Evidence of the Matuyama-Brunhes transition in cave sediment in Central Europe

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Abstract

In this study, we offer significant improvement over previous results that identified the Matuyama-Brunhes magnetic reversal in cave sediments from the Czech Republic in Central Europe. We collected discrete samples from the sedimentary profile in Za Hajovnou cave located in the eastern part of the Czech Republic. The rock magnetism measurements indicated that the magnetic carrier of most of the samples is maghemite. Characteristic remanent magnetization (ChRM) directions and related virtual geomagnetic pole (VGP) paths indicated that the Matuyama-Brunhes transition boundary was within 5.7 cm of the sediment, located in the upper part of the sampled sedimentary section. This result showed a new, more detailed behavior of the polarity transition from that of the central European location. The migration of the paleopole between eastern Africa and western North America was established as a significant marker for the central European paleomagnetic record in terms of global magnetic data. The transition duration was 8.1±0.2 kyr, and the precursor of the reversal occurred 4±0.2 kyr before the transition. In addition, we estimated the sedimentation rate of the studied section (~35 cm) in the cave as 0.7±0.2 cm/kyr.

Keywords: Paleomagnetism; Matuyama-Brunhes; Magnetic reversal; Cave; Sediments

1. Introduction

The Matuyama-Brunhes magnetic reversal occurred approximately 773 kyr ago (Cohen and Gibbard, 2019), as several recent studies have shown (Channell et al., 2010 (773 ± 0.4 ka); Suganuma et al., 2015 (770.2 ± 7.3 ka); Singer et al., 2019 (773 ± 2 ka); Valet et al., 2019 (772.4 ± 6.6 ka); Haneda et al., 2020 (772.9 ± 5.4 ka)). This event was well recorded by sediments that had sufficient sedimentation rates and could be analyzed in detail by paleomagnetism (Channel et al., 2010; Sagnotti et al., 2010, 2014; Suganuma et al., 2010; Jin and Liu, 2011; Giaccio et al., 2013; Kitaba et al., 2013; Pares et al., 2013; Valet et al., 2014; Liu et al., 2016; Okada et al., 2017; Bella et al., 2019).

Sediments acquire remanent magnetization during their deposition. The alignment of magnetic moments of the grains occurs in the direction of the Earth's magnetic field. The acquisition of primary magnetization due to this sedimentation process is called depositional or detrital remanent magnetization (DRM) (Gubbins and Herrero, 2007). Remanent magnetization protected by potential energy barriers can last over geologic time scales. However, due to thermal and/or chemical processes, such as reheating, oxidation, and iron hydroxide formation, over time, secondary magnetizations can occur by crossing potential energy barriers or by the generation of chemical remanences. The new secondary magnetization has an orientation in the direction of Earth's magnetic field at the time of alteration rather than the time of original deposition. Rocks can acquire viscous remanent magnetization (VRM) a long time after their formation due to exposure to the geomagnetic field. VRM contributes to noise in the paleomagnetic data (Butler, 1992; Lanza and Meloni, 2006).

Lock-in-depth affects the nature of the paleomagnetic recording process in sediments. Lock-in-depth is defined as the depth at which the remanent magnetization stabilizes. The position of the lock-in-depth in the sediments is influenced by lithology, the grain-size distribution of the sediment matrix, sedimentation rate, and bioturbation (Bleil and von Dobeneck, 1999; Sagnotti et al., 2005). When assuming a steady sedimentation rate, the lock-in-depth result is a delay in magnetization that corresponds to the time required to accumulate a sediment layer equal to the lock-in-depth. For example, if the sediment has an accumulation speed of 1 mm/kyr and the lock-in depth is 10 mm, the magnetization age is 10 kyr younger than the actual sediment (Sagnotti et al., 2005).

Paleomagnetic analysis of magnetic reversals in cave sediments has been carried out in different locations around the world, including western Europe (Pares et al., 2018), South Africa (Nami et al., 2016), South America (Jaqueto et al., 2016), North America (Stock et al., 2005), southern Europe (Pruner et al., 2010), and eastern Asia (Morinaga et al., 1992). Kadlec et al. (2005, 2014) reported that the central European cave (local name "Za Hajovnou") in the Moravia region of the Czech Republic records the Matuyama-Brunhes transition. The aim of the present study is to analyze the reversal using more detailed paleomagnetic methods and to identify the magnetic carrier of the cave sediment. Here, we obtained a new paleomagnetic dataset from three vertical sediment profiles in this cave. The contribution of the central European paleomagnetic record in cave sediment will be valuable for investigating the characteristic behavior of the Earth's magnetic field during the Matuyama-Brunhes magnetic reversal. Because the sedimentation rate in the cave is not well understood, details of the timing of the transition are not yet known, making our estimation in this study even more crucial.

