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# Analyzing the geomagnetic axial dipole field moment over the historical period from new archeointensity results at Bukhara (Uzbekistan, Central Asia)

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## Abstract

Since the mid-19th century, direct measurements of both intensity and directions of the Earth's magnetic field have been available, allowing an accurate determination of its spatio-temporal variations. Prior to this time, between  $\sim 1600$  and 1840, only direct directional measurements are available. Therefore, the construction of global field models over this period requires either a specific treatment of the axial dipole field component or the use of archeomagnetic intensity data. In this study, we use a regional approach based on the construction of an archeointensity variation curve in Central Asia. We analyze baked clay brick fragments sampled in Bukhara (Uzbekistan),

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dated between the end of the 16th century and the beginning of the 19th century. This city is of particular interest for archeomagnetism due to the well-preserved old buildings accurately dated by documentary archives. A series of archeointensity results is obtained using the Triaxe experimental protocol, which shows a decreasing trend in intensity from  $\sim 1600$  to  $\sim 1750$ , with intensities during the 18th century lower than expected from global geomagnetic field models. These new data appear consistent with other Triaxe data previously obtained in western Europe and western Russia, when transferred to Bukhara using the field geometry of the *gufm1* model. Together, these data are used to recalibrate the axial dipole moment evolution provided by this model. The resulting evolution appears non-linear, with a clear relative minimum in the magnitude of the axial dipole during the late 18th century. We illustrate the fact that at present this evolution can neither be satisfactorily confirmed nor refuted by other datasets available in western Eurasia (as well as at a wider spatial scale), mainly due to the significant dispersion of the data. Our interpretation relies on the accuracy of the field geometry of the *gufm1* model, which appears less reliable prior to  $\sim 1750$ . Nevertheless, the minimum proposed in the 18th century seems to be a true feature of axial dipole behavior.

*Keywords:* Archeomagnetism, Secular variation, Geomagnetic field intensity, Geomagnetic field modeling, Axial dipole moment

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## 1. Introduction

Variations of Earth's magnetic dipole cover a wide range of timescales from a year or less to tens of millions of years. Three different frequency bands

4 are evidenced by analyses of the dipole power spectrum from paleo- and geo-  
5 magnetic data and simulations (Constable and Johnson, 2005; Ziegler et al.,  
6 2011; Olson et al., 2012; Panovska et al., 2013; Bouligand et al., 2016; Lesur  
7 et al., 2018): an ultra-low to low frequency band (UF), a transitional fre-  
8 quency band (TF), and a high frequency band (HF). The UF band comprises  
9 chrons and superchrons and is associated with the thermal evolution of the  
10 outer core. The TF band covers paleo-/archeomagnetic secular variations and  
11 is associated with geodynamo processes. Finally, the HF band contains the  
12 shortest periodicities of the axial dipole’s variations (as observed from satel-  
13 lite data). These bands are separated by two cut-off frequencies  $T_s$  (between  
14 UF and TF) and  $T_f$  (between TF and HF), estimated by Hellio and Gillet  
15 (2018) from recent field statistics as  $T_s = 100$  kyr and  $T_f = 60$  yr, for the pur-  
16 pose of constructing the COV-ARCH model (more on global models below).  
17 The axial dipole’s power spectrum from numerical dynamo simulations cor-  
18 roborates these results (Olson et al., 2012; Bouligand et al., 2016), although  
19 the estimated characteristic timescale  $T_f$  is longer ( $T_f \sim 10^2 - 10^3$  yrs), which  
20 is probably associated with the convective timescale in the outer core of or-  
21 der 150 yr. While secular variations recovered from global archeomagnetic  
22 models are representative of the low-frequency TF band, regional variation  
23 curves spanning the last few millennia based on high-quality archeomagnetic  
24 data could be associated with the high-frequency band, on time scales on the  
25 order of the convective turnover time (e.g. Genevey et al., 2016, 2019).

26 Studying past field variations requires the construction of time-dependant  
27 global field models from the compilation of direct (or instrumental) and/or  
28 indirect geomagnetic field measurements. One of the most widely used mod-

29 els is the *gufm1* model, which covers the past 400 years (Jackson et al.,  
30 2000) from 1590 to 1990, and which was constructed from a large set of di-  
31 rect geomagnetic measurements obtained in land-based observatories and by  
32 mariners during their voyages across the seas (e.g. Jonkers et al., 2003), as  
33 well as from satellite data for the most recent period. However, our ability  
34 to instrumentally measure geomagnetic field intensities only dates back to  
35 the 1830s (Gauss, 1833). To overcome this lack of intensity data, Jackson  
36 et al. (2000), following Barraclough (1974), impose a linear decay rate of  
37 15 nT/yr to the axial dipole component between 1590 and 1840, i.e. a rate  
38 corresponding to a crude extrapolation back in time of the behavior observed  
39 since  $\sim 1840$ . Since it is essential for the construction of the *gufm1* model,  
40 and in general for our knowledge of geomagnetic field behavior during the  
41 historical period, this crude extrapolation has been tested against archeo-  
42 /paleointensity data (i.e. indirect measurements) provided by the study of  
43 the thermoremanent magnetization carried by archeological artifacts and vol-  
44 canic deposits (e.g. Gubbins et al., 2006; Finlay, 2008; Genevey et al., 2009;  
45 Hartmann et al., 2011; Suttie et al., 2011; Poletti et al., 2018). Hulot et al.  
46 (1997) indeed establish that the geomagnetic field can be recovered from  
47 directional data alone, up to a constant multiplier (the uniqueness of the  
48 sought-after solution being guaranteed by the existence of two, and only two,  
49 poles at Earth’s surface). The multiplicative constant is in practice provided  
50 by independent intensity measurements, each Gauss coefficient entering the  
51 mathematical description of the field being renormalized to account for the  
52 intensity measured at the specific location of interest.

53 Gubbins et al. (2006) follow this line of reasoning and this is the first

54 study to use the set of indirect intensity data available between 1590 and  
55 1840 to recalibrate the axial dipole component provided by *gufm1* by the  
56 ratio of measured to predicted intensities at intensity determination sites.  
57 Due to scattered data, they assume that a linear fit is indeed the most rea-  
58 sonable solution prior to 1840, but estimate that the axial dipole component  
59 between 1590 and 1840 had a rate of decrease of  $2.28 \pm 2.72$  nT/yr, which  
60 is significantly lower than that proposed by [Barraclough \(1974\)](#) and used by  
61 [Jackson et al. \(2000\)](#) (15.46 nT/yr and 15 nT/yr respectively).

62 Next, [Finlay \(2008\)](#) combines both direct and indirect geomagnetic mea-  
63 surements to calculate a new geomagnetic field model between 1590 and 1840,  
64 without imposing a linear decrease in the axial dipole during this period (but  
65 with an artificial overweighting of the indirect records). He shows that this  
66 approach does not provide better results than those favoring no change in  
67 axial dipole during the 17th and 18th centuries.

68 [Suttie et al. \(2011\)](#) propose a radically different approach based on the  
69 statistical analysis of errors in the archeo-paleointensity data. In particular,  
70 the dataset available between 1840 and 1990 is used to estimate reasonable  
71 errors in the data, which are best assigned as fractions ( $\sim 15\%$ ) of the field  
72 intensity values expected from *gufm1*. When applied to data prior to 1840,  
73 and again assuming a linear evolutionary trend in axial dipole over this pe-  
74 riod, they find a rate of decay ( $\sim 11.9$  nT/yr) close to what [Barraclough](#)  
75 [\(1974\)](#) found. They further show that if data errors are assigned as fractions  
76 of measured intensities, the decay rate is similar to that proposed by [Gubbins](#)  
77 [et al. \(2006\)](#) and [Finlay \(2008\)](#) (i.e., with either a slight change or no change  
78 at all in the axial dipole component over the 17th and 18th centuries). How-

79 ever, this observation is the result of a bias toward lower field values, as their  
80 uncertainties are lower when given as a proportion of measured intensities.

81 For the different methods above, dispersion of paleo-archeomagnetic data  
82 is such that it prevents overcoming the assumption of a linear evolution of the  
83 axial dipole component over the historical period. In addition, [Suttie et al.  
84 \(2011\)](#) demonstrate that the use of quality criteria on the dataset does not  
85 significantly change the conclusions. More recently, [Poletti et al. \(2018\)](#) also  
86 use a selected global dataset with strict criteria covering the historical period  
87 (1590–2009). After converting intensity data into corresponding axial dipole  
88 moments and performing linear regression computations for datasets covering  
89 various time intervals, they reach a conclusion favoring a linear decreasing  
90 trend of the axial dipole over the historical period of  $\sim 12.5$  nT/yr, thus  
91 close to that advocated by [Barraclough \(1974\)](#) and [Suttie et al. \(2011\)](#).

92 Given the dispersion observed in the global compilation of intensity data  
93 regardless of the selection criteria considered, [Genevey et al. \(2009\)](#) explore  
94 a different approach using a single consistent regional intensity dataset to  
95 recalibrate the  $g_1^0$  coefficient of *gufm1*. The principle remains the same as  
96 above ([Hulot et al., 1997](#)), which assumes that the geometry of the geomag-  
97 netic field as provided by *gufm1* is correct. While it potentially avoids the  
98 problem of global data scatter, and the almost insoluble issue of selecting  
99 only the most reliable data, it does raise the pending issue of which dataset  
100 is sufficiently reliable to be used to recalibrate the Gauss coefficients (an  
101 evaluation that will surely vary from one author to another). [Genevey et al.  
102 \(2009\)](#) use the set of accurate and precisely dated archeointensity results ob-  
103 tained in western Europe (700 km around Paris, France). Instead of a linear

104 decrease of the axial dipole magnitude over the historical period, they find a  
105 significant decrease between  $\sim 1590$  and the second half of the 18th century,  
106 with a minimum magnitude during this period, followed by a moderate in-  
107 crease from  $\sim 1800$  to  $\sim 1840$  and then, the well-established linear decrease  
108 up to the present. As a follow-up to this first study, [Hartmann et al. \(2010,](#)  
109 [2011\)](#) analyse precisely dated architectural brick fragments from southern  
110 and northern Brazil. Despite a significant non-dipole field effect between  
111 these two regions associated with the South Atlantic Anomaly (SAA), the  
112 results obtained appear to support the evolution in dipole field moment pro-  
113 posed by [Genevey et al. \(2009\)](#). As a new development, the present study  
114 carried out in Central Asia (Bukhara, Uzbekistan) focusing on the 1590 to  
115 1850 period aims to further constrain the accuracy of the non-linear dipole  
116 moment evolution deduced from the western European dataset.



## 117 **2. Historical context and sampling**

118 Situated on the Silk Road, Bukhara (39.8°N, 64.5°E, Fig. 1) has long  
119 been an important place for trade, Islamic education and religion in Central  
120 Asia, as evidenced by the many madrasas (or religious schools) and mosques  
121 still standing in the city's historic center. These old buildings were built  
122 throughout the medieval period from fired clay bricks. Their history, and  
123 more generally that of the city itself, is well known through abundant written  
124 testimonies preserved in the state archives of Uzbekistan.

125 [Figure 1 about here.]

