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Review article

Hominin-bearing caves and landscape dynamics in the Cradle of Humankind, South Africa

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ABSTRACT

This paper provides constraints on the evolution of the landscape in the Cradle of Humankind (CoH), UNESCO World Heritage Site, South Africa, since the Pliocene. The aim is to better understand the distribution of hominin fossils in the CoH, and determine links between tectonic processes controlling the landscape and the evolution and distribution of hominins occupying that landscape. The paper is focused on a detailed reconstruction of the landscape through time in the Grootvleispruit catchment, which contains the highly significant fossil site of Malapa and the remains of the hominin species *Australopithicus sediba*.

In the past 4 My the landscape in the CoH has undergone major changes in its physical appearance as a result of river incision, which degraded older African planation surfaces, and accommodated denudation of cover rocks (including Karoo sediments and various sil- and ferricretes) to expose dolomite with caves in which fossils collected. Differentially weathered chert breccia dykes, calibrated with ^{10}Be exposure ages, are used to estimate erosion patterns of the landscape across the CoH. In this manner it is shown that 2 My ago Malapa cave was ~50 m deep, and Gladysvale cave was first exposed; i.e. landscape reconstructions can provide estimates for the time of opening of cave systems that trapped hominin and other fossils.

Within the region, cave formation was influenced by lithological, layer-parallel controls interacting with cross-cutting fracture systems of Paleoproterozoic origin, and a NW–SE directed extensional far-field stress at a time when the African erosion surface was still intact, and elevations were probably lower. Cave geometries vary in a systematic manner across the landscape, with deep caves on the plateau and cave erosion remnants in valleys. Most caves formed to similar depths of 1400–1420 mamsl across much of the CoH, indicating that caves no longer deepened once Pliocene uplift and incision occurred, but acted as passive sediment traps on the landscape.

Caves in the CoH are distributed along lithological boundaries and NNE and ESE fractures. Fossil-bearing caves have a distinct distribution pattern, with different directional controls, a high degree of clustering, a characteristic spacing of 1700 m or 3400 m, and a characteristic bi-model fractal distribution best explained by a combination of geological and biological controls. It is suggested that clustering of fossil-bearing caves reflects a Lévy flight patterns typical for foraging behavior in animals. The controlling element in this behavior could have been availability of water in or near groups of caves, resulting in preferential occupation of these caves with accumulation of diverse faunal fossil assemblages.

The tectonic drivers shaping the dynamic landscape of the CoH did not involve large, seismically active fault lines, but complex interactions between multiple smaller fractures and joints activated in a far field stress controlled by uplift. The landscape of the CoH, with its caves and water sources and dissected landscape provided a setting favored by many animals including hominins. A modern day analog for what the CoH would have looked like 2 My ago is found 50 km east of Johannesburg, near the SE margin of the Johannesburg Dome.

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Contents

1. Introduction

It was in Africa that anthropoid primates gave rise to a lineage of hominins, our ancestors, through a series of evolutionary steps recorded in the fossil record of mainly eastern and southern Africa. Here hominin fossils are preserved along rifts, in the sediments of ancient lacustrine and riverine systems and in caves, raising the fundamental question whether the distribution of these fossils is the result of preferential preservation in sedimentary trapping sites, or whether the fossil distribution reflects a deeper relationship between hominins and the African landscapes in which they lived (e.g. King and Bailey, 2006). Understanding the relationship between the evolving landscape and fossil sites is, therefore, fundamental to understanding hominin evolution. This paper investigates that relationship for the fossil sites in the Cradle of Humankind, South Africa, focusing on the landscapes near Malapa, which hosts the important *Australopithecus sediba* fossils which represent a possible ancestral species to the genus *Homo* (Berger et al., 2010; Pickering et al., 2011a).

The geological record suggests that key evolutionary events in Africa coincided with important changes in the African landscape and climate (e.g. Vrba, 1995; Reed, 1997; Bobe et al., 2002; Bobe and Behrensmeyer, 2004; Passey et al., 2010). Since 35 Ma, the Earth's climate has cooled (Zachos et al., 2008) as the high plateaus in eastern and southern Africa rose to become one of the largest topographic anomalies on the planet's surface characterized by at least 500 m of positive residual elevation, commonly referred to as the African "superswell" (Nyblade and Robinson, 1994). Paleoclimate modeling (Sepulchre et al., 2006) indicates a causal link between uplift of the African plateau, and climate and landscape changes, with strong aridification and an increase in open grasslands since the Miocene (Bobe and Behrensmeyer, 2004; Feakins et al., 2005). Climate change had profound effects on distribution patterns of flora and fauna (deMenocal, 1995; deMenocal and Bloemendal, 1995; Hill, 1995; Kingston et al., 2007) and could have played a key role in the appearance and dispersal of early hominins (White, 1995; Vrba, 1995; Potts, 1998; Trauth et al., 2007; Maslin and Christensen, 2007; deMenocal, 2004, 2011). Attempts have been made to link the onset of extreme climate variability since the mid-Pliocene with fundamental evolutionary changes in hominins, including cranial expansion (e.g. Trauth et al., 2007; deMenocal, 2011; Donges et al., 2011).

Large-scale, long-term climate and vegetation changes in Africa are readily attributed to tectonic drivers, but tectonic effects are rarely considered as influencing hominin evolution within timeframes of less than 1 My and at scales that coincide with the territorial distribution of individual animal groups. Yet, uplift of the African plateau had dramatic impacts on the landscape occupied by hominins at a wide variety of scales (e.g. Reynolds et al., 2011; Bailey et al., 2011). In East Africa uplift was accompanied by the formation of the East African Rift System (Tiercelin and Lezzar, 2002), intimately associated with significant hominin finds (e.g. Tobias, 1985; Asfaw et al., 2002; White et al., 2009). Uplift of the African plateau rerouted large river systems including the Nile, the Congo and the Zambezi (Stankiewicz and deWit, 2006; Roberts et al., 2012), and accommodated the development of ecological corridors as well as biogeographic barriers along rift valleys and their bounding escarpments, creating complex, dynamic landscapes (e.g. O'Brien and Peters, 1999; Bailey et al., 2011), and the sediment traps necessary for the preservation of hominin fossils.

Local dynamic landscapes are created and sustained by active tectonic settings, and involve heterogeneous topography with diverse habitats (Bailey et al., 2000, 2011), with the potential to provide stable water sources, greater biodiversity and ample refuges (e.g. cliffs, gorges, caves etc.) that offer protection to hominins. Tectonically active zones rejuvenate the landscape at regular intervals (i.e. coincident with seismic events) and provide a buffer to long-term negative effects of climate change (e.g. Bailey et al., 1993; King et al., 1994; King and Bailey, 2006). Bailey et al. (2011) argue that dynamic landscape features characterize many of the South African hominid sites, including those in the Cradle of Humankind and suggest that the fossil distribution pattern represents a preference for hominins to occupy these sites, possibly even driving behavioral and morphological adaptations (Reynolds et al., 2011).

