Geophysical and geomorphological investigations of a Quaternary karstic paleolake and its underground marine connection in Cassis (Bestouan, Cassis, SE France)

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Abstract

An original karstic system in a coastal alluvial plain located in a carbonate environment (Cassis, Provence — SE France) was studied using a multi-method approach (electrical resistivity tomography, gravimetric, passive seismic survey), combined with geomorphologic methods, surface observations and drilling. Limestone dissolution and/or cavity collapse led to the development of a polje, which was filled rapidly by erosion of Aptian marls. The combination of several dating and paleoenvironmental proxies indicates that the polje was filled during a glacial period (MIS 6, 8 or 10, i.e. in the 130–360 ka age range). This discovery has implications for the understanding of karst processes. The connection between the polje and the three km-long Bestouan underwater karstic conduit with submarine outlet is strongly suggested by sedimentological studies and geophysical prospections.

Keywords:
Karst
Provence
Geophysical prospection
Submarine spring
Paleolake
Cassis
Bestouan

1. Introduction

Karst areas, which represent about 20% of the land surface, are characterised by a specific landscape, usually devoid of surface runoff but with surface (e.g. sinkhole, canyons) or deep (e.g. caves, underground rivers) karstic structures (Ford and Williams, 2007; de Waele et al., 2009). Karst structures present major economic and societal challenges (e.g. Margiotta et al., 2012). They are adequate for agriculture and ideal sites for human habitation (Siart et al., 2010). However, karst systems have very heterogeneous landforms and it is very difficult to study the extent and geometry of deep structures. Understanding the link between the endokarst and exokarst features and their 3D geometry (Audra and Palmer, 2013) is of particular interest for fresh groundwater management and for geotechnical issues. Karstic sediments are also used for paleoenvironmental studies as they function as terrestrial archives (Delannoy et al., 2009). Water flow is responsible for karst development in surface landscapes and underground carbonate rocks where dissolution open voids. Surface water can also sink below ground, forming an underground river flowing through large karstic conduits (Mocochain et al., 2009).

Poljes are karstic structures in the form of a large and flat-floored enclosed depression. The flat floor is caused by flooding during high flows due to inefficiency of underground drainage. Ford and Williams (2007) recognized three types of polje according to their main function: (1) poljes dominated by allogenic input controls are very common along highlands that can supply abundant clastic load to rivers; (2) structural poljes are dominated by bedrock geological controls; and (3) base-level poljes are dominated by water-table control.

The Mediterranean region is very suitable for the study of karst phenomena due to ubiquitous limestone lithology, mainly of Cretaceous but also of Jurassic age (Audra et al., 2004), in particular in SE France. Limestones in Provence are often massive and deformed by a complex tectonic history (Demory et al., 2011; Mollieux et al., 2011; Lamarche et al., 2012), with significant relief-related energy that is important for karst formation. Karstic areas are shaped by eustatic and/or tectonic forcing as well as hydrological changes. Coastal karst refers to karstic

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features which have been repeatedly flooded and drained by oscillating sea levels. The evolution of coastal carbonate karst depends on many factors such as groundwater hydrodynamic behaviour which is linked to sea-level changes (van Hengstum et al., 2011; Canora et al., 2012). Sea level has a key role in the formation and functioning of coastal karst structures around Mediterranean Sea: 1) the draw-down of sea level during the Messinian Salinity Crisis is responsible for the development of karst networks below glacial period sea level (Audra et al., 2004); 2) the rise of sea level during the Holocene caused the development and flooding of karstic networks active during the glacial period; and 3) the sea water mixing with coastal aquifers (saline intrusion) creates submarine source discharge of brackish water (Fleury et al., 2007).

The purpose of the present study is to describe a peculiar coastal karstic system, by combining geophysical surveys, dating methods, geological surface field observations and underground submarine exploration. The studied area is currently characterised by a lowland in a topographic depression where a polje filled by more than 50 m of lacustrine sediments was discovered. Submarine karst conduits have been explored by speleo-divers from the sea for more than 3 km of development inland, and their topographic survey suggests that they may be linked with the polje. Geophysical methods are useful for studying heterogeneous karst structure (e.g. Chalikakis et al., 2011; Vadillo et al., 2012). Electric resistivity tomography (ERT) permits the characterization of karstic areas thanks to the contrast in electrical resistivity between resistive low permeability rocks and cavities filled with clay-rich sediments (e.g. Siemens et al., 2010). Gravimetric survey is used to investigate sub-surface cavities because the density contrast is large between karstified rocks and the filled cavity (Butler, 1984; Styles, 1984). The Horizontal-to-Vertical spectral ratio (H/V) method is important for mapping the thickness of lacustrine sediments (e.g. Delgado et al., 2000; Guéguen et al., 2007; Méric et al., 2007), with a strong S-wave velocity difference (velocity ratio ≥ 3.5) between the surface layer and deeper material (Rosenblad and Goetz, 2010).

The combined use of different geophysical methods is powerful for characterising karstic structure (e.g. Pueyo-Anchuela et al., 2011; Štěpánčíková et al., 2011), and the comparison of geophysical results with data from drilling allows verification of the interpretations. Moreover, drilled sediments have been dated using complementary absolute and relative dating methods. Identification of reworked nanofossils from sediments allows the determination of the sediment source areas.

### 2. Geological setting

#### 2.1. Geological setting

This study focuses on a small alluvial plain at 86 m altitude in Cassis (Bouches-du-Rhône, Southern France), about half a kilometre wide and 2 km from the Mediterranean Sea (Fig. 1). This depression is currently downstream of a small topographic catchment area of about 8 km², rising to 444 m altitude. The northwestern part of the catchment area is composed of Lower Cretaceous, massive Barremian limestones (Urgonian facies) whilst the southeastern part consists of an Aptian limestone–marl sequence. Structural setting (Debrand-Passard et al., 1984) is characterised by a monoclinal series with a local dip around 8° to the southeast, although specific localized outcrops reach 18° dip (Fig. 1). To the southeast, the catchment area is limited by the Cenomanian sandstone and limestone cliff, delimiting a separate catchment within the Upper Cenomanian marls. A NW–SE orientated fault separates the southern part of the catchment area from an uplifted Barremian limestone ridge, cut by a gorge where the outlet flows to

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**Fig. 1.** Simplified geology and geomorphology of studied area (a) and location of geophysical measurements (b). Relief highlighted following digital terrain model (DTM 5 m, [http://www.crige-paca.org](http://www.crige-paca.org)). The direction of flow from the two tracer injection points, located outside the map, is indicated (see Section 2.2). The trace of the geological section Fig. 8a is indicated on (a) by letters (A, B, C). For further details on sampling site of marls see Table 2.
the sea. Quaternary deposits occur within the depression, which was not identified as a polje until the present study.

