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Palaeomagnetic analysis of the Sterkfontein palaeocave deposits: Implications for the age of the hominin fossils and stone tool industries

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ABSTRACT

Palaeomagnetic analysis was conducted on speleothems from Members 1–5 at Sterkfontein Cave, South Africa. Palaeomagnetic analysis of siltstone and speleothem from the bulk of Member 4 indicate a reversed magnetic polarity that dates the deposits and its *Australopithecus africanus* fossils to between 2.58 and ~2.16 Ma. Further confirmation of this age comes in the form of two short normal polarity events correlated to the Rèunion (~2.16 Ma) and Huckleberry Ridge (~2.05 Ma) events in speleothem capping the bulk of Member 4 and coeval with deposition of the final phase of Member 4, including *A. africanus* fossil Sts 5. At ~2.16–2.05 Ma, Sts 5 is the youngest representative of *A. africanus* yet discovered. Palaeomagnetic analysis of the Silberberg Grotto deposits identifies a single short geomagnetic field event in flowstone overlying the StW 573 *Australopithecus* fossil, which is suggested to represent the Rèunion event at ~2.16 Ma. This further supports the uranium lead age estimates of 2.3–2.2 Ma for the StW 573 fossil. Based on a reversed polarity for the deposits below the skeleton it cannot be older than 2.58 Ma. If StW 573 is considered to be a second species of *Australopithecus* then this indicates that two species of *Australopithecus* are present at Sterkfontein between 2.6 and 2.0 Ma. All of the Member 5 deposits date to less than 1.8 Ma based on a comparison of palaeomagnetic, faunal, and electron spin resonance age estimates. The StW 53 fossil bearing infill (M5A) is intermediate in age between Member 4 and the rest of Member 5 (B-C) at around 1.78–1.49 Ma. The rest of Member 5 (B-C) containing Oldowan and Acheulian stone tools and *Homo* and *Paranthropus* fossils were deposited gradually between 1.40 and 1.07 Ma, much younger than previously suggested.

Introduction

The cave site of Sterkfontein, near Krugersdorp in the Gauteng Province of South Africa (Fig. 1), has so far yielded the remains of several hundred hominin specimens including early *Homo*, *Paranthropus*, and one or more species of *Australopithecus* (Tobias, 2000). Oldowan, Acheulian, and Middle Stone Age assemblages have also been recovered from the site and together with the range of hominin fossils indicates the long life history of the caves, which are still active and forming new passages at the site today. Accurate dating of the site has remained problematic and stratigraphic relationships poorly understood due to the removal of material during mining at the site and a lack of exposed sections of stratigraphy linking the various fossil-bearing deposits. Partridge (1978, 2000) identified six Members (M1-M6) at the site.

A number of dating techniques have been attempted with varying success. Faunal dating has provided very mixed views on the age of the deposits (Delson, 1988; Vrba, 1995; Berger et al., 2002). Palaeoenvironmental interpretations of a wooded environment for Member 4 (Bamford, 1999) in part also led to a general view that this deposit must date to greater than ~2.5 Ma, a period of aridification in eastern Africa (deMenocal, 1995). Preliminary electron spin resonance (ESR) dating suggested ages younger than was generally expected for both Member 4 (M4) and Member 5 (M5) and suggests a high degree of mixing of fossil material or inaccurate identification of temporally distinct deposits has occurred at the site (Schwarcz et al., 1994; Curnoe, 1999). Cosmogenic nuclide burial dating has suggested older dates (4.52–3.72 Ma [4.17 ± 0.14(0.35) Ma]) for the StW 573 bearing layer of Member 2 (M2; Partridge et al., 2003), while uranium–lead (U–Pb) dating has suggested ages much younger than this for the same deposit (2.33–2.06 Ma; Walker et al., 2006).

A number of problems have been documented when conducting palaeomagnetic analysis on breccia deposits at Sterkfontein and other South African hominin sites (Jones et al., 1986; Schmidt and Partridge, 1991; Herries et al., 2006a,b; Adams et al., 2007). Previous palaeomagnetic analysis has been conducted on both the

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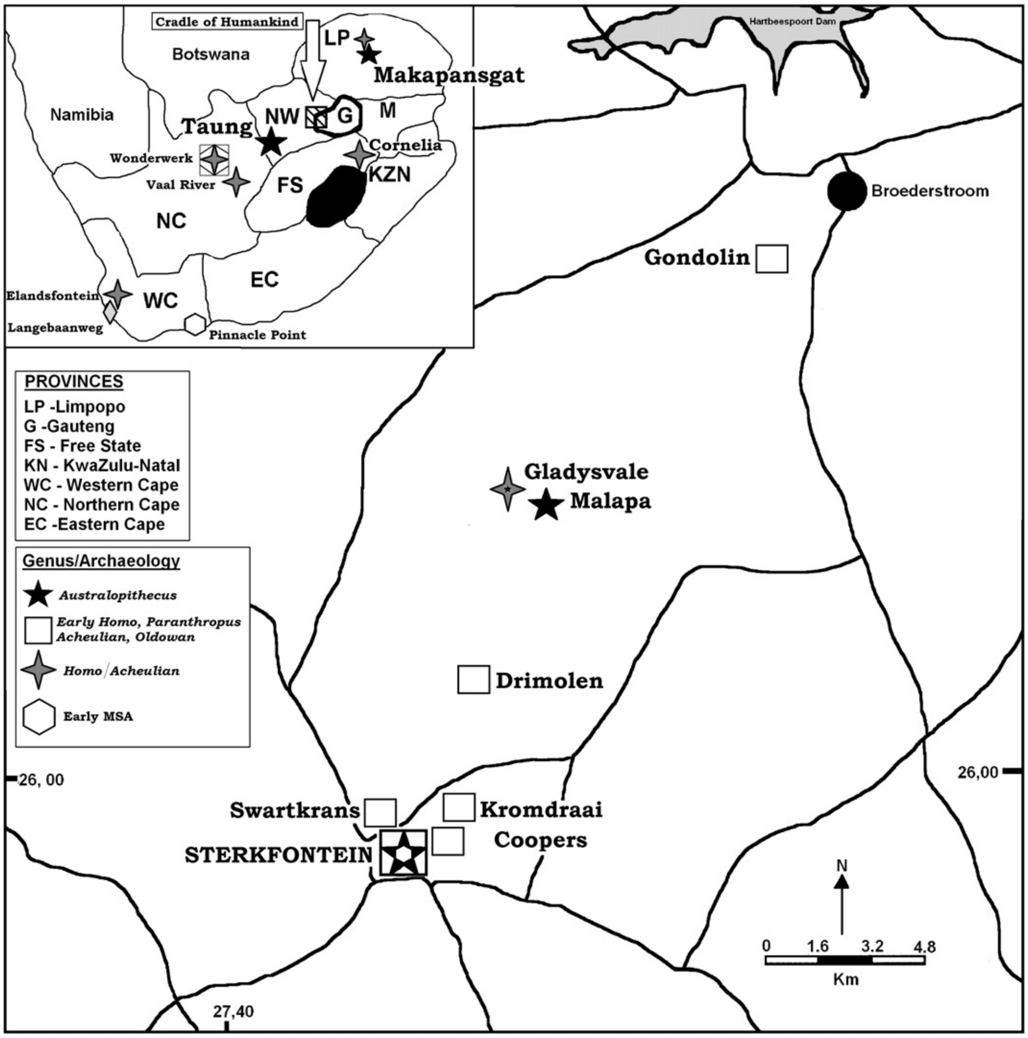

Fig. 1. Locality of the Sterkfontein hominin site in the Cradle of Humankind World Heritage area, Gauteng, South Africa, and its relation to other hominin sites mentioned in the text.

clastic and speleothem deposits at Sterkfontein (Jones et al., 1986; Schmidt and Partridge, 1991; Partridge et al., 1999). Initial analysis of clastic deposits by Jones et al. (1986) suggested that they were generally unsuitable for holding a stable geological magnetic remanence, either due to depositional environments or sediment source. Later analysis of speleothems deposits did prove successful (Partridge et al., 1999) but, as speleothem occurs sporadically in some sections of the sequence, such a magnetostratigraphy will contain gaps of unknown duration. However, deposition of clastic deposits occurs at a much quicker rate than speleothem and so such problems are suggested to be minimal. Brecciated deposits often provide overall random directions of magnetisation due to different clasts within the breccia having different, independent directions of magnetisation (Adams et al., 2007). Moreover, breccia is formed by collapse and so the magnetic grains do not have time to orient themselves with the Earth's magnetic field during the process of deposition. This is thought to explain some of the problems with the earlier work of Jones et al. (1986). As fine clastic sediments in the drip water settle out of suspension they are cemented into position, and magnetic remanence lock-in times are estimated to be of no more than a few years' duration in speleothems, unlike soft sediments and breccias where a post depositional remanent magnetisation or randomising of the magnetic signal can result (Latham and Ford, 1993). After cementation it is very unlikely that modern detrital grains will contaminate these dense, strongly cemented strata, and post depositional movement is less likely to occur unless the whole speleothem structure is moved. The lack of detrital contamination is shown during sectioning for preparing the palaeomagnetic samples and by thin section work conducted for isotopes (Hopley, 2004; Hopley et al., 2009). For this reason, this study was primarily undertaken on the speleothem portion of the deposits as per Partridge et al. (1999). The palaeomagnetic analysis of speleothems alone is based on an assumption that the intervening siltstone and breccia records only very short time periods. However, while the collapse of clastic deposits forming breccia cannot be used for palaeomagnetic analysis, if this material is reworked and deposited by fluvial processes, as at Makapansgat, Gondolin, and Malapa, this material can provide reliable palaeodirections (Herries, 2003; Herries et al., 2006a,b; Adams et al., 2007; Dirks et al., 2010). As such, comparative work has now been undertaken on siltstones associated with the sampled flowstones at Sterkfontein. Recent studies have shown that when breccia deposits are avoided and fine-grained siltstones and speleothem sampled, the fossil direction of the Earth's magnetic field can be recorded from these palaeocave sites and such analysis is consistent with other geochronological methods such as U–Pb (Dirks et al., 2010).

The main issue for palaeomagnetic studies has been the correct understanding of the stratigraphy at the various palaeocave sites.

Initial palaeomagnetic analysis of the Silberberg Grotto deposits estimated an age of between 3.60 and 3.22 Ma with StW 573 dating to ~3.33 Ma based on expected depositional rates (Partridge et al., 1999). This age assessment is thus intermediate between those of the U–Pb and cosmogenic radionuclide burial dating ($^{26}Al/^{10}Be$). However, the palaeomagnetic age assessment of Partridge et al. (1999) is based on an assumption that the deposits were >3.0 Ma due to their depth beneath the surface exposed fossil deposits (Member 4 and 5) and due to faunal comparisons. However, the former argument is based on unsound principles. For a cave system that has been active for a number of millions of years the Sterkfontein system is, laterally and vertically, extremely confined when compared to cave systems throughout the world (pers. obs.) and compared to sites such as the Makapansgat Limeworks (Fig. 2; Latham et al., 1999, 2003). It is a complex multi-level, multi-period system, with the current active system depositing recent material below many much older deposits, causing an inverted age stratigraphy. This indicates the potential flaws of a layer cake-like reconstruction of the deposits. Water appears to have continuously reactivated earlier palaeokarstic conduits eroding out, into, and under older deposits, and this can cause mixing of sediments, fossils, and stone tools from multiple phases. Because the caves are in dolomite, which is much less soluble than calcite or limestone, the water will preferentially re-erode back into the calcified deposits. Examples of resolution of earlier calcified deposits are the Makondo-karren (formed by solution around tree routes beneath soil cover) in the Central Debris Pile deposits at the Makapansgat Limeworks. A similar situation to Sterkfontein can be seen at the Jenolan caves in Australia where Quaternary aged cave passages cross cut much older palaeokarst, preferentially eroding out palaeocave sediments and using existing passages and joint structures (pers. obs.). However, unlike Jenolan, where this is obvious because the limestone has been tilted between karstification phases, there has been no significant uplift or tilting at Sterkfontein. As such, the use of the depth of the deposits within the system as an indicator of age is fundamentally flawed. This very situation is shown by the formation of the M4 and M5 deposits at the same level with the latter eroding into the former.