1.1. The Cradle of Humankind World Heritage Site, South Africa

The late-Pliocene to Quaternary cave deposits in the Cradle of Humankind (CoH), UNESCO, World Heritage area (Fig. 1), South Africa are one of the world's most important geological settings hosting hominin fossils and associated faunal and archaeological remains. From caves including Sterkfontein, Swartkrans, Malapa, Drimolen and Kromdraai (Fig. 1) abundant hominin fossils have been recovered, ascribed to a range of species including *Au. africanus; Au. robustus; Au. sediba* and early *Homo* (e.g. Brain, 1993; Tobias, 2000; Berger et al., 2010). Hominin remains in the caves are encased in clastic, cave-fill deposits situated in stromatolite-rich, dolomite sequences deposited on a late-Archaean continental shelf (Martini, 2006; Eriksson et al., 2006). The dolomite units outcrop along an arch-like, antiformal structure cored by Archaean basement gneiss referred to as the Johannesburg Dome (Fig. 2). Detailed geological descriptions of the most important fossil sites are available (e.g. De Ruiter et al., 2009; Pickering and Kramers, 2010; Dirks et al., 2010; Pickering et al., 2011b), but these studies focus on providing geological context to the immediate environment within which individual fossils are found. Perhaps surprisingly, only few studies (e.g. Partridge, 1973; Martini et al., 2003; Dirks et al., 2010) have looked at the broader geological context of the evolving cave systems, how these systems are expressed within the landscape and how the landscape may have influenced the distribution pattern of the fossils. Several recent, large-scale studies by Bailey et al. (2011) and Reynolds et al. (2011) used satellite imagery to interpret the tectonic geomorphology of the CoH, arguing for fundamental tectonic fault controls, but these studies do not link the features observed from space to the actual geology on the ground; leaving questions regarding the accuracy of their interpretations.

When considering the caves, the distribution of fossils within them and the geology and landscape in which the caves occur, many pertinent questions remain unanswered. Formation, and exposure of the cave systems in the CoH is linked to uplift and denudation of the high-elevation plateau of southern Africa (e.g. Partridge, 1973; Kavalieris and Martini, 1976; Martini, 2006; Dirks et al., 2010). Caves formed in vadose conditions before they were drained by lateral incision of eroding streams of the Crocodile drainage net, allowing phreatic processes to sculpt caves further through collapse and cave propagation along fractures and joints, through accumulation of cave sediments and through erosion allowing access to previously closed cave systems. The caves are an expression of extensional tectonics, which, though less dramatic than the rift valleys of east Africa, also resulted in dynamic, high-relief landscapes (e.g. Partridge and Maud, 2000; Moore et al., 2009; Bailey et al., 2011), with significant uplift since the Pliocene.

But when looking at local scales it is harder to understand the interplay between tectonic drivers and caves, and fossils buried within them. For example, is the presence of fossils near caves a reflection of preservation potential, or does the karst landscape present a habitat enjoyed by our ancestors, like bats in caves?; are the caves simply fortuitous traps in the landscape where fossils including those of hominins became deposited, and by extension, is the distribution of fossiliferous caves random?; or does the distribution of hominid fossils in caves, in some way reflect a fundamental underlying geological or biological process that makes certain caves more suitable then others; e.g. caves frequented by predators and scavengers may collect more bones (Brain, 1981;

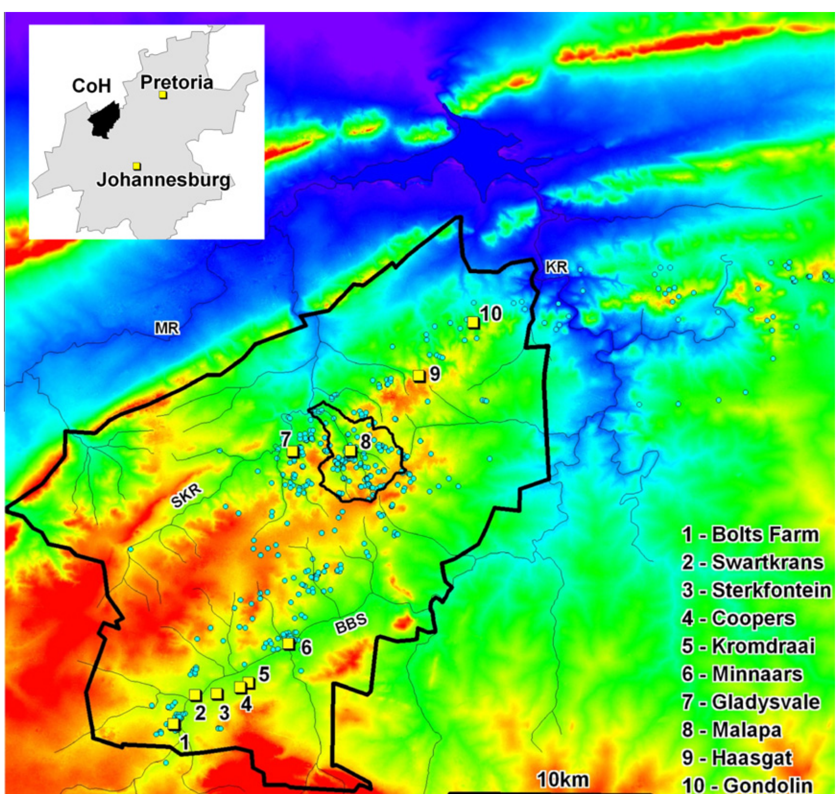

Fig. 1. Location of the Cradle of Humankind (CoH) world heritage area in NW Gauteng province (inset), South Africa, shown on a digital terrain model (red > 1700 m; purple < 1100 m) of the upper reaches of the Crocodile River catchment. Central to the CoH is the catchment of the Grootvleispruit with Malapa (8) at its center. The Grootvleispruit catchment has been used in this paper for detailed landscape reconstructions. Caves in the area are shown (blue circles) and the main fossil sites (yellow squares) have been listed, all of which contain hominin remains except for Minnaars (6) and Haasgat (9). MR = Magaliesberg River; KR = Crocodile River; SKR = Skeerpoort River; BBS = Blaubank River (or Blaubankspruit).