### 2.2. Karst

The Cassis area is well-known for its karstic geomorphology and "Calanques" which are canyons cut into Urgonian limestones by fluvio-karstic processes during the glacial period and presently drowned by sea-level rise. Numerous caves exist in the limestone formations (Early Cretaceous in the western part of the studied area, and Late Cretaceous in the eastern part). Major underground karst features are developed below the current sea level. Two submarine conduits have been explored by speleo-divers: the Port-Miou and Bestouan ‘rivers’ (Arfib and Douchet, 2011). The Bestouan river, ending at the Cassis port entrance, shows a sub-horizontal conduit, which runs at a depth between 0 and 29 m bsl with a cross-section ranging between 10 and 30 m². Its exploration was stopped after 3 km right on the NW–SE fault which borders the southern boundary of the studied depression (Fig. 1) at sea-level, i.e. about 90 m below the soil surface. Port-Miou shows also a sub-horizontal conduit (depth between 0 and 30 m bsl), about ten times larger than Bestouan, over a distance of 2 km from the shoreline followed by a near vertical plunge exposed down to 223 m bsl.

These submarine rivers drain groundwater, forming submarine karstic springs with perennial discharge, which are the main groundwater outlet of the Beausset Unit between Marseille and Toulon. These springs have been extensively studied as they represent a potential resource of freshwater and can be used as proxies for karst evolution and fresh groundwater resources around the Mediterranean Sea (Audra et al., 2004; Blavoux et al., 2004; Arfib et al., 2007; Fleury et al., 2007). The Port Miou and Bestouan springs belong to the same karstic aquifer and are brackish due to seawater intrusion (Tassy et al., 2010). The seawater intrusion occurs at a depth and several kilometres inland, upstream from the terminal explorations by the divers.

The Port Miou and Bestouan springs drain an aquifer probably larger than 400 km², with still uncertain boundaries. A recent dye tracing test showed a rapid connection (transfer velocity about 1.7 km per day) between a sinkhole and the springs in the carbonate–siliciclastic Turonian and Coniacian rocks located 12 km to the east of the studied area (Arfib and Lamarque, 2012) (Fig. 1). It confirms that the springs drain two reservoirs (Arfib et al., 2006): brackish water coming from the deep part of the aquifer through a chimney-shaft, mixed with fresh water flowing in the upper part of the aquifer through vadose, epiphreatic and shallow phreatic conduits. The deep karstic system around the Mediterranean Sea can be related to the Messinian Salinity Crisis (Audra et al., 2004; Mocochain et al., 2006), a sea-level drop of more than 1500 m acting as the base-level for karst development (Krijisman et al., 1999; Rouchy and Caruso, 2006). Later transgressions and Quaternary sea-level oscillations established a new base-level close to the current sea-level. Two submarine rivers drain groundwater, forming submarine karstic springs with perennial discharge, which are the main groundwater outlet of the Beausset Unit between Marseille and Toulon.

### 3. Materials and methods

#### 3.1. Geophysical methods

A 2-D electric resistivity tomography (ERT) was obtained by employing the ABEM Lund Imaging System using an inverse Wenner–Schlumberger configuration with 64 steel electrodes spaced regularly at 10 or 9 m intervals (Fig. 1b). GPS was used to geo-reference datasets. The apparent resistivity was measured using a computer-controlled multi-channel resistivity metre SAS-4000 (Griffiths et al., 1990). RES2DINV software was used to invert ERT data (Lake and Barker, 1995; Rucker et al., 2011). The validity of the obtained resistivity sections is supported by the low RMS error of about 2% for the 6th iteration.

Gravity data were recorded with a relative micro-gravimeter, model Scintrex CG5, with a 50 × 50 m grid, in order to create a high-resolution gravity map (Fig. 1b). The nominal sensitivity is about 1 μGal. Coefficients of measurement points were determined using Trimble GPS in a differential mode with an accuracy of ±0.01 m. Gravimetric data require various corrections (Bonvalot et al., 2008; Merlet et al., 2008; Jacob et al., 2010) to obtain accurate gravity anomalies only corresponding to underground density anomalies. Earth solid tide and ocean loading correction were calculated using TSoft Software (van Camp and Vauterin, 2005). To predict the ocean loading, the tool http://froste.oso.chalmers.se/loading (Bos and Scherneck, 2011) was applied using the FES2004 oceanic model (Lyard et al., 2006). The drift of the gravity metre (Martín et al., 2011) was corrected by applying a linear approximation to the data between two measurements at the base station every 3–4 h. Free-air and Bouguer gravity corrections were performed following Blakely (1996). Finally, the average relative gravity over the survey area is subtracted from the Bouguer anomaly at each station. The Bouguer anomaly data were interpolated using the kriging method (with smoothing distance of 100 m). Oasis Montaj software (GM-SYS, 2001) allows removal of the regional trend and interpolation of a map of Bouguer anomaly. 2D gravity modelling has been performed using the GM-SYS Profile Modelling Program (Ramadass et al., 2006). The model was constrained with several geological and physical data.