2. Geologic Setting, Sampling and Analytical Methods

2.1. Geology of the Cave

The Za Hajovnou cave (49° 40´ N, 16° 55´E) is a former sinkhole located in the Javoricko Karst in the Moravia region of the Czech Republic (Lundberg et al., 2014; Musil, 2014) (Fig. 1). The Javoricko Karst was formed from the dissolution of light-gray-colored massive Devonian limestone that overlies Precambrian phyllite (Lundberg et al., 2014; Musil, 2014). The Spranek and Javoricka Rivers flow through the karst. While Za Hajovnou cave is situated on the northwestern bank of the Javoricka River on the southern slope of Pani Hora hill (Lundberg et al., 2014; Žák et al., 2018), both the Spranek and Javoricka watersheds may have contributed to sediment development in this cave (Fig. 1c).

Za Hajovnou cave is an approximately 500-m-long system (Musil, 2005; Bábek et al., 2015). Previously, ~200 m of the cave was explored, and the cave currently consists of two main parallel corridors with slightly different sedimentological records (Musil, 2014; Musil et al., 2014) (Fig. 1d). One corridor (local name is "Excavated Corridor", which used to be a sinkhole entrance) and the other corridor (local name is "Birthday Corridor") have separate entrances, and they are connected by the Connecting Passage Corridor (Fig. 1d). Sediments from the Excavated Corridor continue to the Birthday Corridor and partially fill the Connecting Passage Corridor (Musil et al., 2014) (Fig. 1d).

The upper sediments in the cave (Fig. 2a and d) were dated by U/Th dating of flowstones from 118 ± 1 to 267 ± 3 ka, and the sediment spans the time of the Cromerian Interglacial Complex in northwestern Europe, which begins with the interglacial period of the marine isotope stage (MIS 19; 773 ka; Cohen and Gibbard, 2019) and the Matuyama-Brunhes reversal (Kadlec et al., 2005, 2014; Musil, 2005, 2014; Musil et al., 2014; Lundberg et al., 2014; Bábek et al., 2015; Žák et al., 2018).

The Matuyama-Brunhes boundary was identified (by Kadlec et al., 2005, 2014) in the upper part of the backwater fine sediments deposited from suspension (total thickness up to 4.3 m) in the flooded cave. These sediments underlie mostly nonfluvial deposits that enter the cave through a steep passage and fill the Connecting Passage Corridor (Kadlec et al., 2014; Lundberg et al., 2014; Musil et al., 2014).

The sedimentary sections studied by Kadlec et al. (2005, 2014) in the Excavated Corridor of Za Hajovnou cave were composed of two parts. The first part, which is 0.8 m thick (Section No. 1, in Fig. 1d) and located approximately 28 m from the cave entrance, was interpreted by Kadlec et al. (2014) to contain the Matuyama-Brunhes transition from reversed to normal polarity. This section corresponds to Bed Nos. 4 and 5 (Fig. 2d) in Profile 1 (Fig. 1d) of Kadlec et al. (2005). The second part, which is ~3.3 m thick (Section No. 2 (Profile 2 of Kadlec et al. (2005)), in Fig. 1d), underlies Section No. 1 and corresponds to the lower part in Profile 1 of Kadlec et al. (2005) (Fig. 2d). Kadlec et al. (2014) indicated that this section only contained sediment with reversed polarity. The interpretation of the paleomagnetic data was difficult because the magnetization of the sediments was too weak for the sensitivity of the Agico JR-5A spinner magnetometer. This difficulty was the motivation for the present research. Here, we collected 44 new oriented discrete sedimentary samples from the Excavated Corridor near upper backwater sedimentary Section No. 1 (Fig. 1d and 2).