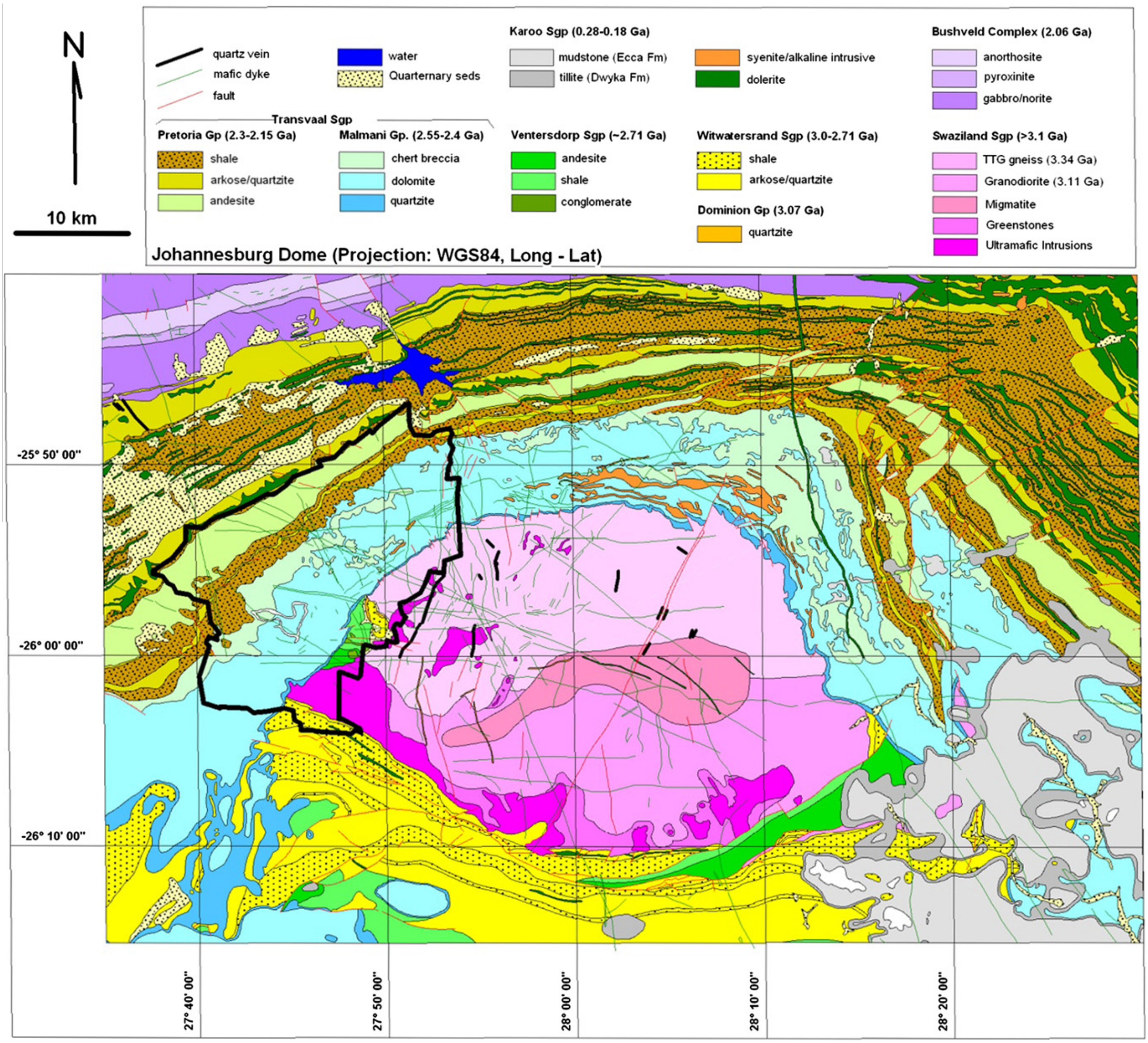

Fig. 2. Geological map of the Johannesburg Dome area. The boundary of the Cradle of Humankind is shown along the western margin of the dome.

De Ruiter and Berger, 2000; Pickering et al., 2004); or perhaps groups of hominins sheltered in caves for security similar to troops of baboons today.

Such questions do not just relate to distribution patterns of fossils but also to timing, i.e. to the dynamic nature of the landscape: how fast did the landscape change?; is the landscape we see today an accurate reflection of the landscape 1, 2 or 3 My ago?; when did caves open to allow animal remains to accumulate?; are erosion and exposure of caves occurring at predictable rates that will allow us to estimate the age of fossils by looking at the position of caves in the landscape?

At the moment we do not know the answers to these fundamental questions. Many attempts have been made to reconstruct what the African Plio–Pleistocene landscape and its habitats looked like in the CoH. Based on faunal, flora and isotope studies a mozaic environment has been reconstructed, which in the early-mid-Pleistocene, was more forested than today, with gallery forests along watercourses and nearby patchy open grasslands or woodland habitats in which Australopiths enjoyed a mixed C3–C4 diet on dolomite or mixed dolomite-shale-granite substrate (e.g. Vrba, 1982; Bamford, 1999; Lee-Thorp et al., 2003; Sponheimer and Lee-Thorp, 2003; Sponheimer et al., 2005, 2006; Reynolds, 2007; Bamford et al., 2010; Copeland et al., 2011). *Au. sediba* lived in the center of this broadly C4 environment, and enjoyed a dedicated C3 diet (Henry et al., 2012), indicating strong variability in hominin behavior. Whilst such reconstructions provide a snap-shot view of the landscape, they do not provide answers to the dynamic relations that may exist between the evolving physical environment and the non-random, evolutionary pressures it exerted on hominins.

The object of this paper is to provide geological and geomorphological context to cave sites in the CoH that contain fossil remains including hominins. By focussing on the distribution of fossil-bearing cave sites, in relation to key geological and geomorphological features, an attempt is made to highlight the dynamic nature of the landscape, and illustrate how physical and/or biological controls exerted a fundamental influence on the distribution of fossil remains. Evidence will be presented for the denudation history of the CoH in the upper Crocodile River catchment along the western margin of the Johannesburg Dome. The Grootvleispruit catchment will be used as a type area (Fig. 1). This well-constrained river catchment in dolomite hosts the important and precisely dated fossil site of Malapa (Dirks et al., 2010; Pickering et al., 2011a), and was visited, if not inhabited, by a family group of *Au. sediba* individuals 1.977 ± 0.002 Ma ago (Berger et al., 2010; Dirks et al., 2010; Pickering et al., 2011a). Malapa was found in August 2008 during a systematic geological survey of the area and differs from other nearby hominin-bearing sites in that a high concentration of hominin remains of at least six individuals occurs in a small outcrop (~60 m^2) of clastic sediment deposited along the lower parts of a now almost entirely eroded cave system that was originally >40 m deep (Dirks et al., 2010).

2. Geological and geomorphological setting

2.1. Geology

The CoH World Heritage area is situated in a region of the Kaapvaal Craton referred to as the Johannesburg Dome (Fig. 2). The Johannesburg Dome consists of a near circular, antiformal structure cored by Mesoarchaean (>3.1 Ga) basement gneiss (Anhaeusser, 2006; Robb et al., 2006) surrounded by outward dipping platform sequences of sedimentary and volcanic rocks of Neoarchaean to Paleoproterozoic (3.0–2.1 Ga) age. These include stromatolite-rich dolomite of the late-Archaean (2.64–2.50 Ga) Malmani Subgroup (Eriksson et al., 2006), which occur to the E, N and W of the 'Dome' and host the cave deposits in the CoH. The layered sequences are intruded by the 2060 Ma Bushveld Igneous complex to the N, and unconformably overlain by horizontally-bedded remnants of Permian sediment of the Karoo Supergroup (Fig. 2).

Dolomite of the Malmani Subgroup has been sub-divided into 5 formations (from base to top: the Oaktree, Monte Christo, Lyttelton, Eccles and Frisco formations) based on stromatolite morphology, chert content and the presence of shale and chert-breccia horizons (Eriksson and Truswell, 1974). In the CoH, the Oaktree formation is 150–200 m thick and consists of chert-poor dolomite that overlies quartzite and conglomerate of the Black Reef formation. The Oak Tree dolomite is interbedded with thin (<2 m thick) shale horizons, that are locally carbonaceous, and it contains two thin (<30 cm) tuff seams near its stratigraphic top. The most prominent tuff layer has been dated at 2585 Ma (Walraven and Martini, 1995) and is well exposed in Sterkfontein cave where it forms the roof to the Elephant Chamber and Milner Hall (Martini et al., 2003). The overlying Monte Christo formation (600–700 m) is a stromatolitic dolomite, rich in chert, with oolite beds near its base. The Monte Christo formation is interbedded with generally thin (<50 cm) shale horizons. The stratigraphic top of the Monte Christo formation is a ~5 m thick layer of sheared chert-breccia in dolomite matrix. It is overlain by the chert-free Lyttelton formation (150–200 m thick), a well-bedded, laminated dolomite topped by a 10–15 m wide chert-dolomite breccia horizon in which chert layers are highly deformed, folded and broken into angular blocks within a dolomite matrix. The overlying, chert-rich Eccles formation is up to 600 m thick with chert breccia near its stratigraphic top interpreted as an erosional feature (e.g. Eriksson et al., 2006) that separates it from the chert-free dolomite of the Frisco formation; a unit not exposed in the CoH. The dolomite units are overlain by chert-breccia of the Rooihoogte formation, which formed above an irregular karstic paleotopography.