The H/V measurements (Nakamura, 1989) were performed at 38 sites, which were located with a GPS (Fig. 1b). Seismic noise was recorded during at least 30 min with a 100 Hz sampling frequency. Each station was equipped with a CSR 24-bit digitiser connected to a LE-3D/5s seismometer. Data were processed with the Sesarray package (http://www.geopsy.org, e.g. di Giulio et al., 2006; Wathelet et al., 2008). Microtremor records were cut into 60 s time windows, for which Fourier spectra were computed and smoothed using methods described by Konno and Ohmachi (1998). At each site, the H/V spectral ratios were computed for all selected time windows, and the mean H/V curve was determined with standard deviations. The H/V curves were interpreted using SESAME (2004) criteria to obtain the fundamental frequency of each site (fᵢ).

#### 3.2. Sediment sample and analysis

Two sedimentary records have been investigated: a 50 m-deep drilling (CAS) in the Quaternary lacustrine sequence from the polje (43°13″33″N, 5°32″32″E, Fig. 1) and a 50 cm-deep percussion core (BST) coming from the Bestouan river. The CAS drilling was undertaken by CEREGE and the ‘Centre de Carottage et de Forage National‘ (C2FN, INSU) using a SEDIDRILL 250–90 drill. Two cores (of 18 and 50 m lengths) were obtained 1 m from each other, with a perfect sedimentary overlap, which allowed us to fill most of the gaps (recovery ratio for lacustrine section: 95%). BST was obtained by divers through inserting a plastic tube into the conduit floor about 50 m from the entrance, at 18 m depth (Fig. 8). CAS and BST drillings were characterised using sedimentological analysis in order to establish relationship between the karstic structure and the submarine spring. Four shallow soundings were also performed by trenching or percussion coring (Cobra system) in four sites around the CAS drilling to confirm the presence of lacustrine deposits.

The global mineralogy of bulk sediment and the mineralogical composition of clay fraction were determined by X-Ray diffractometry (XRD). Samples of 0.5 or 1 g were dried and crushed in an agate mortar. Bulk samples were analysed without pre-treatment. For clay samples, as the signal strength of carbonates masked the signal of clay, carbonate and organic matter were removed by chemical treatments. Then only the fraction finer than 2 μm was kept. Three glass slides were prepared: an untreated control slide, a heated slide (490°C, 4 h) and a slide treated with ethylene glycol. Identification of clay mineralogy is based on changes in the spectral signature after physical or chemical treatment, compared to an untreated control (Moore and Reynolds, 1989).
Marriner et al., 2011). Semi-quantitative estimations (±5–10%; Biscaye, 1965) of the main clay species were based on the height of specific reflections measured on ethylene glycol pattern. Clay and bulk sample XRD patterns were recorded using a Philips PW 1729 X-Ray generator using Kα Co radiation (λ = 1.79 Å) at 40 mA and 40 kV. Data were collected by the step-counting method (step size 0.05°) between 2.5° and 35° (2θ) with a counting time of 5 s per step.

The Ctot, Corg, N and carbonate compositions were determined using a CNS NA 1500 series 2. The standard used for Corg was an Acetanilide powder (C8H9NO, containing 71.09% carbon). Samples of about 30 mg, dried and crushed in an agate mortar, were burnt at 1024 °C. The measurement of degassed CO2 and Nx allows determination of %Ctot and %N. Then samples between 10 and 15 mg were decarbonated with 250 μl of HCl (1 M) at 40 °C over two days. The determination of degassed CO2 provides %Corg. %CaCO3 content was estimated using the following equation: %CaCO3 = (%Ctot − %Corg) * 8.33.

3.3. Dating and paleontological methods

Radiocarbon dating (6 charcoal samples) was performed at the Poznań Radiocarbon Laboratory (Poland) and at the ARTEMIS facility (Saclay, France). Charcoals and wood fragments were identified by B. Talon at IMBE (Aix-en-Provence, France) before dating. Magnetostratigraphic relative dating is based on characteristic features of the magnetic field recorded during sedimentation (polarity chron and excursions; e.g. Demory et al., 2005). Natural Remanent Magnetization (NRM) was measured with a 2G Enterprises DC-SQUID cryogenic magnetometer 2G 760R equipped with on-line alternating field demagnetization. Discrete 8 cm³ cubic samples were used from BST whilst a continuous record was obtained from CAS using U-channels. Direction of characteristic NRM was determined by principal component analysis of the AF demagnetization curve (Kirschvink, 1980). Optically Stimulated Luminescence (OSL) dating was carried out on quartz mineral grains at the Luminescence Dating Laboratory, Research Laboratory for Archaeology and the History of Art, University of Oxford (England). Three samples for OSL dating were collected from half cores which had previously been opened in daylight. Core sections showing a 5–10 cm sandy layer plus a few centimetre of the upper and lower clays were selected for dating. In order to avoid the inclusion of bleached mineral grains, the exposed surface of the cores (circa 1.5 cm) was removed under low intensity filtered LED lighting (emitting at ~588 nm). OSL measurements were performed on coarse sand sized grains (180–255 μm) extracted from the sediment using standard preparation procedures (Mejdahl, 1985). OSL measurements were performed on an automated Risø luminescence reader (Bøtter-Jensen, 1997; Bøtter-Jensen et al., 2000) using a SAR post-IR blue OSL...
measurement protocol (Murray and Wintle, 2000; Banerjee et al., 2001; Wintle and Murray, 2006). Dose rate calculations are based on Aitken (1985). The OSL age estimates include an additional 3% systematic error to account for uncertainties in source calibration. One sample for cosmogenic nuclide dating (26Al/10Be burial method) sample was prepared at CEREGE-LN2C and measured at ASTER, the French accelerator mass spectrometry (AMS) facility (Aix-en-Provence, France). Quartz (~50 g) obtained after decarbonation of the fine sandy fraction (100–400 μm) from a thick sand layer near the core base was then treated following the protocol of sample preparation for 26Al and 10Be measurement (Braucher et al., 2000).

The pollen record (10 samples) was studied in IMBE (France). Because of the poor concentration of sporo-pollinic material, samples were prepared using the method of Takeshi (Nakagawa et al., 1998) using heavy liquid (density = 1.96). Calcareous nanofossils were identified at the University of Lyon (LGLTPE, France) in order to characterise the source of karstic sediment in BST. Samples were prepared as smear slides following the technique described in Bown and Young (1998), and slides were analysed under a Leica optical microscope, with crossed polars, and 1000× magnification. Taxonomic determinations and age assessments were based on Bown (1998).