2.2. Sampling of Section No. 1

The upper part of Section No. 1, from 0 cm to ~12 cm (Fig. 2), which is called Bed No. 1 (Kadlec et al. 2014), is composed of fine backwater sediment of brown clayey silt with white angular clasts of weathered limestone and bone fragments (Bed No. 1) (Kadlec et al., 2014). The lower part of the section, from ~12 to ~35 cm, consists of brown silty clay without white clasts (Bed No. 2) (Kadlec et al., 2014). Although our "Section No. 2" is the same as "Profile 2" of Kadlec et al. (2005) and "Section No. 2" of Kadlec et al. (2014), our "Section No. 1" is not the same as "Profile 1" of Kadlec et al. (2005) but it is the same as "Section No. 1" of Kadlec et al. (2014). The depths in the present study are not the same as those in Kadlec et al. (2005, 2014).

Our sample collection involved the larger sediment part below the transition and the smaller part above the transition according to the position of the reversed polarity in Kadlec et al. (2014). We chose three sets of overlapping boxes (Fig. 2) to completely characterize the transition. The 35.1 cm of exposed sediment at the base of Section No. 1 was flattened to a clean vertical face, and the samples for the paleomagnetism measurements were taken by pushing the plastic boxes (2x2x2 cm; 8 cc) into the sediment (Fig. 2). We used a Brunton geological compass to measure the azimuth and tilt of the boxes. In addition, another 4 samples were collected for rock magnetism measurements (Fig. 2c), which corresponded with the paleomagnetic samples (13_0P, 7_7P, 17_2M, 22_0M).

We examined the sediment structure near the walls of the plastic sediment holder and observed that the process of pushing the box into the sediment created deformation structures along the walls of the boxes. The structure was on the order of 0.05 mm thick. With a box volume of 8000 mm^3 (20x20x20 mm) and a structurally modified layer volume of 80 mm^3 (4x 0.05x20x20 mm), the volume modified by pushing is 1/100 of a fraction of the unmodified volume. Even if this moment would be organized in (e.g., perpendicular direction), it would only deflect the magnetic remanence by $<5\%$.

2.3. Rock Magnetism Measurements

High-temperature magnetic susceptibility measurements (( ) mass normalized) were performed up to 635 °C using an Agico Kappabridge MFK1-FA susceptibility meter to determine the magnetic minerals in the samples. Isothermal remanent magnetization (IRM) acquisition was performed at 25 mT intervals from 0 to 100 mT, 50 mT intervals from 100 to 400 mT, and 100 mT intervals from 400 to 1000 mT using a Magnetic Measurements MMPM10 pulse magnetizer. Stepwise AF demagnetization of IRM was performed at 5 mT intervals from 0 to 40 mT and 10 mT intervals from 40 to 50 mT using an Agico LDA 5 AF demagnetizer. All the remanent magnetizations were measured using an Agico JR-6 spinner magnetometer after each step. The samples for the rock magnetism measurements were chosen according to the paleomagnetic data.

2.4. Demagnetization Measurements

A stepwise alternative field (AF) demagnetization method was applied to clean the secondary magnetizations from the sedimentary samples. The demagnetization method was carried out using a 2G Enterprises Cryogenic Magnetometer on 44 samples divided into 3 different sequences. The first sequence of 17 samples (shown in the leftmost column of Fig. 2c) was demagnetized at 1 mT intervals from 0 to 49 mT and 10 mT intervals from 50 to 100 mT. The second sequence of 14 samples (shown in the middle column of Fig. 2c) was demagnetized at 2 mT intervals from 0 to 48 mT and 10 mT intervals from 50 to 100 mT. The third sequence of 13 samples (shown in the rightmost column of Fig. 2c) was demagnetized at 0.5 mT intervals from 0 to 39.5 mT and 10 mT intervals from 40 to 100 mT. Demagnetization data were interpreted with Remasoft software (Agico Company; Martin Chadima and Frantisek Hrouda).

ChRM directions and maximum angular deviation (MAD) values were determined from principal component analysis (PCA) (Kirschvink, 1980) on the Zijderveld diagram (Zijderveld, 1967). ChRM intensity values show the intensity of the samples at ~15-20 mT AF demagnetization according to the PCA. Virtual geomagnetic pole (VGP) latitudes and longitudes were calculated using PMGSC software (Randy Enkin). Supplementary Table S1 shows paleomagnetic data for each sample. Examples of AF demagnetization results for the Matuyama and Brunhes intervals (two examples each) are shown in Fig. 5. The rest of the samples are shown in Supplementary Table S2 and Supplementary Fig. S3.

