calculated using equation 1 of Hellstrom (2006) with the ^{230}Th-^{234}U decay constants of Cheng *et al.* (2013) and an initial $(^{230}Th/^{232}Th)_i$ of 1.5±1.5. Samples were analysed in batches of 24 with eight accompanying HU-1 Harwell Uraninite measurements, and reported activity ratio uncertainties are expanded by the observed standard deviations of the standards. Age uncertainties are approximated as symmetric, except for the oldest samples where separate upper and lower uncertainty bounds were calculated.

4. Results

In general, the speleothem ages produced realistic radiometric ages, with samples possessing a high content of ^{238}U and, for the most part, a high $^{230/232}Th$ activity ratio

(^{230}Th/^{232}Th; Table 1). ^{238}U is never below 290 ng/g, with maximum concentrations exceeding 3000 ng/g, and with an average value of 1206 ng/g. (^{230}Th/^{232}Th) averages ~8000, but it correlates with the age of the speleothems. The Holocene samples report the lowest values (<10), followed by speleothems deposited during the early last glacial (100-200). In older samples (i.e. last interglacial or older) this ratio is two to three orders of magnitude greater. Initial ^{232}Th content has an important role in determining the 2σ uncertainty associated with the final age. Young samples possess relatively small amounts of authigenic ^{230}Th, thus (^{230}Th/^{232}Th) and corrected age uncertainty are highly sensitive to thorium derived from non-authigenic sources. For this reason, the uncertainties relative to the Holocene ages are, in percentage terms, greater than the other samples.

The final corrected ages and the associated errors are reported in Table 1. A comparison of speleothem age with the main climatic stages over the last ~800 ka is provided in Figure 4. The compilation follows the subdivision of the marine isotopic stages (MIS; Emiliani, 1955) on the global 'LR04' δ^{18}O benthic stack proposed by Lisiecki and Raymo (2005). Four speleothems grew during the last two millennia of the Holocene (MIS1) (RTy, Sp1, SpS, and RB3) while seven speleothems span from the middle (Mor2, RB1, RBT, and P3) to the early Holocene (GO and P2). Flowstones 3A, PP, and A50 were deposited during shorts intervals of the period ranging from ~110 ka to ~70 ka, the transitional phase from the last interglacial (MIS 5e) to the last glacial, characterised by the Greenland Interstadial (GI) and Stadial (GS) oscillation (NGRIP project members, 2004). Speleothem 3A formed from 108.86 ± 0.98 to 106.29 ± 7.23 ka, PP from 87.80 ± 0.70 to 75.21 ± 1.44 ka and A50 from 77.89 ± 6.06 to 74.69 ± 0.68 ka, correlated respectively with GI 24 and 21-20

(Columbu *et al.*, 2015). Two speleothems formed during the last interglacial: RTf and Ba_Big. They were mainly deposited during the climatic climax (~130-120 ka, MIS 5e), although the upper portion of the Ba_Big stalagmite grew across the LIG-glacial transition (top age 112.36 ± 1.09 ka). Flowstone SpD started to grow during the last phase of the glaciation corresponding to the MIS 8, but was mostly formed during the MIS 7e, from 253.90 ± 4.41 to 239.34 ± 4.30 ka. Two flowstones from the Monte Mauro system, MM4 and MM2, reported bottom ages much older than the top ages. MM4 bottom is at 468.00 $^{+130}/_{-42}$ ka (MIS 13 considering the average age, MIS 12 to 15 considering the error) and the top at 313.44 ± 14.27 ka (MIS 9), MM2 bottom is at 316.17 ± 12.65 ka (MIS 9) and the top at 239.08 ± 4.46 ka (MIS 7e). Although intermediate ages between the bottom and the top of these last two speleothems were not done, petrographic evidence shows that the deposition of the carbonate was not continuous but characterised by growth interruptions. Ba1 and Ba2, belonging to the same karst system, report only one age each respectively at 689 ± 369.10 ka (bottom age) and 378.26 ± $^{+29}/_{-20}$ ka (top age). Although the large uncertainty, we are 95% confident that Ba1 is older than 580 ka, and might coincide with MIS 15. Ba2 instead might fit with the latest part of MIS 11.