126 For the period covered by our study, the sampled buildings were erected  
127 during three successive dynasties that ruled Bukhara from the mid-16th cen-  
128 tury to the beginning of the 20th century: the Shaybanid dynasty during the  
129 16th century, the Djanid dynasty from the 17th to the mid-18th century, and  
130 the Manghit dynasty from the mid-18th century to the early 20th century.  
131 The Shaybanid dynasty, which claimed to be descended from Genghis Khan,  
132 conquered Bukhara from the Timurids in the early 16th century and founded  
133 the khanate of Bukhara. Their domination for  $\sim 100$  years was interrupted  
134 by the Djanid dynasty (which also claimed to be descended from Genghis  
135 Khan), which then established its rule over Bukhara for about a century and  
136 a half. The Shah of Iran (Nader Shah) conquered the khanate around the  
137 mid-18th century, but the collapse of his empire a few years later led to the  
138 establishment of the Manghit dynasty. This dynasty was of Uzbek origin  
139 and ruled the Emirate of Bukhara until 1920 when Soviet Red Army troops  
140 invaded the city.

141 Throughout the above period, the city of Bukhara was divided into sev-  
142 eral small social units called *guzars*, whose history is well documented in  
143 the archives. Each guzar was led by a chief (*aqsaqal*) nominated by the el-  
144 ders and had its own mosque around which the community was structured  
145 (Khalid, 1991). Sukhareva (1976) (see Khalid, 1991) has done considerable  
146 work in compiling the oral and written testimonies on the guzars, providing  
147 extremely valuable information on the dating of even the smallest madrasas  
148 and mosques built in Bukhara over the past millennium.

149 Our archeomagnetic sampling was focused on several major and some  
150 minor buildings in and around Bukhara (Fig. 2, Table 1). Among the most  
151 important are three madrasas (with a group of fragments collected for each  
152 of them): two were built during the reign of Abdullah Khan (1583-1598),  
153 one of the most famous rulers of the Shaybanid dynasty. One was built for  
154 his own glory (Madrasa Abdullah Khan; BK03, Fig. 2a), and the second  
155 was to glorify his mother (Madrasa Modari Khan, built around 1561; BK04,  
156 Fig. 2b). The third madrasa was erected in  $\sim 1651 - 1652$  by the Khan Abd  
157 al-Aziz of the Djanid dynasty (Madrasa Abd al-Aziz Khan; BK01). We also  
158 carried out a sampling in the Chor Bakr necropolis, built near Bukhara at  
159 the time of the Shaybanid dynasty, and at the location of older tombs dating  
160 from the 10th century. There, two groups of fragments associated with the  
161 tomb (or *khazira*) of Khwādja Saad, son of Khwādja Islām Juybārī, leader of  
162 the Sufi order (i.e., a mystical order of Islam), erected just before his death at  
163 the end of the 16th century, have been collected (Khwādja Saad tomb's wall  
164 and floor, BK05 and BK06 respectively; Fig. 2d; Table 1). The sampling also  
165 included the Ark citadel. This ancient fortress, which was last destroyed in

166 1920 and rebuilt several times during the history of the city, was inhabited  
167 by the rulers. It comprised several buildings surrounded by an imposing wall,  
168 among which we sampled the *kānaqāh*, which is a dwelling place for dervishes  
169 (adherents of Sufi orders) dating from the mid-18th century (BK14, Fig. 2f).  
170 In addition, we sampled fragment groups from three minor mosques (Mosque  
171 Dostum Chor Oghasi, BK09; Mosque Magoki Kurpa, BK12; and Mosque  
172 Kemuhtagaron, BK13) and three madrasas from smaller neighborhoods (i.e.  
173 the *guzars*; Madrasa Kunjak, BK07; Madrasa Rakhmanqul, BK08; Madrasa  
174 Rashid-al-Din, BK11, Table 1). It should also be mentioned that special  
175 care was taken to avoid restored wall segments and/or recycled bricks, which  
176 would result in an inaccurate dating.

177 [Figure 2 about here.]

178 In total, our archeomagnetic study is based on 13 groups of architectural  
179 brick fragments. The three above-mentioned dynasties are equally sampled :  
180 five buildings belong to the Shaybanid dynasty, four buildings to the Djanid  
181 dynasty and four buildings to the Manghit dynasty. For each of the two frag-  
182 ment groups BK08 and BK11, the samples were collected in different rooms  
183 of the same building; in this case, each sub-subset has been identified but  
184 all fragments are considered to come from the same ensemble (e.g. BK08A  
185 or BK08B; see supplementary Table S1). Particular care was taken with the  
186 available dating constraints and we selected those buildings that have age  
187 uncertainties of less than  $\pm 25$  years, but in most cases these are less than 15  
188 years. For some of these buildings, the construction is very well constrained  
189 by archives due to the social prominence of the people they were built for. In



198 **3. Methods**

199 *3.1. Archeointensity determinations*

200 All experiments were conducted in the paleomagnetic laboratory of the  
201 Institut de Physique du Globe de Paris (IPGP). The archeointensity deter-  
202 minations are based on the experimental protocol developed for the Triaxe  
203 magnetometer. This unique magnetometer allows continuous magnetization  
204 measurements (every  $\sim 5^\circ \text{C}$ ) of a small individual specimen (less than  $1 \text{ cm}^3$ )  
205 at high temperatures and under controlled field conditions, both in intensity  
206 and direction (Le Goff and Gallet, 2004).

207 The Triaxe procedure consists of five measurement series automatically  
208 performed between a low temperature  $T_1$ , usually  $150^\circ\text{C}$ , and a high tem-  
209 perature,  $T_2$ , at which most of the magnetization carried by the specimen is  
210 erased:

- 211 – Step 1: After rapid heating from room temperature to  $T_1$ , the speci-  
212 men is heated in a zero field from  $T_1$  to  $T_2$  to demagnetize its natu-  
213 ral remanent magnetization (NRM). The corresponding magnetization  
214 measurements give the  $M_1$  series;
- 215 – Steps 2 and 3: The specimen is cooled from  $T_2$  to  $T_1$  (step 2) and  
216 next heated from  $T_1$  to  $T_2$  (step 3), both steps in a zero field, to allow  
217 characterization of the thermal variability of the NRM fraction that  
218 remains blocked at  $T_2$ . The magnetization measurements give the  $M_2$   
219 and  $M_3$  magnetization series, respectively;
- 220 – Step 4: The specimen is cooled from  $T_2$  to  $T_1$  in a laboratory field, the  
221 intensity of which is chosen close to the expected ancient field intensity,

222 and its direction is automatically adjusted so that the direction of the  
 223 newly acquired laboratory thermoremanent magnetization ( $\text{TRM}_{\text{lab}}$ ) is  
 224 parallel to the direction of the original TRM (i.e., NRM). This step  
 225 therefore leads to the acquisition of a new TRM with unblocking tem-  
 226 peratures between  $T_2$  and  $T_1$  (magnetization series  $M_4$ );

227 – Step 5: The specimen is then heated again between  $T_1$  and  $T_2$  to  
 228 demagnetize the  $\text{TRM}_{\text{lab}}$  (magnetization series  $M_5$ ).

229 The procedure ends with rapid cooling of the specimen to room temperature.

Intensity determinations are based on the ratio between the NRM and  $\text{TRM}_{\text{lab}}$  fractions unblocked between  $T_1$  and a temperature  $T_i$ , varying from  $T_1$  to  $T_2$ . At any  $T_i$ , these fractions are respectively determined by:

$$\Delta'_1(T_i) = (M_1(T_1) - M_1(T_i)) - (M_3(T_1) - M_3(T_i)); \quad (1)$$

$$\Delta'_5(T_i) = (M_5(T_1) - M_5(T_i)) - (M_3(T_1) - M_3(T_i)). \quad (2)$$

And the intensity value at  $T_i$  is given by:

$$R'(T_i) = H_{\text{lab}} \frac{\Delta'_1(T_i)}{\Delta'_5(T_i)}. \quad (3)$$

230 An intensity value is obtained for each specimen from the average of the  
 231  $R'(T_i)$  data derived from all temperatures  $T_i$  between  $T_1$  and  $T_2$  (see more  
 232 details and discussion in [Le Goff and Gallet, 2004](#)). Since intensity values  
 233 should only be determined on the primary and single-vector magnetization  
 234 acquired during the manufacture of the archeological artifacts, if a secondary

235 magnetization is observed above  $T_1$  up to  $T_1'$  (but below  $T_2$ ) from the anal-  
236 ysis of the NRM demagnetization data, then the reference temperature  $T_1$   
237 can be increased to  $T_1'$ .

238 Compared to more conventional paleo/archeointensity methods that rely  
239 on stepwise demagnetizations and magnetization measurements carried out  
240 at room temperature, the Triaxe procedure has several advantages, including  
241 the fact that the  $TRM_{lab}$  is acquired under thermal and field conditions rela-  
242 tively similar to those that led to the NRM acquisition. This helps to mitigate  
243 possible spurious effects that would result from the presence of multi-domain  
244 grains. The fact that the direction of the  $TRM_{lab}$  is parallel to that of the  
245 original TRM eliminates the need for anisotropy correction on TRM acqui-  
246 sition (Le Goff and Gallet, 2004). In addition, experiments have shown that  
247 the use of  $R'(T_i)$  data allows to overcome the effect of the cooling rate on  
248 TRM acquisition (Le Goff and Gallet, 2004; Genevey et al., 2009; Hartmann  
249 et al., 2010, 2011, see also a more general discussion on TRM anisotropy and  
250 cooling rate effects in Genevey et al., 2008).

251 The intensity data are then examined according to a set of quality criteria,  
252 which have remained the same as in previous studies in which archeointensity  
253 data obtained using the Triaxe procedure are reported (e.g. Genevey et al.,  
254 2013, 2016, 2019; Gallet et al., 2014, 2015, 2020). At the specimen level,  
255 the  $R'(T_i)$  data must involve at least 50% of the NRM still blocked at  $T_1$   
256 (or  $T_1'$ ) and the relative variations of  $R'(T_i)$  between  $T_1$  (or  $T_1'$ ) and  $T_2$   
257 must not exceed 10%. A mean intensity value is determined at the fragment  
258 level when successful results are obtained from at least two specimens (note  
259 that in our study, we increase this number to three different specimens). In

260 addition, the fragment-mean value is retained only if its standard deviation  
261 does not exceed 5% of the corresponding mean-intensity value. Finally, a  
262 mean intensity value is calculated at the level of a group of fragments when  
263 results are available from a minimum of three different fragments meeting  
264 the above criteria. The error is given as the standard deviation computed  
265 from the set of the retained intensity values at the fragment level. These  
266 criteria applied successively at the specimen, fragment, and fragment-group  
267 levels have proven to be extremely effective in constraining the quality and  
268 consistency of the intensity values obtained using the Triaxe method (e.g.  
269 [Gallet and Le Goff, 2006](#); [Genevey et al., 2009](#); [Hartmann et al., 2010, 2011](#);  
270 [Hervé et al., 2017](#)).