The Rooihoogte formation has a close spatial association with caves and consists of a layer of clast-supported, chert breccia in a sandstone matrix (Figs. 2 and 3). This layer is variable in thickness, as a result of syn-sedimentary block faulting of the underlying dolomite. In the CoH, the envelope to the Rooihoogte formation forms a gently NW dipping layer that slopes at 3–10° i.e. less than the underlying dolomite, resulting in a low-angle angular unconformity of ~6° (Fig. 4). In the CoH, the Rooihoogte formation overlies the upper Eccles formation along the NW contact of the Malmani dolomite, and is transgressive onto Monte Christo formation dolomite to the SE (Figs. 2 and 3). Rooihoogte formation breccias are locally reworked, presumably along paleo-river channels to form lenses of chert conglomerate (referred to as the Bevets member).

A characteristic feature of the Rooihoogte formation is its irregular, karstic lower contact associated with orthogonal to polygonal networks of normal fractures that developed open chasms on the Paleoproterozoic landscape. These fractures were in-filled with chert breccia in a sandstone matrix, to appear as sedimentary dykes in dolomite that penetrate up to 80–100 m below the paleosurface defined by the base of the Rooihoogte formation. The Rooihoogte formation probably developed as a result of extension of the paleo-shelf and opening of the basin in which the Pretoria Group sediments were subsequently deposited.

Overlying the Rooihoogte formation is shale of the Timeball Hill formation, which forms the basal unit of largely clastic sediments of the Paleoproterozoic Pretoria Group (Eriksson and Truswell, 1974; Eriksson et al., 2006). The dolomite units in the CoH experienced greenschist facies metamorphism and bedding-parallel shear, recording top-to-the-north movement (Gibson et al., 1999; Alexandre et al., 2006) along mylonite zones in dolomite and interbedded shale units (Alexandre et al., 2006). Shearing is locally associated with north-verging folding and foliation development (Gibson et al., 1999). The dolomite was intruded by metamorphosed diorite sills and dykes of unknown age.

2.2. Geomorphology

The geomorphology of Africa is characterized by well-developed, erosion surfaces that formed as a poly-genetic landform in response to a series of regional uplift and planation events since Cretaceous breakup of Gondwana (e.g. Burke and Gunnell, 2008; Partridge, 2010). King (1949) recognized relics of successive erosion surfaces of different ages in southern Africa, including the senile, continent-wide, African surface, later interpreted as a deeply weathered, low-altitude, largely flat, tropical plain formed at the end of the Crateceous (e.g. Burke, 1996). Uplift of this plane occurred across broad, domal flexures or "epeirogenetic uplift axes", which warped the African surface upward rejuvenating it from time to time to form younger planation surfaces (the Post-African I (Miocene) and Post-African II (Pliocene) surfaces, e.g. Partridge, 2010), and causing head-ward erosion and incision of streams along the escarpment edges. In southern Africa it has been suggested that the landscape was uplifted during the Oligocene (30 Ma, Burke, 1996; Burke and Gunnell, 2008), early-mid Miocene (15 Ma) and Pliocene (2–5 Ma; Partridge and Maud, 1987; Partridge, 2010). Based on outcrops of Miocene and Pliocene marine sediments (Partridge et al., 2006), and offshore unconformities and sedimentary fill along the continental margins of southern Africa (e.g. McMillan, 2003), estimates for Miocene uplift are 200–300 m, whilst uplift since the Pliocene may have been as much as 900 m east of the Drakensberg escarpment (Partridge, 2010); although the amount of this uplift is contested (e.g. Erlanger, 2011). Uplift in the Johannesburg area was probably less, but could still have been significant.

The area of the CoH and the Johannesburg Dome, positioned just north of the "epeirogenetic uplift axis" warping the African erosion surface (e.g. Partridge and Maud, 1987), show a variety of landscape elements of the various African erosion surfaces incised by the headwaters of the Crocodile River (Figs. 1 and 2). The southern margin of the Johannesburg Dome is characterized

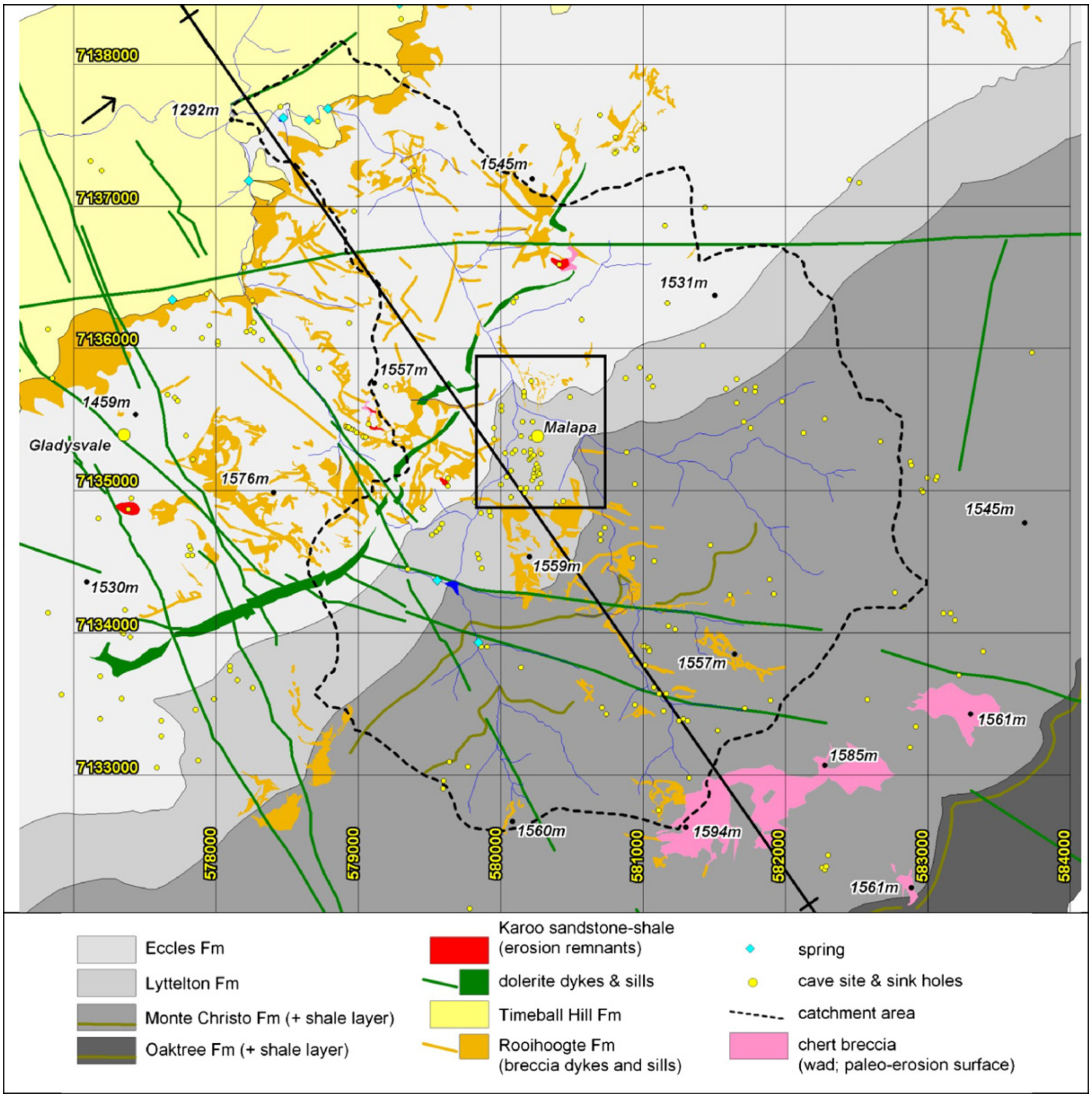

Fig. 3. Detailed geology of the Grootvleispruit catchment area and immediate surroundings. The position of this catchment is shown in Fig. 1. The cross section (bold line) is shown in Fig. 4. The detailed geology around Malapa (the inset box) is shown in Fig. 5.