4. Sedimentary and geophysical results

4.1. Sedimentary record from Cassis and Bestouan

Detailed analysis of the sedimentary sequence is beyond the scope of the present publication. The information relevant to the geomorphological interpretation will thus be briefly summarized. In the polje CAS core the first 4.5 m corresponds to Holocene alluvium, whilst from 4.5 to 50 m finely laminated lacustrine sediment was observed (Fig. 2). At the bottom, the contact with massive Cretaceous marl was sampled over a few centimetre. The upper 5 m of lacustrine deposits was yellowish brown and oxidized because of redox phenomena due to the vertical oscillation of the watertable and pedogenesis occurring since lake filling. Alteration of the sediment prevents the identification of lamina in that upper part. Lamination is (sub)horizontal, except in the last few metres close to the bottom. The up to 15° dip of bottom laminae is probably related to subsidence of the bedrock at the beginning of sedimentation. The lacustrine sediments show considerable homogeneity in colour, texture and composition. Grey sediments are indicative of a reducing environment (sulphides have been identified). The lacustrine sediments are carbonate-rich and poor in organic matter (Table 1). XRD analyses undertaken on bulk samples indicated that global mineralogy is dominated by calcite (CaCO₃), dolomite (Ca, Mg (CO₃)₂), quartz (SiO₂) and clay. XRD of the clay fraction shows the presence of smectite, illite and kaolinite.

The BST core can be divided in four parts. P1, from 0 to 17 cm depth, is composed by laminated and ocher clay. Colour and low CaCO₃ content at the top of P1 (from 3 to 19%) point towards oxidation and decarbonation of the sediment. From 0 to 10 cm (P1a) laminae are only due to changes in texture (clay/sand) whereas from 10 to 17 cm (P1b) laminae are caused by changes in colour (few grey laminae in ocher sediment). P2, from 17 to 23 cm, is composed by laminated clay. Clay is mostly grey with some ocher levels and has a higher percentage of CaCO₃ (from 25 to 35%). From 17 to 20 cm (P2a), clay is dark grey whereas from 20 to 24 cm (P2b) clay is light grey. P3, from 23 to 34 cm, is
Fig. 4. Map of Bouguer anomalies (a) and modelled profiles GRAV1 (b) and GRAV2 (c). The grid cell of the map of Bouguer anomalies is 100 m. 2D models (a, b) were computed using three formations of different density: $d_{\text{host}} = 2690 \text{ kg/m}^3$ (cretaceous bedrock), $d_{\text{sed}} = 2181 \text{ kg/m}^3$ (lacustrine sediments and Holocene deposits), and $d_{\text{mix}} = 2440 \text{ kg/m}^3$ (mixture between sediments and bedrock). Drilling “CAS” (star) and fault (dotted line) are localized on GRAV1.

Fig. 5. H/V results grouped according to calculated interface depth (a). The area with depths above 40 m is highlighted by a ellipse. Two examples of H/V amplitude ratio versus frequency with resonance peak identification are shown (b, c). Drilling “CAS” is represented by a star. Topography is indicated by isolines (m), rivers are in continuous lines and submarine river in dotted line.
composed by ocher clay with numerous angular pebbles and large sand and gravel fractions. P4, from 34 to 49 cm, is composed by ocher clay, with some sandy laminae.

This succession implies that a high energy event was recorded during P3, with a subsequent fine sediment input (P2) of composition and origin different than the previous and later input (P1 and P4). The observed difference (higher carbonate content and reduced environment) indicates rapid sedimentation of a sulphide- and/or organic-rich material.

4.2. ERT survey

Six ERT profiles with NW–SE orientation were measured (Fig. 3). The topographic variations, which were also considered in the inversion model, are quite small (dh = 5 m). According to Samouëlian et al. (2005) the range of resistivity of unaltered and altered limestone is very wide, between 50 and 5000 Ω·m whereas lacustrine sediment and marl have low resistivity (5–70 Ω·m) (e.g. Metwaly and Al-Fouzan, 2013; Nordiana et al., 2013). Apart from very porous or highly fractured saturated limestones, not present in our study, clay-poor slightly fractured limestone resistivity is assumed >100 Ω·m. ERT is unable to discriminate between formations having similar electrical properties, and the resistivities of marl and Quaternary lacustrine sediment may be too close to distinguish the boundary between both formations. ERT data are only able to locate the boundary between Barremian limestone and the overlying marls or Quaternary formations.

2D pseudo-sections are described from north to south. The description is done using the distance from the electrode located at the NW end (designated 0). The NW end of profiles, the massive limestone levels are nearly outcropping and should plunge about 10° below the Aptian marl or any Quaternary filling. The expected top of limestone layer without the impact of karst/dissolution phenomena is indicated by dotted line in Fig. 3. Taking the 100 Ω·m value as the lower limit for limestones, one could see in CAS11 that a limit between low resistivity and high resistivity material can fit with the expected top of limestone

Fig. 6. Comparison between ERT (geometric patterns), gravimetric (map) and H/V (crosses) results. The parts of the ERT profiles represented by a straight line correspond to a gradual transition between the two depth classes which surround it. Drilling "CAS" and five shallow (~ 3 m) additional drilling are located by a star.
layer. In other ERT profiles, limits do not match with expected the limestone tops, thus suggesting fracturing and/or dissolution of the limestones and infilling by Quaternary deposits in karstic structures. Major infillings down to more than 80 m depth are identified between electrodes 300 and 350 m in CAS 8, 13 and possibly 14.

Two distinct parts are visible on profile CAS11, the closest profile to the drilling. There is a deep continuous resistivity layer (from 100 to 440 $\Omega\cdot$m) which is covered by a less resistive layer ($< 50 \, \Omega\cdot$m). The more resistive layer has a dip characteristic of the regional limestone substratum. Locally, the dip is not constant and it sharply increases around the electrode located at 320 m. The limestone is covered by marl and clay (less resistive). The two layers are observed on the CAS08 profile but their morphologies change. The more resistive layer is cut by the lower resistivity layer between electrodes located at 290 and 340 m. It forms a narrow funnel-shaped structure in the substratum which is filled with a low resistivity material. The depth of the structure is greater than 80 m. CAS10 profile shows a general rise of the high-resistivity material and particularly between electrodes located at 270 and 320 m and between electrodes located at 185 and 230 m. On CAS13 and CAS14 profiles, structures and resistivity variations are similar.