Partridge (1978, 2000) identified six Members (M1–M6) and subunits (Member 4 and 5 both have subunits A–C) based mainly on variations in sedimentological characteristics, stratigraphic associations, and depth within the lithostratigraphic column that he developed for the site (Fig. 3). Additionally, Kuman and Clarke (2000) recognised a post-Member 6 infill with MSA and the deposits of Lincoln Cave. Partridge's interpretations were based on five small circumference boreholes taken across the site (Partridge and Watt, 1991; Figs. 3 and 4). As Wilkinson (1985) noted, the relationship of the various deposits and in some cases their characterisation is uncertain and not all the deposits were included and so the formation of a composite formal stratotype was perhaps premature. A Formation with a Member system should consist of a series of sequential deposits of known age relationship, Member 1 being the oldest. Research at Makapansgat (Latham et al., 1999, 2003) has shown how the use of a Member system can become complicated and confusing when this is not known with certainty. Despite these problems the exposed M4 and M5 deposits are on the whole well-defined, and it is the relationship of deposits outcropping at different levels of the cave system, those that do not outcrop except in bore holes and the idea of a layer cake-like stratigraphy that continues to be a matter of debate (as depicted in Fig. 3; note that this sequence is not exposed anywhere at the site). The Member terminology of Partridge (2000) is used here in a form that denotes no suggestion of temporal relationships of Members in the classical sense of a Member system. It should be noted, however,

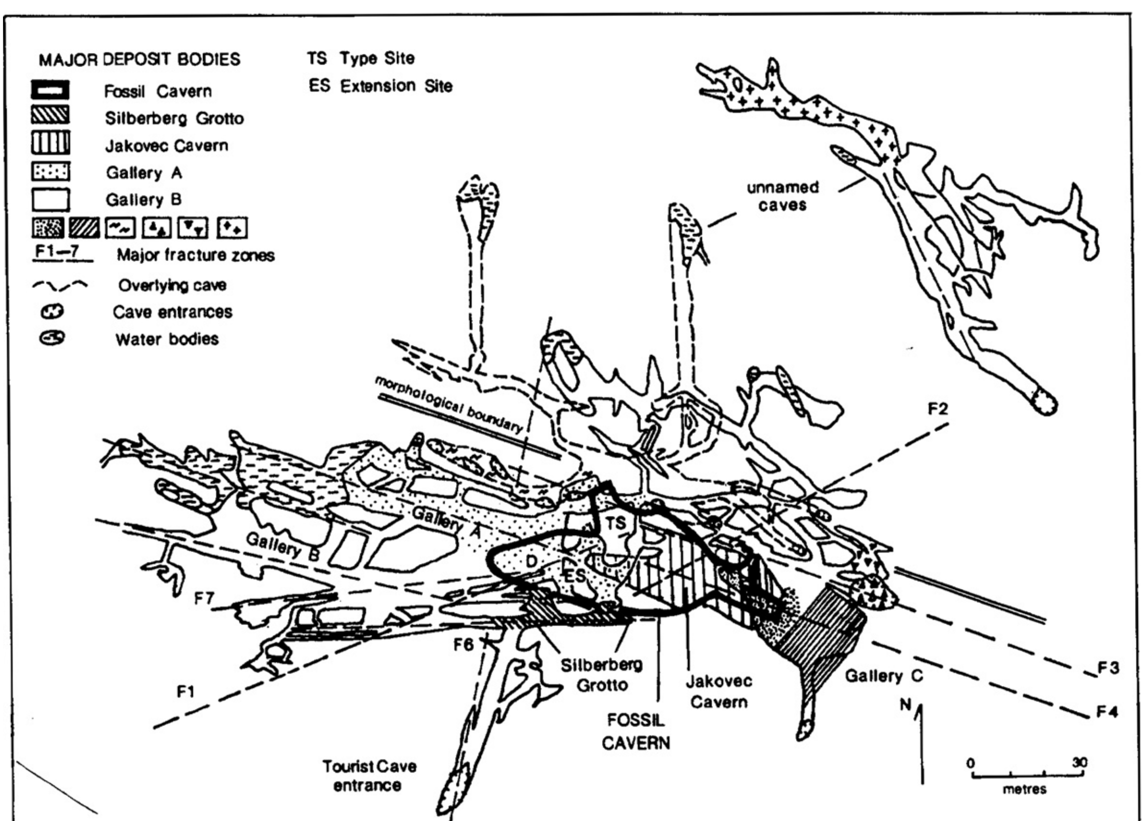

Fig. 2. Plan of the Sterkfontein System after Wilkinson (1985).

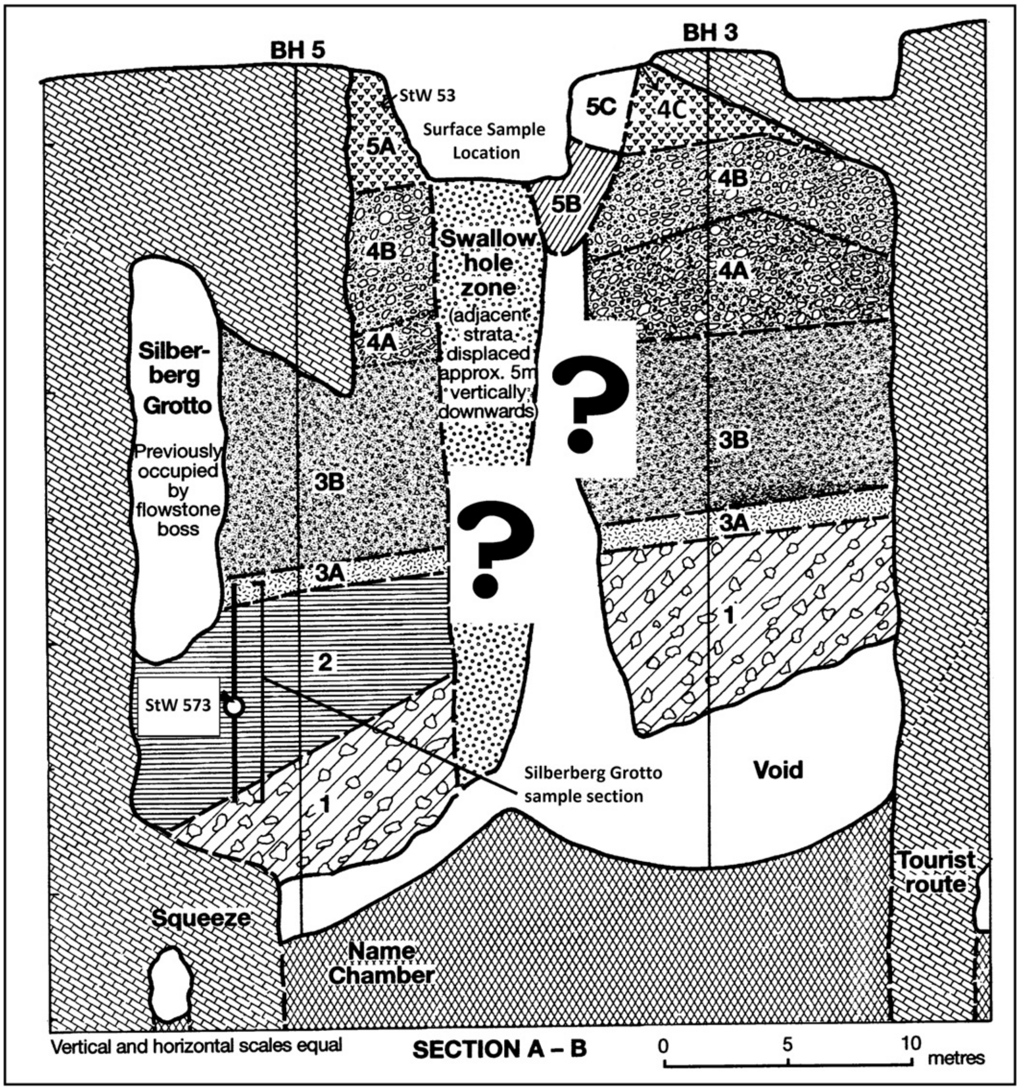

Fig. 3. Proposed section through the Sterkfontein deposits after Partridge (2000). The location of boreholes, exposure of Members, and sampling location of the palaeomagnetic study by Partridge et al. (1999) are shown.

that Kuman and Clarke (2000) use a different system of naming the various deposits, with M5A termed the 'StW 53 infill', M5B is termed the 'Oldowan infill,' and M5C is split into 'M5 east' and 'M5 west.'

From the point of view of geological exposures the Sterkfontein system can be separated on the basis of surface exposed deposits (Fossil Cavern on Fig. 2; Figs. 3 and 4) and subterranean exposed deposits (Silberberg Grotto and Jackovec Cavern in Figs. 2–4). Members 1–3 (M1–3) are exposed in the subterranean exposures and Members 4–6 (M4–6) in the surface exposures (Partridge, 1978). Based on these five bore cores taken across the site, Partridge and Watt (1991) reconstruct the cave as an essentially layer cake-like deposit with the youngest deposits occurring higher in the sequence and older deposits lower in the sequence (Fig. 3). Interpretations from such cores are difficult given their small diameter, and there is significant potential for vast changes in deposition across the site to go unrecorded. As Wilkinson (1985) notes, deposits of a similar depositional character are noted in both the surface exposures and in the subterranean caverns and yet such deposits are separated horizontally by $\sim$30 m. The complexity of cave formation and deposition makes it uncertain if the similar looking deposits identified in exposed sections and in the cores are indeed the same deposits. Similar processes can occur in different parts of the same cave system both at the same and different time periods creating deposits of a similar character.

The likelihood of a layer cake-like deposition occurring is decreased by the complexities of large speleothem formations and vertical fissuring. The vertical fissuring forms steep dolomite walls that results in disconnected sediment traps which inhibit similar-aged, laterally extensive deposition at the site. Such a situation is also seen with Partridge's (1979) classification of Member 2 at Makapansgat. While the red siltstone deposits of the eastern and western quarries are sedimentologically similar, they are laterally separated by over 100 m as well as the entirety of the massive CDP and a speleothem ridge. Furthermore, at Sterkfontein Stratford (2008) has recently described stone tool bearing deposits within the Name Chamber, which is located deep beneath the surface exposures (Fig. 3). These deposits have been eroded from the M5 deposits and deposited in deeper chambers as a series of talus cones. Similar situations can be seen all over the Sterkfontein system and as such a similar situation could be envisaged for the Silberberg Grotto deposits. Such examples show the danger of classification of deposits merely on sedimentary character. Most caves have more than one entrance that can deposit sediment, and similar

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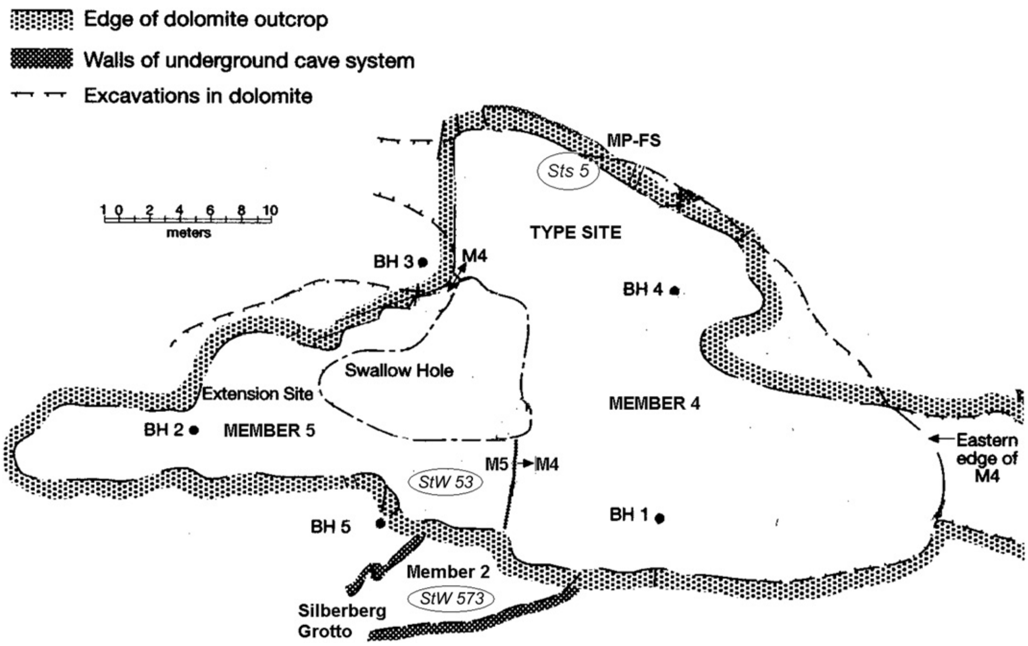

Fig. 4. Plan of Sterkfontein showing the positions of boreholes, Member exposures (M2–5), and hominin fossil recovery locations (StW 573, StW 53, Sts 5; modified from Partridge, 2000).

depositional processes may occur in different parts of the cave at different time periods. Such a model has already been suggested for the Jakovec Cavern at Sterkfontein (Partridge et al., 2003).