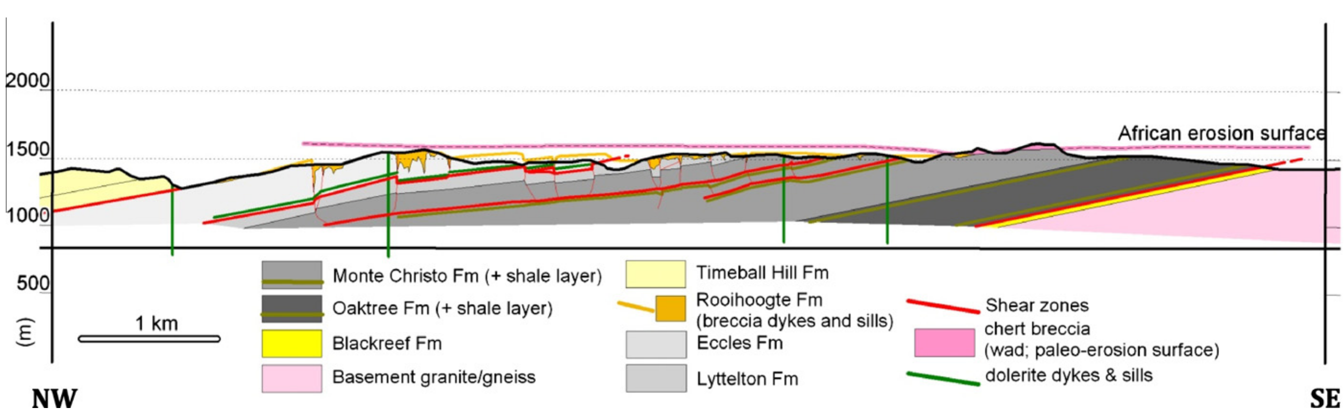

Fig. 4. NW–SE trending, geological cross section across the Grootvleispruit catchment. Details are discussed in the text (position of section line is shown in Fig. 3).

by an escarpment up to 1900 m high, formed by quartzite of the 3.0–2.8 Ga Witwatersrand Supergroup, which forms the watershed between the Crocodile-Limpopo River system draining N and E, and the Orange River system draining to S and W. South of the escarpment is the start of a flat plateau mostly underlain by near-horizontal sedimentary rock sequences.

To the E, N and W of the Johannesburg Dome outward dipping sediments of the Transvaal Supergroup form undulating hills accentuated by quartzite and chert ridges, bounded to the N by a prominent escarpment, the Magaliesberg, composed of coarse-grained quartzite through which the Crocodile River has incised a narrow gorge, now dammed. In the past Karoo sediments would have covered most of the Johannesburg Dome as attested by erosion remnants (e.g. Wilkins et al., 1987).

In the CoH several parallel ridges of sandstone and chert crop out, and the landscape rises from N to S, drained by two tributaries of the Crocodile River: the Skeerpoort River to the N, and Blaubank River to the S. These two tributaries host all major fossil-bearing caves within their catchments (Fig. 1). Chert breccia of the Rooihoogte formation is weathering resistant and underlies a 1–2 km wide, dissected dip slope that formed along the upper contact of the dolomite along sections of the Skeerpoort River in the center of the CoH. A number of NW draining tributaries to the Skeerpoort River, including the Grootvleispruit, which will be discussed in detail below, have cut narrow, steep-sided gorges through this dip slope. Prominent nick points characterized by waterfalls, and in places spring sites have developed where creek beds cut across the Rooihoogte formation; i.e. all nick points in this area are lithologically controlled.

3. The evolving landscape of the CoH

The ephemeral Grootvleispruit in the center of the CoH forms a catchment, 15.51 km^2 in size, situated within the Malapa Nature Preserve on the farm Diepkloof (Fig. 1). Malapa cave with its spectacular fossils of *Au. sediba* occurs towards the geographical center of the Grootvleispruit catchment approximately 15 km NNE of Sterkfontein. Numerous caves, several springs and complex landscape features occur within this catchment (Fig. 3), which provides a well-constrained setting to reconstruct the landscape at the time *Au. sediba* was alive, by using geological structures, geomorphological features and cave distribution patterns.

3.1. The Grootvleispruit catchment

The Grootvleispruit is fed by a spring 1.5 km SSW of Malapa at the point where a graphitic shale band in the Monte Christo formation crosses the valley. Water from this spring disappears into subterranian cave systems 500–800 m below its source. Water reappears in a second, larger spring 2.6 km NW of Malapa near the upper contact of the dolomite, where a thin (<2 m thick) layer of chert breccia of the Rooihoogte formation is overlain by shale of the Pretoria Group. Below this spring active tufa formation occurs as the river passes onto a sandstone-shale substrate across a series of rapids where dissolved CO_2 is lost from the water.

The Grootvleispruit has eroded into the African Erosion surface (Partridge et al., 2006), which in this part of the CoH occurs at $\sim$1550 m. The drainage divide of the catchment occurs along flat-topped hills with summits at a constant altitude of 1545–1560 m (Fig. 3). The greatest altitude of $\sim$1600 m is reached along a ridge that borders the catchment to the SE, whilst the confluence with the Skeerpoort River in the NW occurs at 1292 m. The Grootvleispruit drains into a NW direction at near-right angles to the regional strike of bedding within the Malmani Subgroup. Large sections of the creek are relatively straight and trend parallel to a prominent set of NW striking joints and fractures, which are locally intruded by dolerite dykes. The catchment area is underlain by layered dolomite of (from SE to NW) the Monte Christo, Lyttleton and Eccles formations, which dip at $\sim$10–15° to the NW, unconformably overlain by chert of the Rooihoogte formation (Figs. 3 and 4), which dips 5–10° to the NW. Rooihoogte chert overlies the Eccles formation in the NW and central parts of the Grootvleispruit catchment, and the Monte Christo formation in the SE (Fig. 3).

The site of Malapa represents an erosion remnant of a deroofed, in-filled cave with no visible open chambers (Dirks et al., 2010). It occurs in layered dolomite, which dips gently (8–13° to the NNW) at an angle 2–5° shallower then the valley slope where a series of caves are exposed along a fracture system, stratigraphically bound to the top 40 m of the Lyttelton Formation (Figs. 4 and 5).

3.2. Breccias and Paleo-erosion surfaces

Within the Grootvleispruit catchment, several types of sedimentary breccia occur that are associated with paleo-erosion surfaces. Three breccias types can be used to constrain the evolving Pleistocene landscape and cave distribution in the CoH: 1. Breccias associate with pedogenic erosion remnants underlying the African (or Post-African) erosion surface; 2. Blocks of Karoo sandstone some contained within the pedogenic erosion remnants of (1), and some associated with faults and paleo-(Permian?) sinkholes to form Karoo erosion remnants on the landscape; and 3. Breccia of the Rooihoogte formation on top of a faulted Paleoproterozoic karst topography.

Apart from these three breccia types, breccias also occur as fossil-bearing, Pleistocene cave fill (Dirks et al., 2010), and as tectonic crush zones along shear zones and fault planes. These latter two types will not be discussed in detail in this paper.