In the near-surface layers of Holocene sediments one can see a steady decrease of average resistivity from NE to SW. The decreasing resistivity corresponds to a more water-rich subsurface layer, in agreement with the lower altitude and accumulation of water, as well as thicker Holocene deposits.

4.3. Gravimetric survey

The Bouguer anomaly map (Fig. 4a) shows that after corrections gravity has a small amplitude range of variation (1.2 mGal) from −0.7 to 0.5 mGal. A central band, oriented SW–NE, is highlighted by a negative anomaly which indicates the presence of a low density material, corresponding to the lacustrine sediments. This negative anomaly is elongated SW–NE and is about 400 m long and 200 m wide. It is flanked to the east and west by two positive anomalies, indicating the presence of high density material near the surface corresponding to the Cretaceous bedrock (marl and massive limestone).

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4.4. H/V data

Quaternary sediments and their Cretaceous marls and limestones substratum are characterised by S-wave velocity of 200–600 m/s (Bièvre et al., 2011) and 2000–3500 m/s (Dvorkin et al., 2001; Fournier et al., 2011), respectively. This velocity contrast is large enough to provide a significant H/V peak (Rosenblad and Goetz, 2010). Treatment of H/V curves can integrate all horizontal components or use a specific azimuth in order to minimize 2D effects. Such effects are linked to surface and interface topographies. The NE–SW azimuth, which

![Diagram](a)

**Fig. 8.** Interpretative geological section (a) and chronological evolution of karstic system (b). The explored conduit of Bestouan is indicated by a continuous line in the explored part and a dotted line in the hypothetical upstream. Location of geological section is shown in Fig. 1a by letters (A, B, C). Geological legend is n4: Barremian (Urgonian), n5-6: Aptian and Fy-E: Quaternary alluvium. Location of cinematic section is shown by the rectangle. Drilling “CAS” is represented by a star and drilling “BST” by a triangle.

### Table 1

<table>
<thead>
<tr>
<th>ID</th>
<th>Depth (cm)</th>
<th>Level</th>
<th>Smectite (%)</th>
<th>Illite (%)</th>
<th>Kaolinite (%)</th>
<th>Chlorite (%)</th>
<th>Water (%)</th>
<th>%C tot</th>
<th>%Corg</th>
<th>%CaCO3</th>
</tr>
</thead>
<tbody>
<tr>
<td>BST 1</td>
<td>3</td>
<td>P1</td>
<td>33</td>
<td>45</td>
<td>11</td>
<td>11</td>
<td>40</td>
<td>0.46</td>
<td>0.14</td>
<td>2.64</td>
</tr>
<tr>
<td>BST 2</td>
<td>11</td>
<td>P1</td>
<td>59</td>
<td>30</td>
<td>5</td>
<td>6</td>
<td>46</td>
<td>2.84</td>
<td>0.53</td>
<td>19.17</td>
</tr>
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<td>BST 3</td>
<td>17</td>
<td>P2</td>
<td>32</td>
<td>46</td>
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<td>11</td>
<td>55</td>
<td>3.67</td>
<td>0.71</td>
<td>26.32</td>
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<tr>
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<td>20.5</td>
<td>P2</td>
<td>48</td>
<td>35</td>
<td>11</td>
<td>6</td>
<td>38</td>
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<td>33.44</td>
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<tr>
<td>BST 5</td>
<td>41.5</td>
<td>P4</td>
<td>20</td>
<td>64</td>
<td>16</td>
<td>0</td>
<td>28</td>
<td>2.00</td>
<td>0.13</td>
<td>15.64</td>
</tr>
<tr>
<td>BST 6</td>
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<td>P4</td>
<td>44</td>
<td>36</td>
<td>8</td>
<td>12</td>
<td>27</td>
<td>1.95</td>
<td>0.09</td>
<td>15.46</td>
</tr>
<tr>
<td>CAS</td>
<td>Mean</td>
<td></td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>α</td>
<td>–</td>
<td>70</td>
<td>20</td>
<td>10</td>
<td>0</td>
<td>0</td>
<td>21</td>
<td>6.17</td>
<td>0.49</td>
<td>47.26</td>
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<tr>
<td>Max</td>
<td>–</td>
<td>90</td>
<td>33</td>
<td>22</td>
<td>0</td>
<td>0</td>
<td>26</td>
<td>7.80</td>
<td>0.70</td>
<td>62.00</td>
</tr>
<tr>
<td>Min</td>
<td>–</td>
<td>46</td>
<td>6</td>
<td>4</td>
<td>0</td>
<td>0</td>
<td>18</td>
<td>4.38</td>
<td>0.30</td>
<td>38.00</td>
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<tr>
<td>Mf</td>
<td>–</td>
<td>29</td>
<td>38</td>
<td>33</td>
<td>0</td>
<td>&lt;1%</td>
<td>9.04</td>
<td>0.32</td>
<td>72.60</td>
<td></td>
</tr>
<tr>
<td>Ma</td>
<td>–</td>
<td>76</td>
<td>13</td>
<td>11</td>
<td>0</td>
<td>0</td>
<td>22</td>
<td>7.62</td>
<td>0.22</td>
<td>61.64</td>
</tr>
</tbody>
</table>

Clay percentages are normalised at 100%; CAS data are reported as a mean of 15 samples with standard deviation (σ). For more explanation about “level” refer to Section 4.1. Sampling location of fresh marls (Mf) and material resulting from the weathering of marls (Ma) are located in Fig. 1a.
corresponds to the direction of the valley and elongation of gravimetric data, was chosen for data processing. Data are presented in Fig. 5. The peak values of H/V ratio range between 2.2 and 21.96. Eight sites (over a total of 38) were discarded after treatment, either because they yielded no well-defined peak or because the geometry of the peak did not fulfil the SESAME criteria (width and height). The frequencies obtained vary between 2 and 32 Hz. Most data are below 5 Hz (44%) or between 13 and 24 Hz (48%), with a single site showing a higher frequency. The period of the day during which measurements are performed systematically decreases with depth. The period of the day during which measurements are performed varies between 2 and 32 Hz. The frequency (Vs) is deduced with the aim of controlling the validity of peaks measured during the day. S4N is close to the CAS14 drilling, and allowed determination of the H/V depth of 9 m determined on the CAS14 profile. The frequency (Vs) is deduced from peaks at 2.9 and 2.9 Hz, respectively.