Hydrologically, Sterkfontein is part of a series of caves (e.g., Kromdraai, Coopers), now mostly inactive, that occur along either side of the Bloubank River Valley. Analysis of the active part of the cave system by Wilkinson (1985) shows that despite the highly fractured nature of the dolomite, which can cause the false impression that a flat water table exists at the site, a piezometric gradient does occur across the site with the water level fluctuating along passages and from chamber to chamber. Such processes can cause the formation of caverns at different levels at the same period. Moreover, active caverns may occur in lower areas when older upper caverns are collapsing and infilling with surface sediment forming inverted stratigraphies. In such circumstances the similar-aged sediments may have markedly different sedimentological characteristics. This is epitomised at the Makapansgat Limeworks where large clast supported breccia fills the middle of the site (M4/Central Debris Pile; Partridge, 1978, 2000) at the same time that siltstone and *Australopithecus africanus* bearing Member 3 deposits are infilling the western Central Quarry area (Latham et al., 1999, 2003) between 3.0 and 2.6 Ma (Herries, 2003; Hopley et al., 2007a; Herries et al., in press). The complexity of these South African karstic systems can also be seen at Gladysvale, where a more recent cave system has formed within the older palaeocave deposits. This relationship can be seen at the very bottom of the current series of cavities where the walls of the cave are made entirely of breccia (Herries, 2003). A similar situation can again be seen at Peppercorn's Cave at Makapansgat (pers. obs.). For these reasons the current study views the site as a series of disconnected short sections with no pre-determined stratigraphic relationships while using the terminology of the Partridge (1978, 2000) lithostratigraphy.

Methods and sampling

The location of palaeomagnetic samples (STER-1–24 and A1–A8) from Sterkfontein is shown in Figure 5 (and Table 1) and was compiled by Tim Partridge. The samples cover M2 and M3 from the Silberberg Grotto and M4 and M5 from the surface exposures. This includes a thick speleothem layer that formed at the end of M4 deposition when *A. africanus* fossil Sts 5 was deposited. This is referred to as the Mrs Ples Flowstone (MPFS). Each subunit of M4 is capped by a speleothem lens made up of multiple calcite rafts, indicating the existence of localised pools at the close of this cycle of sedimentation. Speleothem samples were taken from throughout the sedimentary deposits of M4 (STER-10-15) and particular emphasis was placed on the MPFS deposit (STER 12, 15 and STER-A1–A8) located in the Type site area (marked Sts 5 in Fig. 4). Only a limited number of samples could be taken from M5, one from M5A (STER-17) and two from M5C (STER-18–19). For a comparison and to make sure the deposition of flowstone and sediments were coeval, a series of samples were also taken from fine-grained sediments attached to the speleothem blocks.

Two methods of sampling were used at the site, block sampling and drill coring. A magnetic compass provided orientation for both methods. Subsequent corrections were made for the declination of the local field according to the International Geomagnetic Reference Field accessed through the British Geological Survey (available at http://www.geomag.bgs.ac.uk/gifs/igrf.html). Due to a potential loss of orientation and the possible creation of drilling induced remanent magnetisations, as shown by previous work (Partridge et al., 2000; Herries, 2003), the results from these cores (which sampled inaccessible parts of the deposit) were only used as a guide. Those from M3 have been discarded due to a lack of stratigraphic association and the nature of some of the speleothem from the core suggests it is a dripstone that may have formed on a roof or wall rather than a floor deposited flowstone deposit. The context of those from the floor of the Silberberg Grotto (STER-20, 21) is more certain, and these were used to extend the base of the Partridge et al. (1999) sequence, whose blocks were also resampled and the original date reanalysed. (These blocks [STER-01-09] were stored in the basement of the U. Liverpool Geomagnetism Laboratory). In total, 27 block samples and 5 bore hole core samples were recovered. From these, 158 standard 25 × 20 mm palaeomagnetic subcores were drilled vertically in a zero magnetic field environment. This was undertaken so that

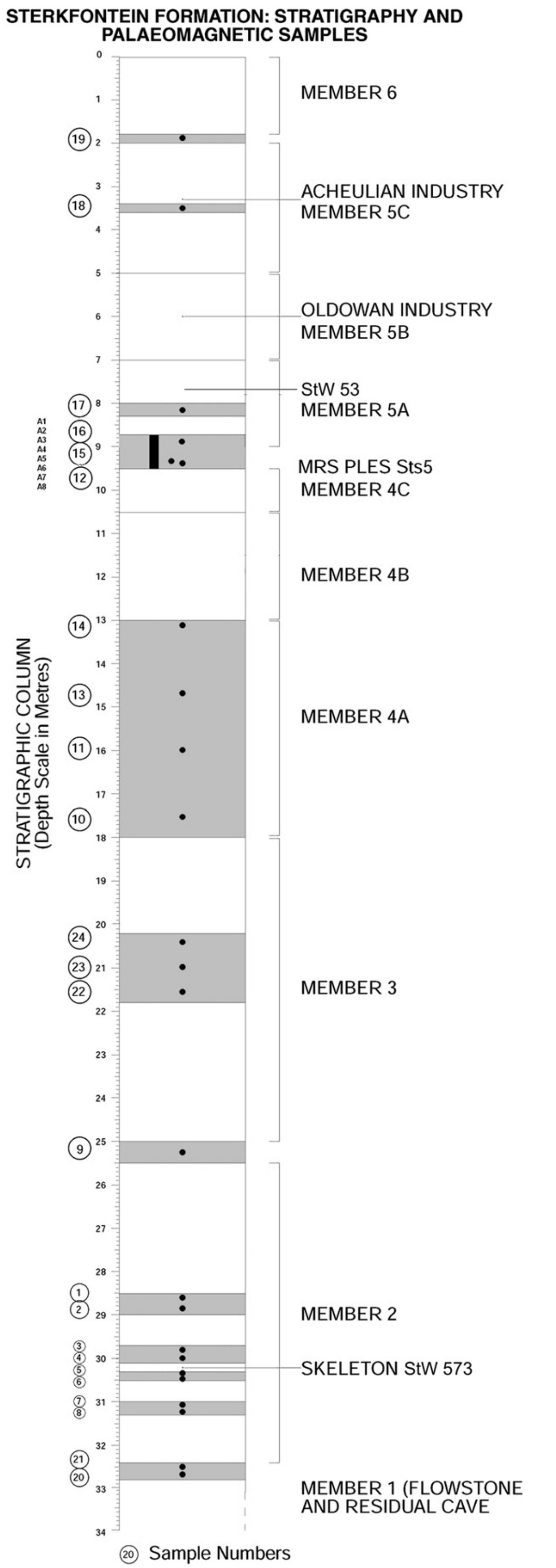

Fig. 5. Location of palaeomagnetic samples (STER-1–24 and A1–A8) based on composite stratigraphy compiled by Partridge (2000).

variations through the depth of each speleothem block could be examined in the context of the overall magnetostratigraphy in order to potentially reveal changes in polarity within a single block. If possible, two to three sets of samples were taken from each level in each block. This resulted in the measurement of an average of 4–6 subsamples per original block sample and at least 2–3 from each sublevel. In some cases the remaining stratigraphy and weakness of some samples did not permit more than two subsamples to be measured from each layer. In such circumstances the overall consistency of the block sample from which it came were used as an additional guide for reliability.

Measurements were made using an in-house modified FIT high temperature SQUID-based magnetometer and a dual speed JR6 magnetometer. Due to the fact that speleothems were being studied, magnetic cleaning to identify the characteristic remanent magnetisation (ChRM; main remanence preserved in the sample after the removal of any secondary magnetisation) was primarily undertaken by stepwise alternating field demagnetization (AFd), rather than thermal demagnetisation, which can cause expansion of the calcite and cracking at medium to high temperatures. AFd was undertaken in 2.5–5 mT steps using a laboratory-built reverse tumbling alternating field demagnetizer capable of imparting fields as high as 100 mT. Siltstone samples were also subjected to thermal demagnetization (THd) in 40–50 °C temperature steps up to 700 °C. After magnetic cleaning, ChRMs were determined using principle component analysis (Kirschvink, 1980) with vector and stereographic projections to determine declination (orientation in horizontal plain) and inclination (orientation in the vertical plane). Samples were considered good if they had a MAD (maximum angular deviation) value of <10 but were accepted with values of <15. The polarities of subsamples were assigned to normal (N), reversed (R), or intermediate (I) polarity according to their palaeopole positions as determined by the program Fish98. This produced a sequence of polarity intervals and reversals that were then correlated to the Geomagnetic Polarity Time Scale (GPTS; Ogg and Smith, 2004) and other well established geomagnetic polarity events and excursions (see Kidane et al., 2007; Dirks et al., 2010) to produce potential age ranges for the various deposits and site as a whole.

Mineral magnetic measurements were undertaken on siltstone attached to the edge of the speleothem block samples to determine the magnetic mineralogy, magnetic grain size, and concentration of remanence-carrying minerals in the Sterkfontein clastic deposits (see Walden et al. [1999] for a more detailed description of the methodology). Understanding the mineralogy of the samples is important for helping to understand the origin of the magnetic polarity preserved within the deposits; that is, primary (formed at the time of deposition), secondary (formed from secondary chemical alteration, i.e., chemical remanent magnetisation; CRM), or from potential overprinting due to relaxation of low coercivity, viscous magnetic grains that do not hold a stable remanence over the time period represented by the age of the deposits (a viscous remanent magnetisation; VRM). This work was then contrasted with demagnetisation characteristics of the palaeomagnetic samples to determine if the same mineralogy of clastic inclusion likely occurred within the speleothem samples. Magnetic susceptibility measurements (K_{LF}: Low Frequency; K_{HF}: High Frequency; K_{LT}: Low Temperature) were undertaken using the Bartington MS2B and χ/T system for frequency dependant room temperature and low temperature analysis down to -196 °C. Measurements were not corrected for weight as this was primarily influenced by the weight of diamagnetic calcite inclusions and so are volume specific (10^{-5} SI). Isothermal Remanent Magnetisation (IRM) acquisition curves, and backfields, hysteresis loops, and Curie curves were run on a Magnetic Measurements Variable Field Translation Balance (VFTB).