3. Results

3.1. Rock Magnetism Results

According to the high-temperature magnetic susceptibility measurements, the transition of maghemite to magnetite can be seen with an increase in susceptibility values at approximately 250-350 °C (Fig. 3). The samples have Curie temperatures between 530 and 550 °C, which may be a sign of titanium in the minerals and a new sulfite phase created from decomposing the maghemite and incorporation of the sulfur from the surrounding clay. The increase in susceptibility values in the cooling curves corresponds to the percentage of maghemite decrease after heating (Fig. 3a, c and d). Two different drops in susceptibility values at 410 °C and 530 °C in Fig. 3c show the existence of maghemite and magnetite together in the sample. The IRM results show that the samples in Fig. 4a-d were saturated at 400-500 mT and demagnetized at 50-60 mT, indicating low coercivity. Samples in Fig. 4e-h are those that could not be AF demagnetized up to 100 mT in section 2.4. These samples did not reach saturation up to 1000 mT and were not demagnetized up to 100 mT, indicating high coercivity (e.g., hematite). Two samples (17_2M and 22_0M) have Curie temperatures of approximately 540° (Fig. 3c and d), which shows the presence of both low- and high-coercivity minerals in these samples.

3.2. Paleomagnetic Results

The samples were generally demagnetized up to 20 mT (for details, see Supplementary Fig. S3), which removed the VRM component, causing a change in the direction of remanent magnetization. This soft VRM component has mean D: 13.8° and I: 56.8° values, which is close to the present day Earth's magnetic field direction for the Czech Republic (D: 4.4° and I: 66.8°) (see Supplementary Fig. S4). Some samples (01_8M, 04_2M, 17_2M, 17_9M, 22_0M) could not be demagnetized to 100 mT.

The intensity of the natural remanent magnetization (NRM) of the samples varies between 8.5×10^{-3} and 34.1×10^{-3} A/m. Median destructive field (MDF) values, where samples lost half of their magnetization, range between 5 and 8 mT. The NRM intensity and MDF values of the samples are shown in Supplementary Fig. S4.

Fig. 6 shows the data in comparison with published studies, which consisted of cave sediments (Bella et al., 2019; Ge et al., 2021; Shaar et al., 2021; Muttoni et al., 2017), marine sediments (Liu et al., 2016; Okada et al., 2017; Valet et al., 2014), and other types of sediments (Giaccio et al., 2013; Sagnotti et al., 2014; Jin and Liu, 2011). The MAD values for the Matuyama and Brunhes sections are between 0.3° and 5.4° (Fig. 6m). These values for the transition section are between 0.7° and 5.3°, which is relatively reliable for detecting the migration of the paleomagnetic vector from reversed to normal polarity (Fig. 6m). The trend of the MAD values in our data increases before and during the transition (shown with dashed lines and dots in Fig. 6m-o) between 23.1 and 7.1 cm depth. This increase can also be seen in other studies, while these values are higher than those in our study (Fig. 6m-o).

Our paleomagnetic data showed inclination values that change by approximately 90° (shown with empty and filled arrows in Fig. 6e-h) from 12.8 to 7.1 cm depth (Fig. 6e). This result revealed the transition nature of the Matuyama-Brunhes magnetic reversal in Za Hajovnou cave. The change can be seen in other datasets from negative to positive inclination (Fig. 6f-h). Between 12.8 and 11.8 cm depth, inclination changes to a positive value (shown with circular arrows in Fig. 6e-h) just before the transition in our data. Additionally, in other studies, the change is larger in other types of sediments (Fig. 6h) than in cave and marine sediments (Fig. 6f and g). Below this depth, the Matuyama section has inclination fluctuations (shown with dashed lines in Fig. 6e-h) between -6.3° and -89.3°. In other datasets (Fig. 6f-h), these fluctuations are seen less frequently in other types of sediments (Fig. 6h). Above the transition, the inclination angle changes between 25.2° and 65.9° for the Brunhes section in our data (Fig. 6 e).