S4N is close to the CAS drilling, and allowed determination of the velocity of the surface layer. At this point, the depth of the bedrock (h) is 50 m and frequency $f_h$ is 2.90 Hz. The velocity (Vs) is deduced with the following formula: $f_h = V_s / (4 \times h)$. So, the calculated velocity Vs is 581 m/s. This value is in accordance with the known velocities of the following formula: $f_0 = V_s / (4 \times h)$, where $f_0$ is 50 m and frequency $f_0$ is 2.90 Hz. The velocity (Vs) is deduced with the aim of controlling the validity of peaks measured during the day. S4N is close to the CAS14 drilling, and allowed determination of the H/V depth of 9 m determined on the CAS14 profile. The frequency (Vs) is deduced from peaks at 2.9 and 2.9 Hz, respectively.

The aim of the study was to delineate the morphology of a buried karstic structure and understand the mechanisms of its formation and connection with a submarine karst network. Comparing the results from three geophysical methods based on different properties improves the interpretation of the data. None of the three methods used alone can clearly determine the boundaries between the three formations present: Quaternary sediments, limestones and Cretaceous marl bedrock. Due to similar electrical resistivities, marl and clay are not distinguished on ERT profiles. Gravimetric study does not differentiate between marl and limestone because of similar densities. The H/V method, like gravimetry, gives the depth of bedrock and lacustrine sediment boundary. The area covered by the ERT survey is smaller than that covered by gravimetric survey and H/V methods, but the comparison of the results shows that the three methods give coherent results. The results of the ERT profiles and Bouguer anomaly map overlap well, qualitatively (Fig. 6). However, the ERT depth is usually greater than the modelled depth from gravimetry, especially towards the NE. This may be attributed to marls and suggests that the thickness of marl increases towards NE, moving away from the fault. Furthermore, it is possible to interpolate the theoretical depth of the surface contact between marl and limestone, not taking into account karstic structures. With the hypothesis of a clear interface between marl and limestone and a dip of $-8\times10^8\;E$ (Fig. 3), the depth of marl/limestone boundary observed is about 45–57 m below the ground surface within the polje area (at drilling CAS).

In the southwest, the data from the three methods agree well within uncertainties. All H/V-derived depths greater than 30 m are concentrated in the deeper area indicated by gravimetry, with the largest values fitting well with the gravity minimum. The only discrepancy concerns a H/V depth of 9 m determined on the CAS14 profile whilst gravity and ERT indicate a deep interface. This may be due to the presence of a discontinuity within the Quaternary sequence or unsuitable H/V spectra.

Geophysical observations can be used to determine the general shape of the bedrock of the paleolake, and so to give an estimation of the volume of sediment required to fill the polje. Based upon gravimetric modelling, the shape of the polje can be approximated using simple geometric forms: 1) a narrow vertical structure, representing a collapsed sinkhole or any alternative way to connect the surface to the karstic network at sea level within the massive limestone; and 2) a larger volume, representing the extension of the paleolake related to the erosion of marl through the sinkhole. The general shape of the karstic structure is a similar to a large funnel (Fig. 8), deeper at the centre than at the edges (Fig. 6) and filled by lacustrine sediments. The polje dimensions are about 340 m long, 300 m wide and 40 m average depth. Using these dimensions the volume of sediment filling the polje is about $4 \times 10^8\;m^3$ (neglecting the central sinkhole volume). Although this simplified calculation minimizes the complexity of the structure of karst depression, it should provide a correct order of magnitude. Moreover, geophysical studies and sedimentary observations give information about paleolake water depth. Lacustrine sediments contain abundant Cretaceous microfossils but are very poor in Quaternary...
organisms such as ostracods or diatoms. There are no traces of bioturbation (except one at 11 m). The lack of diatoms, which are very common in lacustrine sediments, is due to the carbonate-rich nature of the sediment (e.g. Lamb et al., 2007). The grey colour of the sediment is characteristic of anoxic sediment, rich in sulphides (van Geen et al., 2003). The sediment lamination also indicates the lack of currents (e.g. waves-induced currents) which would homogenise the sedimentary deposits and oxygenate the water column, and is concordant with sediment anoxia. The sedimentology thus indicates a deep depositional environment (e.g. Tudryn et al., in press). Based on the near-surface presence of lacustrine sediment in all coring points, the surface of the paleolake was once at a present altitude of at least 90 m a.s.l (altitude of lacustrine sediment cored in the most northward point), thus constraining the altitude of the paleo-outlet.

