Magnetotactic bacteria produce chains of magnetite\(^1\)\(^2\) and/or greigite\(^3\)\(^5\) crystals within their cell bodies called magnetosomes that are permanently magnetized\(^7\). They use these magnets to navigate along geomagnetic field lines to reach their preferred habitat\(^7\). Greigite magnetosomes have been well documented in modern sedimentary environments, but their identification in the fossil record remains controversial. Here we use transmission electron microscopy, electron diffraction patterns and rockmagnetic analyses to assess the origins of nanometre-scale greigite crystals found in Pliocene claystones from the Carpathian foredeep of Romania. We find that, like modern magnetosomal greigite grains, the crystals are single domain\(^4\), with few crystallographic defects and an overall shape consistent with an intracellular origin. We suggest these crystals are magnetosomal in origin, which would place them among the oldest greigite magnetofossils identified so far. The crystals also carry a primary magnetic signal, which has remained stable since its acquisition 5.3–2.6 million years ago. We suggest that greigite magnetofossils could therefore provide reliable records of ancient geomagnetic field variations, and that they could also be used as a proxy to assess palaeoenvironmental conditions in low-oxygen sedimentary environments.

Robust fossil evidence for magnetite magnetofossils dates back to the Cretaceous period\(^8\), whereas claims of magnetofossils extend back to the early Proterozoic era\(^9\) (\(~2\) Gyr). Today, these bacteria are cosmopolitan in distribution and easy to identify in modern environments, where they contribute to the biogeochemical cycling of important elements, including iron, nitrogen, sulphur and carbon\(^7\). These crystals exhibit high chemical purity, specific crystal morphologies and exceptionally narrow grain-size distribution\(^8\)\(^11\).

Magnetotactic bacteria achieve directional sensing using magnetosomes, which are membrane-bounded chains of ferrimagnetic crystals. Two magnetic minerals have been unequivocally recognized to be produced by magnetotactic bacteria: magnetite\(^1\)\(^2\) \((\text{Fe}_3\text{O}_4)\) and greigite\(^3\)\(^5\) \((\text{Fe}_2\text{S}_3)\). Magnetosomal minerals dominantly form around the oxic–anoxic interface in aquatic habitats\(^7\), reflecting the palaeo-redox conditions and are of interest for geochemistry and geobiology. They are of importance also for palaeomagnetism because, when preserved, they can significantly contribute to the primary natural remanent magnetization (NRM) of sedimentary rocks.

Magnetosomal magnetite is often well preserved and regularly observed in geological archives (for example, see refs 9,10). Greigite magnetosomes have been uncontestedly identified only in recent soils and lake sediments, although their occurrence has also been claimed in Miocene rocks of the Western Carpathian foredeep\(^12\).

Authigenic greigite has been known since 1964 (ref. 13) and is preserved in rocks at least as old as the Cretaceous\(^14\). However, the reliability of palaeomagnetic data from greigite-bearing rocks is frequently questioned\(^15\) because greigite is thermodynamically metastable\(^16\) and the timing of NRM acquisition by greigite is not well constrained because of its diagenetic formation\(^15\). It was further assumed that greigite would not last long in the geological record because excess sulphur would cause transformation to pyrite\(^16\). A recent re-evaluation of greigite’s thermodynamics\(^17\), however, suggests that, despite its metastability, greigite may preserve a primary NRM for geological times\(^14\).

Recent greigite-based magnetostratigraphies straightforwardly correlate to the geomagnetic polarity timescale\(^18\)\(^19\) and support the formation and preservation of ancient greigite in sedimentary rocks\(^20\). These records come from the Carpathian foredeep of Romania (Fig. 1), which was part of the Eastern Paratethys (see Supplementary Information, Fig. S1), a large semi-isolated fresh to brackish-water domain that comprised the present-day Black Sea and Caspian Sea regions\(^18\). Large quantities of detrital material derived from the uplifted orogen and active volcanic sources were deposited in shallow-water environments with ostracods and molluscs indicative of the euphotic zone\(^21\)\(^22\). Posfai et al.\(^12\) previously conducted a transmission electron microscopy (TEM) study of greigite from one sample from Poland’s Miocene Carpathian foredeep and concluded, solely on the basis of analysis of particle size and shape distributions, that the sediment might contain crystals produced by the multicellular magnetotactic prokaryote.