271

### 272 *3.2. Magnetic mineralogy characterizations*

273 In addition to the archeointensity experiments, we also performed dif-  
274 ferent analyses on the retained fragments to identify the magnetic minerals  
275 present in the samples and to further assess whether this magnetic miner-  
276 alogy alters during heating. Analyses include, for all retained fragments,  
277 low-field susceptibility vs. temperature measurements (using a KLY3 kap-  
278 pabridge from Agico coupled with a CS3 furnace) and for at least two frag-  
279 ments from each retained group, the acquisition (using a Vibrating Sample  
280 Magnetometer Model 3900) of isothermal remanent magnetization (IRM)  
281 and hysteresis loop measurements as well as the thermal demagnetization of  
282 three-axis IRM acquired (using a MMPM10 pulse magnetizer) in orthogonal  
283 fields of 1.5, 0.6 and 0.2 T ([Lowrie, 1990](#)). Additionally, for a selection of  
284 representative samples, complementary low-temperature magnetization mea-



285 surements are carried out using a magnetic property measurements system  
286 (MPMS XL-5 EverCool). The latter experiments include the following re-  
287 manent magnetization measurements: 1) temperature cycling of an IRM  
288 acquired at room temperature (RT-SIRM) in a 2.5 T field down to 10 K and  
289 return to room temperature in a zero field (less than  $\pm 500$  nT), and 2) the  
290 thermal demagnetization from 10 K to 300 K of an IRM acquired at 10 K in  
291 a 2.5 T field following a zero-field cooling (ZFC) and 2.5 T-field cooling (FC)  
292 pre-treatments from 300 K to 10 K. Both the RT-SIRM and ZFC-FC mea-  
293 surements were duplicated in a second series of experiments where the initial  
294 IRMs acquired in 2.5 T are demagnetized in a 300 mT using the MPMS's  
295 superconducting magnet in a field oscillation mode, a method introduced and  
296 validated in [Lagroix and Guyodo \(2017\)](#). The objective of the second series  
297 of experiment is to remove the contribution from low coercivity minerals to  
298 the total magnetization.

## 299 4. Archeointensity results

### 300 4.1. Magnetic mineralogy

301 IRM acquisition curves are reported in Fig. 4a. They show that satura-  
302 tion of the magnetization is often not completely achieved at 1T, but a clear  
303 inflexion in the magnetization curves is observed at  $\sim 0.1$  T. The thermal  
304 demagnetization of three-axis IRM further indicates that the magnetization  
305 is mostly carried by low-coercivity minerals with unblocking temperatures  
306 below  $600^{\circ}\text{C}$ , which is consistent with the presence of (titano)magnetite  
307 (Fig. 4b-e). Fig. 4b-e also shows the presence of high-coercivity minerals  
308 whose unblocking temperatures do not exceed  $\sim 550^{\circ}\text{C}$ , being sometimes as  
309 low as  $\sim 200^{\circ}\text{C}$  or with an inflexion around this temperature (Fig 4b-d).  
310 Fine grained hematite, lowering its unblocking temperature (e.g. [Özdemir  
311 and Dunlop, 2014](#)), and/or epsilon iron oxide, a magnetic phase often ob-  
312 served in archeological artifacts (e.g. [Genevey et al., 2016](#); [López-Sánchez  
313 et al., 2017](#); [Kostadinova-Avramova et al., 2019](#)) are mineral phases com-  
314 patible with the above observations. A duality of low and high coercivity  
315 minerals is observed in a few hysteresis loops displaying slight constrictions  
316 (Fig. 5a). However, most of the hysteresis loops are not wasp-waisted, al-  
317 ways exhibiting a monotonic decrease in loop opening with increasing field  
318 (Fig. 5b-c).

319 [Figure 4 about here.]

320 Low-field susceptibility versus temperature (heating and cooling) curves  
321 yield two main observations (Fig. 5d-i). First, heating and cooling curves  
322 are reversible or very nearly, which attests to the stability on heating of the

323 magnetic mineralogy in the temperature range used for intensity determi-  
324 nations. Second, most susceptibility curves show a clear inflexion around  
325 300°C, arising from a range of susceptibility evolutions from rapid rates of  
326 change (Figs. 5g,i) to slower monotonic ones (Figs. 5d,e,f,h), in addition to  
327 a higher temperature inflexion above 500°C. At this stage, we could propose  
328 that the inflexions indicate the presence of two families of (titano)magnetite  
329 differing by their grain size, their titanium content and/or their oxidation  
330 state (see below).

331 [Figure 5 about here.]

332 Low-temperature magnetization measurements bring additional insight  
333 into the magnetic mineralogy. Compared to the classic RT-SIRM and ZFC-  
334 FC experiments (left panels in Fig. 6 and 7 respectively), their 300 mT  
335 demagnetized counterparts highlight the temperature dependent behaviour  
336 of the high coercivity minerals (right panels in Fig. 6 and 7 respectively).  
337 Comparing the two provides information on the relative contribution of low or  
338 high coercivity minerals to the total remanence (Lagroix and Guyodo, 2017).  
339 The lack of a Verwey transition in ZFC-FC data (Fig. 7) and primarily re-  
340 versible RT-SIRM curves (Fig. 6a and c) or temperature suppressed Verwey  
341 transition (Fig. 6b) are compatible with titanomagnetite (Kakol et al., 1994;  
342 Moskowitz et al., 1998; Muxworthy and McClelland, 2000). Hematite is un-  
343 ambiguously identified in the demagnetized RT-SIRM data (right panels of  
344 Fig. 6a and c) from the observed Morin transition, which displays a rema-  
345 nence loss and partial recovery over a wide temperature range ( $\sim 235$  K to  
346 170 K) compatible with fine grained (0.1 to 1  $\mu\text{m}$ ) hematite (Özdemir et al.,

347 2008; Özdemir and Dunlop, 2014). Another noteworthy observation is the  
348 kink seen at  $\sim 70$  K in both ZFC and FC curves which is also compatible  
349 with Ti-rich (50 to 60 % Ti; see Moskowitz et al., 1998) titanomagnetite  
350 and the persistence of the kink after 300 mT demagnetization (except for  
351 BK04-10) finds an explanation in the significant increase in coercivity at low  
352 temperature of Ti-rich titanomagnetite (see for example fig. 15c in Almeida  
353 et al., 2014). The 70 K kink could alternatively be related to the epsilon iron  
354 oxide phase (López-Sánchez et al., 2016, 2017). Lastly, behaviour suggestive  
355 of nanogoethite (Guyodo et al., 2003), which would be of weathering origin,  
356 is occasionally observed (right panels of Fig. 6b and Fig. 7a).

357 [Figure 6 about here.]

358 [Figure 7 about here.]

#### 359 4.2. *New archeointensity data*

360 We analyzed a total of 160 fragments (532 specimens) from 13 different  
361 archeological (historical) contexts. Most often, the magnetization of the sam-  
362 ples is comprised between  $\sim 30$  and  $\sim 140 \times 10^{-8} \text{Am}^2$  (with a maximum of  
363  $\sim 500 \times 10^{-8} \text{Am}^2$ ) and 42 of them are too weak to be measured with the  
364 Triaxe magnetometer ( $< 30 \times 10^{-8} \text{Am}^2$ ), which has a measurement sensitiv-  
365 ity on the order of  $\sim 10^{-8} \text{Am}^2$  (Le Goff and Gallet, 2004). Among the 118  
366 remaining fragments, 70 fragments are rejected due to non-linear or com-  
367 plex behavior compared to the nominal behavior described in Le Goff and  
368 Gallet (2004) (see also Genevey et al., 2009; Hartmann et al., 2010, 2011)  
369 and because of scattered magnetization measurements. In addition, 12 frag-  
370 ments are rejected because satisfactory results are obtained from only one

371 specimen for each of them (whereas a minimum number of three specimens  
372 is required). Finally, 36 fragments from nine groups of fragments (112 speci-  
373 mens) provide archeointensity results that meet our set of selection criteria,  
374 while four groups are rejected (i.e. BK01 [1642 – 1652], BK02 [1735 – 1759],  
375 BK03 [1578 – 1590], BK09 [1580 – 1586]). This corresponds to a low suc-  
376 cess rate of 31% compared with the 118 fragments actually analyzed with  
377 the Triaxe. Details of the successful data are presented in Table S1 in the  
378 supplementary material.

379 Three representative thermal demagnetization diagrams are shown in  
380 Fig. 8 (left panels), together with the corresponding  $R'(T_i)$  data (right pan-  
381 els). In general, the specimens are fully demagnetized at relatively low tem-  
382 peratures, below 450–500°C. A single magnetization component is essentially  
383 isolated, even though a small secondary component probably of viscous ori-  
384 gin is identified in most cases at low temperatures (below 150°C) but also in  
385 some cases having slightly higher temperatures (around 200°C).

386 [Figure 8 about here.]

387 Of the nine groups of fragments, the data from six groups are shown  
388 in Fig. 9 (with one panel each). In this figure, each curve represents the  
389  $R'(T_i)$  data obtained for a specimen. It is also worth recalling that for each  
390 group of fragments, the  $R'(T_i)$  data are first averaged at the specimen level  
391 (over the temperature range between 140°C-260°C and 385°C-525°C), then  
392 at the fragment (brick) level, and finally all the fragment-mean values are  
393 averaged at the level of each group of fragments. Six groups of fragments are  
394 defined by data obtained from three different fragments (with a total of nine

395 specimens) and the maximum number of fragments is seven (group BK11  
396 with 21 specimens; Table 1). Fig. 9 also illustrates the overall consistency of  
397 the data obtained for each group of fragments, resulting in small standard  
398 deviations. They range from  $0.9 \mu\text{T}$  (BK08) to  $2.0 \mu\text{T}$  (BK11), or between  
399 1.7% (BK06) and 4.5% (BK11) of the mean intensity values.

400 [Figure 9 about here.]

401 The new archeointensity data obtained at Bukhara cover a time interval  
402 of  $\sim 250$  years, between the mid-16th century and  $\sim 1800$  (Table 1; Fig. 10).  
403 A significant decrease in intensity values by  $\sim 14 \mu\text{T}$  is observed from  $\sim 1560$   
404 to  $\sim 1725$ , leading to an average rate of change of  $\sim -0.1 \mu\text{T}/\text{yr}$ . The second  
405 half of the 18th century is then marked by an increase of  $\sim 4 \mu\text{T}$  until the  
406 early 19th, leading to a variation rate of  $\sim 0.05 \mu\text{T}/\text{yr}$ . This rate of change is  
407 fairly comparable to that of the present-day field in Bukhara ( $\sim 0.08 \mu\text{T}/\text{yr}$ ).

408 [Table 1 about here.]