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Table 1
Palaeodirectional results from Sterkfontein Cave shown as per their stratigraphic position[a]

Sample	Member	Declination (degrees)	Inclination (degrees)	Mean MAD	*K*	No.	P. Lat	Polarity	Depth (m)
Surface									
STER-19	M5/6	185.6	−45.5	2.6	8.1	3	−36.8	I	1.9
STER-18	M5C	205.1	30.5	6.1	24.6	4	−64.7	R	3.5
STER-17	M5B	170.7	38.5	4.0	55.7	6	−80.4	R	8.0
STER-A01U	MPFS	210.4	50.1	9.6	195.0	2	−62.9	R	8.7
STER-16U	MPFS	230.9	17.8	11.0	480.3	2	−39.0	I	8.75
STER-16L	MPFS	170.3	−44.8	12.4	166.8	2	−36.7	I	8.8
STER-A01L	MPFS	313.4	−16.4	9.9	98.2	2	42.4	I	8.85
STER-A02	MPFS	348.4	−52.6	3.1	33.4	3	77.6	N	8.9
STER-A03	MPFS	10.3	−37.3	8.1	36.4	3	79.2	N	8.95
STER-A04	MPFS	355.5	−34.8	8.9	12.1	3	82.0	N	9.0
STER-A05U	MPFS	150.9	−42.8	11.2	65.5	2	−31.9	I	9.05
STER-A05L	MPFS	166.9	30.7	9.1	77.6	2	−77.3	R	9.1
STER-A06	MPFS	48.6	−12.2	10.5	24.2	3	39.6	I	9.15
STER-A07	MPFS	66.6	−48.3	10.1	18.2	3	31.7	I	9.2
STER-12U	MPFS	4.0	−48.0	10.2	51.5	3	77.3	N	9.25
STER-15U	MPFS	348.2	−22.8	10.3	36.1	3	72.0	N	9.3
STER-12M	MPFS	308.6	14.0	12.7	44.7	3	30.2	I	9.35
STER-15L	MPFS	155.1	31.8	2.7	14.6	3	−65.2	R	9.4
STER-A08	MPFS	178.6	25.3	7.5	63.9	3	−77.2	R	9.45
STER-12L	MPFS	158.6	35.4	9.7	41.6	3	−83.1	R	9.5
STER-15C	MPFS	153.9	30.2	2.5	55.3	2	−60.1	R	9.55
STER-14C	M4B	150.6	42.2	12.2	174.9	3	−63.4	R	12.8
STER-14S	M4B	177.1	42.9	5.0	210.1	4	−80.7	R	13.0
STER-13	M4A	153.7	27.9	12.1	55.7	4	−63.0	R	14.7
STER-11U	M4A	209.2	38.8	3.1	32.2	3	−63.0	R	15.9
STER-11L	M4A	342.5	24.9	10.7	3.2	2	42.4	I	16.1
STER-10C	M4A	174.4	46.3	5.4	691.7	4	−87.1	R	17.5
STER-10S	M4A	199.1	32.9	3.2	222.8	3	−81.7	R	17.6
Silberberg									
STER-9+	3	3.6	−58.7	9.0	35.6	4	76.2	N	25.0
STER-1+	2D	355.5	−54.7	3.6	81.4	4	80.1	N	28.5
STER-2+ (C&S)	2D	357.2	−62.2	2.4	69.7	3	72.4	N	29.0
STER-3+	2C	187.6	49.3	1.1	161.9	3	−82.1	R	29.7
STER-4T+	2C	24.7	−49.5	0.8	44.5	3	70.1	N	29.9
STER-4M+	2C	2.1	36.5	14.7	50.2	2	44.0	I	30.0
STER-4 B+	2C	147.9	52.8	6.4	88.0	3	−61.3	R	30.1
STER-5+	2B	78.4	−58.0	19.8	4.0	4	24.5	I	30.2
STER-6+	2B	219.6	56.2	3.9	25.9	3	−54.8	R	30.5
STER-7+(C&S)	2A	187.0	35.4	1.7	108.6	3	−80.9	R	31.0
STER-8+	2A	170.4	32.3	8.7	71.2	3	−77.7	R	31.2
STER-21	1B	196.8	20.0	4.1	1177.2	3	−67.7	R	32.4
STER-20	1A	198.9	27.6	3.9	17.1	3	−69.0	R	32.7

[a] Re-evaluation and addition to data from Partridge et al. (1999). Samples STER-12 and 15 come from different exposures of the MPFS, which formed at the same time as parts of M4C, and are considered to sample the same time period. No. = subsamples per block sample. C and S designate sediment and calcite; L, M, U designate lower, middle, and upper; P.Lat = Palaeolatitude.

Palaeomagnetic results

Directions of magnetisation at the site are presented in Table 1. NRMs ranged between 2.78×10^{-6} Am^2/kg for calcite samples, with detrital crust surfaces to below the measurable limit of the cryogenic SQUID-based spinner magnetometer (3×10^{-7} Am^2/kg) for pure speleothem samples. About a quarter of the subsamples were too weak to hold any measurable NRM. Variability in the demagnetisation spectra is quite high (Figs. 6 and 7). Some samples have a more curved vector plot and show great arc circles that suggest the incorporation of more than one overlapping component formed in grain sizes of a very similar size (Figs. 6 and 7). This is likely due to variation in the multiple layers of speleothem and detrital crusts that needed to be sampled in some cases to get strong enough remanence properties to be measured as well as due to secondary overprints formed after deposition of the deposits. Some samples have more than one component of magnetisation, while others have little evidence of such overprints. Some sample NRMs contained only a weak soft viscous remanent magnetisation (VRM) or isothermal remanence formed by transport of the samples from South Africa to the laboratory in the UK and other modern processes acting on low coercivity unstable grains. A comparison of the demagnetisation behaviour in samples that were run immediately on arrival in Liverpool, samples that had been sitting in a zero field cage for a couple of months, and samples run a number of years later confirms this overprint to be very recently acquired isothermal and viscous magnetisations (Herries, 2003). In the majority of samples, AF demagnetisation to around 8 mT removed this VRM (Fig. 7). In others, the primary remanence had been more severely overprinted due to relaxation of viscous grains and numerous field changes (reversals) since the time of deposition. In these cases, the harder VRM was only removed by fields of 10–12 mT. Further stepwise demagnetisation in 2–5 mT stages to between 30 and 100 mT then permitted the identification of a primary ChRM that was considered to represent the primary remanence formed at the time of deposition (Fig. 7).

Thermomagnetic curves gave Curie temperatures (T_c) with a mean of 585 °C, indicating that the main remanence-carrying mineral is magnetite, although maghaemite is also indicated in some cases with a drop in magnetisation after heating and a small phase transition at low temperatures (Fig. 8). IRM unmixing curves indicate the presence of three distinct populations representing low, medium, and high coercivity minerals. These are interpreted as viscous-single-domain (vSD) grains and stable-single-domain grains (SSD) of magnetite,

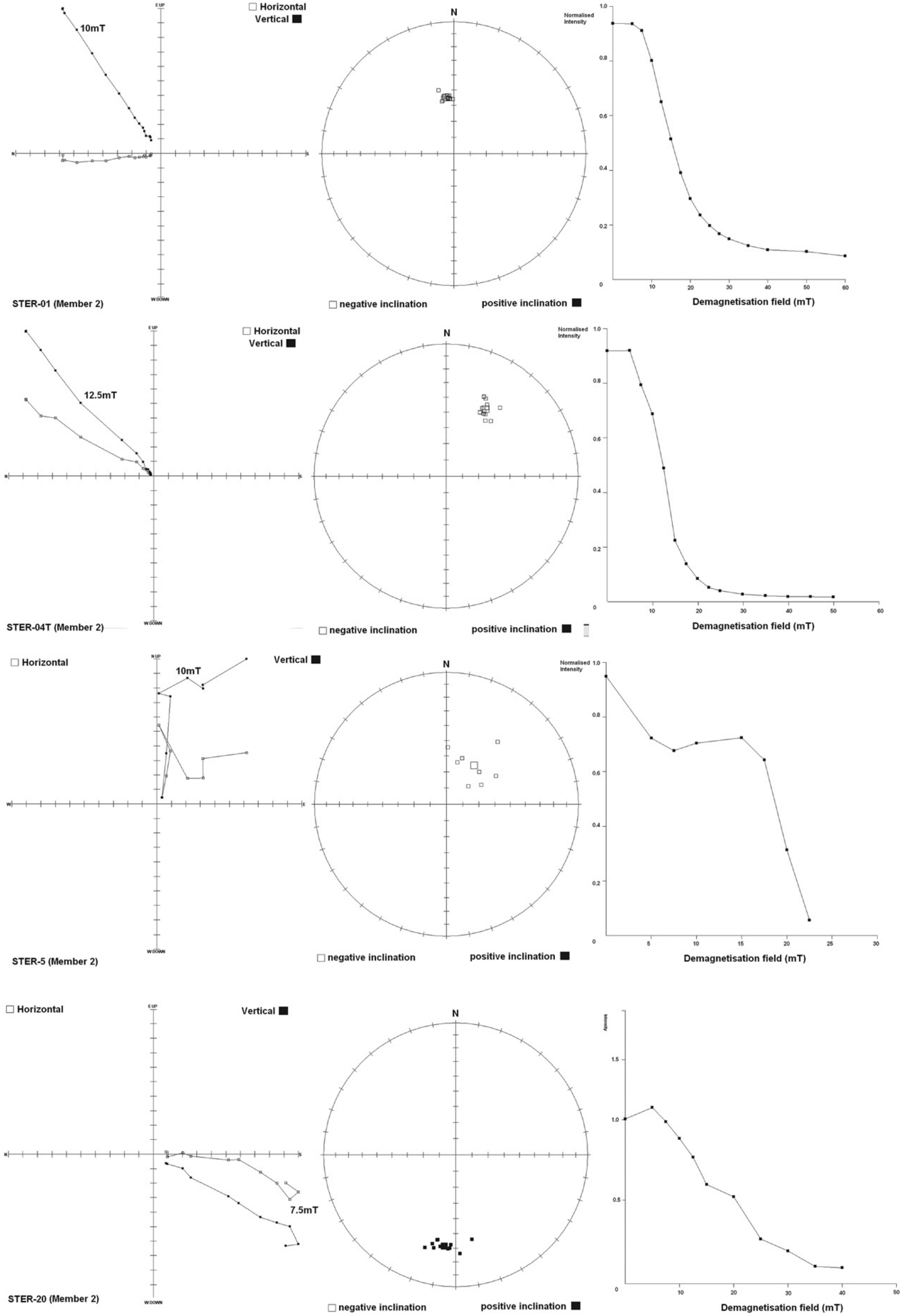

Fig. 6. Demagnetisation spectra for samples from the subterranean exposures at Sterkfontein (M1–3).