Thermomagnetic measurements in air, acquired gyro-remanence during alternating field demagnetization (see Supplementary Information, Fig. S2) and scanning electron microscopy showed that greigite is the key magnetic mineral in the sedimentary rocks of the Romanian Carpathian foredeep\(^20\). Positive reversal tests, a positive fold test and the occurrence of inclination shallowing (Fig. 1e,f) provided further evidence for an early acquisition of the NRM\(^20\).
Figure 1 Greigite-based magnetostratigraphy of the Bădislava valley (central plots). Circles and triangles represent high-temperature (HT) and low-temperature (LT) components, respectively. Shaded bands indicate intervals of delayed low-temperature acquisition. a, Thermomagnetic run for greigite-bearing samples. b–d, Demagnetization diagrams illustrating the transition from reversed (d) to normal polarity (b) via a sample (c) recording two antipodal directions. e, f, Histograms of inclinations on using the elongation/inclination (E/I) correction method. Coloured lines refer to expected (Inc\text{GAD}, yellow), original (Inc\text{org}, blue), unflattened (Inc\text{EI}, green) and most frequent bootstrapped (red, with 95% error bounds: dashed red lines) mean inclinations.

An earlier scanning electron microscopy study on the Romanian rocks revealed octahedral greigite crystals of an inorganic origin that range 400–1,000 nm in size. The number of greigite particles, however, was remarkably small when compared with the high initial intensity of the samples (10–65 mA m\(^{-1}\)). Consequently, the presence of an extra magnetic carrier in a smaller (magnetofossil) grain-size range was suspected. To identify the precise nature of greigite formation in these rocks, we used a combination of mineral magnetic methods and TEM. In addition, we evaluated our results using the six criteria for magnetofossil identification of Kopp and Kirschvink: (1) the contextual and palaeomagnetic evidence for a primary origin, (2) the presence of a significant single-domain magnetic phase, (3) size and shape distributions characteristic of magnetosome crystals, (4) evidence for chains of crystals, (5) evidence for chemical purity and (6) high-resolution TEM (HRTEM) evidence for crystallographic perfection. Our earlier palaeomagnetic and palaeoenvironmental research of the Carpathian foredeep fulfilled the first criterion. The other five criteria require an extensive TEM study combined with rockmagnetic experiments. For TEM imaging, we used magnetic
Figure 2 Fossil elongated prismatic greigite magnetosomes. a, TEM micrograph of a prismatic, slightly elongated \{100\} + \{111\} greigite crystal, typical of magnetosomes, very close to the idealized crystal habit (inset). b, HRTEM detail of the greigite crystal. The inset shows well-ordered lattice fringes from the area in the dashed square. c, EDS from the crystal with distinct Fe and S peaks; O, Al and Si are from the background signal of the clay flake; Cu peaks originate from the TEM grid. d, Selected-area electron diffraction pattern with the same orientation as b, b and d were recorded with 41°-tilt difference from a.

concentrates obtained from approximately 20 g of powdered rock samples (see Supplementary Information).

TEM analyses on these extracts identified numerous nanoparticles (Figs 2a, 3a,b and Supplementary Information, Figs S3,S4) that seemed to be chemically pure and have few crystallographic defects, passing two criteria (chemical and crystallographic perfection)\(^2\). The crystals span a grain-size range of 20–75 nm, implying that they are magnetically single domain, although smaller, probably superparamagnetic grains were also distinguished. Many of the particles are irregular in shape, but exhibit strong diffraction (Fig. 3a), indicating that they are highly crystalline. Other particles show elongated (Fig. 2a and Supplementary Information, Fig. S3) and hexagonal (Fig. 3a) outlines when viewed in projection, which is characteristic of elongated prismatic and truncated cuboctahedral grains and points to cubic crystal symmetry. Energy-dispersive X-ray spectroscopy (EDS) analysis shows that most of particles consist of iron and sulphur (Figs 2c and 3c). HRTEM (Fig. 2b) and single-crystal selected-area electron diffraction patterns (Fig. 2d) indicate that the measured \(d\) spacing corresponds to those of greigite. The size distribution of 20–75 nm, the elongated prismatic or cuboctahedral crystal morphologies and the elemental composition all indicate that these greigite crystals have a magnetosomal origin. The roughly prismatic (Fig. 2a and Supplementary Information, Fig. S3) and cuboctahedral (Fig. 3a,b) magnetosomes are magnetically single domain (see Supplementary Information, Fig. S4) and would have been responsible for the magnetotactic reaction of the living organism\(^4\). Isothermal remanent magnetization component analysis revealed a very small dispersion parameter of approximately 0.10–0.15 log units\(^2\), which is also indicative of a magnetosomal origin of the magnetic crystals\(^1\). The single-domain greigite with the grain-size distribution and shape typical of magnetofossils, having truncated-edge crystal morphology, passes two more magnetofossil identification criteria\(^2\).