5. Discussion

Messinian gypsum caves in South Spain were formed in a semi-confined aquifer since the lower Pleistocene (Calaforra and Pulido Bosch, 2003). Marly strata intercalated with gypsum beds drove the speleogenesis creating the first proto-conduits under phreatic conditions, which later developed into a multi-level cave system following river entrenchment and vadose erosion. Several field evidences testify this process, such as “V” shaped passages, meandering braided pendants, ceiling channels and gypsum layer breakdown in the cave roofs as relict of the ancient speleogenetic evolution (Calaforra and Pulido Bosch, 2003). The situation is quite different in our area of study. All caves levels are ~horizontal, while the gypsum sequence is tilted (from 20˚ to 45˚, De Waele and Pasini, 2013). This means that marly strata did not controlled speleogenesis. We recently demonstrated that the formation of caves composing the Monte Tondo karst system was related to climate-driven base level oscillations over the last ~130,000, occurred in an unconfined aquifer (Columbu *et al.,* 2015). The formation of epigenic caves in macrocrystalline gypsum is only possible once the soluble rocks are emerged and in contact with undersaturated fresh water. The *Vena del Gesso* gypsum formation was originally covered by marine silts and clays of the *Argille Azzurre* (sky-blue clays) formation (Amorosi *et al.*, 1998), deposited from ~5.3 Ma to ~1.8 Ma, followed by the *Sabbie Gialle* (yellow sands) formation (Antoniazzi *et al.*, 1993; Cyr and Granger, 2008) (Figure 5). The latter is mostly comprised of littoral marine sandstone (Marabini *et al.*, 1995) deposited during Early Pleistocene phases of sea-level high stands. The top age of the *Sabbie Gialle* is attested at 780-820 ka, provided by electron spin resonance and palaeomagnetic calculations (Antoniazzi *et al.*, 1993; Falgueres, 2003; Muttoni *et al.*, 2011). This age (~800 ka) marks the beginning of the so-called “continental Quaternary” (Benini *et al.*, 1999; Amorosi *et al.*, 2015), which in the local sequence stratigraphy comprises all sediments deposited after the definitive regression of the Adriatic Sea. From a speleogenetic perspective, it defines the potential maximum age for the beginning of the development of the oldest epigenic caves. The emergence of the Apennine piedmont resulted from the rapid uplifting of the area, calculated at 0.2-0.3 mm/year around the middle Pleistocene (Cyr and Granger, 2008; Picotti and Pazzaglia, 2008). The retreat of the shoreline was possibly fuelled by the Middle Pleistocene Transition (Pisias and Moore, 1981; Muttoni *et al*., 2003; Maslin and Ridgwell, 2005), after which the severity of glaciations increased.

It is unlikely that speleogenesis started as soon as the sea retreated. All caves show epigenic morphology, characterised by typical gypsum through-flow conduits (Klimchouk, 2000). A sub-horizontal cave tunnel is excavated parallel to the piezometric level at the same altitude as the base level. This implies that: i) the majority of the sediments originally covering the gypsum karst terrain (i.e. Argille Azzurre clays and Imola sands) were removed before the commencement of the epigenic speleogenesis; and ii) the superficial drainage system already developed a certain status of maturity, sustaining the valley incision and creating the altitudinal gradient necessary for the flow of the groundwater (Figure 5). The oldest river terrace (Qt0) preserved along the Bidente and Reno valleys, located respectively to the north and to the south of the study area, has no numerical ages (Figure 4). Through stratigraphical correlation with the equivalent Po Plain foredeep sediments, it is tentatively assigned to one of the MIS 22, MIS 20, or MIS 18 glaciation peaks (Cyr and Granger, 2008; Picotti and Pazzaglia, 2008; Wegmann and Pazzaglia, 2009), respectively at ~870, ~800 and ~740 ka. Thus, proto valleys already developed at least by ~740 ka, which is 60-80 kyrs after the definitive emergence of the area. However, the hydrological network was possibly fully efficient between 400 and 500 ka, when the first alluvial fans were deposited in the Padana foredeep (Gunderson *et al*., 2014; Amorosi *et al*., 2015).