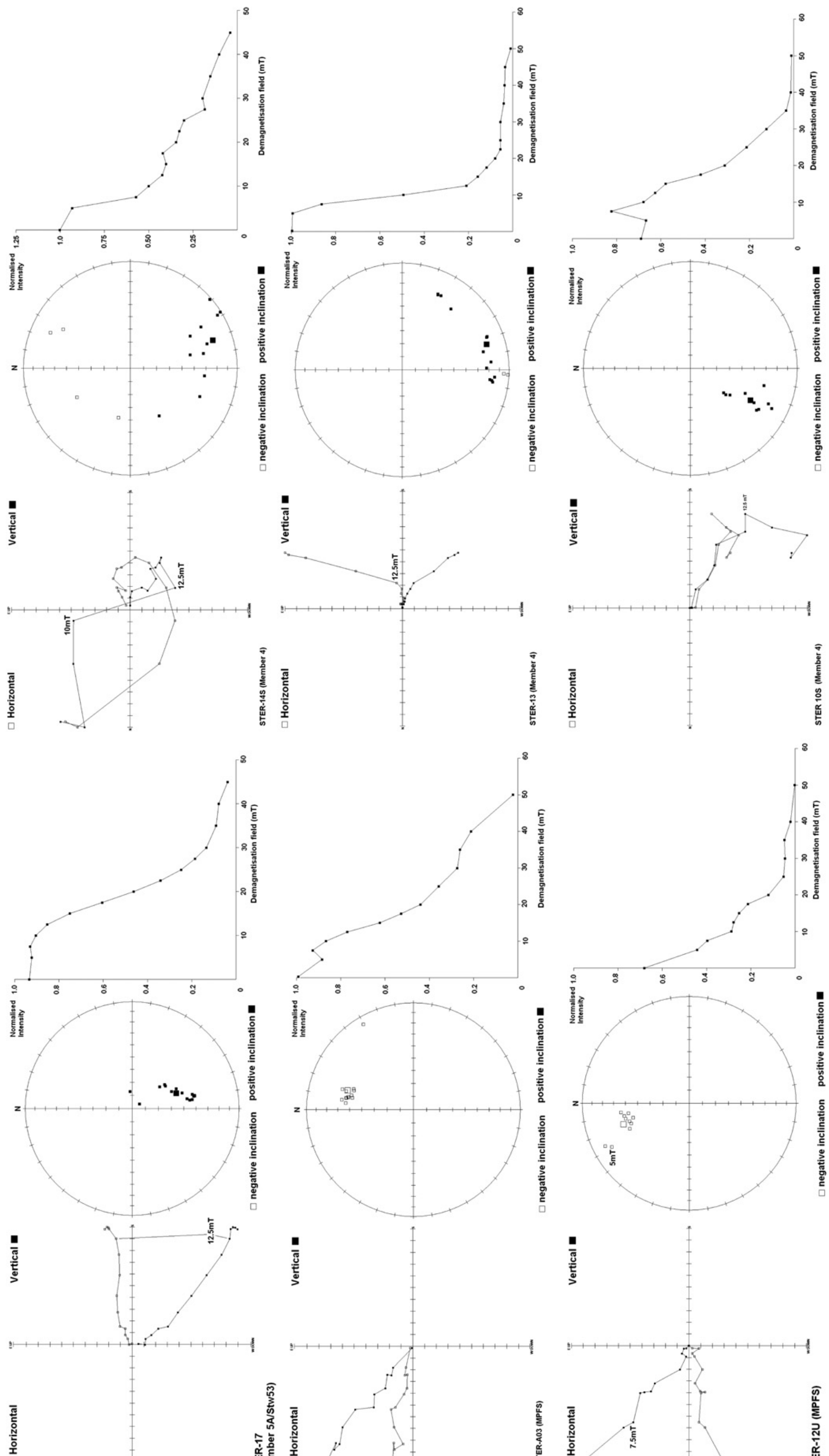

Fig. 7. Demagnetisation spectra for samples from the surface exposures at Sterkfontein (M4–M5).

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maghaemite, and pigmentary haematite in decreasing proportions. Backfield measurements indicate that the ferrimagnetic grains lie in the SSD to pseudo-single-domain (PSD) grain size region (Supplementary Online Material [SOM] Fig. 1). Modified Lowrie–Fuller (LF) tests (as per Bailey and Dunlop, 1983) indicate either the presence of a pure SD grain or mixed SD/MD mineralogy for different samples. However, Herries et al. (2007, 2008) have noted that the presence of fine-grained SD (vSD) grains close to the superparamagnetic to single-domain boundary can produce a similar LF plot to much larger multi-domain (MD) grains when mixed with SD grains. The two grain sizes produce similar coercivity spectra and as such this will also effect IRM based measurements. All samples have a high percentage (mean $\chi_{FD}\% \sim 10\%$) of frequency-dependent values of magnetic susceptibility, which indicates that they have a high proportion of fine vSD grains (Table 1). The presence of ultra-fine superparamagnetic (SP) grains and larger-single-domain grains is confirmed in low temperature magnetic susceptibility curves where there is no evidence of a MD peak (SOM Fig. 1). As such, there is no conclusive evidence of MD grains from the mineral magnetic experiments, as suggested by the work of Schmidt and Partridge (1991).

Both MD and vSD grains will cause similar relaxation in the primary remanence; however, if enough SSD grains exist the primary remanence can still be isolated successfully, as shown in these and other South African samples. Mineral magnetic work suggests that the Sterkfontein clastics that were studied fall closer to material from Gladysvale and Gondolin (Lacruz et al., 2002; Herries, 2003; Herries et al., 2006a; Adams et al., 2007), which have stable magnetic carriers, than samples from some levels at Buffalo Cave and the Makapansgat Limeworks (Herries, 2003; Herries et al., 2006b), which have less stable magnetic carriers of remanence. Mineral magnetic work suggests this is due to the relationship of SSD and vSD grains. A high proportion of vSD grains are often an indicator of sediment source with more alluvial deposits having high vSD percentages and colluvial sources having lower percentages (Herries, 2003). The high percentage of vSD grains in South African soils is thought to be due to long term burning of the South African landscape through bushfires (Herries, 2009). Work carried out on the calcified sediments attached to some of the speleothem samples from M4 gave a consistent magnetic polarity, suggesting that random orientations noted by

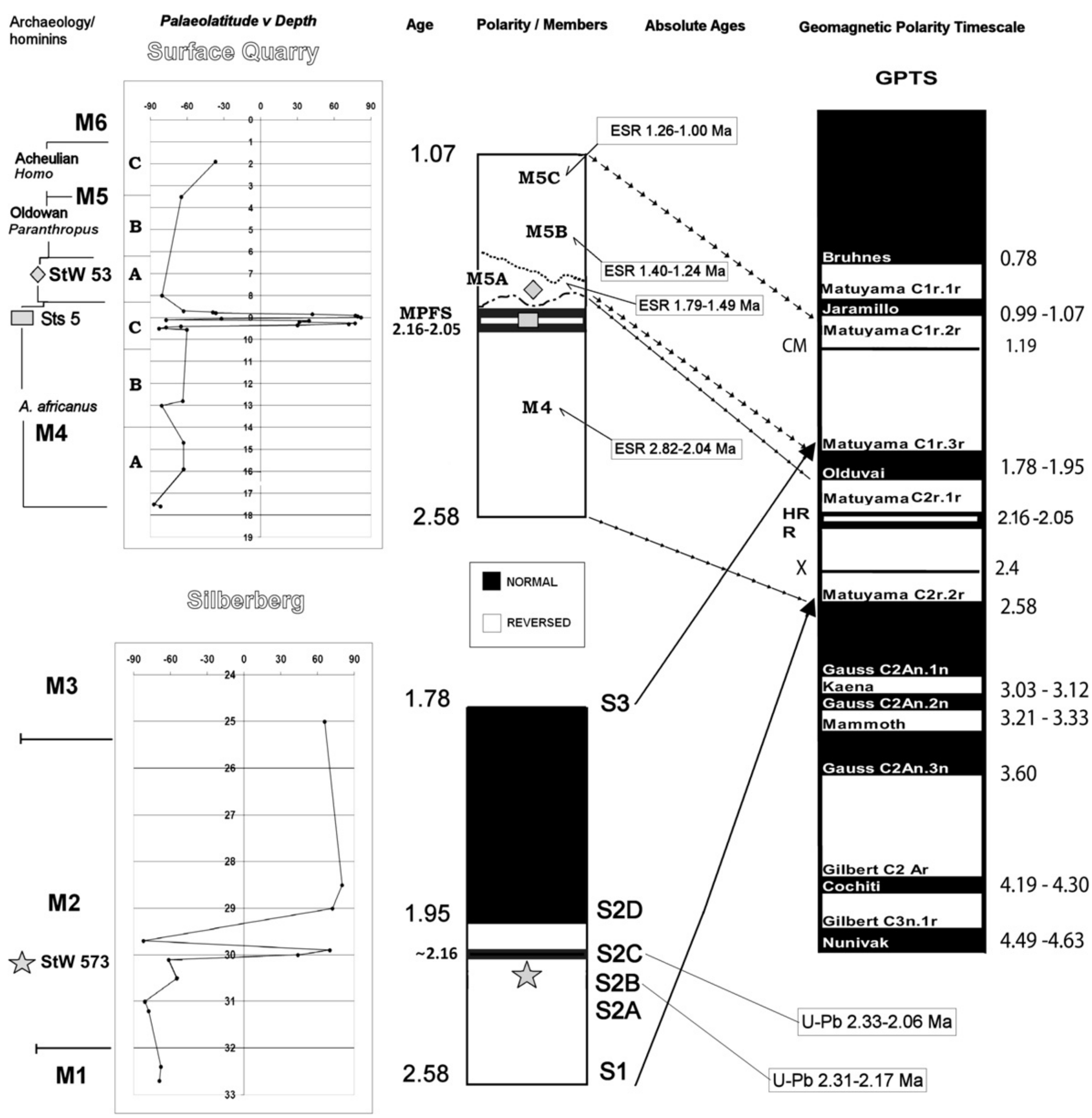

Fig. 8. Magnetostratigraphy for Sterkfontein Cave. Palaeolatitude (degrees) versus depth (metres) plotted against: archaeology, hominin fossils, polarity, major depositional units (Members), and the Geomagnetic polarity timescale (GPTS). O = Olduvai event, CM = Cobb Mountain Event, R = Reunion event, HR = Huckleberry Ridge event.

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Jones et al. (1986) are not characteristic of all such deposits at Sterkfontein and suggests that the speleothem and siltstone were deposited at very similar time periods. The consistency of the magnetic polarity from within and between blocks from the same unit (Table 1) suggests that their primary remanence has been successfully isolated. In conclusion, this detailed analysis indicates that the samples containe magnetic grain sizes that can successfully fossilise a geomagnetic field over geological time. Moreover, the Sterkfontein samples successfully hold a detrital remanence that can be isolated and represents the geomagnetic field direction at the time of their deposition.

The palaeomagnetic data for Sterkfontein is presented in Table 1 and Figure 8. Almost all of the samples from M4 and M5 record a reversed direction of magnetisation. The exceptions are STER-19, which is at the interface of M5 and M6, STER-11 from M4 which records intermediate directions of magnetisation, and samples from the MPFS partly capping M4. As such, the majority of M4 and M5 deposits cannot be distinguished by palaeomagnetism alone. However, the MPFS records two short periods of normal and intermediate magnetic polarity (STER-12M to STER-A06 and STERA-05u to STER-16U). The STER-11 sample was extremely weak, with a very low K value (this shows the degree of dispersal of different samples and the higher the value the less scatter there is between subsamples from the same block). As such, it is unlikely to record an accurate field direction. STER-19 also has a low *K* value. The Sterkfontein surface exposures therefore record a long period of reversed polarity separated by two short normal polarity events (Fig. 8).

Clarke (2007) divides the M2 sequence into a lower siltstone deposit, which contains the flowstones 2A (samples STER-5, 6), 2B (samples STER-7, 8), and an upper breccia deposit containing *Australopithecus* fossil StW 573. This is capped by flowstones 2C (sample STER-3/4) and 2D (samples STER-1, 2). The sequence in the Silberberg Grotto is dominated by reversed polarity (STER-20 to 21 and STER-3 to 8) at its base and normal polarity at the top of the section (STER-9 and STER-1 to 2). In speleothem layer STER-4 a layer of short normal magnetic polarity is noted towards the end of the reversed polarity period, whereas in sample STER-5 intermediate directions were recorded. In the original analysis by Partridge et al. (1999) the STER-5 sample was interpreted as normal polarity. However, in a reanalysis of the primary data, Herries (2003) expressed the need for some caution in interpreting the normal polarity episodes in block STER-5 (S2B) in light of potential resetting of the magnetic signal due to shock induced magnetisation from mining because a mining shot hole is present in the side of this block sample. Also, additional samples and reanalysis suggest this block records an overall intermediate direction of magnetisation trending towards a normal polarity rather than a well-defined normal polarity direction. As such, the reliability of this normal polarity episode is questionable. In addition, the MAD for this sample is very high (>20) and the *K* value is extremely small. This is because it is both weak and shows a high degree of variation in demagnetisation behaviour and overall direction. There is a suggestion of multiple components of magnetisation in the vector plots and great arc circles moving towards a reversed direction (Fig. 7). As such, the direction has been rejected. Similar behaviour is not noted in any other specimens from Sterkfontein and so such shock induced magnetisation from mining does not appear to have been a major issue at the site. However, such factors should be accounted for when sampling at such sites. The sequence in the Silberberg Grotto therefore records a change from a long period of reversed polarity to a period of normal polarity with a short normal polarity episode occurring just before the reversal.