The last and most difficult to fulfil criterion in rock records is the presence of magnetosomal chains, because both diagenesis and magnetic extraction techniques can contribute to chain disruption\(^\text{24}\). We identified a few single-domain magnetite magnetosomes (see Supplementary Information, Fig. S5), although their contribution to the NRM is minimal because
the samples are demagnetized below 400 °C. We conclude that our greigite magnetofossils pass five out of six criteria for magnetosome identification.

Our Carpathian samples thus comprise two distinctly different types of greigite, generated by two different formation mechanisms (Fig. 1a): (1) small (20–75 nm) slightly prismatic elongated and cuboctahedral crystals of magnetosomal origin and (2) larger octahedral grains (400–1,000 nm) of authigenic origin. Close to polarity reversals, the thermal demagnetization diagrams show the presence of two different (even antipodal) NRM components in a single specimen (Fig. 1b–d). Unexpectedly, the hysteresis loop and the first-order reversal curves (see Supplementary Information, Fig. S6) indicate only a unimodal coercive force distribution, which is unusual for a specimen recording two different directions. We therefore fitted the isothermal remanent magnetization acquisition curve, with two magnetic components having approximately the same mean acquisition field (B_{acq}) but a significantly different dispersion parameter (see Supplementary Information, Fig. S6). This translates into two different grain-size distributions, having the same mean coercivity. Magnetosomes are known for their narrow coercivity switching field distribution (low dispersion parameter) and therefore we tie the narrow dispersion parameter in our samples to the greigite magnetosomes. The wider distribution would thus be related to the authigenic phase of greigite. We acknowledge that authigenic greigite has a narrow grain-size distribution (when compared with other magnetic minerals), but magnetosomal greigite has even lower dispersion parameter values than the authigenic greigite phases. First-order reversal curve analysis of thermally treated samples shows that greigite survives up to 350 °C and is replaced at ~360 °C, by a non-magnetic phase. This behaviour is consistent with the thermal demagnetization spectra of our samples, indicating that greigite was indeed the NRM carrier.

Authigenic or early diagenetic greigite formed later than the magnetosomes, deeper in the sediment and therefore acquired a later magnetic field. The delayed NRM acquisition is easily observable and directionally traceable from the demagnetization diagrams, especially in the intervals that straddle polarity reversals (Fig. 1). The low-temperature component must be attributed to diagenetic greigite because it records the delayed component (Fig. 1c and Supplementary Information, Fig. S7). The high-temperature component represents greigite magnetosomes, which formed close to the sediment–water interface or in the water column at the time of the deposition and thus record the magnetic field without significant delay. Generally, larger grains resist thermal demagnetization and, in the case of greigite, thermal alteration longer. Here, the ~10 times smaller magnetosomes with their high chemical purity and few crystallographic defects seem to persist to the highest temperatures. The larger, diagenetic pseudo-single-domain greigite particles would have a less stable magnetization, explaining partially why the larger fraction is less resistant to thermal demagnetization than the smaller one. The origin of the two components is furthermore confirmed by their different inclination values distinguishable in the thermal demagnetization diagrams. We obtain mean inclinations of 52.0° for the high-temperature (Fig. 1e) and of 60.4° for the low-temperature component (Fig. 1f). This implies that inclination shallowing has significantly affected the high-temperature component (Fig. 1e), caused by dewatering and compaction of the top sediment layer. In contrast, no significant inclination error is observed for the low-temperature component (Fig. 1f), because the authigenic greigite forms later in the sediment, when compaction had largely come to an end. Applying the inclination error correction method using the field model TK03.GAD (ref. 29) on both data sets corrects the high-temperature and low-temperature inclinations to 65.5° and 64.9°, respectively. Both corrected values are remarkably similar and indistinguishable from the expected inclination at the site latitude (I_{GAD} = 63.6°).
Palaeomagnetism is widely used in earth sciences for plate tectonic reconstructions, for dating and correlation of marine and continental sequences and for studying geomagnetic field behaviour. Here, we show that magnetosomal greigite can survive geological times and that a primary NRM component can be extracted, noticeably enhancing the value of greigite for palaeomagnetic studies, including records of rapid geomagnetic variations. We emphasize the importance of small demagnetization steps in the 300–360 °C temperature range, because magnetosomal greigite survives heating up to 350–360 °C, whereas authigenic greigite is removed at 290–300 °C.