## 409 5. Discussion

### 410 5.1. Comparison of the new archeointensity data with model predictions at 411 Bukhara

412 The new archeointensity data obtained in Bukhara are compared in Fig. 10  
413 with the variations in intensities predicted by several geomagnetic field re-  
414 constructions (see Section 1). For the historical period, this is the *gufm1*  
415 model (Jackson et al., 2000), and the models of Gubbins et al. (2006); Fin-  
416 lay (2008); Suttie et al. (2011) derived from *gufm1* and calibrated for the  
417 1590 – 1840 time interval from a global compilation of archeointensity mea-  
418 surements. Over this time interval, they predict the same pattern of variation  
419 but with various amplitudes corresponding to the different rates of decay im-  
420 posed on the axial dipole component (recall the related commentary in the  
421 introduction). The comparison is also extended to geomagnetic models cov-  
422 ering longer time intervals (between 3000 and 14,000 years): A\_FM (Licht  
423 et al., 2013), pfm9k.1 (Nilsson et al., 2014), SHA.DIF.14k (Pavón-Carrasco  
424 et al., 2014a), CALS10k.2 and ARCH10k.1b (Constable et al., 2016), COV-  
425 ARCH (Hellio and Gillet, 2018) and BIGMUDI4k.1 (Arneitz et al., 2019)  
426 constructed using global archeomagnetic datasets. A number of these mod-  
427 els are constrained by *gufm1* over the historical period (i.e. CALS10k.2,  
428 ARCH10k.1b, SHA.DIF.14k) and the corresponding predictions fall within  
429 the range of *gufm1*-recalibrated models. The pfm9k.1 model predicts a sim-  
430 ilar evolution, although the predicted dipole moment is higher compared to  
431 the other models for this period. The authors interpret this overestimation  
432 as resulting from the introduction of directional sedimentary data (Nilsson  
433 et al., 2014). As this study is focused on the field variations over the his-

434 torical period, the models constrained by *gufm1* (CAL510k.2, ARCH10k.1b,  
435 SHA.DIF.14k) and those integrating sedimentary data (pfm9k.1) are not  
436 represented. A<sub>FM</sub> predicts an intensity evolution in Bukhara very close to  
437 the prediction from *gufm1*, with higher values than the observed intensities  
438 during the 18th and early 19th century. On the other hand, BIGMUDI4k.1  
439 predicts a different intensity evolution, with a quasi-constant intensity during  
440 the 17th century and a well-marked intensity peak during the 18th century.  
441 Unlike the other models, BIGMUDI4k.1 is built from the simultaneous inver-  
442 sion of both direct and indirect data. The authors note a significant decrease  
443 in the dipole energy associated with an increase of the non-dipole energy  
444 around  $\sim 1600$ . According to [Arneitz et al. \(2019\)](#), this is mainly due to the  
445 large increase in the amount of data at the onset of the historical geomag-  
446 netic era, rather than a true geomagnetic feature. The subsequent increase  
447 in dipole energy is therefore probably artificial, as is the resulting intensity  
448 peak. The modeled field behavior during this period should therefore be  
449 considered with caution. The COV-ARCH model, integrating only archeo-  
450 magnetic data, predicts an evolution close to those of [Gubbins et al. \(2006\)](#)  
451 and [Finlay \(2008\)](#). This evolution shows a minimum intensity at the end of  
452 the 18th century, slightly later than the minimum intensity observed from  
453 our data. Interestingly, in this model, the high-frequency range of the axial  
454 dipole variations is constrained by a timescale on the order of the convective  
455 turnover time (recall Introduction).

456 However, regardless of the model, none of the expected intensity evolu-  
457 tions reach the low intensity values observed in the 18th century from the new  
458 Bukhara archeointensity data, with a minimum overestimate of  $\sim 5 - 6 \mu\text{T}$ .



459 Prior to this period, the Bukhara data show a rapid decrease in intensities  
460 between 1550 and the early 18th century with a rate quite similar to that of  
461 *gufm1* (see also the model of [Suttie et al., 2011](#)). On the other hand, the  
462 new data require an increase in intensities during the first half of the 19th  
463 century, which is either absent or much more limited in model predictions.

464

## 465 *5.2. Dispersion of archeointensity results in regional datasets*

466

[Figure 10 about here.]

467 Comparison with other archeointensity results previously obtained in  
468 the Bukhara area, as well as elsewhere in western Eurasia, raises a criti-  
469 cal problem related to the dispersion that generally characterizes the regional  
470 datasets. Here we distinguish four geographical areas within a 700 km-radius  
471 around the cities of Bukhara (Uzbekistan), Moscow (western Russia), Tbilisi  
472 (Georgia), and Thessaloniki (Balkans). In each zone, the data are reduced  
473 to the latitude of the corresponding city. Most of these data are rather old  
474 and were described in the ArcheoInt compilation ([Genevey et al., 2008](#)). In  
475 our study, they are selected using at first the same minimalist criteria as in  
476 [Genevey et al. \(2008\)](#) (referred to as G2008 below). A location map of the  
477 selected data is given in supplementary material (Fig. [S1](#)). These criteria  
478 were originally proposed to allow the discussion of old data acquired without  
479 all the quality criteria now considered necessary for any new study, whereas  
480 applying modern criteria would eliminate most (if not all) of them. They do  
481 not consider the intensity methods directly but instead rely on 1) the error  
482 (most often a standard deviation) on the average intensity, which must be

483 known and less or equal to 15%; 2) the number of intensity determinations  
484 (Nint) used to derive an intensity mean. Nint is required to be greater than  
485 or equal to three when no pTRM-check was implemented or when this test  
486 does not apply. Otherwise, Nint must be greater than or equal to two. For  
487 objects recognized to be strongly anisotropic (such as pottery or tiles), Nint  
488 is required to be greater than or equal to three if anisotropy effects on TRM  
489 acquisition were not taken into account. In a second step, we consider stricter  
490 criteria requiring pTRM-check (when this test does apply) and intensity av-  
491 erage derived from at least three independent fragments. In both cases, we  
492 select data whose age uncertainties are less or equal to  $\pm 50$  years because we  
493 are interested in fairly rapid variations over a short time interval of  $\sim 300$   
494 years. Further note that practically none of the available data have been  
495 corrected for the cooling rate effect on TRM acquisition.

496 Intensive work by Russian archeomagnetists in the 1970s and 1980s (S.  
497 Burlatskaya, I. Nachasova and K. Burakov) resulted in three regional datasets,  
498 in Uzbekistan, around Moscow (Moscow, Gor'kiy, and Vologda) and in Geor-  
499 gia. These data share common features. They were acquired from analyzed  
500 baked bricks. The number of intensity determinations corresponds to the  
501 number of independent bricks studied. The method used is the original  
502 [Thellier and Thellier \(1959\)](#) protocol with the use of pTRM-check along the  
503 measurements cycle. This key element was, however, not specified in the  
504 articles (mainly published in Russian journals) but given by S. Burlatskaya,  
505 I. Nachasova and K. Burakov in a personal communication (2004). A sin-  
506 gle dataset was obtained both using the [Thellier and Thellier \(1959\)](#) pro-  
507 tocol and an original method developed by [Burakov and Nachasova \(1978\)](#),

508 so-called thermal curves, derived from [Wilson \(1961\)](#)'s method (the TRM  
509 anisotropy being also taken into account). Both datasets are considered in  
510 our paper (identified by two different symbols), although we should point out  
511 that [Nachasova and Burakov \(1996\)](#) argue that the thermal-curves dataset  
512 is more reliable.

513 The results from Uzbekistan, more precisely obtained in the cities of  
514 Bukhara ([Burlatskaya et al., 1977, 1986b](#)), Samarkand ([Burlatskaya et al.,](#)  
515 [1969, 1986b](#)), and Khiva ([Burakov and Nachasova, 1978](#)) are of particular  
516 interest as they allow a direct comparison with the new archeointensity data  
517 reported in the present study (Fig. 11a). For Bukhara, the sampled sites  
518 are unfortunately not indicated in the original Russian publication (nor in  
519 the compilation of [Burlatskaya et al., 1986b](#)), but their estimation of the  
520 age of the buildings does not correlate with our new data, and the buildings  
521 sampled in the present study are therefore probably different. Despite some  
522 scatter, a fairly satisfactory agreement could be found for all results dating  
523 from  $\sim 1700$  to  $\sim 1850$ , with values often lower than the intensities expected  
524 in Bukhara from the models. However, this satisfaction must be tempered  
525 by the fact that the older data (before  $\sim 1700$ ) appear systematically weaker  
526 than our own intensity values. The discrepancy would be even larger if  
527 a cooling rate correction (for instance 5% as suggested by [Genevey et al.,](#)  
528 [2008](#)) was applied to the data. This ambiguous information is not improved  
529 by considering stricter selection criteria as most of the Uzbek data meet those  
530 criteria (see Fig. S2 in supplementary data).

531 [Figure 11 about here.]

532 A fairly large scatter is observed for the data around Moscow, whether

533 these data are selected using the G2008 set of criteria or stricter criteria  
534 (remember that only the number of fragments per site,  $N_{int} \geq 2$  or  $\geq 3$ , is the  
535 difference between the two selections, Fig. 11b and Fig. S2b). This dispersion  
536 questions the reliability of at least part of these data, as previously discussed  
537 in Salnaia et al. (2017a,b). The data seem consistent with a decreasing trend  
538 in intensities over the historical period. Nonetheless, as the cooling rate effect  
539 was not evaluated in the old Russian studies (Nachasova, 1972; Burlatskaya  
540 et al., 1986b), these data could also agree with lower than predicted values  
541 during the 18th century. Comparing the scant data obtained in Georgia with  
542 the model predictions leads to another contrast (see Fig. 11c). While fairly  
543 consistent agreement is observed for the data up to  $\sim 1700$ , the results are  
544 more scattered from  $\sim 1700$  to  $\sim 1850$ , and in general the results are less  
545 consistent with the expected intensity values. With an arbitrary correction  
546 of the cooling rate effect of 5%, some of the latter results would be too low,  
547 in particular those dating from  $\sim 1800$  to  $\sim 1850$ .

548 Finally, the Balkan area incorporates results from Greece and Bulgaria  
549 (Aitken et al., 1989a; Spassov et al., 2010; Kovacheva et al., 2009, 2014).  
550 Contrary to the Russian datasets, whether pTRM checks are implemented  
551 or not is critical for distinguishing between the two selected datasets based  
552 on the G2008 versus the stricter set of criteria. (Fig. 11d and Fig. S2d). With  
553 the loose selection (Fig. 11d), the data available between  $\sim 1550$  and  $\sim 1700$   
554 appear relatively scattered, with the Bulgarian data in particular generally  
555 higher than the values expected from the models. A limited decrease to  
556 account for the cooling rate effect would improve the agreement. However,  
557 when stricter criteria are applied, all but one data point from this time

558 interval are eliminated (Fig. S2d). Still considering this selection, five data  
559 points remain for the 18th century (including two obtained by Spassov et al.  
560 (2010) from the same volcanic event) and only one for the 19th century. We  
561 may note that two Bulgarian results from the second half of the 18th century  
562 do not seem to indicate lower values than those predicted by the models. In  
563 contrast, this would be the case for the only result dating from the early 19th  
564 century. At this stage, it is therefore difficult to draw a firm conclusion from  
565 these rare data as well as from the entire Balkan dataset.

566 Overall, Fig. 11 shows that the data available in each of the four areas  
567 discussed above are too scattered to show any consistent pattern of intensity  
568 variations, at least at the century scale. As pointed out by many authors,  
569 determining a set of selection criteria that allows for significant reduction in  
570 the regional data scatter is a challenge. In the present case, increasing the  
571 strictness of the criteria does not alter our conclusion (at most, it leads to the  
572 rejection of most data in the Balkans), as this was also previously observed  
573 and discussed for western Europe (Genevey et al., 2009, 2013, 2019).