Magnetostratigraphy and ESR ages

As stated previously, the main complicating factor in interpreting the sequence of polarity zones and reversals at the site is in understanding the complex stratigraphy. On such short sections of stratigraphy fitting the sequence to the GPTS is impossible without a guide from other sources, be it a radiometric age on a particular layer or from the fauna.

The surface exposures (M4, M5, and M6)

Previous faunal analyses by Vrba (1982) and McKee et al. (1995) suggested an age range for M4 of 3.0–2.6 Ma, based on comparisons with faunas from East African sites. The occurrence of a more wooded environment at the site (Bamford, 1999) has also led many (Kuman and Clarke, 2000) to believe that M4 was deposited before a major period of cooling and aridification noted in East Africa between 2.8 and 2.5 Ma (deMenocal, 1995). However, an age of between 3.0 and 2.6 Ma is inconsistent with the reversed polarity directions recorded from M4 in this study. Other researchers (Vrba, 1975, 1995; Delson, 1984, 1988; Berger et al., 2002) have suggested an age of less than 2.5 Ma based on fauna, and most researchers suggest (McKee et al., 1995; Vrba, 1995, 2000) that Sterkfontein is younger than Makapansgat. Recent magnetobiostratigraphic analysis of the *A. africanus* bearing deposits at Makapansgat (Herries, 2003; Hopley et al., 2007a; Herries et al., in press) indicate that they date to between 3.03 and 2.58 Ma and likely less than 2.85 Ma based on the fauna. In addition, preliminary ESR dates suggested a likely age of between 2.66 and 2.08 Ma for M4 (Schwarcz et al., 1994), a period of reversed magnetic polarity. Together this suggests that M4 in fact dates to the beginning of the Matuyama reversed polarity Chron between 2.58 and 1.95 Ma.

Further support for this interpretation comes in the form of a double spiked reversal event identified in the MPFS capping the majority of M4 and contemporaneous with Sts 5. This event(s) may represent one or more of the X-event at ~2.4 Ma, the Rèunion event at ~2.17 Ma, or the Huckleberry Ridge event at ~2.05 Ma (Fig. 8). The reversals are too short to be the younger Olduvai event (1.95–1.78 Ma) but could represent precursors to this event during instability in the Earth's magnetic field just prior to or during its reversal at 1.95 Ma. Such an event is documented by Braun et al. (2010) from a site in Kenya and perhaps represents the 'Pre-Olduvai' excursion dated to ~1.98 Ma (Roberts, 2008). The sequencing and age of these events and excursions has been a matter of great debate (see Kidane et al., 2007) and numerous events have been postulated for this time period (see Walker et al., 2006). However, much of the confusion is likely related to different chronological methods being used and the event being recorded in different sedimentary environments (volcanic, ocean sediments, etc). Kidane et al. (2007) identify only two events, dated at 2.09–2.05 Ma and 2.21–2.16 Ma, in this time period from their sequence in Kenya. Baksi and Hoffman (2000) identified only one reversal on Rèunion itself and this was dated between 2.16 and 2.12 Ma. Roger et al. (2000) also date a normal polarity reversal in the French Massif central to between 2.17 and 2.11 Ma. Quidelleur et al. (2010) have recently re-dated the Rèunion sequence and suggest two distinct events at 2.17–2.13 Ma and 2.06–2.02 Ma. Taken together this suggests a good age estimate for the Rèunion subchron of ~2.16 Ma. Dirks et al. (2010) identified a short geomagnetic reversal event from the Malapa *Australopithecus sediba* bearing palaeocave near Sterkfontein and this has been dated by U–Pb to 2.026 ± 0.021 Ma (2.05–2.01 Ma). This correlates with ages for the youngest reversal noted by Kidane et al. (2007) in Kenya between 2.09 and 2.05 Ma, by Lanphere et al. (2002) for the Huckleberry Ridge ash itself at 2.09–2.05 Ma, and by Baksi and Hoffman (2000) for a reversal event in the Afar Depression of Ethiopia at 2.07–2.03 Ma. Taken together with the recent age of 2.06–2.02 Ma by Quidelleur et al. (2010), an age of ~2.05 Ma would be suggested as an approximate age estimate for the Huckleberry Ridge event. The occurrence of two events very close to each other would suggest that the MPFS geomagnetic reversals most likely represent the Rèunion and Huckleberry Ridge events and that the MPFS most likely formed in a period between roughly 2.16 and 2.05 Ma. *A. africanus fossil* Sts 5 was deposited at the same time and is therefore estimated to date to between 2.16 and 2.05 Ma.

The rest of M4, which contains the majority of the *A. africanus* fossils, records a reversed magnetic polarity and lies underneath the MPFS. It is interpreted as dating to the beginning of the Matuyama Chron between 2.58 and the Rèunion subchron at ~2.16 Ma. The occurrence of *Equus* in the M4 deposits (Kuman and Clarke, 2000) suggests that it potentially dates to <2.33 ± 0.03 Ma (2.36–2.30 Ma; FAD of *Equus* in Member G of the Shungura Formation; Brown et al., 1985). This would also rule out the X-event at ~2.4 Ma for the reversal events in the MFPS and may further constrain the M4 deposit containing the fossils to between 2.33 and 2.05 Ma. Kuman and Clarke (2000), however, consider *Equus* to be intrusive to M4 due to mixing while blasting the deposits. Work at Lincoln Cave, where Acheulian style cores of M5 origin are mixed with MSA material, suggests such processes have also occurred naturally (Reynolds et al., 2007). ESR data for M4 has also been used to support this mixing hypothesis (Kuman and Clarke, 2000). Despite this, *Equus* is documented to occur in M4, and the younger ages suggested by this study suggest that its occurrence would not be unexpected. *Equus* certainly occurs at Malapa at slightly younger than 1.95 Ma (Dirks et al., 2010).

Herries et al. (2009b) suggest that the younger ESR ages for M4 (<2.0 Ma), which have been used to suggest mixing, may in fact be due to sampling areas where the stratigraphy is less well understood. Figure 9 indicates the vertical and lateral coordinates of teeth sampled for ESR analysis by Curnoe (1999), and the ESR date is presented in the SOM. The M4 derived samples are in open triangles. Three samples (1348, 1347, 1352) are consistent in having linear uptake ages of between 3.09 ± 0.29 Ma and 2.06 ± 0.18 Ma and one sample (1349) is much younger with ages for different parts of the teeth between 1.23 ± 0.16 to 1.04 ± 0.09 Ma. The younger sample (1349) comes from an area at the interface of M5 and is consistent with other ESR ages for samples taken from this level within M5 at 1.04 ± 0.22 Ma (1351). When sample 1349 is removed from the M4 data of Curnoe (1999) (Table S1) the weighted mean age for M4 is 2.42 ± 0.38 Ma (2.80–2.04 Ma), which is consistent with the interpretation made from palaeomagnetic analysis (2.58–2.05 Ma). However, the ESR data do suggest that there is some potential for parts of M4 to be as old as ~2.8 Ma. This suggests that linear uptake ESR age estimates are providing relatively reliable indicators of age.

The limited palaeomagnetic samples from M5A (StW 53 infill) and M5C (Acheulian infill) all indicate a reversed polarity and so are indistinguishable from those from M4. However, the M5 samples must date to after the Huckleberry Ridge event (~2.05 Ma) identified in the MPFS. As such, the reversed polarities would date the M5 deposits to either side of the Olduvai event (1.95–1.78 Ma) and suggest an age range of between 2.05 and 1.95 Ma and/or 1.95 and 1.07 Ma. Partridge (2000) suggests that significant time occurred between M4 and M5 based on erosion and subsequent infill of the M4 deposits by later M5 deposits and so an age of between 2.05 and 1.95 Ma would seem unlikely. However, Clarke (2007, 2008) believes M5A to be a remnant of M4 and the StW 53 fossil to be *A. africanus*. Faunal interpretations for M5 and age ranges based on them have been mixed. Pickering (1999) notes that only a few specifically identified faunal remains have been described from

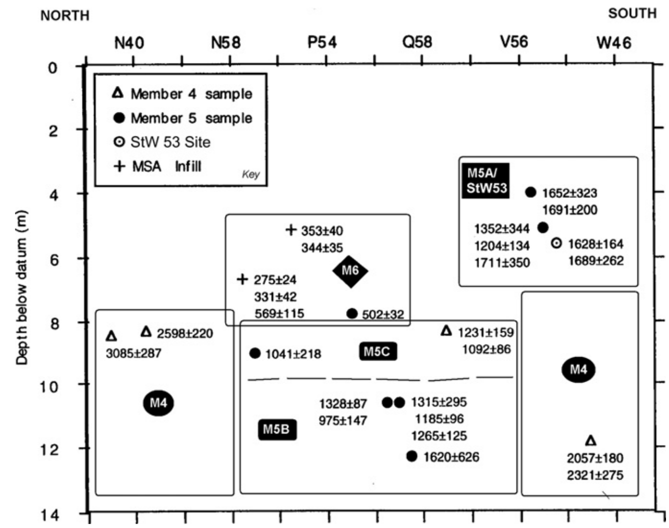

Fig. 9. Location of Electron Spin Resonance samples from Member 4 and Member 5 after Curnoe (1999).

each of the three M5 assemblages, providing a very limited chronological context for the deposits. O'Regan (2007) further indicates the assignment of the material to specific submembers is extremely complex. Kuman and Clarke (2000) suggest that the presence of *Theropithecus oswaldi* as indicating an age of between 2.6 and 2.0 Ma. However, as Herries et al. (2009b) note, this species is found in the Okote Member at Koobi Fora dated to between 1.63 and 1.51 Ma and therefore does not suggest an age older than 2.0 Ma. Moreover, it is found in all three members at Swartkrans, which are also generally thought to be less than 2.0 Ma and perhaps as young as 0.6 Ma, as well as the obviously much younger Sterkfontein M5B (Herries et al., 2009b). In fact, it is fauna from the supposedly younger M5B and M5C deposits that suggest greater antiquity for M5 than suggested by the ESR dates of Curnoe (1999) and Herries et al. (2009b). Cooke (1994) noted the similarity between *Metridiochoerus modestus* molars and specimens from Olduvai Bed 1 (2.03–1.75 Ma) and *Dinofelis barlowi* has come from M5C and would seemingly suggest an age between 3.0–1.9 Ma (Herries et al., 2009b). This is extremely unlikely given that this deposit contains Acheulian artefacts, and as such this may represent further evidence for mixing, be it geological or anthropogenic. Fossils could equally have been mistakenly excavated from M4 or have been eroded from older deposits, as suggested for the younger material found in M4. Vrba (1982) originally suggested an age of ~1.5 Ma for M5, and O'Regan (2007) concludes that a reanalysis of the carnivores from M5 and lack of extinct species makes the deposits seem much younger than previously suggested.