In some conditions, greigite magnetosomes may have a greater preservation potential than magnetite magnetosomes because the latter ultimately dissolve under anoxic conditions whereas the former persist—having been formed under such conditions. Greigite magnetofossils might be expected to be more abundant in higher productivity, more sulphidic sediments, but these environments are still insufficiently studied\(^3\). The greigite-producing bacteria prefer reduced conditions and are probably anaerobic sulphate reducers. The high preservation capacity of greigite magnetosomes may help to detect environmental variations, expressed by biogeochemical changes in sedimentary basins. The magnetofossil record may serve as tracers of local changes in oxygen level and provide an underexploited archive of the long-term evolution of marine redox stratification important in characterizing anoxic/euxinic sedimentary environments such as the Oceanic Anoxic Events.

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Author contributions
I.V. initiated the project, undertook the analyses and provided the interpretation. C.F. and J.D.M. assisted and advised on TEM microscopy. C.G.L. assisted with the NRM analyses and the TK03 GAD correction. M.I.D. and W.K. advised and assisted throughout.

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Supplementary Information to accompany

Putative greigite magnetofossils from the Pliocene epoch
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Supplementary Methods: Palaeomagnetism and Rock Magnetism

To assess the magnetic mineralogy, we performed thermomagnetic runs in air with a modified horizontal translation type Curie balance\(^1\) with a sensitivity of approximately \(5 \times 10^{-9} \text{ Am}^2\). Runs during different heating (solid lines) and cooling (dashed lines) cycles were performed at a rate of 10 °C min\(^{-1}\). Approximately 70 mg of powdered sample was put into a quartz glass sample holder held in place by quartz wool. Total magnetization is plotted in a series of runs to increasingly higher temperatures (100, 200, 300, 350, 450, 500 and 700 °C, respectively). Each six seconds a data point was recorded, equivalent to one degree Celsius. The applied field was cycled between 150 and 300 mT. An alternating gradient magnetometer (Princeton Measurements Corporation, MicroMag Model 2900 with 2T magnet, noise level \(2 \times 10^{-9} \text{ Am}^2\)) was used to successively measure hysteresis loops, FORC diagrams, IRM acquisition (after having used the demagnetization option of the MicroMag) and backfield demagnetization curves, all at room temperature. The saturation magnetization \((M_s)\), the saturation remanent magnetization \((M_{rs})\) and coercive force \((B_c)\) were determined from the measured hysteresis curves, after applying the correction for the paramagnetic contribution on a mass-specific basis. To further assess the magnetic domain state, the effects of magnetic interactions and the magnetic mineralogy, FORC diagrams were measured\(^2\). IRM acquisition curves, acquired with the MicroMag System, were decomposed into coercivity components using the fitting program of Kruiver\(^3\), limited to symmetric distributions in the log-space. Palaeomagnetic results have been previously documented for the studied sedimentary rocks\(^4-7\) and extensive rock magnetic investigations were presented elsewhere\(^7\).
Supplementary Methods: Electron Microscopy

For TEM analyses, magnetic extracts were obtained using the extraction procedure described by Dekkers\textsuperscript{8}. The extraction was performed on 20 grams of ground rock samples. The resulting powders were well dispersed by ultrasonic agitation for forty minutes in demineralised argon-purged water. Sodium polyphosphate [Na\textsubscript{4}P\textsubscript{2}O\textsubscript{7}\cdot10H\textsubscript{2}O] was used as a peptising agent to keep the clay mineral particles dispersed. The obtained suspension was put on the extracting column for three hours. The prolonged ultrasonic agitation was necessary because the rock samples were strongly lithified clays. The resulting extracts were washed with demineralised water in three steps using a ‘magnetic finger’\textsuperscript{9} to purify the grains from remaining clay flakes. To further remove adhering clay mineral coatings we used the cleaning procedure described by Franke et al.,\textsuperscript{10}. After a second agitation in the ultrasonic bath for another two minutes to disperse the magnetic concentrate, carbon coated TEM copper grids were dipped into the extract and subsequently dried in air. For efficiency we used a magnetic finger attached to the exterior of the vial that attracted the particles mostly on one side of the copper grid. We took care not to attract too much of the coarser particles. For all TEM analyses, a \textit{FEI Tecnai 20 FEG} transmission electron microscope was used at an acceleration voltage of 200 kV in bright field mode, at high resolution, or for recording electron diffraction patterns. Elemental compositions were identified by using energy dispersive spectroscopy.
Geological setting and sedimentary environment