574 Our purpose is not to analyze and discuss in detail all datasets currently  
575 available worldwide (see for instance Poletti et al., 2018). As observed in  
576 western Eurasia, data at the regional spatial scale are generally either too  
577 scant or too scattered to draw a clear evolution of intensities, which could  
578 lead to a meaningful comparison with the intensity values derived from the  
579 geomagnetic field models (see for instance Tema et al., 2017; Gogutchiaichvili  
580 et al., 2018; Kapper et al., 2020, for Hawaii, Mesoamerica and West Africa,  
581 respectively). Due to the general dispersion of the archeointensity data at  
582 the regional scale, a linear evolution of the dipole moment as constrained

583 over the entire historical period, notably between 1600 and 1800, by a re-  
584 calibration of  $g_1^0$  with these data, is a simple and reasonable approximation  
585 (e.g. [Gubbins et al., 2006](#)). However, this does not demonstrate that the  
586 axial dipole moment evolution is actually linear; the scatter and the small  
587 amount of data leave room for more complex, possibly hidden variations in  
588 axial dipole moment.

589

590 *5.3. A non-linear evolution of the axial dipole moment over the historical*  
591 *period*

592 One might consider two options for explaining the dispersion of the data,  
593 either the frequent presence of biased results masking the “true” regional  
594 field intensity evolution or an inherent limitation in archeointensity determi-  
595 nations. In other words, due to their lack of resolution (and/or underesti-  
596 mation of their uncertainties), archeointensity data could not reliably detect  
597 and describe century-scale intensity variations. It is worth pointing out that  
598 this (dull) option is in clear contradiction with the convincing detection in  
599 western Europe of century-scale intensity variations over the past  $\sim 1500$   
600 years ([Genevey et al., 2009, 2013, 2016, 2019](#)). For the historical period,  
601 western Europe benefits from a fairly dense archeointensity dataset showing  
602 a smooth evolution, with reduced dispersion (see description of the data in  
603 the mentioned studies). This leads [Genevey et al. \(2009\)](#) to explore a dif-  
604 ferent approach for the recalibration of *gufm1* Gauss coefficients by using a  
605 limited but consistent regional dataset.

606

[Figure 12 about here.]

607 Fig. 12a shows a direct comparison between the new data from Bukhara  
608 and the western European results recently upgraded and summarized in [Genevev et al. \(2019\)](#). Most of these results share the same (Triaxe) experimental  
609 methodology and obey the same set of selection criteria. This comparison  
610 methodology and obey the same set of selection criteria. This comparison  
611 takes into account the geomagnetic field geometry given by the *gufm1* model.  
612 Following [Gubbins et al. \(2006\)](#) and [Genevev et al. \(2009\)](#), a ratio of mea-  
613 sured to predicted intensity is determined for each data point of [Genevev](#)  
614 [et al. \(2009, 2013, 2019\)](#). It is then used to recalibrate all Gauss coefficients  
615 from *gufm1*, allowing the computation of a new field intensity prediction at  
616 Bukhara. This procedure applied to all the western European Triaxe data, to  
617 which we add three Triaxe data obtained in Russia by [Salnaia et al. \(2017a,b\)](#),  
618 allows for the determination of a consistent dataset (Fig. 12a). The results  
619 show a clear intensity decrease between  $\sim 1600$  and the first half of the 18th  
620 century, followed by an increase up to 1850. This intensity pattern is further  
621 evidenced by the computation of a mean intensity variation curve and its  
622 credible interval using the transdimensional Bayesian method recently devel-  
623 oped by [Livermore et al. \(2018\)](#). Based on this consistency, we also use the  
624 same dataset to recalibrate the axial dipole component ( $g_1^0$ ) given by *gufm1*  
625 (Fig. 12b). While the  $g_1^0$  values (provided in Table S2) are rather compat-  
626 ible with those of the models during most of the 17th century, significant  
627 differences are then observed with all models, with smaller recalibrated val-  
628 ues, throughout the entire 18th century and the early 19th century. On the  
629 other hand, our study also shows that there is currently no dense regional  
630 archeointensity dataset in western Eurasia, as elsewhere considering also the  
631 dispersion of the data, that could clearly contradict this  $g_1^0$  evolution. As

632 previously suggested by [Genevey et al. \(2009\)](#) and now based on a larger col-  
633 lection of results, [Fig. 12b](#) strongly militates for a non linear evolution of the  
634 axial dipole field moment over the historical period, with a distinct minimum  
635 of  $|g_1^0(t)| \approx 29400$  nT during the 18th century. The average rate of decrease  
636 of  $|g_1^0(t)|$  during the 17th and the late 18th reaches  $\sim -26$  nT/yr, while the  
637 increase during the first half of the 19th reaches a rate of  $\sim 34$  nT/yr. These  
638 two variation rates are higher than the one observed over the past 150 years  
639 ( $\sim 15$  nT/yr; e.g. [Barraclough, 1974](#); [Jackson et al., 2000](#)).

640 The regional approach used above is based on the reliability and accu-  
641 racy of the geomagnetic field geometry of the *gufm1* model ([Jackson et al.,](#)  
642 [2000](#)). The implication on the axial dipole field moment's evolution between  
643 1590 and 1850, as mainly constrained by the available Triaxe archeointen-  
644 sity data, therefore depends on this reliability. However, several studies have  
645 highlighted discrepancies between archeomagnetic directional data and the  
646 directions predicted by the *gufm1* model (see for instance [Tanguy et al., 2011,](#)  
647 for the western Indian Ocean). In France, [Le Goff and Gallet \(2017\)](#) have  
648 also shown that while satisfactory consistency is observed after  $\sim 1675$ , the  
649 *gufm1* predictions differ significantly from most direct directional measure-  
650 ments prior to this date.

651 The relatively low reliability of the *gufm1* model during the 17th century  
652 and part of the 18th century should not be surprising, given that very few, if  
653 any direct inclination data are available before 1700 – 1750, and more gen-  
654 erally, given the poor spatial and temporal coverage of historical directional  
655 measurements between 1590 and  $\sim 1700$  (e.g. [fig. 1a,e Jackson et al., 2000](#)),  
656 in particular in Central Asia (see [fig. 8 to 14 in Jonkers et al., 2003](#)). This



657 calls for caution when interpreting recalibrated variations of axial dipole mo-  
658 ments as shown in Fig. 12b. However, at the scale of Western Eurasia (from  
659 Western Europe to Central Asia), the satisfactory modelling of the non-dipole  
660 effects in *gufm1* is evidenced by the good consistency, upon recalibration, of  
661 the Triaxe archeointensity datasets shown in Fig. 12a. In addition, it should  
662 be noted that this consistency is poorer when the data from Western Eu-  
663 rope and Russia are transferred to the latitude of Bukhara using a purely  
664 axial dipole field approximation (Fig. S3). Nevertheless, the clear non-linear  
665 dipole moment evolution deduced using the *gufm1* model in a region across  
666 which significant non-dipole field effects are not expected to occur over the  
667 historical period (e.g. Pavón-Carrasco et al., 2014b) does not demonstrate its  
668 truly dipole origin. For this, we need a large set of reliable and geographically  
669 distributed archeointensity data from around the world.

670 Away from western Eurasia, Hartmann et al. (2010, 2011) obtained in  
671 southern and northern Brazil coherent historical archeointensity data using  
672 the Triaxe protocol (so far this is the only Triaxe data obtained outside  
673 western Eurasia) thus sharing the same criteria as before. As pointed out  
674 by Hartmann et al. (2011), these results also show the *gufm1* model's lack of  
675 reliability for the period before  $\sim 1750$ . Note that based on an archeointen-  
676 sity result obtained in Ethiopia dated  $\sim 1615$ , Osete et al. (2015) also arrive  
677 at the same conclusion. On the other hand, using the same recalibration  
678 method as previously used by Genevey et al. (2009), the Brazilian data are  
679 consistent with a minimum of the axial dipole magnitude  $|g_1^0(t)|$  around the  
680 late 18th century, as shown by the western Eurasian Triaxe data (Fig. 12).  
681 This feature could thus represent a true dipole feature contradicting a linear

682 evolution of the  $g_1^0$  term over the entire historical period. At this stage, how-  
683 ever, we recognize that its global (dipole) nature has yet to be confirmed by  
684 the acquisition of new high quality archeointensity data.

## 685 6. Conclusions

686 The acquisition of nine new archeointensity data from Bukhara, Uzbek-  
687 istan using the Triaxe experimental protocol allows for reconstruction of the  
688 geomagnetic field intensity variations in Central Asia from the mid-16th to  
689 the beginning of the 19th century. The evolution derived from the new data  
690 is marked by a rapid decrease of the intensities by  $\sim 14 \mu\text{T}$  from  $\sim 1560$  to  
691  $\sim 1725$  followed by an intensity minimum during the late 18th century and  
692 then by an increase from the mid-18th to the beginning of the 19th century.  
693 Using the field geometry provided by the *gufm1* model, we show that these  
694 results are consistent with other Triaxe data previously obtained in western  
695 Europe and in northwestern-central Russia.

696 When these data are used to recalibrate the axial dipole coefficient given  
697 by the *gufm1* model, the resulting evolution appears non-linear over the his-  
698 torical period, with a clear minimum in magnitude of  $\sim 29400 \text{ nT}$  during the  
699 18th century. This trend contrasts with the linearity assumed by most global  
700 models so far. The validity of the global, dipolar nature of this analysis is  
701 contingent upon the reliability and accuracy of the field geometry provided  
702 by *gufm1*, both of which are well established from 1750 onward. The trend  
703 we find for  $g_1^0(t)$  can neither be satisfactorily confirmed nor refuted by the  
704 other regional datasets available in western Eurasia due to their dispersion.  
705 The sole data confirming the low of  $|g_1^0(t)|$  during the second half of the 18th  
706 century are Triaxe data from Brazil.

707 The conclusions we can draw from this study are twofold: first, it shows  
708 again that the acquisition and analysis of archeomagnetic data can provide  
709 useful information on the temporal behavior of the geomagnetic dipole on

710 those time scales close to the convective turnover time (around the junction  
711 between the TF and HF frequency bands discussed in the introduction);  
712 second, it stresses that that information could be particularly useful to better  
713 constrain the geomagnetic secular variation during the historical period prior  
714 to the observatory era.