Two teeth (1337, 1343) from M5A (StW 53 infill; Curnoe, 1999) gave a weighted mean ESR age estimate of 1.64 ± 0.15 Ma (1.79–1.49 Ma). A third tooth (1338) gave an age of 1.35 ± 0.34 Ma (1.69–1.01 Ma) for one part of the tooth and 1.20 ± 0.13 Ma (1.33–1.07 Ma) for a second part of the tooth. While one part of the tooth (1338a) gave an age consistent with the other two teeth (1337, 1343) from M5A, the second part (1338b) gave much younger ages and this has been used to suggest that the dating is unreliable (Gilbert and Grine, 2010). It is such suggestions that have led to the distrust of ESR dating at these sites. Also, ESR ages are often presented as weighted means of all samples (see Curnoe et al., 2001) even if some have spurious ages, and this in turn leads to a misrepresentation of the true potential age and again mistrust in the method. However, when the context of the samples from M5A is investigated, there is a clear reason for this difference in the age estimates of the teeth studied. The two consistent samples from M5A (1337 and 1343; Table 2) came from calcified deposits, while the third sample (1338) came from decalcified deposit where the uranium uptake and leaching history will have been extremely complex. As such, this sample should be discounted rather than averaged together with the two other teeth. The fact that linear uptake ESR ages are consistent with palaeomagnetism for M4 means that there is no reason to discount them for M5 and at other sites when contextual information is known with certainty. An ESR age estimate of 1.64 ± 0.15 Ma (1.79–1.49 Ma) is therefore suggested for M5A and hominin fossil StW 53 based on the in situ recovery of teeth from the calcified deposits. As in M4, this is consistent with the palaeomagnetic data which record reversed magnetic polarity and so cannot be in the 1.95 to 1.78 Ma time range (Olduvai normal polarity Chron). These data also fit with the stratigraphic interpretations of Partridge (2000), who suggested a large depositional break between M4 and M5. Due to this the Olduvai event is not recorded in the surface deposits.

The ESR samples from M5B and C show an age increase with depth through the deposit and an internal consistency (SOM Table 1; Fig. 9). Sample 1351 was originally interpreted as belonging to M5 but, again, its position at the interface of M6 and its age (~500 Ka) suggest it belongs to M6 rather than M5. The ESR samples from M5B (Oldowan infill) have a weighted mean age of 1.32 ± 0.08 Ma (1.40–1.24 Ma) and those for M5C (Acheulian infill) have a weighted mean age of 1.13 ± 0.13 Ma (1.26–1.00 Ma). As such, there is only a small difference in age between M5B and M5C and so the Oldowan and Acheulian assemblages. Again, this is consistent with the palaeomagnetism from M5C, although an age greater than the Jaramillo event (1.07–0.99 Ma) is perhaps most likely. Combining the palaeomagnetism and ESR suggests that M5C likely dates to between 1.26 and 1.07 Ma. M6 was not sampled for palaeomagnetism but ESR dates (SOM Table 1) suggest the deposit dates to between 470 and 289 ka, slightly older than or perhaps partly contemporaneous with the oldest deposits from Lincoln Cave (~300–100 ka; Reynolds et al., 2007).

The Silberberg Grotto (M2, M3)

The previous magnetostratigraphy for the Silberberg Grotto suggested that there were five changes in polarity, within 5 speleothem layers (2A–D and 3), thought to cover the end of the Gilbert reverse Chron from around 4.18 Ma, to some time before the end of the Gauss normal Chron at around 2.58 Ma (Partridge et al., 1999). This was based in part on the presence of two short period normal geomagnetic polarity events occurring within the basal period of reversed polarity and also in part on interpretations from fauna and depth within the cave system. *Chasmaporthetes silberbergi* (Partridge et al., 2003; Clarke et al., 2003) in particular was used to suggest an age >3.0 Ma as this has some similarities to a species found at Langebaanweg at ~5.0 Ma (Franz-Odendaal et al., 2002) As previously stated, the depth of deposits within a cave system is not necessarily an indicator of their age, and modern sediments are now being deposited below the palaeodeposits creating an inverted stratigraphy. Moreover, Berger et al. (2002) subsequently suggested a younger age for the fauna from this area, although Pickering et al. (2004) then noted that many specimens previously thought to have been part of the M2 assemblage come from other areas of the Silberberg Grotto and are not necessarily the same deposit. As such, the assemblage from this area is generally small and of limited use for chronological assessment, although most species also occur in M4 or at younger sites, including *C. silberbergi* (Turner, 1997; Kibii, 2004).

The magnetostratigraphy for the site as a whole would perfectly fit the geological interpretations of Partridge (1979, 2000) in that the normal polarity at the top of the Silberberg Grotto sequence would represent the end of the Gauss normal polarity Chron between 3.03 and 2.58 Ma, lying below the reversed polarity of the M4 deposits between 2.58 and ~2.05 Ma. If this were the case you would expect the majority of the Silberberg Grotto sequence to be normal polarity with short reversed polarity events representing the Kaena and Mammoth event as envisaged by Partridge et al. (1999). However, this is clearly not the case. The Silberberg sequence is dominated by reversed magnetic polarity with a single short normal polarity event and then followed by a longer period of normal magnetic polarity. Several attempts have been made to reanalyse the magnetostratigraphy in light of recent incompatible cosmogenic nuclide burial ages (4.17 ± 0.14(0.35) Ma; Partridge et al., 2003) and U–Pb age estimates (2.33–2.06 Ma; Walker et al., 2006, 2006; Berger et al., 2002; Muzikar and Granger, 2006). The identification of only one short normal polarity episode in the Silberberg Grotto as suggested by this study has a significant impact on the magnetostraigraphic interpretation as per Partridge et al. (1999), Berger et al. (2002), and Walker et al. (2006) and means that all are inaccurate.

When the new polarity sequence is compared to the cosmogenic burial ages, the sequence could represent the end of the Gilbert Chron between 4.49 and 3.60 Ma. The normal polarity event in the 2C flowstone would have to represent the Cochiti event at 4.30–4.19 Ma, making StW 573 between 4.49 and 4.29 Ma.

Table 2
Age ranges for the Sterkfontein deposits and fossils based on a combination of U–Pb, ESR, and palaeomagnetism

Deposits	Max Age (upper) (Ma)	Max Age (lower) (Ma)	Suggested Age Range (Ma)	Archaeology	Hominin Genus
post-M6	0.47	0.29	0.5–0.3	MSA	*Homo*
M5C	1.26	1.07	1.3–1.1	Acheulian	*Homo*
M5B	1.40	1.24	1.4–1.2	Oldowan	*Paranthropus*
StW 53 infill (M5A)	1.78	1.49	1.8–1.5	Oldowan?	*Homo*
Silberberg (M2)	2.58	1.78	2.6–1.8		*Australopithecus*
M4C	2.16	2.05	2.2–2.0		*Australopithecus*
M4B	2.58	2.16	2.6–2.2		*Australopithecus*
M4A	2.80	2.16	2.8–2.2		*Australopithecus*
Key Fossils	Max Age (upper) (Ma)	Max Age (lower) (Ma)	Suggested Age Range (Ma)	Member	Species
StW 585	0.47	0.29	0.5–0.3	Post-M6	*Homo* sp.
StW 80	1.26	1.07	1.3–1.1	M5C	*Homo* sp.
StW 584	1.40	1.24	1.4–1.2	M5B	*P. robustus*
StW 566	1.40	1.24	1.4–1.2	M5B	*P. robustus*
StW 53	1.78	1.49	1.8–1.5	M5A	*Homo* sp.
Sts 5	2.16	2.05	2.2–2.0	M4C	*A. africanus*
StW 573	2.58	2.16	2.6–2.2	M2	*Australopithecus* sp.
Sts 14	2.58	2.16	2.6–2.2	M4B	*A. africanus*
Sts 431	2.58	2.16	2.6–2.2	M4B	*A. africanus*
StW 505	2.58	2.16	2.6–2.2	M4B	*A. africanus*
Sts 71	2.58	2.16	2.6–2.2	M4B	*A. africanus*

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However, this interpretation is inconsistent with expected depositional rates in the Silberberg sequence as there is only a short section between the normal polarity event in the 2C flowstone and the overlying normal polarity zone in flowstone 2D. Moreover, maximum U–Pb ages (not taking account for ^{234}U disequilibrium) for the SC and 2B speleothems lie in the range of 3.1–2.6 Ma (Walker, 2005), much younger than the cosmogenic ages. Cosmogenic burial dating can be subject to inaccuracy and over estimation of age if mixing of quartz from different deposits has occurred and if the geomorphic history of the deposit is not known with certainty and previous burial of older quartz has occurred (Granger, 2006). The inverted cosmogenic ages in this sequence may suggest such potential reworking.

It has been noted by Walker (2005) and Clarke (2007) that the stratigraphy in the Silberberg Grotto is extremely complex and that the relationship of the flowstone layers to the fossil bearing sediment is not entirely certain. For example the 2C flowstone was formed through the deposit after it had slumped, and the flowstone cuts the fossil in half. As such, it is definitely younger than the fossil, and its age of ~2.2 Ma provides a minimum age for StW 573. It could be suggested that the same thing occurred for the other flowstone layers, which may have all originated from the flowstone boss that caps the deposit and that this formed over much later sediment as suggested by the cosmogenic ages. However, a comparison of the palaeomagnetic polarity of the flowstones and associated sediments suggests they are of the same polarity and therefore were most likely deposited synchronously. Moreover, there is a trend through the section with the upper speleothem layers being primarily normal polarity and the lower layers reversed polarity. As such, they cannot be the same age.

If no disequilibrium of ^{234}U occurred in the flowstones, which is extremely unlikely, then the maximum ages would place them between 3.01 and 2.6 Ma, a period of stable normal magnetic polarity. As such, this is impossible when compared against the palaeomagnetic sequence. However, when the U–Pb ages are corrected for ^{234}U disequilibrium, a mean age of 2.17 $\pm$ 0.17 Ma (2.33–2.00 Ma; Walker et al., 2006) is obtained for flowstone 2C. This correction is based on the measurement of modern speleothems at Sterkfontein, which suggests an initially very high disequilibrium, as did excess ^{234}U detected in all the samples (Walker, 2005). When the corrected U–Pb age is compared with the palaeomagnetic data the sequence fits extremely well with expected depositional rates. In this scenario the normal polarity episode in the 2C flowstone overlying the StW 573 fossil would represent one of the short normal polarity episodes (Rèunion or Huckleberry Ridge) also identified in the MPFS between ~2.16 and ~2.05 Ma. This precise fit between the U–Pb ages and the occurrence of similarly aged geomagnetic polarity events noted in other speleothems from Sterkfontein and Malapa (Dirks et al., 2010) lend further weight to this correlation. Moreover, recent redating of the 2B flowstone below fossil StW 573 (2.24 $\pm$ 0.09 Ma; Walker et al., 2006) by Pickering et al. (2010) confirms a ~2.3 Ma age estimate for this flowstone. Sample SKA3 from layer 2C above StW 573 has an age of between 2.33–2.19 Ma. As this age is virtually indistinguishable from the age for layer 2B (2.33–2.17 Ma) that is below StW 573, the fossil likely dates to this time period between roughly 2.3 and 2.2 Ma. This makes it contemporaneous with or perhaps very slightly older than *A. africanus* fossil Sts 5 from M4. As the reversal within layer 2C is in the lower central portion of the flowstone and the youngest estimated age from the individual samples is 2.06 Ma (Walker, 2005), the reversal most likely represents the older Rèunion event (~2.16 Ma) rather than the younger Huckleberry Ridge event (~2.05 Ma).

When the palaeomagnetism for the rest of the Silberberg Grotto sequence is analysed the base cannot be older than the beginning of the Matuyama Chron at 2.58 Ma and the top of the sequence is slightly younger than the beginning of the Olduvai normal polarity subchron at 1.95 Ma (Fig. 8). If both the 2B and 2C flowstones did represent the same phase of speleothem formation as suggested by Clarke (2007) then the fossil could be slightly older, but not older than 2.58 Ma. The different polarity of the flowstones at the top and bottom of the Silberberg Grotto sequence confirms that they were not all formed at the same time after the fossil bearing sediment was deposited. As such, a conservative age estimate for StW 573 is between 2.6 and 2.2 Ma, although it is more likely around 2.3 to 2.2 Ma.