5.2. Sedimentary record chronology

Samples for radiocarbon dating obtained from the lacustrine sediments (Table 2a) are all beyond the limit of the technique, about 49 ky (Donahue, 1995). One sample (RC23 at 23.13 m depth) provides a date of 48 ± 3 ky which could be due to minor, more recent 14C contamination. Therefore, the entire period of lacustrine sedimentation, and moreover karstic structure formation, can reasonably be assumed to be older than 50 ky. Continuous paleomagnetic investigations (Romey, 2013) reveals normal polarity throughout the sequence, indicating a Brunhes age (~780 ka) without any record of geomagnetic excursion. Three OSL dates obtained from 34.04, 41.45 and 48.40 m depth provided age estimates of 321 ± 29 ky (MIS 9–10), 353 ± 37 ky (MIS 9–10) and 209 ± 21 ky (MIS 6–7), respectively (Table 2b). Optical dating is used to determine burial ages of sediments that were exposed to sunlight prior to deposition. However, OSL dates can be biased by many factors (Bradley, 1999) and one of the main limitations is the possibility of insufficient resetting of the luminescence signal. Partially bleached sediment may contain a mixture of grains carrying different levels of residual signals, and this can lead to an age overestimation (Böse et al., 2012). In this study, the small size of the watershed (~8 km²) would imply a short transport time which could reduce the likelihood of a full resetting of the quartz OSL signal at deposition. The associated low 26Al/10Be ratio (1.58 ± 0.26) measured in a sample of quartz from a depth of 46 m suggests a minimum burial age of 2.9 My, which is inconsistent with other data within the core. This implies that the quartz grains have been buried in a Tertiary layer, rapidly eroded and redeposited in the lake, i.e. without enough time to reset the cosmogenic nuclides signals. These findings point towards a more complex depositional scenario, involving multiple surface exposures and burial sequences. The OSL dates are likely to represent age overestimates (hypothesis supported by catchment area dimension, overestimation of 26Al/10Be dating and no stratigraphic order of the OSL ages) but when considered in combination with the 14C dating suggest that lacustrine sedimentation is likely to have taken place between 50 (lower limit of 14C dating) and 350 ky (upper limit of OSL dating).

Concerning pollen analysis (Fig. 7), the presence of Artemisia (up to 28.5%), other steppe herbs, pioneer shrubs and the dominance of Pinus (up to 80%) in the pollen assemblages are typical of glacial and late-glacial vegetation from the Upper to Middle Pleistocene periods in Western Provence (Triat-Laval, 1978). No Tertiary exotic taxa were observed in pollen slides. As the extinction of those taxa dates to the end of MIS 11 in Western Europe, the lacustrine infilling of Cassis is younger than MIS 11 (ca 400 ka) and dates from Middle to Upper Pleistocene.

The different elements of dating indicate an age (1) greater than 50 ky, (2) less than 400 ky, and (3) from a cold period. Moreover, due to the sedimentary homogeneity with the absence of discontinuities, lacustrine sediments probably correspond to a unique continuous cold period. The intersection of these data can highlight the corresponding periods. MIS 6 (~130 to 190 ky), MIS 8 (~244 to 301 ky) and MIS 10 (~334 to 364 ky) can match these criteria (Bassinot et al., 1994; Lisiecki and Raymo, 2005). Therefore, the sedimentary sequence of Cassis would represent a maximum period of deposition representing circa 30 to 60 ky.

Moreover, the analysis of calcareous nanofossils (Table 3) shows three main ages for the sediment source: Cenomanian (~93–99 My), Aptian (~112–125 My) and Valanginian–Hauterivian (~130–140 My), with a dominant contribution of the Aptian. So the carbonate sedimentation is mostly the result of the erosion of Aptian marl coming from the eastern part of the catchment area. However, Aptian marls may not have supplied the sandy levels. The presence of Cenomanian fossils could confirm that sandy levels come from the erosion of the Cenomanian lithologies located to the east of the study area. The outcrop area of Aptian marls in the catchment area is about ~4 km² (Fig. 1) and the volume of the lacustrine sediment is about 4 × 10⁶ m³. So one needs to erode on average about 1 m of Aptian marls to totally fill the karst depression (assuming total deposition of eroded material).

The maximum duration of sedimentation suggested here (30 to 60 ky) would provide minimum denudation rates ranging from 17 to 33 m/My. A comparable denudation rate of 37 m/My was obtained on the nearby Cap Canaille cliff summit, which shares the same climate and slope as the catchment area (Braucher et al., 2011). Cap Canaille lithology being a hard conglomerate, one may expect a greater denudation rate in the Aptian marls, i.e. a sedimentation duration below 30 ky. This also agrees with studies in mountainous areas that showed that it is

### Table 3

<table>
<thead>
<tr>
<th>Sample</th>
<th>Index species</th>
<th>Known range</th>
<th>Inferred probable age</th>
<th>Associated species</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Brunodosphaera bigelowi</td>
<td>Cenomanian-present</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>Nanmoons tusuki</td>
<td>Late Aptian-Campanian</td>
<td>Aptian-Albian</td>
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<tr>
<td>N. quadriangularis</td>
<td>Late Aptian-Albian</td>
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<td></td>
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<tr>
<td>Lithaphides carniolensis</td>
<td>Late Berriasian-Maastrichtian</td>
<td>Valanginian-Hauterivian</td>
<td>Valanginian-Hauterivian</td>
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<tr>
<td>Eiffithus striatus</td>
<td>Late Valanginian-Late Hauterivian</td>
<td>Early Valangian-Late Barremian</td>
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<tr>
<td>Calcidacithus abollgata</td>
<td>Early Valangian</td>
<td>Early Barremian</td>
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<td></td>
</tr>
<tr>
<td>Tubidicus vereneae</td>
<td>Late Berriasian</td>
<td>Late Hauterivian</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BST 2</td>
<td>Kokia curvata</td>
<td>Valanginian</td>
<td>Valanginian</td>
<td>L. carniolensis, B. constans, various species of Watznaueria, various species of Zechstein,</td>
<td>Dominant contribution of upper Cretaceous sediments</td>
</tr>
<tr>
<td></td>
<td>Exophyllum octopus</td>
<td>Turonian</td>
<td>Turonian</td>
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</tr>
<tr>
<td></td>
<td>Biscutum magnum?</td>
<td>Santonian-Maastrichtian</td>
<td>Santonian-Maastrichtian</td>
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<tr>
<td>BST 4</td>
<td>Prediscophora poncaica</td>
<td>Albain-Maastrichtian</td>
<td>Cenomanian-Maastrichtian</td>
<td>C. conicus, R. infinitus, various species of Watznaueria</td>
<td>Watznaueria species are quite dominant</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BST 6</td>
<td>(P1)</td>
<td>Albain-Maastrichtian</td>
<td>Cenomanian-Maastrichtian</td>
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<tr>
<td>(P2)</td>
<td></td>
<td></td>
<td></td>
<td>W. britannica, W. manitivi,</td>
<td></td>
</tr>
<tr>
<td>(P3)</td>
<td></td>
<td></td>
<td></td>
<td>W. prheure</td>
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</tr>
</tbody>
</table>