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Figure 1: General location map of Uzbekistan and Bukhara



Figure 2: Example of buildings sampled in Bukhara: a) courtyard of Madrasa Abdullah Khan (BK03, 1578 – 1590), b) BK04: façade of Madrasa Modari Khan (1556 – 1567), c) BK08: façade of the Madrasa Rakhmanqul (1790 – 1795), d) BK05: tomb of Khwādja Saad in Chor Bakr (1589 – 1615), e) BK13: Mosque Kemuhtagaron (1700 – 1750), and f) BK14: *kānaqāh* inside the Ark citadel (1758 – 1785)

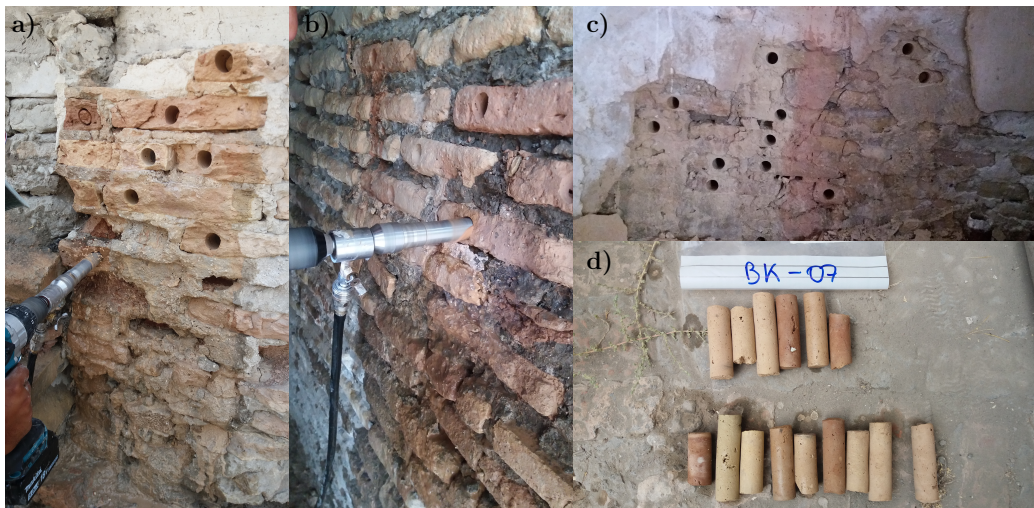


Figure 3: Examples of archeomagnetic sampling carried out in Bukhara: a) sampling of the Mosque Magoki Kurpa (BK12, 1631 – 1637), b) sampling of the Madrasa Rakhmanqul (BK08, 1790 – 1795), c) sampled wall in Madrasa Modari (BK04, 1556 – 1567), and d) cores sampled in the Madrasa Kunjak (BK07, 1700 – 1722).

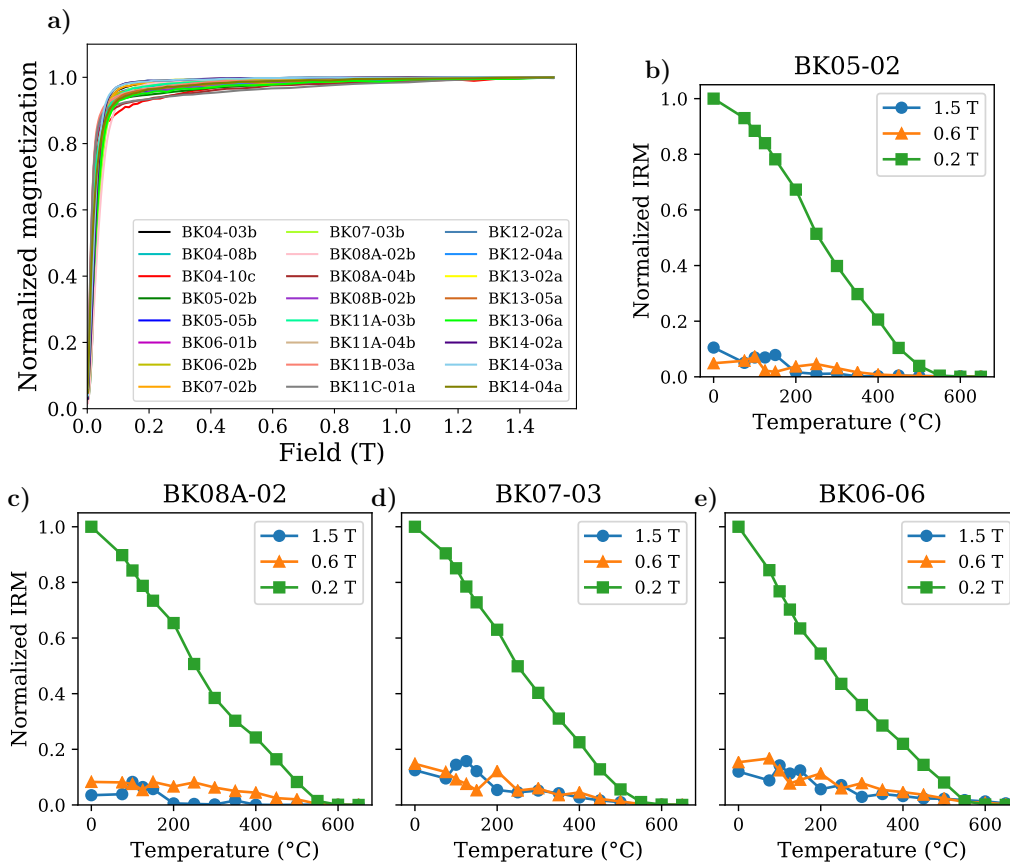


Figure 4: a) Normalized IRM acquisition obtained up to 1.5 T for 24 representative fragments of the selected groups, b-e) four examples of thermal demagnetization of 3-axis IRM acquired in orthogonal fields of 1.5 T (blue dots), 0.6 T (orange triangles), 0.2 T (green squares).

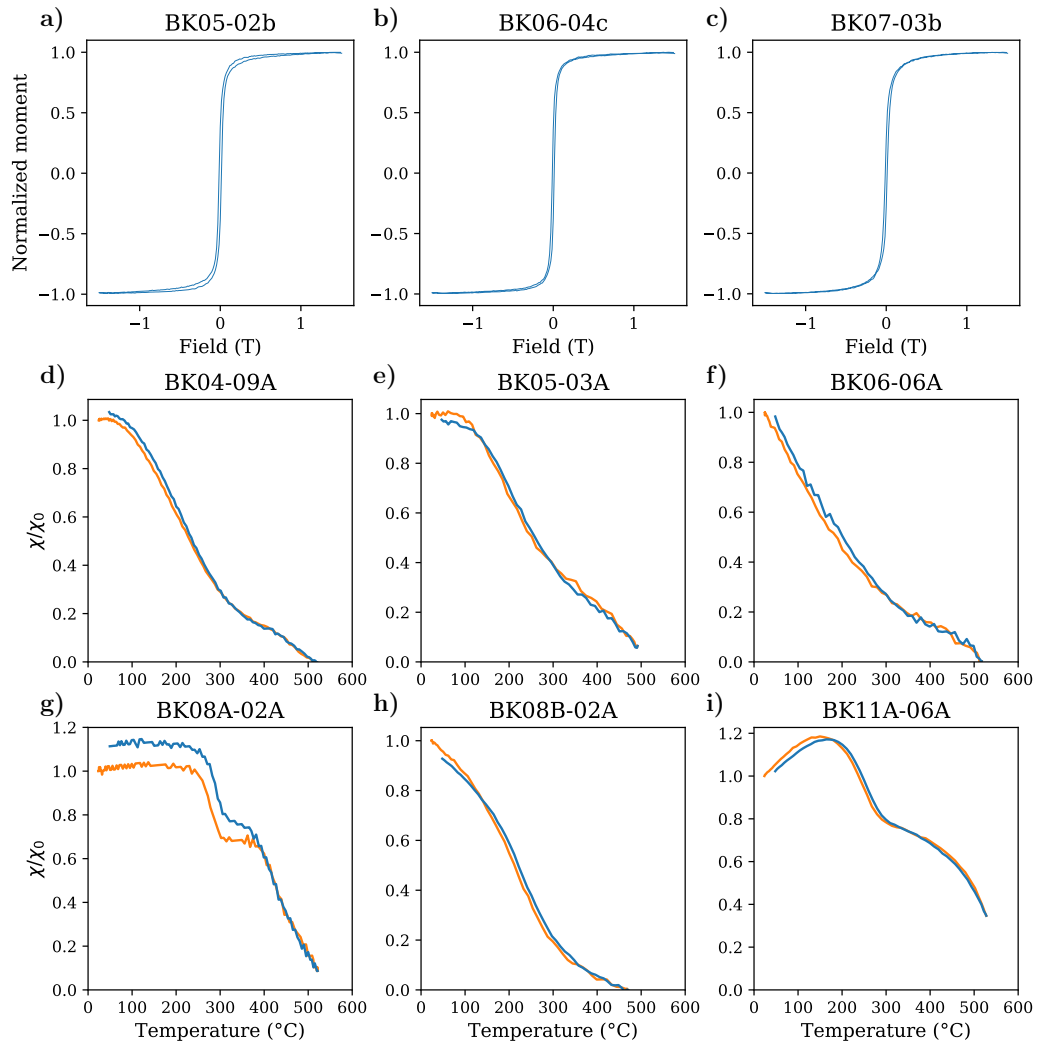


Figure 5: Examples (a-c) are hysteresis loops obtained for selected fragments where a) is an example of slightly constricted behavior, b-c) are examples of common behavior with narrow but open loops with a squared shape. d-i) are normalized low-field susceptibility vs. temperature curves obtained for some of the selected fragments up to  $\sim 500^{\circ}\text{C}$ . The orange curves (resp. blue) show the behavior during the heating (resp. cooling) step.