Our declination data show more frequent fluctuations for the whole sediment section (Fig. 6i). The change between 3.0 and 9.2 cm depth (shown with empty and filled arrows in Fig. 6i-l) can be seen with a larger difference in other studies. Below the transition, frequent fluctuations (shown with dashed lines in Fig. 6i-l) with a large declination change (shown with square arrows in Fig. 6i-l) between 25.3 and 23.1 cm depth were observed in other studies. These fluctuations in cave sediments (Fig. 6j) are more frequent than in marine and other types of sediments (Fig. 6k and l).

Despite the fluctuations, the intensity values of ChRM, which can depend on the concentration variation in magnetic carriers of every individual sample, decreased for the Matuyama section from the bottom to the transition between 35.1 and ~15 cm depth in our data (Fig. 6d). After the transition from reversed to normal polarity, these values increased in the Brunhes section between 7.1 and 0 cm depth (Fig. 6d). Even though there are some differences in absolute values due to the changes in the paleomagnetic data depending on the location and sediment type, comparisons of this dataset with other studies showed that fluctuations and frequency of fluctuations in our data are consistent with other datasets and serve as a supporting argument for the Matuyama-Brunhes magnetic reversal in Za Hajovnou cave.

3.3. VGPs and Pole Migration

The VGP shows the position of the geomagnetic paleopole (Lanza and Meloni, 2006). The VGP latitudes from this dataset show fluctuations ranging from -64° to -1° before the transition in the Matuyama section, which are similar to the data from Haneda et al. (2020) with fluctuations from -85° to -32° (Fig. 7a and b). These values indicate a 90° change between 7.1 and 12.8 cm depth (5.7 cm thickness) during the transition due to pole migration (Fig. 7a). A 75° change in VGP values at 11.8 cm depth (Fig. 7a) shows the precursor of the reversal according to Valet et al. (2012) (Fig. 7c). In addition, we plotted the VGP path using VGP latitudes and longitudes based on the ChRM directions of our data (Fig. 8). VGP locations for the Matuyama section are in the Southern Hemisphere (Fig. 8). During the transition from reversed to normal polarity, the magnetic pole fluctuates east of Africa in the Southern Hemisphere and then migrates to the area west of North America in the Northern Hemisphere (Fig 8a). The same occurrence of this migration of the paleopole compares well with the M/B transition section from Okada et al. (2017) recorded in marine sediments near Japan (Fig. 8b). After the geomagnetic transition, paleopoles fluctuate around the geographic north pole (Fig. 8a).

3.4. Estimation of the Sedimentation Rate and Transition Duration

To estimate the sedimentation rate of the studied part (~35 cm) in Section No. 1, we compared the thickness of the transition section in our study (cm) (5.7 cm section from 7.1 cm to 12.8 cm depth) with the duration of the transition in published studies (kyr) from European cave sediments (Pares et al., 2013; Muttoni et al., 2017; Bella et al., 2019; Zupan Hajna et al., 2021; Gibert et al., 2016). Then, the sedimentation rate ranges between 0.5 and 1.1 cm/kyr. Thus, the average sedimentation rate is 0.7±0.2 cm/kyr. The transition duration calculated from the sedimentation rate estimates ranges between 5.2 and 11.4 kyr. The average sedimentation rate shows an 8.1±0.2 kyr transition duration. Comparisons between the sedimentation rate and transition duration estimates and published studies are shown in Table 1.

4. Discussion

Even though some samples could not be demagnetized up to 100 mT, our data show that minerals with low coercivity are responsible for cave sediment magnetization. This finding is supported by rock magnetism results that indicate the behavior of maghemite for most of the samples.

Our data indicate that the Matuyama-Brunhes transition boundary constitutes 5.7 cm between 7.1 and 12.8 cm depth in the sampled sedimentary section of Za Hajovnou cave. The magnetic reversal is characterized and represented by frequent fluctuations in the inclination angle (Fig. 6e) and VGP latitude (Fig. 7a). We think that fluctuations in the declination data indicate instability in the Earth's magnetic field and remanent magnetization. On the other hand, similarities seen in previous studies (Fig. 6e-o) demonstrate the reliability of the data.