3.2.1. Pedogenic breccia underlying an African erosion surface

Along the ridge bounding the Grootvleispruit catchment to the SE, at altitudes over $\sim$1570 m, red-colored breccia caps occur as erosion remnants along the crest line (Fig. 3). The breccias are clast-supported, and consist of angular chert blocks and, to a lesser degree, rock fragments of shale and sandstone cemented by a fine grained, brick-red matrix of Mn- and Fe-oxides, subsequently impregnated with siliceous fluids resulting in silicified, weathering-resistant rocks (Fig. 6a). Some rock fragments are up to 4 m in diameter and include red, fine-grained, un-metamorphosed, but silicified sandstone, probably of Karoo age. This breccia type is referred to as ‘wad’ (e.g. Martini, 2006) and formed as a residual mantle along a paleo-erosion surface, left after dissolution of dolomite (Eriksson and Truswell, 1974). The wad marks a paleo-erosion surface, presumably the remnants of the African erosion surface (Partridge and Maud, 1987; Fig. 4). Around Malapa, small remnants of wad can be found along the crests of some of the flat-topped hills on either side of the Grootvleispruit catchment, indicating that in these places the African erosion surface passed only 10’s of meters above current erosion levels at a height of about 1560 m. Similar remnants of wad also occur elsewhere in the CoH; e.g. on the hill top 300 m SE of Coopers where an erosion cap of wad remains above 1500 mamsl (Fig. 6b). Although it has been suggested that the African erosion surface dips gentle N across the CoH with a drop of 5 m per km (Partridge, 1973; Partridge and Maud, 2000), this is not consistent with the much more irregular distribution of Wad, in terms of its height in mamsl, observed on the landscape in the CoH as part of this study, and it is not clear whether all outcrops of wad actually represent parts of the same erosion surface, since none of these units has actually been dated.

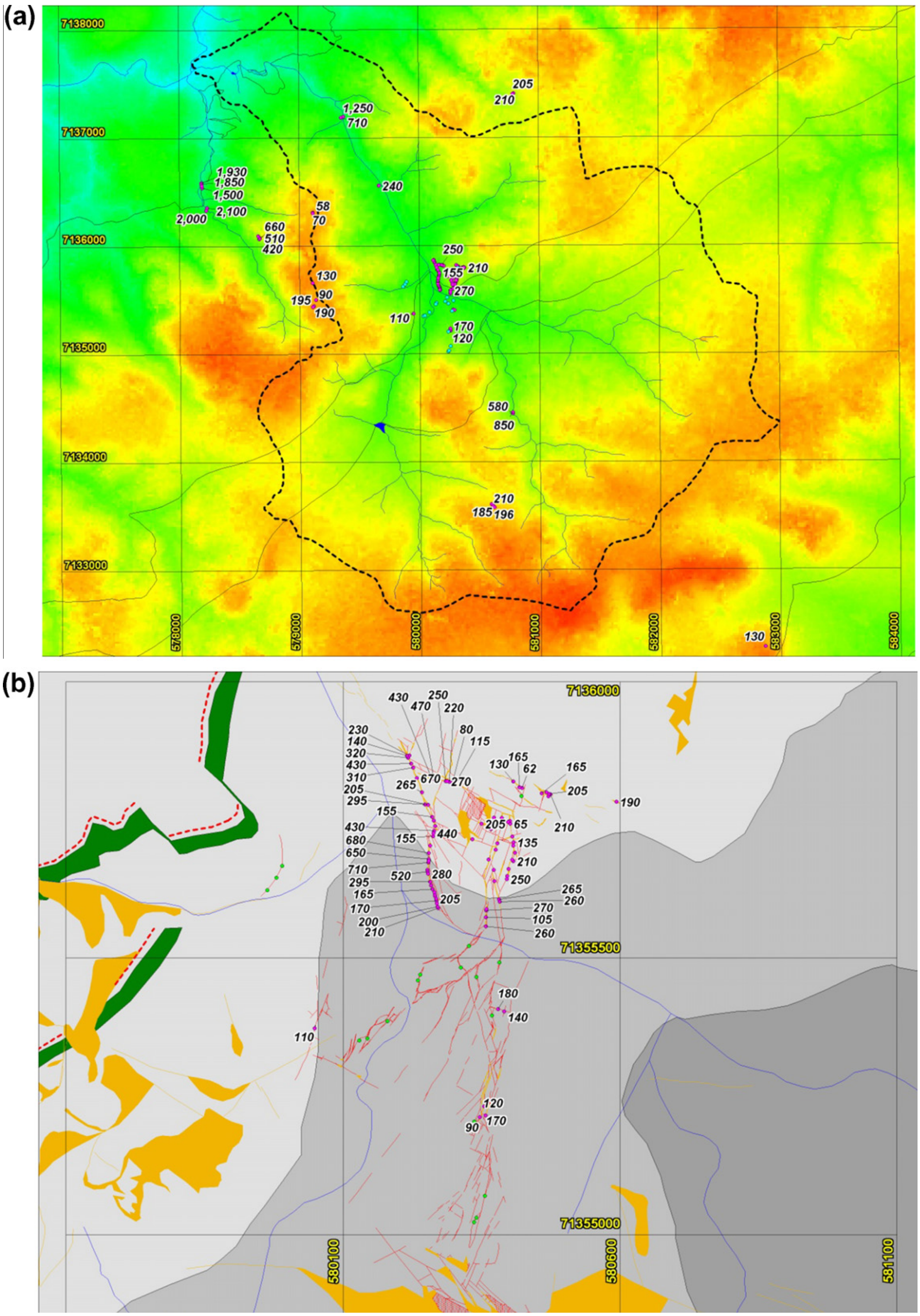

Fig. 5. (a) digital terrain model (red > 1600 m; light blue < 1300 m) for the Grootvleispruit catchment. Locations of chert breccia walls (maximum height in cm) are shown as purple dots. The location of chert vein stockwork walls are shown in blue dots. (b) Detailed geological map of the Malapa area (legend and position of map is shown in Fig. 3). Locations of chert breccia walls (maximum height in cm) along the network of orthogonal fractures are shown as purple dots. Locations of chert vein stockwork walls are shown as green dots.

3.2.2. Karoo remnant from an earlier cover sequence

Along the ridge line on either side of Malapa, small erosion remnants of well-bedded, cross-stratified, unmetamorphosed, red sandstone can be found, which probably represent remnants of fluvial sandstone of the Vrijheid formation (i.e. the northern deltaic facies of the basal Ecca Group of the Karoo Supergroup; Johnson