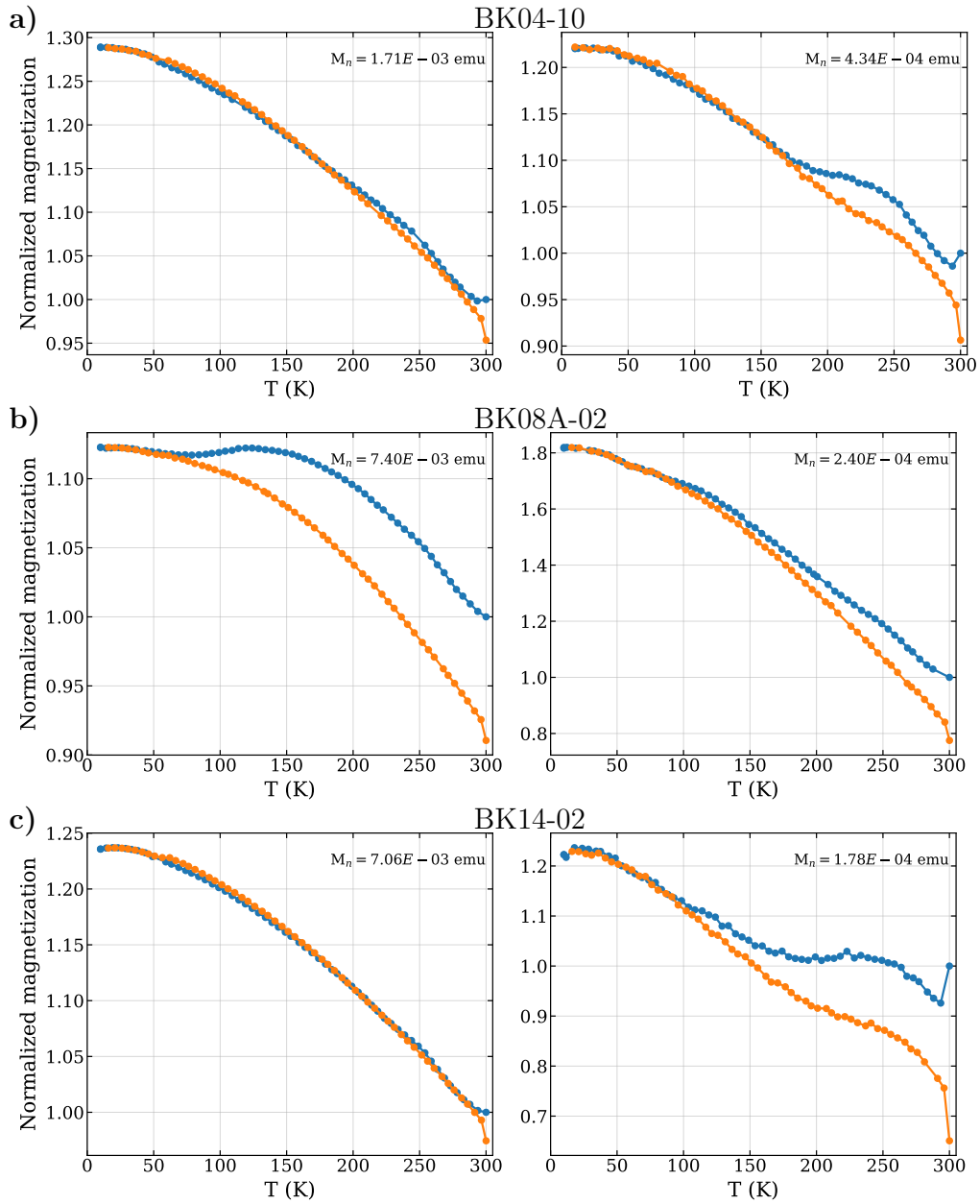


Figure 6: Representative examples of RT-SIRM cycles of a 2.5 T field IRM (left panels) and a 2.5 T IRM partially demagnetized (right panels) with a 300 mT field generated by the superconducting magnet operating in oscillation mode. The blue (resp. orange) dots correspond to the cooling (resp. heating) step. The results are normalized to  $M_n$  (corresponding to the initial RT-SIRM at 300 K).

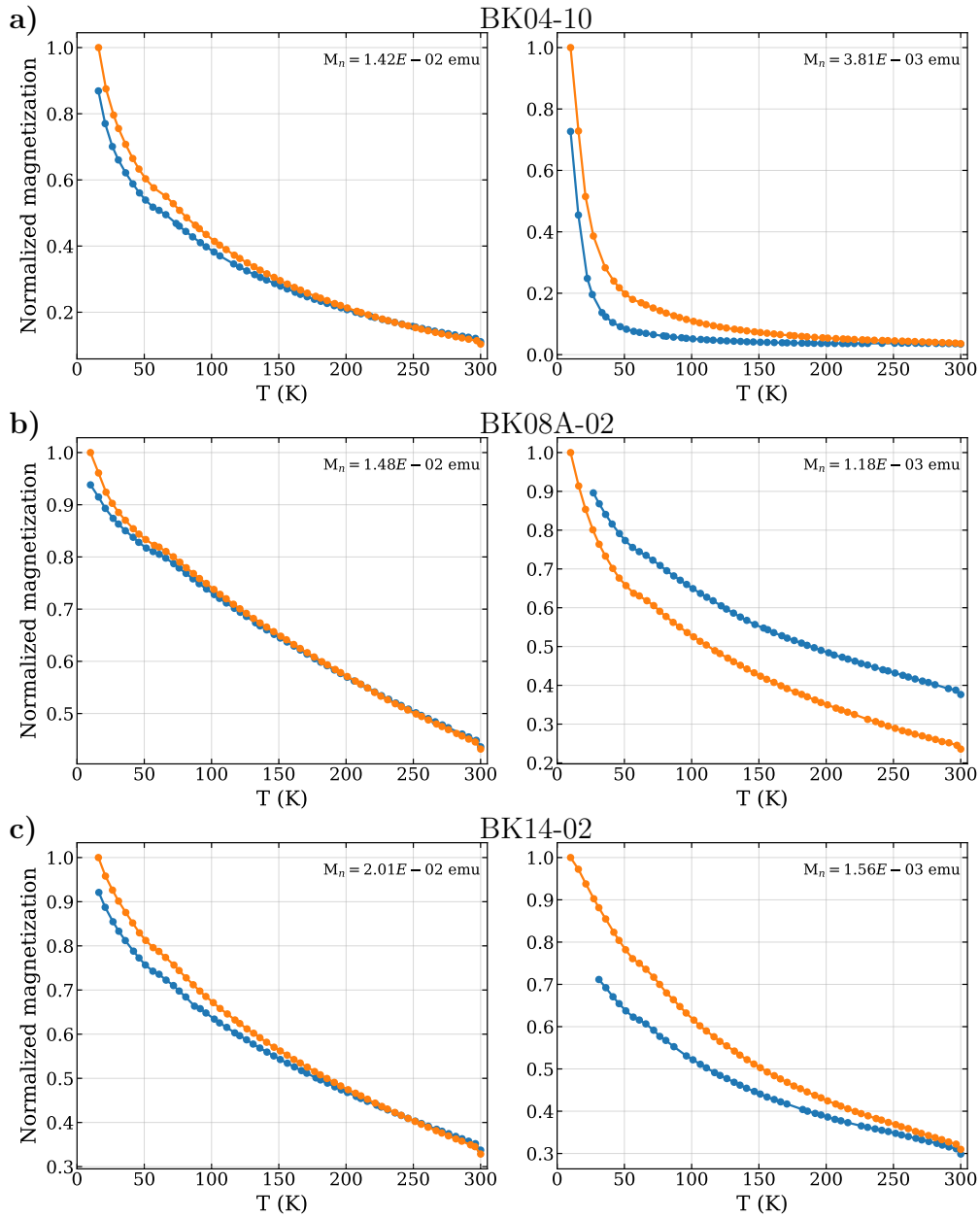


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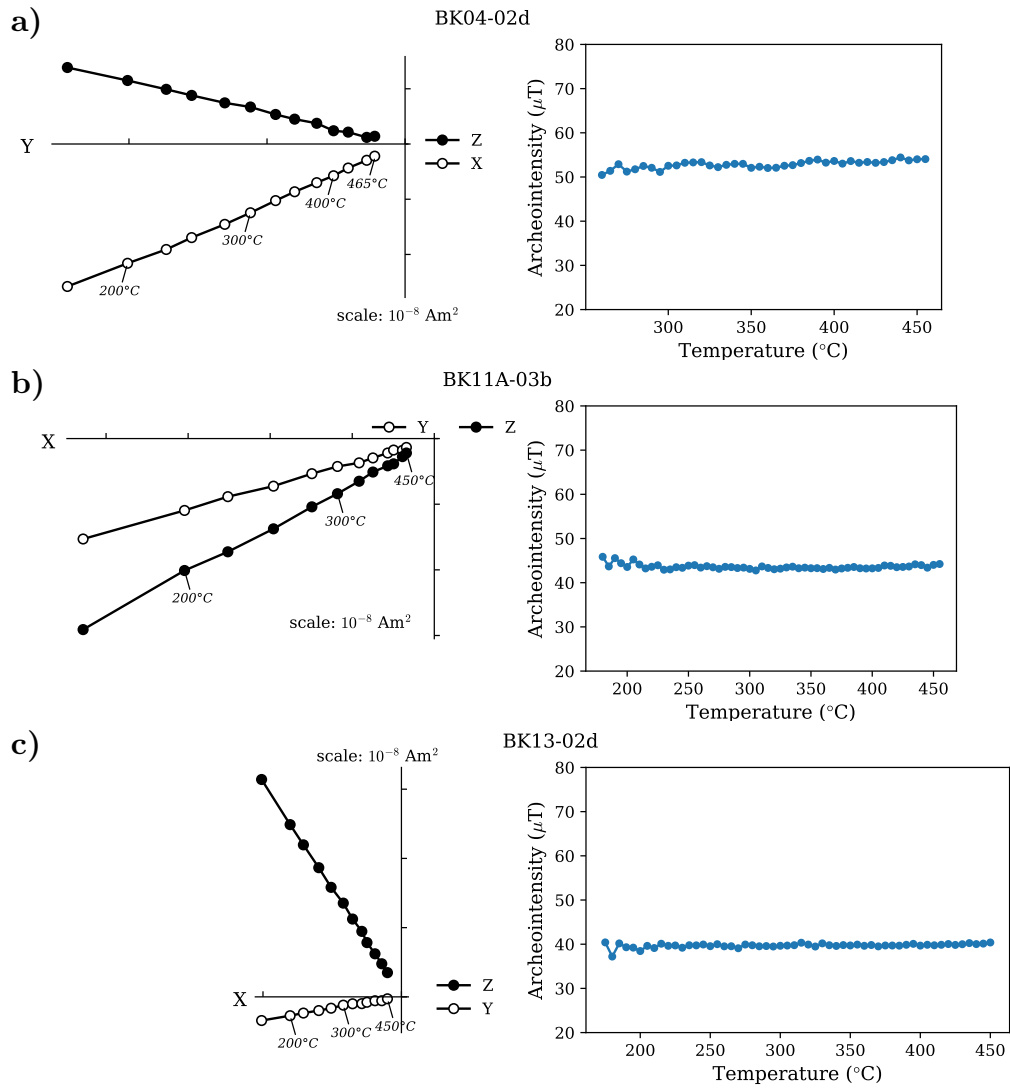


Figure 8: Left panels: thermal demagnetization data obtained for three different specimens. Open (close) symbols refer to the inclinations (declinations). Right panels: corresponding  $R'(T_i)$  datasets obtained from the same specimens (see in [Le Goff and Gallet, 2004](#))