The migration of the magnetic North pole from eastern Africa to western North America is a key point for the behavior of the magnetic field during the transition. Although the data in this study and Okada et al. (2017) belong to geographically different locations and sediment types, the similarity during polar migration (Fig. 8) shows that the reversal was a dipole transition, and the non-dipole field component was less significant (Oda et al., 2000; Mochizuki et al., 2011; Simon et al., 2019).

Note that most of the sediment section contains samples from the polarity transition. The data show that the magnetic field was already unstable in reversed polarity for our oldest sample. This observation agrees with Haneda et al. (2020), who showed that the magnetic pole was unstable a long time before the reversal boundary (Fig. 7b), and the magnetic field started to fluctuate almost 20 kyr before the actual transition in their relative paleointensity (RPI) data. We think that our data illustrate the same instability, which is why no paleomagnetic samples have VGP latitudes that deviate less than 25° from the reversed position. We provide a more detailed explanation of the reversed VGP behavior in our data, showing reversed polarity fluctuation well before the actual magnetic reversal.

4.1. Precursor Event

Valet et al. (2012) showed a 90° change in VGP during reversed polarity before the transition (Fig. 7c). According to this model, the precursor prior to the magnetic reversal has a 2.5 kyr duration, and it occurs ~3.5 kyr before the actual transition (midpoint), which has a 1 kyr duration (Fig. 7c). The model showed another 90° change as the rebound with a 2.5 kyr duration after the transition (Fig. 7c). Sagnotti et al. (2014) reported Valet et al.'s (2012) precursor with 140° change in VGP, 0.7 kyr duration, and 5 kyr prior to the actual transition. In our VGP data (Fig. 7a), an ~75° change between 13.6 and 11.8 cm depth shows a 2.6±0.2 kyr duration (according to 0.7±0.2 cm/kyr average sedimentation rate estimation) that can be interpreted as the precursor of the Matuyama-Brunhes magnetic reversal. The pick point of the precursor (11.8 cm depth) is 4±0.2 kyr before the actual transition (9.0 cm depth). The actual transition duration is 0.6±0.2 kyr between 9.2 and 8.8 cm. These values are consistent with Valet et al.'s (2012) model and show the unique behavior of the Earth's magnetic field during the reversal time. The rebound after the transition in the model is not seen in our data since a VGP change between 7.1 and 3.2 cm is not enough to interpret it as the rebound.

4.2. Sediment Deposition, Sedimentation Rate, and Transition Duration

The M/B event occurred during the interglacial period (MIS 19) following the glacial period (MIS 20) (Cohen and Gibbard, 2019). In the case of cave sediments, we see coarser grains at a greater depth (below 12 cm depth; Bed No. 2 in Fig. 2c; Matuyama section) and finer grains at a shallower depth. Since cave sedimentation took place at the time of glaciation, the cave itself was not completely frozen, which means that seasonal variation induced a thaw-freeze cycle that typically generates physical weathering and a source of coarser sediment. This phenomenon is observed in our sediment. The change from glacial to interglacial is supported by finer grain size sediment due to the lesser influence of the thaw-freeze cycle. Therefore, our observations support frozen surfaces and later warming with smaller sediment availability for sedimentation, supporting the transition from glacial to interglacial periods.

According to Lundberg et al. (2014), the cave was filled with water during sedimentation, which was continuously active without any significant color change or hiatus with the exception of a slight change toward finer grains. This provides the continuous magnetic record of the reversal in the cave sediment and allows the sediment to acquire and keep the primary magnetization without the possibility of secondary mineralization. Furthermore, there are no obvious signs of breaks in deposition (e.g., lithological boundaries or desiccation cracks) in the studied sediment section, other than a slight change in grain size.

Our sedimentation rate estimation (0.7±0.2 cm/kyr) seems to be similar to the sediments from other European cave studies (Table 1). While the duration of the M/B transition was reported to last between 4 and 13 kyr (Suganuma et al., 2010; Valet et al., 2014; Okada et al., 2017), the average sedimentation rate of 0.7±0.2 cm/kyr in this study suggests a transition duration of 8.1±0.2 kyr (7.1-12.8 cm transition section) and thus supports the reliability of our sedimentation rate and paleomagnetic record estimates. King and Channell (1991) suggested that large lock-in-depths are associated with interparticle rigidity and strength, characteristic of clayey low accumulation rate sediments (<1 cm/kyr), which results in delays of magnetic acquisition. This shows that magnetic polarity reversal could have a large (25 kyr) apparent age offset between sediments with high and very low accumulation rates (King and Channell, 1991).