When comparing the minimum ages of the samples with global climatic variations over the last ~800 ka (Lisiecki and Raymo, 2005), it is worth noting that the presence of carbonate speleothems coincides with stages of climate optima (Figure 4), in line with our previous work based on the speleothems from Monte Tondo (Columbu *et al*., 2015). Speleothems were mainly formed during the peak interglacial periods (Holocene, MIS5e, MIS7e, MIS9, MIS11), with two of them potentially linkable to MIS13 (MM4 flowstone) and MIS 15 (Ba1). Although interglacials do vary in terms of duration, average temperatures and rainfall dynamics (Tzedakis *et al.*, 2009), these periods are recognised as the emblems of considerably warm and wet climate (Sirocko *et al.*, 2006). The influence of climate on the deposition of the speleothems is also evident in an intra glacial/interglacial timescale. GO, PP and A50 flowstones formed during the Greenland interstadials (GIS) 24, 21, and 20 respectively (Figure 4). GIS are rapid returns to an interglacial-like climate (although generally less warm) during an otherwise glacial period (Dansgaard *et al.*, 1993). Enhanced rainfall and humidity is the most suitable condition for the development of pervasive vegetation in the piedmont hillslopes. From a geochemical point of view, this means that the overabundance of biogenic CO_2 released by the pedogenic layers at the surface has a key impact in the production of carbonate speleothems once the waters percolate into the gypsum caves (Borsato *et al*., 2015).

From a geomorphological perspective, a thicker vegetation cover protects the hillslopes from surface erosion during these warmer and wetter periods. The availability of sediments on the slopes is reduced when compared to relatively colder climates. In southern Italy, pollen data indicate that glacial climates favoured the expansion of steppe-like and bush vegetation (Allen and Huntley, 2009). In this circumstance, the bedrock is largely exposed to weathering processes; the resulting regolith is gravitationally conveyed toward the bottom of the slopes, usually constituted by the active base level valley (Simoni *et al.*, 2013). Considering the latitudinal difference between the southern side of the peninsula and the northern Apennines, it is likely that in the studied area this vegetation persisted beyond the peak of the glaciations, prolonging conditions conducive to maximum regolith production.

The fluvial terraces preserved in the main rivers draining the northern Apennines (Bidente and Reno rivers) were all formed during these periods of abundant regolith production (Wegmann and Pazzaglia, 2009) (Figure 4). Besides the aforementioned Qt0 generation, terrace Qt1 correlates with MIS16, Qt2 with MIS12 or MIS10, Qt3 with MIS6, Qt4 with MIS4 and Qt5 with MIS2 (younger fluvial terraces are also known from the Holocene, but are not reported in Figure 4). The presence of a terrace deposit testifies that the river maintained a ~stable palaeo-altitudinal position necessary for the aggradation of the slope sediments in the river trunk. These periods of stability were longer than 1,000 years (Wegmann and Pazzaglia, 2009); the alluviated valleys were incised by a lowering of the base level under a different climate regime. The epigenic karst systems are strictly connected to the local base-level variation. When the base level lowers, karstification proceeds vertically, in the same way rivers entrench the valley bottoms. The Apennine gypsum karst shows evidence of this process through deep and narrow shafts; in the

Monte Tondo cave network, the *Pozzo Pellegrini* shaft is more than 30 metres deep (De Waele *et al.*, 2013). In contrast, when the river establishes a new altitudinal position, underground waters run according to the new piezometric level. The accumulation of slope sediments forms alluvial valleys that, at the millennial scale, act as new stable base levels for the karst systems. The main sub-horizontal cave tunnels are thus excavated during these stable time intervals.