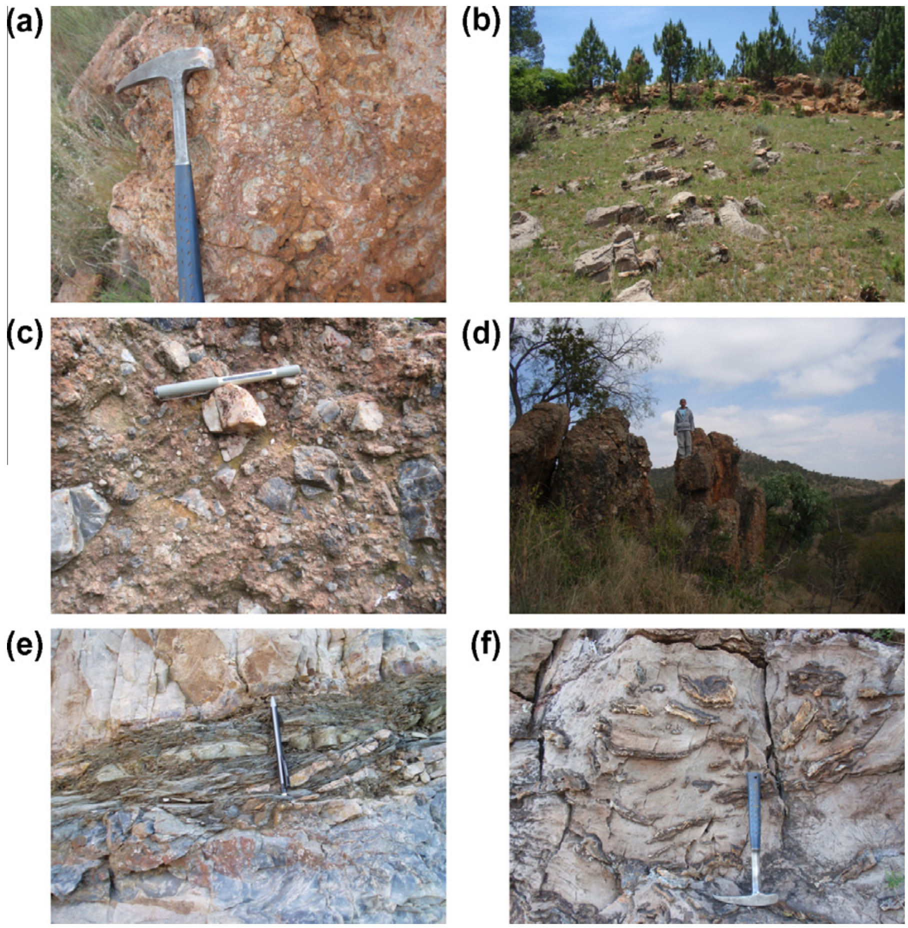

Fig. 6. (a) Clast-supported breccia of angular chert blocks and rock fragments of shale and sandstone cemented by a fine-grained, brick-red matrix of Mn and Fe oxides that is heavily silicified. The breccias are referred to as 'wad'. (b) An erosion cap of 'wad' on the hill top 300 m SE of Coopers cave (Fig. 1). The 'wad' overlies a paleo-erosion surface that marks the African erosion surface of Partridge and Maud (1987), which in the area of the photo occurs at a height of ~1500 mamsl. (c) Chert breccia in a sandstone matrix near the base of a fracture that formed during deposition of the Rooihoogte formation 50 m east of Malapa pit. The breccia is clast supported with angular breccia clasts that consist almost exclusively of chert derived from surrounding dolomite units embedded in a sandstone matrix, subsequently cemented with chert. (d) The chert breccia-filled fractures of the Rooihoogte formation form conspicuous, wall-like landscape features where they weather out differentially from surrounding chert. The wall in this photograph occurs 300 m NNW from Malapa and reaches a height of 6.8 m. (e) Layer-parallel shear zone in a shale band near the base of the Monte Christo formation. Caves commonly form along shear zones like this. (f) Layer-parallel shear zone in dolomite near the top of the Lyttleton formation. The deformational nature of this horizon can be deduced from intense folding and dismemberment of chert horizons.

et al., 1997, 2006; Fig. 3). Karoo remnants occur in several forms. At GR579110–7135440 and 579090–7135560 (all grid references are in UTM WGS84), meter-sized blocks of sandstone occur as red breccia fragments in a silicified sandstone matrix locally cemented with Fe–Mn oxides; i.e. these Karoo blocks occur in wad-like erosion remnants, at the base of an African erosion surface. The Karoo material is arranged in NW-trending, 10 m wide, elongated outcrops along a fracture zone running along the top of the hill. To the SE, At GR579610–7135060 (Fig. 3), a third outcrop with large blocks of red Karoo sandstone in a silicified sandstone matrix occurs along a 10–25 m wide, NW trending corridor exposed in the side of the hill between altitudes of 1530–1490 mamsl. The Karoo breccia contains several meter large, cross-bedded sandstone blocks, and appears to represent a cavity fill along a fault that formed at a time when Karoo rocks still covered the dolomite. At GR580430–7136590 (Fig. 3), a further erosion remnant of red sandstone breccia in wad occurs as a several meters thick, sheet-like erosion cap near the top of the hill.

A synformally folded Karoo remnant of white siltstone and shale overlying coarse boulder beds (Dwyka formation diamictite?) occurs in a 170 m $\times$ 80 m, oval-shaped outcrop just below the ridge line at GR 577380–7134880 (1480–1500 mamsl; Fig. 3), 500 m S of Gladysvale, and 2.5 km W of Malapa. This Karoo outcrop represents fill of a sinkhole or de-roofed cave system, above a paleokarst surface of Permian age, similar to Karoo cave fill described from the Lyttelton quarry south of Pretoria (Marker, 1974; Wilkins et al., 1987).

The erosion remnants of Karoo illustrate that Karoo sediment blanketed the catchment area in the past. At the time when the African (or Post-African) erosion surface formed, Karoo-aged rocks still existed as a thin blanket in the Grootvleispruit area resulting in blocks of Karoo sediment contained in wad. The stratigraphic base of the Karoo cover cannot have been much above the current altitude of the erosion remnants at ~1550 mamsl.

The distribution and age of fossils in caves in the CoH is linked to the erosion history of the Karoo-aged cover rocks. Progressive removal of the veneer of Karoo sediment allowed exposure of the caves, granting access to fauna, including hominins (e.g. Martini et al., 2003); it is thus logical that the earlier the Karoo was removed and caves could develop or cave openings were exposed, the older the fossils that can be expected to accumulate in the caves.

3.2.3. Rooihoogte breccia, erosion remnants and down-cutting rates

The breccia–sandstone filled fracture networks along the base of the Rooihoogte formation (Figs. 3 and 5) have geometries that