The samples have been taken from riverbeds where the rock surfaces were freshly cleaned by the stream. The sedimentary sequence consists of an alternation of blue to grey sandstones, siltstones and clays. Our sections start stratigraphically in deposits of late Sarmatian age and end at Romanian (Eastern Paratethys nomenclature\textsuperscript{11}). In the Supplementary Figure 1 we show the uppermost Miocene and Pliocene, greigite-bearing, part of the section. The older (upper Miocene part) has magnetite as a magnetic carrier. The appearance of greigite in Chron C3r (between 6.0 and 5.5 Ma) is most likely related to regional tectonic and/or climatic events that reshaped the basin configuration and consequently changed drastically the palaeoenvironmental conditions. However, the presence of benthic ostracods\textsuperscript{12,13} and molluscs\textsuperscript{13,14} throughout the entire upper Miocene-Pliocene time interval demonstrates that the lowermost water column remained sufficiently oxygenated for these organisms to live. Anoxic conditions favouring greigite formation could therefore only have been present within the sediments, related to degradation of organic matter during rapid burial. Greigite formation in the Romanian Carpathian foredeep palaeoenvironment occurred in a very high sedimentation rate setting (60-150 cm/kyr\textsuperscript{4}), a situation similar to south-western Taiwan\textsuperscript{15-18} and eastern New Zealand\textsuperscript{19,20}. High sedimentation rates accompanied by rapid burial of organic matter will lead to a completely anoxic diagenetic environment close to the sediment/water interface\textsuperscript{21,22}. In such settings, the authigenic greigite in the Eastern Carpathian foredeep likely has formed through bacterial mediation during early diagenesis, up to kyrs after deposition.
Supplementary Figure 1
Palaeogeography of the study area. (a) Schematic palaeogeographic map of the late Miocene/early Pliocene, showing the Paratethys area and the present-day land contribution; (b) map of the study areas modified after Vasilev et al., 2007 with the indication the geographic position of the sections in front of the Romanian Carpathians. The polarity zones with palaeomagnetic data for the studied river sections: Badisla and Rimnicu Sarat. Black (white) denotes normal (reversed) polarity. The indicators for the most important subchrons are displayed (C = Cochiti, S = Sidufjall, N = Nunivak, T = Thvera). The names of the substages are according to the Eastern Paratethys nomenclature\(^1\). Mean latitudes and longitudes (in decimal degrees) are given in the header of the magnetostratigraphy columns.
Supplementary Figure 2.