Five superimposed cave levels are particularly well preserved in the Monte Tondo-Re Tiberio karst system. The connection between the fluvial dynamics and speleogenetic processes is also testified by the presence of paragenetic canyons in several of the cave ceilings (Pasini 1966, 1967, 1973). When the supply of slope material exceeds the maximum bed load sustainable by the river flow, aggradation occurs in the valley bottom pushing the local base level slowly upward. In response, underground conduits also accumulate sediments and the running waters erode the cave ceilings through so-called antigravitative erosion process (Pasini, 2009). Taking into account the cyclical stability of the base level witnessed by the river terraces and the stacked-like nature of the gypsum karst systems carved in the north Apennine piedmont, we attribute the excavation of the sub-horizontal cave tunnels to periods of relative colder climate (Figure 4 and 6). River terraces are present at the same altitude as the main cave levels of the studied area (Columbu *et al.*, 2015). Although they have not always been dated, their formation should correspond to the same genetic dynamics controlling the hydrogeology of the area (Wegmann and Pazzaglia, 2009), being deposited during cold stages. However, the ages of the speleothems provide an indication of the minimal age of the cave tunnels (and thus the related terraces). When terraces and caves are in the formation stage, the result is underground passages completely filled by water, and vadose speleothem deposition is not possible. Once these new caves are drained, following renewed entrenchment of the river, speleothem formation can start (Columbu *et al.*, 2015). Because cold-periods favouring the excavation of the cave voids are followed by warm periods favouring speleothem deposition, it is feasible to assign the speleogenesis of a certain cave level to the first cold-period preceding the oldest speleothem age sampled in that cave (Figure 6).

It could be argued that speleothems might have formed even thousands of years after the drainage of the cave, confounding our understanding of the age of speleogenesis. This cannot, of course, be excluded in this study. However, there is a good agreement between the antiquity of the speleothems and the altitude of the caves in which they were found (Figure 7). In the Monte Tondo-Re Tiberio system, RT flowstone – recovered at ~340 metres a.s.l. - was deposited ~130 kyr years ago (Columbu *et al.*, 2015). The speleothems sampled close to the actual base level in the cave (~95 metres a.s.l.) were Holocene in age. Those speleothems sampled in between these two cave levels report intermediate ages. In general, all the oldest speleothems come from the highest-altitude cave levels. The speleothems older than 300 ka, coming from the Monte Mauro system, have all been recovered at altitudes higher than 400 metres a.s.l., the most elevated portions of the Apennine foothills (Figure 7). Thus the chronology of the speleothem production follows an altitudinal gradient, which in turn mirrors the progressive base level-driven downward migration of the speleogenetic processes.

The oldest speleothems in our collection provides a basal age of >580 ka (Ba1) and 468.00 $^{+130}/_{-42}$ ka (MM4 flowstone). This suggests that caves already existed in the area at least since MIS15 (i.e. ~600 kyrs ago, testified by Ba1 age and the oldest possible MM4 age). Assigning the timing of speleogenesis to the coldest period just before MIS15 (MIS16), the excavation of the first caves likely occurred at least 630 kyrs ago, ~200 kyrs after the complete emergence of the area. This timespan (~200 kyrs) appears sufficiently long for several reasons. First, there is adequate time for the erosion of the majority of the marine sediments covering the area and the gypsum beds, fuelled by the rapid uplifting of the area. Currently, in many watersheds, these sediments are visible as uplifted fluvial terraces (Picotti and Pazzaglia, 2008; Wegmann and Pazzaglia, 2009). Second, it is a reasonable timeframe for the formation of a drainage network, which will later develop into deep river valleys. The age of the caves at ~630 ka is considerably close to the time of the formation of the first alluvial fan in the Padana foredeep (Gunderson *et al*., 2014; Amorosi *et al*., 2015), symptomatic of the general maturity of the fluvial system. Finally, there is sufficient time for triggering the geomorphological phenomenon known as "relief inversion" (Pain and Oilier, 1995) (Figure 5). Nowadays, gypsum beds stand out as prominent cuesta-like ridges because of the differential erosion of the adjacent formations (De Waele *et al.*, 2012b) (Figure 2B). Whereas in the non-soluble terrains surface erosion leads to a progressive lowering of the ground level, enhanced by the rapid uplifting of the area, in gypsum the bulk of the erosion is transferred underground. When gypsum was first exposed, the rivers descending from the upper part of the foothills penetrated the karst rocks, initiating blind valleys at the contact between soluble and non-soluble terrains