The location of samples is shown in Fig. 2. 38 species were identified in CAS samples, which show a perfect overlap of species, and 22 species in BST samples. Levels (see Fig. 2), in which BST samples were collected, are shown within brackets. "P" indicates an uncertain identification.
5.3. Polje formation and link to Bestouan river

The paleolake is located above karstified, fractured and faulted limestones (Fig. 8b (T0)). This context is very favourable for large cavity formation. The funnel shape of the subsurface karst as inferred from gravity data strongly suggests that the paleolake formation was controlled by two factors: the erosion of marls by surface flow and the formation of a karstic cavity in limestones, located at the centre of the depression (Fig. 8b (T1)). The karstic cavity was formed by the dissolution of limestones by water flow. The progression of ongoing limestone dissolution led to the collapse of the cavity or the creation of a vertical conduit, which created a karst sinkhole establishing a connection between the surface and the Bestouan conduit (Figs. 1, 8b (T2)). Constant erosion of marls and drainage into the sinkhole with subsequent transport towards the now submerged outlet, combined with the high clay content of marls, caused a gradual clogging of the depression. Connection between surface and underground karstic network was thus interrupted (Fig. 8b (T3)). Once the cavity clogged, the depression created can operate as a lake and gradually be filled by sediments from the catchment (mostly Aiptian marls) (Fig. 8b (T4)). The sequence of events described above is in agreement with other results obtained from studies of karst lakes (e.g. Kindinger et al., 1999; Salvati and Sasowsky, 2002). Kindinger et al. (1999) detailed four progressive developmental stages of karstic lake formation: 1) the collapse of karst (active phase), 2) the filling of the cavity by sediments and the subsidence of the lake bottom (transitional phase), 3) the filling of the depression and 4) a phase of sinkhole development, associated with structural features (not observed in Cassis paleolake). Another formation hypothesis may be suggested by the fault located south of the polje. The fault brought into contact compact limestone and impermeable marls. This preferential drainage area could allow the establishment of a sinkhole, as commonly observed in karst area. However, the centre of the depression (Figs. 4, 6) is about 100 m to the north of the fault and thus has no close relationship with the fault. If there had been a karstic sinkhole close to the fault, sediment thickness should be maximal, whilst geophysical measurements give a thickness of about 10 m. Because of this configuration, it is unlikely that the fault is related to the formation of the polje. Even if the fault is not related to the formation of the polje, it could create a preferential drainage area that could have functioned as a minor outlet of the polje. The high homogeneity of sedimentary facies and the absence of unconformities and large deformation indicate that subsidence effects did not significantly affect the lacustrine sedimentation. The presence in the sedimentary record of a few microfaunas (displacement less than a few centimetre) and a significant dip at the base could correspond to differential compaction. Moreover, from 45 to 50 m depth they are layers containing Aiptian marls gravel from 10 to 20 cm thickness and abundant charcoal, whereas from 0 to 45 m depth, the core is poor in charcoal and devoid of gravels. The presence of marl pebbles and charcoal could be related to the instability of the slopes of the polje after its formation.

Concerning the BST core, there is no way to obtain chronological constraints, apart from the consistent normal polarity of paleomagnetic signal (Fig. 9), supporting an age younger than 780 ka.

**5.3. Polje formation and link to Bestouan river**

Fig. 9. Bestouan magnetic data. Orthogonal projection plots (a) of stepwise alternating field (AF) demagnetization data of Bestouan samples (location of samples is shown in Fig. 2). Open and solid symbols represent projections on two perpendicular planes whose intersection is the horizontal axis. The samples show a stable vertical component above 10–20 mT, converging to the origin. AF demagnetization steps are given in mT. NRM means Natural Remanent Magnetization. Stereoplot (b) shows a stable inclination for all samples (solid points), ranging from 30° to 65°. Circles indicate MAD (maximum angular deviation) values. The variation in declination (from 45° to 140°) is not significant since the core is not oriented and is likely to have been twisted on the gravel layer.
correspond to levels P1 and P2, respectively. Given the absence of dating, it is not possible to be absolutely sure of the correlation between levels A–B and P1–P2. However, if a correlation between levels A–B and P1–P2 is assumed, the events described above are consistent with the sedimentation of BE1. In addition, the study of pollen indicates that lacustrine sedimentation took place during a cold period, i.e. a sea level lower than present (Antonioli et al., 2004), about 60 to 120 m bsl. Bestouan sedimentation of P3 layer is estimated to shortly predate lacustrine sedimentation, thus likely occurring also in a cold period. The Bestouan conduit is located today approximately between 10 and 30 bsl and was thus emerged during lacustrine sedimentation. Karstic sedimentation occurs in flooded caves, implying that the karst conduit acted as a perched sump in the low permeability carbonate matrix of cretaceous limestones, fed by the polje outlet and further karst drainage.

6. Conclusions

Multi-method geophysical investigations allow the delineation of the shape of a polje filled by lacustrine sediments, in Cassis (SE France) in a coastal carbonate environment. A few metres of Holocene deposits, referred as an alluvial plain at about 90 m above sea level, hide a huge infilling in a karstic depression. The karstic features of the region are ubiquitous, such as the Calanques karstic landforms, or a large polje (Cuges) 12 km north of the studied area, but in the studied area of Cassis only submarine karstic springs were known. The combination of several dating methods, albeit of limited efficacy, and paleoenvironmental proxies indicates that the polje was filled during a glacial period (MIS 6, 8 or 10, i.e. in the 130–360 ka range). The 3 km-long conduit of Bestouan already existed, acting as an underground bypass for surface flow through a sinkhole before the filling of the polje. The Messinian Salinity Crisis is commonly taken into account to explain the development of karst at a great depth above the current sea level around the Mediterranean sea. However, Quaternary karstic evolution is highlighted in this paper and shows that karst hydrogeological functioning or geotechnical issues must account for the successive stages of karst development or infilling, hidden under coastal plains.

Acknowledgements

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