Gyroremanent magnetisation in greigite. (a) and (b) are tilt corrected Zijderveld diagrams of typical alternating field (AF) demagnetisation behaviour; open (closed) circles are projections on the vertical (horizontal) plane. Initial NRM intensities (Int$_i$) are given. The values represent milliTesla (mT). AF demagnetisation was performed by an in-house developed robot, which let the samples pass-through a 2G Enterprises SQUID magnetometer (noise level 10$^{-12}$ Am$^2$). After 45 mT AF demagnetisation, the samples were increasingly acquiring a gyroremanent magnetisation much larger than the initial NRM. (a) BD 110.1B is the sister sample of BD 110.1A (Fig. 1c); note the remarkable difference between thermal and AF demagnetisation. (b) BD 114.1B is the sister sample of BD 114.1A which has identical thermal demagnetisation behaviour as sample BD 115.1A (Fig. 1b).
Supplementary Figure 3.
**Greigite prismatic magnetofossil.** (a) TEM micrograph of elongated prismatic greigite crystal with well-defined truncated edges. There is non-uniform contrast indicating changing thickness as expected from the outline of the crystal. (b) High magnification of the area indicated in panel a. Three sets of lattice fringes: set 1 corresponds to the (200) plane, sets 2 and 3 correspond to the (111) plane. Set 2 forms angles of 68 ° and 53 ° with sets 3 and 1, respectively. (c) High magnification of the lattice image of the area indicated in panel a. The bands between the lines indicate a d-spacing of 5.7 Å, corresponding to (111) greigite reflection. (d) High magnification of the lattice image from a truncated edge of the crystal.
Supplementary Figure 4. Magnetofossil greigite plotted on the SD stability field of magnetite. (a) The diagram is redrawn after Kopp and Kirschvink, 2007 and is a function of shape factor (width/length ratio) and length. For details see. Shaded regions mark size and shape of crystals from magnetotactic bacteria. The shape fields in the diagram are based upon the specific crystal morphologies produced by living magnetite-producing magnetotactic bacteria, not greigite-producing ones. The elongated prismatic greigite crystals in our samples fit on the cuboidal area of Kopp and Kirschvink. We still interpret them as prismatic because they are different from the cuboctahedral crystals in our samples. In addition, the cuboctahedral crystals are smaller than all existing data so far. The importance of the magnetic interaction at these critical sizes has been shown to be very important. (b) TEM micrograph of a cluster of cuboctahedral greigite crystals; (c) the size distributions of the greigite crystals from panel b. These measurements are subject to bias because, at this size, even with the excellent resolution, it is difficult to obtain the true size values of the crystals. Therefore, we chose to count the particle three times, then stacked these measurements (in total 443) and run the statistics. The mean size value of the particles is 24.61 nm with 3.71 as standard deviation. The histogram shows a clearly skewed (to the left) distribution, which indicates a magnetosomal origin of these crystals.
Supplementary Figure 5.
Fossil magnetite magnetosomes. (a) TEM micrograph of prismatic magnetite crystals. In the upper left part of the image they are aligned in a (disrupted) chain and in the lower right part they are clustered in no preferred order. (b) Energy-dispersive X-ray spectrum from the crystals of the chain in the image showing the distinct peaks for Fe and O of iron oxide. Al and Si peaks are caused by the background signal of the clay mineral flake; the C and Cu peaks originate from the carbon coated copper TEM grid.
Supplementary Figure 6.
Palaeomagnetic and rock magnetic measurements performed on sample RR 122. (a) Zijderveld diagram (after tilt correction) with the direction of the low (blue arrow) and high (red arrow) temperature component. Correspondingly coloured numbers are the temperature steps taken to calculate the directions of the two temperature components. See also caption to figure 1. (b) Thermomagnetic runs during different heating (red solid lines) and cooling (blue dashed lines) show an irreversible decrease in magnetization caused by the greigite decomposition reaction chain, between 200 and 400 °C. A slight increase in the total magnetization is visible at ~500 °C, indicating the formation of new magnetic minerals because of oxidation of an iron sulphide. This temperature is higher than the ones known for alteration of iron sulphides. (c) Normalized decay curve represented as absolute (blue diamonds) and difference vector sum (red circles) values. There are two major inflections in the decay curves arising from the combined effect of greigite alteration and magnetic unblocking. These extrapolated lines (to ~290 °C and ~350 °C) indicate two magnetic carriers, represented by blue and red areas. The hatched area indicates a possibly viscous (e.g. laboratory induced) magnetisation removed up to 100 °C. (d) Hysteresis loop shown between ±300 mT. (e) FORC diagram, indicating (highly interacting) SD particles. (f) IRM acquisition curve, decomposed into coercivity components using a log-normal fitting program. Open blue squares are measured data points. The components are marked with green and purple lines on the linear acquisition plot (LAP) and equivalently coloured shading on the gradient acquisition plot (GAP). Both components have the same coercivity (B1/2), but different dispersion parameters (DP) indicating different grain size distribution. For additional details, see also the supplementary palaeomagnetic and rock magnetic methods section.
Supplementary Figure 7.

Characteristic thermal demagnetization behaviour. On the left hand side of each panel, Zijderveld diagrams are plotted after tilt correction. Blue and red arrows as in caption to Supplementary Fig. 1. On the right hand side are the corresponding normalized decay curves. We show examples from two valleys that generated greigite-based magnetostratigraphy (RR and RM from Rimnicu Sarat and TP from Topolog sections). (a) Sample with different normal HT and LT components showing a decay curve suggesting two main magnetic carriers for the respective components; (b) sample showing dual polarity recorded within the same sample at the transition from normal to reverse. The delayed LT component was acquired during the later reversed field. From Topolog valley: (c) sample TP 22 was magnetized in a reversed magnetic field and has again two distinguishable components; (d) sample TP 12 shows dual polarity recorded within the same sample at the reversal from normal to reversed.
Supplementary References