(Figure 5). The Rio Stella blind valley is today an excellent example of this style of surface versus subterraneous drainage (De Waele, 2010). Sinkholes also facilitated the penetration of the water into the karst systems. The large Spipola doline close to Bologna, with a diameter of more than 500 metres, is one of the most striking examples (Forti and Sauro, 1996) (Figure 2A). Around six hundreds years ago, the first proto-caves transferred surface water into the downstream main river throughout karst springs, at least in the Monte Mauro and Banditi caves where MM4 and Ba1 flowstone were found. Large portions of these first caves have been now disaggregated. At the same time the flowstone collected in the Spipola doline (close to Bologna), dated at ~250 ka, testifies to the formation of the caves in this area from at least 260,000 years ago; these old caves, probably forming the upper levels of the current levels constituting the Spipola-Acquafredda system, have also been eroded.

6. Conclusions

This study reports the U-Th radiometric dating of twenty carbonate speleothems sampled in the Messinian gypsum karst areas of the northern Apennines. In agreement with previous work based on speleothems recovered from the Monte Tondo-Re Tiberio karst system, the age of the samples fits with periods of warm and wet climate. Specifically, the speleothems correlate with the current (Holocene) and previous interglacials (MIS 5e, 7e, 9, 11, 13 and probably 15) and the Greenland interstadials 24, 21 and 20. The basal age of the speleothems provides an indication on the minimum age of the cave passages in which they are found. The age of the speleothems found at the surface testify to the presence of past cave levels that have been partially or completely destroyed by surface denudation. The main sub-horizontal cave tunnels constituting the explored karst system formed when the palaeo-base level stabilized at a new altitudinal position during cold-dry climate stages. In these periods, vegetation-free hillslopes saw high amounts of regolith transported downslope to the trunk river valley, after which subsequent incision formed river terraces at the same altitude of the cave levels. The available ages of the river terrace formations in the area corroborate the correlation between speleogenesis and periods of relatively cold and dry climate. Considering the age of the speleothems, the formation of the associated cave passages were assigned to the cold-dry climate stage immediately before the warm-wet phase attributed to speleothem growth.

This study has important implications for understanding the timing of underground and surface drainage systems development in rapidly uplifting areas. Firstly, it reviewed the duration of speleogenesis in Northern Italian gypsum terrains, which has been underestimated until now. In the Monte Mauro area, caves were already forming at least ~630,000 years ago, 200,000 years after the emergence of the Apennine piedmont. It follows that this time period was necessary for the erosion of most of the sediments covering the gypsum sequence, the incision of the proto valleys and the creation of the karst sinkholes, the latter introducing the majority of surface waters into the subterranean voids. Furthermore, the carving of epigenic caves was enhanced once the phenomenon of relief inversion exposed the gypsum sequence as a karstificable ridge among non-soluble terrains. In the Spipola area close to Bologna, caves were present at least 260,000 years ago. These old cave conduits have mostly been destroyed by surface denudation of the gypsum bedrock, while the carbonate speleothems that decorated their walls/floors have been better preserved and can still be found scattered across the surface.

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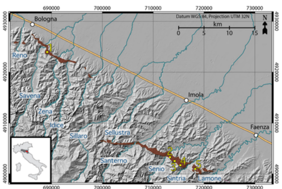