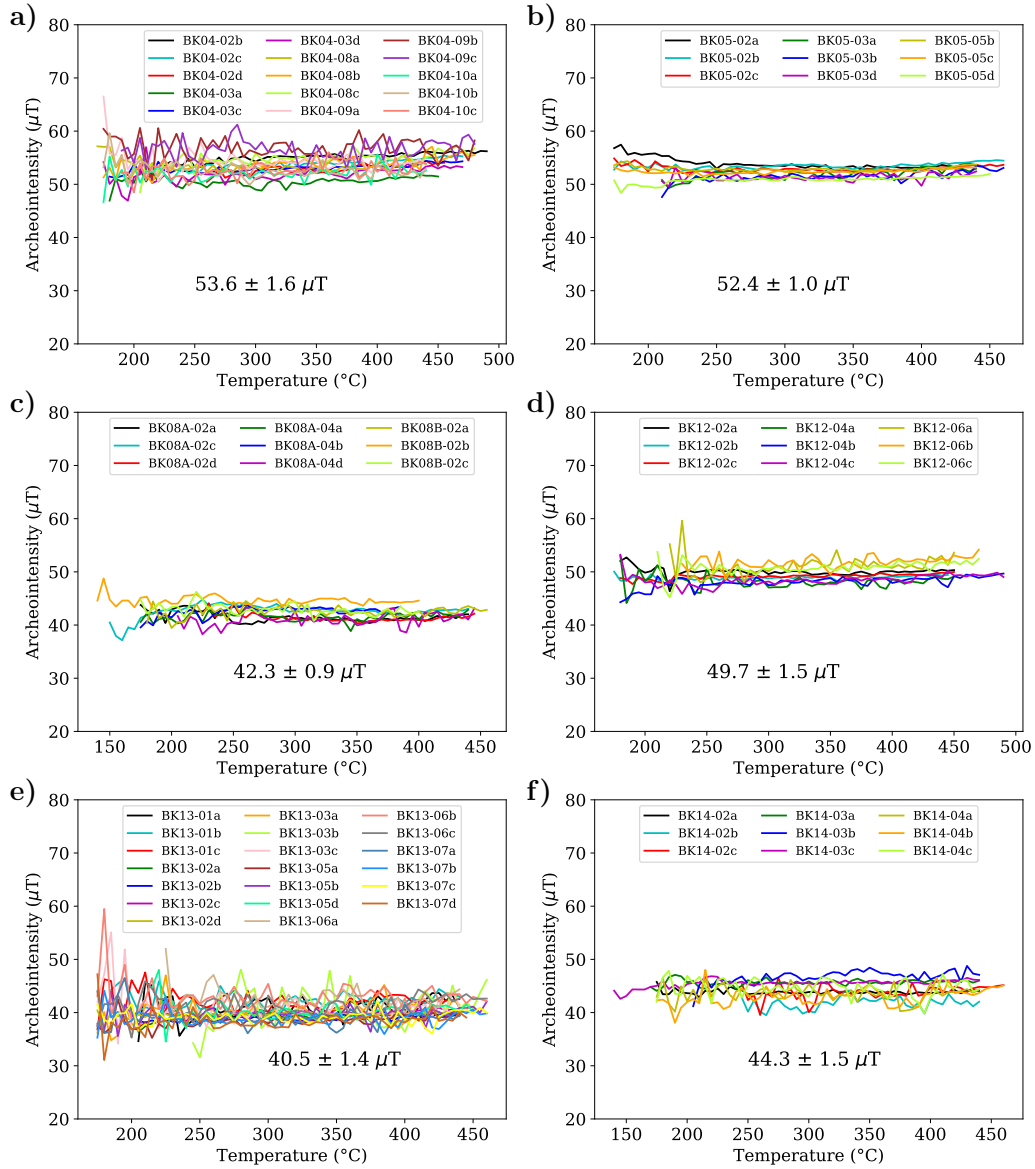


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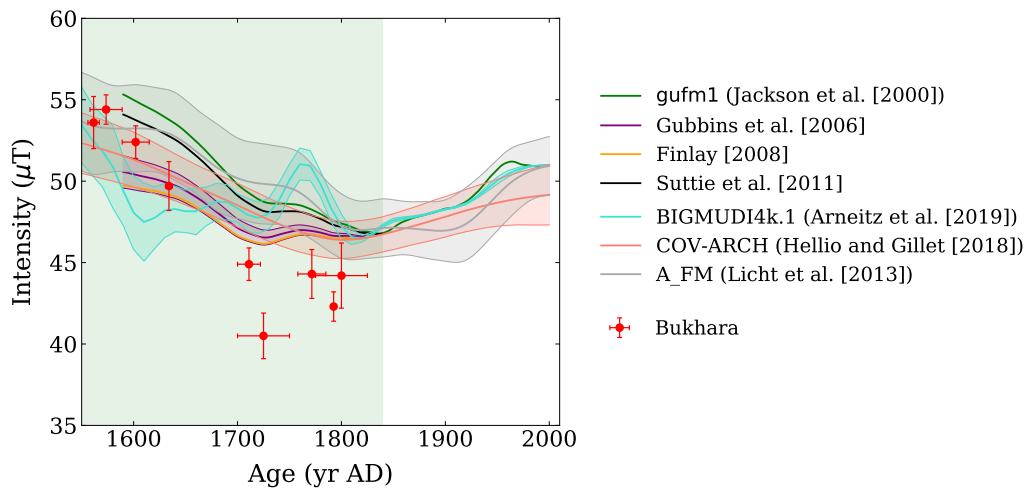


Figure 10: Archeointensity data obtained in Bukhara (red dots). These data are compared with intensity values predicted from different global field models (continuous lines, errors are given as two standard deviations by shaded areas; see legend in the figure).

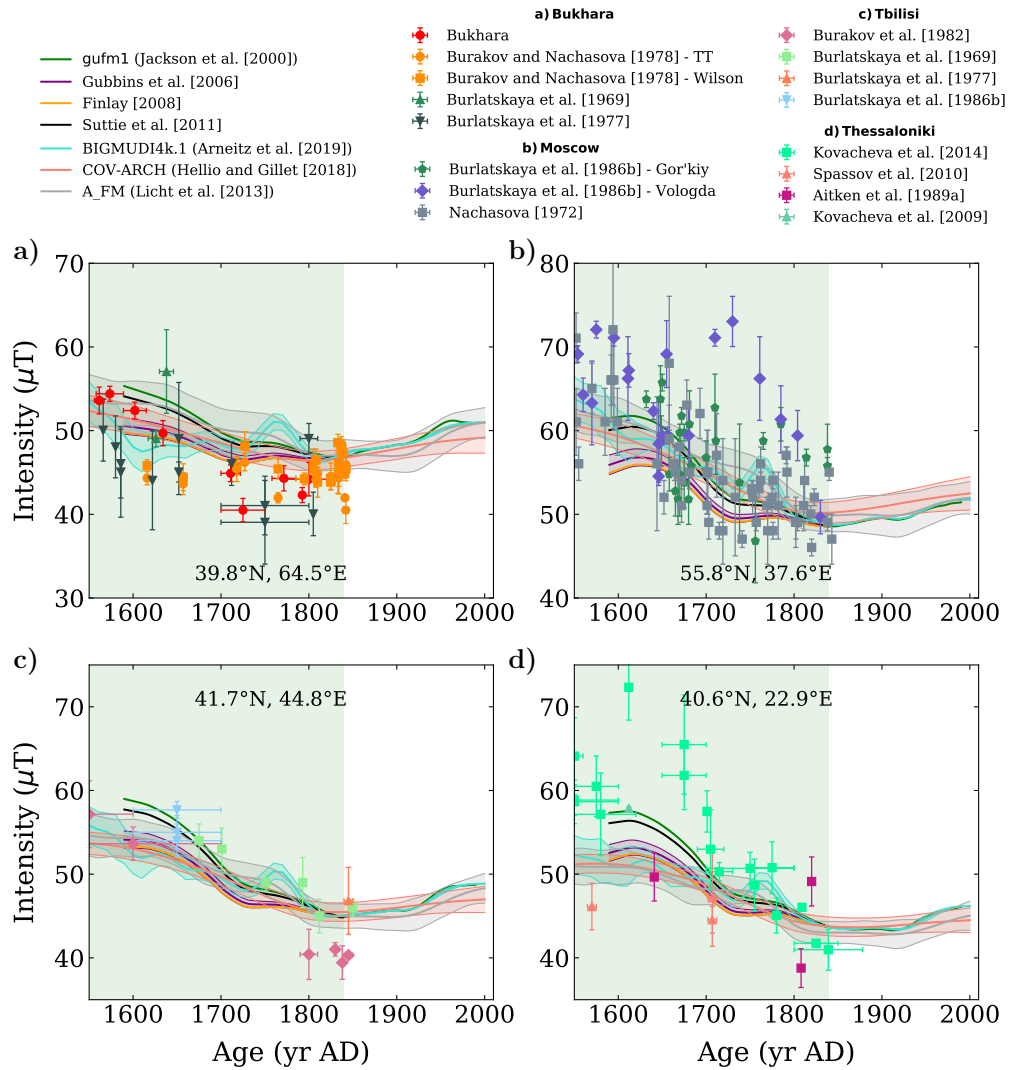


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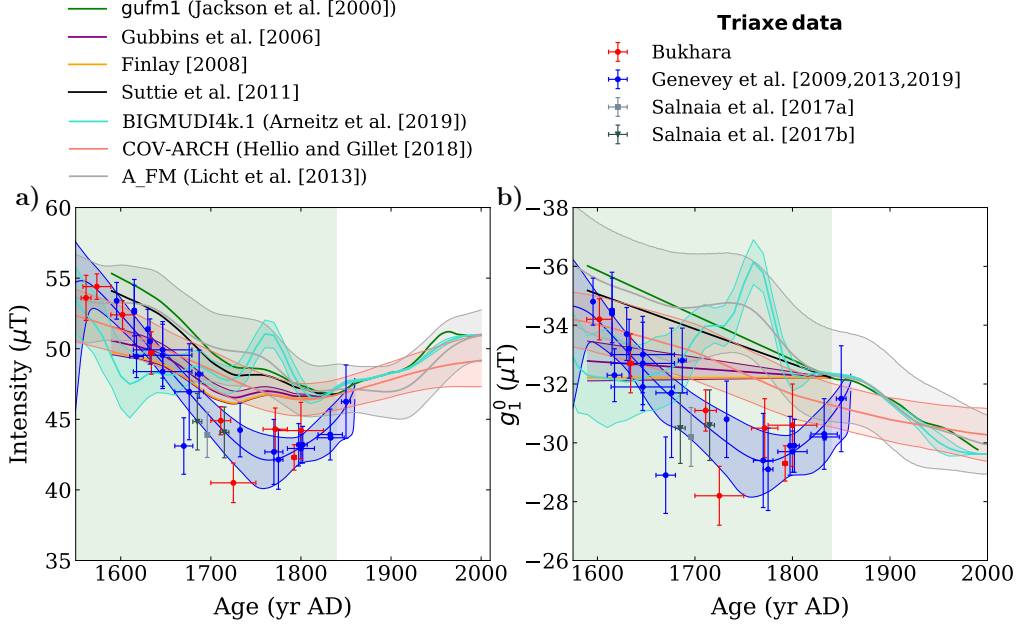


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Table 1: Mean archeointensity data obtained from nine groups of fragments collected at Bukhara. The historical context is indicated in the first column. The archeomagnetic reference of the groups of fragments is given in the second column. The dating of the context/group is provided in the third column. The number N of successful fragments (n specimens) used to compute the intensity value for each group is specified in the fourth column. The last column contains the corresponding mean archeointensity values.

Archeological site	Label	Age (yr AD)	N frag. (n spec.)	$F_{\text{mean}} \pm \sigma F$ ( $\mu\text{T}$ )
Madrasa Modari Khan	BK04	1556 – 1567	5(15)	$53.6 \pm 1.6$
Chor Bakr - Khwādja Saad tomb's wall	BK05	1589 – 1615	3(9)	$52.4 \pm 1.0$
Chor Bakr - Khwādja Saad tomb's ground	BK06	1558 – 1589	3(10)	$54.4 \pm 0.9$
Madrasa Kunjak	BK07	1700 – 1722	3(9)	$44.9 \pm 1.0$
Madrasa Rakhmanqul	BK08	1790 – 1795	3(9)	$42.3 \pm 0.9$
Madrasa Rashid-al-Din	BK11	1775 – 1825	7(22)	$44.2 \pm 2.0$
Mosque Magoki Kurpa	BK12	1631 – 1637	3(9)	$49.7 \pm 1.5$
Mosque Kemuh-tagaron	BK13	1700 – 1750	6(20)	$40.5 \pm 1.4$
Ark - k̄anaqāh	BK14	1758 – 1785	3(9)	$44.3 \pm 1.5$