Conclusions and implications

The combined chronological evidence (fauna, ESR, palaeomagnetism, U–Pb, U–Th) outlined above suggests that there were a series of cave fillings at Sterkfontein between 2.6 and 2.0 Ma (Member 4 and Silberberg), 1.8–1.5 Ma (M5A, StW 53 infill), 1.4–1.1 Ma (M5B-C, Oldowan and Acheulian infill), and 500–100 ka (post-M6 deposits and Lincoln Cave; Table 2). This analysis of the ESR and palaeomagnetic data suggests that a clear sequencing of events can be seen in the surface exposures, and this is partly illustrated by Figure 9. The bulk of M4 was deposited between 2.6 and 2.2 Ma. M4 continued to form but with a thick flowstone forming in one area of the site between 2.2 and 2.0 Ma. A long hiatus occurred followed by the deposition of M5A (StW 53 infill) between 1.8 and 1.5 Ma. M5A is distinct from the remainder of M5 in that it occurs at a much higher level and is seemingly an earlier remnant. This is perhaps why Clarke (2007) has mistaken it for part of M4. M5A is partly eroded before the infilling of M5B (Oldowan infill) between 1.4 and 1.3 Ma. This is closely followed by the infilling of M5C (Acheulian infill) between 1.3 and 1.1 Ma. As such, while M5A is an earlier remnant distinct from M5B-C as suggested by Clarke (2008), it is not as old as Member 4, as suggested by Kuman and Clarke (2000). The M5A deposit should perhaps therefore be removed from inclusion in the M5 deposit and referred to as the 'StW 53 infill' as per Kuman and Clarke (2000).

The implications for these dates are that the Sterkfontein deposits, hominins, and stone tool industries are all much younger than previously suggested. The oldest Sterkfontein M4 deposits and *A. africanus* fossils are perhaps contemporaneous with M3 from the Makapansgat Limeworks (2.85–2.58 Ma; Herries et al., in press) based on the upper age limit of the ESR dates (2.8 Ma). While specimens of *A. africanus* have been recovered from all of the subunits of M4, most, including the partial skeletons Sts 14 and 431, come from M4B (Fig. 3). As such, the majority of M4 *A. africanus* fossils appear to be younger than 2.6 Ma, making them younger than the Makapansgat Limeworks. Fossils Sts 5 and StW 573 are contemporaneous in the time range of 2.3–2.0 Ma although StW 573 may be slightly older, while Sts 5 may be as young as ~2.0 Ma. Sts 5 therefore represents the youngest *A. africanus* fossil yet described. If StW 573 is not *A. africanus* as suggested by Clarke (2008), then there were definitely two species of *Australopithecus* present in the Sterkfontein area around 2.3–2.0 Ma. Moreover, these fossils are only slightly older than *A. sediba* fossils from Malapa at slightly younger than 1.95 Ma (Dirks et al., 2010) and the early *Homo* and *Paranthropus* fossils from Swartkrans M1, which may be as old as ~2.1–1.9 Ma based on ESR and faunal age estimates (Herries et al., 2009b). However, Swartkrans M1 is seemingly more complex than previously described and likely contains more than one age of deposit (pers. obs.).

The youngest deposits in the Silberberg Grotto formed during the Olduvai event (1.95–1.78 Ma) and are contemporaneous with Malapa (Dirks et al., 2010), Gondolin GD2 and GD1 (1.9–1.8 Ma; Herries et al., 2006a; Adams et al., 2007), Kromdraai-B M2

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(1.95–1.78 Ma; Thackeray et al., 2002), the Oldowan deposits at Wonderwerk Cave (Chazan et al., 2008), as well as perhaps parts of Swartkrans M1 (Herries et al., 2009b).

The StW 53 fossil from Sterkfontein M5A is estimated to date to around 1.8–1.5 Ma. If this fossil is *A. africanus* as suggested by Clarke (2007) then it is an extremely young specimen of one, making it younger than *A. sediba*. However, most researchers have described it as a form of early *Homo* (see Curnoe and Tobias, 2006; Table 2 of Smith and Grine, 2008; Curnoe, 2010) StW 53 is contemporaneous with other *Homo* and *Paranthropus* fossils from Kromdraai-B M3 (1.78–1.65 Ma; Herries et al., 2009b), Coopers D (1.6–1.4 Ma; de Ruiter et al., 2009), and perhaps also Swartkrans M2 (1.65–1.07 Ma; Balter et al., 2008) and the Goldsmith's faunal and archaeological locality (2.0–1.4 Ma; Mokokwe, 2005).

Based on the Sterkfontein M5 dates it appears that only the Oldowan layers from Wonderwerk (1.95–1.78 Ma) and perhaps Swartkrans M1 have evidence for very early stone tools in South Africa. While no stone tools have been definitively identified from M5A, the StW 53 fossil does have cut-marks that suggest their presence at this time (Pickering et al., 2000) and therefore suggests they occurred in the Sterkfontein area between 1.8 and 1.5 Ma. Examples of these may be the quartz artefacts of the Name Chamber described by Stratford (2008), which are thought to have been eroded from the area containing both the M5A and M5B deposits.

The StW 53 fossil deposits are potentially the same age as the LCA Acheulian bearing deposits of the Rietputs Formation of the Vaal River (Gibbon et al., 2009). These deposits have maximum ages between 1.89 ± 0.19 Ma and 1.34 ± 0.22 Ma (2.08–1.12 Ma) and minimum ages between 1.72 ± 0.16 Ma and 1.29 ± 0.21 Ma (1.88–1.08 Ma). Gibbon et al. (2009) suggest that these deposits show coeval development of Acheulian technology across the African continent at 1.6 Ma. No ages were obtained from the pit in which the stone tool sample was collected (ex-situ), although bifacial technology is noted to occur in all the mining pits. The closest age comes from Pit 1, where a single age suggested the LCA deposits in which the stone tools occur is between 2.08–1.56 Ma. This is suggested to be consistent with the last appearance date of *Metridiochoerous andrewsi* from the lower part of the Okote Member (1.63–1.51 Ma; McDougall and Brown, 2006). However, *M. andrewsi* specimens are known from all three Members of Swartkrans, the youngest of which may be less than 1 Ma (Herries et al., 2009b). In contrast, the ages for the LCA layers in Pit 2 range between 1.59 and 1.12 Ma (maximum ages) and 1.59 and 1.08 Ma (minimum ages), and multiple ages are consistent with depth. In contrast to Pit 1, these dates suggest that the Acheulian from Pit 2 could be as old as 1.6 Ma but could also be as young as 1.1 Ma. The study appears to suggest that the Acheulian in different parts of the Rietputs may be of quite varying age and that the gravel was deposited over more than half to as much as a million years. The LCA is noted to be as much as 4 m deep in some pits (up to 7 m overall) and as such a single date from a pit may not be reflective of the true age of the stone tools if they come from different layers within the gravel. As such, Acheulian artefacts as old as 1.6 Ma may occur in the Rietputs but much of, or potentially all of, the Acheulian may also be much younger. The Acheulian deposits at Wonderwerk Cave (1.78–1.07 Ma; Chazan et al., 2008) may also be contemporaneous with the StW 53 infill. Again, Chazan et al. (2008) suggest an age of ~1.6 Ma for the beginning of the Acheulian at Wonderwerk. However, the Acheulian first occurs in sediments dated to between 1.78 and 1.07 Ma and as such may again be much younger than 1.6 Ma, especially given apparent depositional breaks in the lower part of the sequence. Therefore, as with the Rietputs formation, the Acheulian of Wonderwerk Cave has the potential to be much younger than 1.6 Ma and thus contemporaneous with that at Sterkfontein between 1.3 and 1.1 Ma as described below.

Sterkfontein M5 (B-C) and perhaps the stone tool bearing deposits from the Name Chamber formed between 1.4 and 1.1 Ma and are contemporaneous with Swartkrans M2 (Balter et al., 2008), Coppers D upper deposits (<1.4 Ma; de Ruiter et al., 2009). An age of <1.4 Ma is much younger than expected for these Oldowan and *Paranthropus* bearing deposits (2.0–1.7 Ma based on typology and fauna; see Kuman, 2007). However, the Oldowan does occur in eastern Africa until this time period and indeed much of the same technology is still utilised throughout the Acheulian. Based on the limited number of hand axes, the Acheulian and *Homo* bearing M5C deposits are also a lot younger than expected at 1.3–1.1 Ma. In contrast, based on typology these have always been considered to be earliest Acheulian at 1.7–1.4 Ma (Kuman and Clarke, 2000; Kuman, 2007). However, it is possible that the Acheulian is not extensively seen, if at all, prior to 1.4–1.3 Ma in southern Africa. This is not that surprising given the fact that the Acheulean is equally not prolific in eastern Africa until after 1.5 Ma. Hopefully future work on both the stone tools and geochronology will resolve this issue and a more accurate age for Swartkrans is a key step. What is more certain is that the Sterkfontein M5B-C, Wonderwerk, and Rietputs Acheulian is seemingly older than the other dated southern African Acheulian bearing deposits such as Cornelia (~1 Ma; Herries et al., 2009a), Elandsfontein (1.0–0.6 Ma; Klein et al., 2007), and Gladysvale Internal Deposits (>0.78 Ma; Herries, 2003; Hall et al., 2006).

The Sterkfontein post-M6 MSA bearing deposits (470–289 ka) seem to be slightly older than the MSA bearing material in Lincoln Cave between ~290 and 107 ka (Reynolds et al., 2007), which is contemporaneous with late middle Pleistocene deposits at Florisbad (Grün et al., 1996) and Pinnacle Point (Marean et al., 2007). Younger material also occurs at Plovers Lake and Swartkrans within the last 110,000 years or so (de Ruiter et al., 2008; Sutton et al., 2009). Future work will hopefully recover material from the middle Pleistocene, which is underrepresented in South Africa, occurring at only a few sites such as Gladysvale between 780 and 580 ka (Lacruz et al., 2002) and perhaps the Cave of Hearths (Herries and Latham, 2009) and Elandsfontein (Klein et al., 2007).

These geochronological data suggest that a number of different australopithecine species occurred in southern Africa between 2.6 and 1.8 Ma and perhaps the first evidence of *Homo* and *Paranthropus* from Swartkrans M1, although at present this is one of the less well-dated hominin sites in South Africa. It appears that the first occurrence of most fossils attributed to *Paranthropus* and *Homo* is not until after 1.8 Ma, with *Paranthropus* teeth from Gondolin (~1.8 Ma), StW 53 *Homo* from Sterkfontein M5A (1.8–1.5 Ma), and *Paranthropus* from Kromdraai (<1.8 Ma) and Coopers (<1.6 Ma). Dupont et al. (2005) and Weigelt et al. (2008) suggest that major environmental change and aridity did not occur until ~2.1 Ma in southwest Africa, much later than in eastern Africa where it occurs from about 2.8–2.5 Ma (deMenocal, 2004). The palaeoenvironmental evidence appears to support this at Sterkfontein with wooded environments in M4 (2.6–2.0 Ma) and more open environments in M5 (1.8–1.1 Ma; Bamford, 1999). Hopley et al. (2007b) also note increased evidence for aridity in speleothems from Buffalo Cave in northern South Africa between 1.78 and 1.69 Ma that may relate to the beginning of the Walker circulation. These events appear to have had a major effect on hominin populations with the end of *A. africanus* and the StW 573 australopithecine species at around 2.3–2.0 Ma and the occurrence of new hominin species, first *A. sediba* at ~1.9 Ma and then *Paranthropus* and early *Homo* after 1.8 Ma. As such, a major increase in aridity appears to have occurred roughly 700–500 ka later in southern Africa than it did in east Africa, and this may also account for the later occurrence (<1.8 Ma) of *Homo* and *Paranthropus* in southern Africa, which notably also first occur soon after the onset of aridification in eastern Africa around 2.7–2.5 Ma.

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Appendix. Supplementary material

Supplementary data related to this article can be found online at doi:10.1016/j.jhevol.2010.09.001.

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