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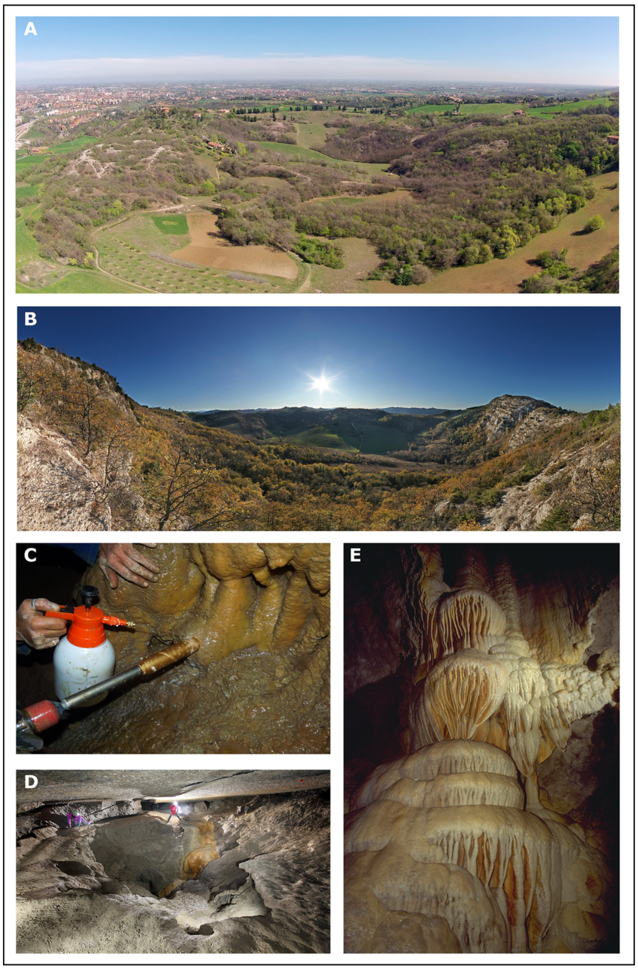

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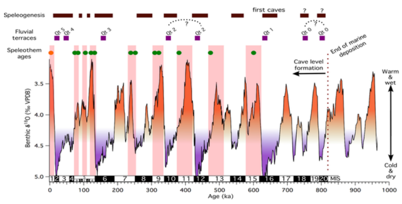

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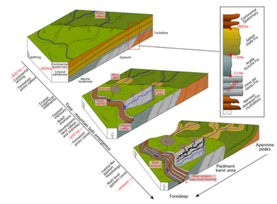

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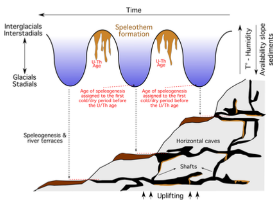

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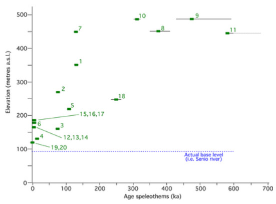