5. Conclusions

We compared new paleomagnetic data with the published magnetic reversal record, used the detailed magnetic characteristics of the cave sediment, and inferred the specific magnetic reversal (Matuyama-Brunhes). This comparison is possible due to the nature of the magnetic reversal. Note that the paleopole resided east of Africa and then quickly reappeared west of North America. We consider this occurrence to be an important signature for dating the central European paleomagnetic record from this time period.

The precursor event in our data is a significant anomaly to identify the behavior of the Matuyama-Brunhes magnetic reversal. Additionally, we were able to estimate the accumulation rate of the studied section (~35 cm) in Za Hajovnou cave.

Author contributions

HU and GK designed the research and wrote the paper. JK contributed to the fieldwork, provided detailed information about the cave sediment, and commented significantly.

Data availability

All data are incorporated into the article and its online supplementary material.

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Fig. 1. Location of the study area in a) Central Europe and b) the Czech Republic. Map in c) shows regional details of the Za Hajovnou cave placement (modified after Lundberg et al., 2014; Musil, 2014). d) Map of Za Hajovnou cave (modified after Kadlec et al., 2014; Lundberg et al., 2014; Musil, 2014) (m a.s.l.: meters above sea level).

Fig. 2. Za Hajovnou cave sediments. a) Age diagram of the cave; b) sampled sedimentary Section No. 1; c) discrete samples for the paleomagnetism measurements and the rock magnetism samples (numbers show the sample name); d) the stratigraphic correlation (dashed lines) scheme of Section No. 1 with Profile 1 of Kadlec et al. (2005) and Section No. 2 (Profile 2) of Kadlec et al. (2005, 2014) (modified after Kadlec et al., 2005, 2014; Lundberg et al., 2014). Orange dashed lines show boundaries between Bed Nos. 1 and 2 in Section No. 1 and Bed Nos. 4 and 5 in Profile 1. Currently, all the sedimentary sections in the cave were excavated, except for Section No. 1.

Fig. 3. High-temperature magnetic susceptibility measurement results (: mass normalized magnetic susceptibility, T: temperature in Celsius).

Fig. 4. Results of acquisition (purple dots) and AF demagnetization (black dots) of IRM.

Fig. 5. Changes in magnetization directions on the Zijderveld diagram and Wulf stereonet during the AF demagnetization method and demagnetization curve for typical samples (see Supplementary Figs. S3 for all other samples). a) Normal polarity from the Brunhes section (12_0P, 13_0P); b) reversed polarity from the Matuyama section (08_0M, 21_5M).

Fig. 6. The data in a-d) show the inclination, declination, MAD values, and intensity of ChRM from this study. e-o) Comparisons of inclination, declination, and MAD values from this study with published studies in cave, marine, and other types of sediments. Cave sediments (brownish silty clay): Bella et al. (2019) (Slovakia, 0.6 cm/kyr sedimentation rate), Ge et al. (2021) (China, 0.2 cm/kyr sedimentation rate), Shaar et al. (2021) (South Africa, 0.13 cm/kyr sedimentation rate), Muttoni et al. (2017) (Bulgaria, 1 cm/kyr sedimentation rate). Marine sediments: Liu et al. (2016) (China, 9 cm/kyr sedimentation rate), Okada et al. (2017) (Japan, 61 cm/kyr sedimentation rate), Valet et al. (2014) (Indian Ocean, 5 cm/kyr sedimentation rate). Other types of sediments: Giaccio et al. (2013) (Italy, lacustrine sediments, 26 cm/kyr sedimentation rate),

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Fig. 7. VGP latitudes of a) this study and b) Haneda et al. (2020) (Japan, marine sediments, 89 cm/kyr sedimentation rate). c) the precursor model of Valet et al. (2012).

Fig. 8. VGP path of a) this study and b) VGP path of transition section from Okada et al. (2017) (Japan, marine sediments, 61 cm/kyr sedimentation rate). Dashed lines show the migration of the paleopole from east of Africa to the area west of North America for both studies.

Table 1. Sedimentation rate and transition duration estimation for Za Hajovnou from published studies in European cave sediments.