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Cave system	Cave	Sample ID	Mass (g)	^{238}U (ng/g)	(^{230}Th/^{238}U)$_A$	2σ	(^{234}U/^{238}U)$_A$	2σ	(^{232}Th/^{238}U)$_A$	2σ	(^{230}Th/^{232}Th)$_A$	Corrected Age (ka)	2σ
		RT-AI r	0.037	1483	0.6542	0.0049	0.946	0.004	0.000454	0.000001	1440	130.58	2.39
		RT-A1 bis	0.015	848	0.6541	0.0043	0.945	0.002	0.000657	0.000006	995	131.10	1.87
		RT-A1	0.106	1023	0.6558	0.0012	0.945	0.002	0.000848	0.000007	774	131.74	0.75
		RT 2015 1	0.020	1554	0.6490	0.0035	0.944	0.002	0.000402	0.000003	1616	129.04	1.53
		RT 2015 5	0.019	1656	0.6572	0.0039	0.956	0.002	0.000607	0.000003	1082	128.67	1.65
		RT-B1	0.009	934	0.6541	0.0044	0.956	0.003	0.000239	0.000005	2733	127.57	1.83
		RT 2015 7	0.019	1623	0.6510	0.0028	0.957	0.003	0.000730	0.000004	892	125.77	1.28
		RT-B I	0.010	1797	0.6479	0.0052	0.952	0.008	0.000125	0.000001	5174	126.24	3.20
		RT-B2	0.008	1068	0.6541	0.0039	0.962	0.003	0.000517	0.000006	1266	125.87	1.64
	Mezzano	RT-C1	0.010	1004	0.6562	0.0046	0.965	0.003	0.000599	0.000007	1095	125.58	1.85
		RT-CI	0.012	1748	0.6498	0.0055	0.958	0.002	0.000075	0.000002	8677	125.35	2.11
		RT-C m1	0.013	3127	0.6472	0.0025	0.957	0.002	0.000098	0.000001	6606	124.95	1.08
		RT-C m2	0.017	1619	0.6422	0.0020	0.949	0.002	0.000295	0.000003	2176	125.14	0.91
		RT-CII	0.012	1669	0.6389	0.0053	0.951	0.002	0.000250	0.000003	2552	123.45	2.02
		RT 2015 10	0.020	1696	0.6508	0.0032	0.963	0.002	0.000146	0.000001	4448	124.36	1.32
		RT-C2	0.015	972	0.6450	0.0028	0.953	0.002	0.000215	0.000003	3004	125.17	1.23
		RT-DI	0.033	1764	0.6461	0.0049	0.955	0.001	0.000682	0.000003	948	124.74	1.85
Monte Tondo - Re Tiberio		RT-D III	0.050	1899	0.6484	0.0048	0.961	0.003	0.000684	0.000003	948	123.94	1.94
		3A	0.014	2970	0.5932	0.0025	0.946	0.002	0.002052	0.000027	289	108.86	0.98
	3 Anelli	3A-2016-2	0.045	534	0.6016	0.0024	0.960	0.002	0.005345	0.000025	113	107.69	1.28
		3A-t	0.020	1157	0.6223	0.0037	0.968	0.003	0.038551	0.000421	16	106.29	7.23
		A501	0.008	1681	0.5176	0.0025	0.970	0.002	0.033461	0.000349	15	77.89	6.06
	Abisso 50	A502	0.011	1867	0.4843	0.0024	0.965	0.002	0.002576	0.000037	188	76.10	0.75
		A503	0.010	955	0.4829	0.0024	0.975	0.002	0.001806	0.000019	267	74.69	0.68
		D3	0.100	513	0.5378	0.0021	0.973	0.002	0.002398	0.000013	224	87.80	0.70
		PP	0.015	295	0.5207	0.0045	0.976	0.003	0.002766	0.000019	188	83.14	1.24
	Pozzo Pollini	PP1	0.009	795	0.5201	0.0029	0.970	0.002	0.018004	0.000116	29	81.18	3.23
		D2	0.100	652	0.4886	0.0009	0.950	0.001	0.008241	0.000022	59	77.90	1.40
		D4	0.100	670	0.4912	0.0013	0.957	0.001	0.008062	0.000030	61	77.70	1.40
		PP2	0.017	1165	0.4859	0.0028	0.968	0.002	0.007278	0.000078	67	75.21	1.44
	Grotta Oliver	GO1	0.012	1217	0.0921	0.0009	0.949	0.002	0.023011	0.000284	4	7.10	4.10
		GO2	0.010	924	0.0962	0.0007	0.943	0.002	0.026957	0.000186	4	6.96	4.92
		GO-2016-2	0.044	1205	0.0670	0.0004	0.939	0.002	0.017181	0.000321	4	5.03	3.09
	Galleria Principale	RTy 1	0.047	1511	0.0052	0.0002	0.960	0.001	0.000451	0.000008	12	0.52	0.08
		RTy 2	0.045	1432	0.0035	0.0002	0.960	0.001	0.000613	0.000014	6	0.29	0.11
		RTy 3	0.051	1548	0.0042	0.0001	0.982	0.002	0.002122	0.000050	2	0.11	0.36
		Spd-2016-1	0.047	924	1.0480	0.0042	1.126	0.002	0.000147	0.000002	7124	253.90	4.41
		SpD-E	0.120	586	0.9303	0.0097	1.026	0.009	0.000416	0.000004	2238	252.10	14.77
		SpD-D	0.118	442	0.9061	0.0104	1.009	0.010	0.002357	0.000026	384	246.63	16.16
		SpD-C	0.049	1003	1.0197	0.0064	1.112	0.004	0.000020	0.000001	50924	243.58	6.52
		SpD-B	0.051	3052	1.0215	0.0051	1.111	0.004	0.000004	0.000000	277192	245.55	5.64
Spipola - Acquafredda	Spipola	SpD-A	0.050	1971	1.1318	0.0055	1.207	0.004	0.000465	0.000009	2432	246.34	5.31
		SpD b	0.063	429	0.9245	0.0035	1.028	0.002	0.000264	0.000002	3506	243.53	3.97
		Spd-2016-2	0.048	484	0.8964	0.0035	1.007	0.002	0.002002	0.000039	448	239.34	4.30
		Sp1-b	0.038	736	0.0170	0.0004	1.029	0.003	0.007400	0.000186	2	0.63	1.19
		Sp1-t	0.034	731	0.0071	0.0002	1.024	0.003	0.003287	0.000071	2	0.23	0.53
		SpS-b	0.044	1012	0.0143	0.0003	1.045	0.002	0.001267	0.000025	11	1.30	0.20
		SpS-t	0.045	1062	0.0027	0.0002	1.031	0.002	0.000901	0.000015	3	0.14	0.14
		P2-b	0.039	989	0.0818	0.0007	0.984	0.003	0.009378	0.000174	9	7.89	1.58
	Peroni	P2-t	0.036	1120	0.0389	0.0004	1.048	0.003	0.000289	0.000005	135	4.08	0.06
Castelnuovo		P3-B	0.035	1435	0.0479	0.0009	1.014	0.002	0.000412	0.000011	116	5.22	0.12
		P3-T	0.044	1368	0.0246	0.0005	1.057	0.002	0.000071	0.000001	347	2.56	0.05
	Moring	MOR2-b	0.020	1314	0.0424	0.0006	0.880	0.002	0.012795	0.000034	3	2.99	2.44
		MOR2-t	0.021	2313	0.0218	0.0003	0.857	0.003	0.001254	0.000005	17	2.57	0.24
		RBT-b	0.049	1076	0.0588	0.0005	0.893	0.003	0.013927	0.000202	4	4.84	2.64
		RBT-t	0.052	1102	0.0432	0.0005	0.887	0.003	0.008033	0.000094	5	3.95	1.51
Stella - Rio Basino	Rio Basino	RB3-b	0.050	511	0.0252	0.0005	0.937	0.003	0.004679	0.000060	5	2.15	0.83
		RB3-t	0.050	492	0.0089	0.0004	0.950	0.003	0.003228	0.000060	3	0.47	0.56
		RB1-b	0.034	569	0.0612	0.0013	0.935	0.002	0.011802	0.000275	5	5.29	2.11
		RB1-t	0.042	494	0.0282	0.0008	0.953	0.002	0.010104	0.000227	3	1.53	1.77
		MM2-b	0.039	731	0.9225	0.0042	0.981	0.003	0.019454	0.000541	47	316.17	12.65
	Monte Mauro	MM2-t	0.034	1367	0.8783	0.0031	0.990	0.003	0.001359	0.000028	646	239.98	4.46
		MM4 b	0.021	556	0.9782	0.0062	0.994	0.003	0.001776	0.000010	551	468.00	+130/-42
		MM4 t	0.023	393	0.9375	0.0060	0.995	0.003	0.002130	0.000009	440	313.44	14.27
Monte Mauro		BA 1.1	0.050	1071	0.9988	0.0051	0.998	0.004	0.111054	0.000331	9	> 580	-
		BA 2.1	0.050	804	0.9569	0.0045	0.991	0.004	0.001508	0.000023	635	378.00	+29/-20
	Banditi	BA_BIG_1	0.075	1837	0.6804	0.0026	0.976	0.004	0.000010	0.000000	69553	131.29	1.46
		BA_BIG_2	0.069	886	0.6776	0.0030	0.987	0.003	0.000011	0.000001	62589	127.07	1.40
		BA_BIG_3	0.066	706	0.6563	0.0029	0.990	0.003	0.000039	0.000001	16700	119.01	1.26
		BA_BIG_4	0.070	979	0.6363	0.0026	0.990	0.003	0.000462	0.000007	1377	112.36	1.09