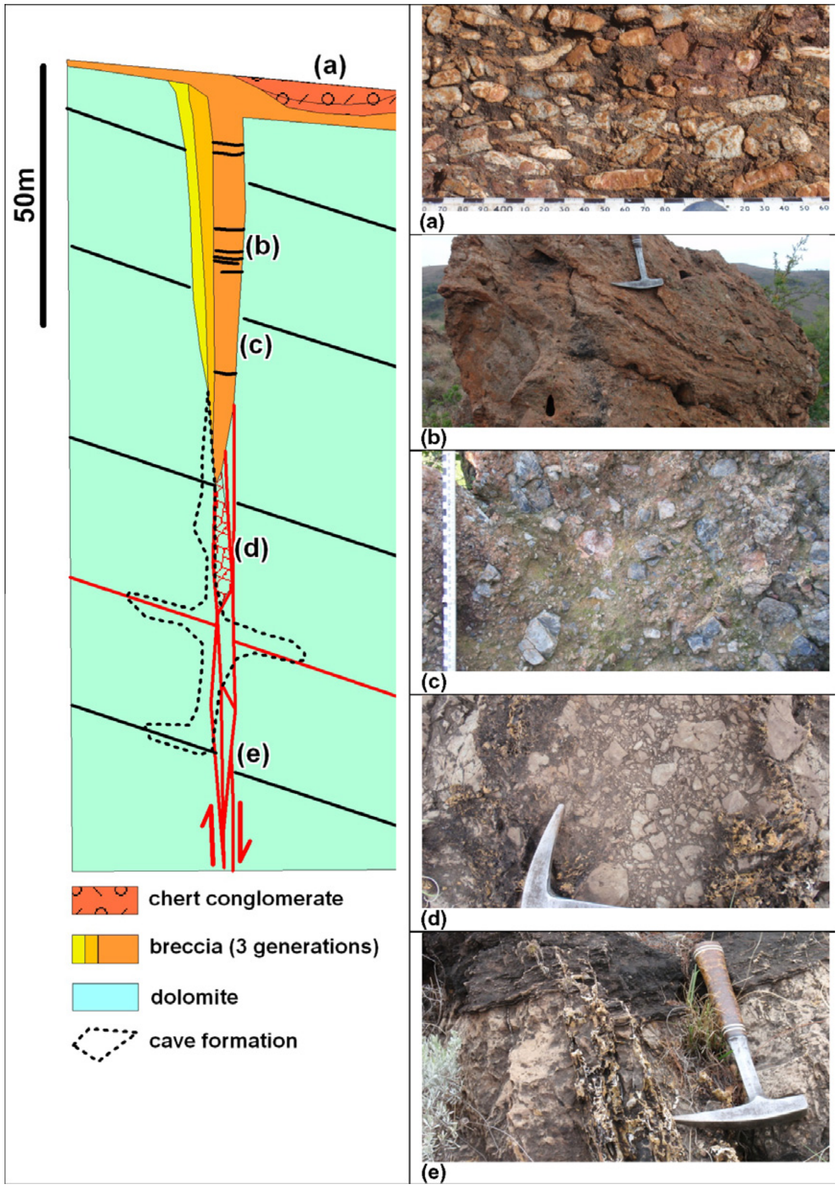

Fig. 7. Cartoon section of a breccia–sandstone-filled fracture system along the base of the Rooihoogte formation. The character of the fracture and its fill changes as a function of stratigraphic level. The chert-breccia-in-sandstone sheets are wedge-shaped and widen upward. Towards the stratigraphic top of the Rooihoogte formation, lenses of chert conglomerate occur (a). At lower levels, within the top of the fracture systems fracture fill may consist of sandstone with few angular to rounded chert clasts preserving horizontal layering and graded bedding (b). At lower levels in the fracture, breccia is clast supported, and angular clasts consist almost exclusively of chert derived from surrounding dolomite (c). Near the base of the sediment-filled fractures, hydrothermal breccia zones in dolomite are locally preserved (d). These breccia zones merge with 0.001–3 m wide, near-vertical, sheet-like, chert vein, stockwork systems that developed along normal faults below the base of the sediment filled fractures (e). Caves developed preferentially along these fracture systems.

vary with depth below the paleo-erosion surface on which they formed. In an outcrop 50 m east of Malapa, the base of a breccia-filled fracture can be seen in the form of a 20–30 cm wide chert breccia sheet (Fig. 6a and b). Moving up-hill N and S of Malapa, the dyke-like, breccia sheet merges into thick, irregular bodies of breccia at a height of about 1500 mamsl (Fig. 5). Continuous chert-breccia caps occur above an altitude of about 1520 mamsl, where the steeply dipping fractures merge with the layered Rooihoogte formation.

Based on outcrops in the Grootvleispruit catchment, Fig. 7 illustrates how the morphology of the fractures changes as a function of stratigraphic level. Malapa occurs near the base of one of the fracture systems, which acted as extensional normal faults at the time of deposition of the Rooihoogte formation. Along the fractures near Malapa (height 1442 mamsl), hydrothermal breccia zones in dolomite merge with 0.001–3 m wide, near-vertical, sheet-like, chert-vein-stockwork systems. At higher levels the fault zones created planar cavities in-filled with sheet-like bodies of poorly-sorted, chert breccia/conglomerate in a sandstone matrix with chert cement. The chert-breccia-in-sandstone sheets are wedge-shaped and widen upward. Some sheets display meter-scale, vertical layering reflecting progressive fracture opening and infill. The composition of clasts in the sediment fill changes along single fractures from deeper to shallower stratigraphic levels. At lower levels breccia is clast-supported, and angular clasts consist almost exclusively of chert derived from surrounding dolomite units. At higher levels the sand matrix component increases together with the percentage of allochtonous fragments including shale and vein quartz, and the degree of rounding of individual fragments increases. With the increase in sandstone matrix within fracture

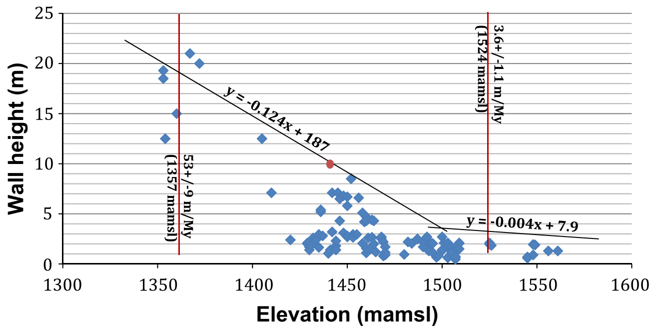

Fig. 8. Height of erosion remnants of chert breccia walls in and near the catchment of the Grootvleispruit, plotted against surface elevation along the base of the walls (observational locations are shown in Fig. 5). The position of Malapa is shown as a red dot. The maximum wall height decreases in a linear manner with elevation. Above 1500 mamsl the rate of decrease is small (~0.4 m per 100 m elevation gain) reflecting lowering of the African erosion surface at a rate of ~4 m/My (Dirks et al., 2010); between 1350 and 1500 mamsl the rate of decrease is larger (~12 m per 100 m elevation gain) reflecting valley incision. The linear decrease in wall height, and measured denudation rates at 1357 mamsl and 1525 mamsl (Dirks et al., 2010) can be used to derive a simple relationship between elevation and erosion rate (see text for details).

fill towards the stratigraphic top of individual fracture systems, horizontal layering, graded bedding, and cross bedding can be found indicating water flow through the fractures at the time of deposition (Fig. 7).

Fig. 3 shows the distribution of the Rooihoogte formation in the Grootvleispruit catchment including the vertical breccia sheets, which locally cross the valley. Breccia sheets broadly trend WNW and ENE, but at scales of <100 m different orientations occur along complex networks of orthogonal fractures (e.g. around Malapa, Fig. 5b).

In the western part of the Grootvleispruit catchment, chert-breccia sheets penetrate the Eccles formation to a vertical height of ~1380 mamsl. Around Malapa, the base of the breccia sheets occurs within the Lyttelton formation at a height of ~1440 mamsl. Near the SE margin of the catchment, the base of the breccia sheets occurs at 1540 mamsl near the contact of the Oaktree and Monte Cristo formations. In section, the dolomite units dip WNW at an angle of ~10–15° (with steeper dips occurring near faults), whereas the envelope of the Rooihoogte formation dips NW at about 5° (Fig. 4).

Apart from the obvious Palaoproterozoic, tectonic significance of the fracture systems, the fracture-filled, chert breccia sheets are of great significance for reconstructing the Plio–Pleistocene landscape for two reasons: 1. Many caves in the CoH formed in close spatial association with chert breccia of the Rooihoogte formation (e.g. Fig. 5b; this will be discussed in more detail below); and 2. The chert breccia is highly silicified making them more resistant to weathering than the surrounding dolomite (Fig. 6d). The latter point has two important implications in terms of assessing the evolving landscape. Firstly, weathering-resistance of chert has resulted in the formation of dip slopes of Rooihoogte formation, forming the SE valley slopes of the Skeerpoort River between 1300 and 1500 mamsl (Figs. 3 and 4). The dip-slopes developed because overlying shale of the Timeball Hill formation was removed through lateral down-cutting of the river. Dirks et al. (2010) report a rate of vertical down-cutting of the Skeerpoort River below Gladysvale cave (at GR576678–7135760) of 53 ± 9 m/My. From a point along the river at 1339 mamsl, i.e. several 100 m downstream from where the cosmogenic sample was taken, the dip slope of Rooihoogte formation chert breccia is intact and can be traced to the top of the hill (at 1459 mamsl) that forms the roof of Gladysvale cave ~850 m to the SE (Figs. 3 and 4); i.e. the dip slope dips at ~8°. Thus, a vertical down-cutting rate of ~55 m/My along the river translates into a NW-directed horizontal shift of the river along the dip slope of ~400 m/My, as shale beds of the Timeball Hill formation were progressively removed, mainly through undercutting by the Skeerpoort River. In this framework, a simple calculation suggests that Gladysvale cave was first exposed by the Skeerpoort River ~2 My ago, and it is unlikely that Gladysvale will contain fossils that are much older than ~2 Ma; a fact consistent with work at the site (e.g. Lacruz et al., 2002; Pickering et al., 2007).

The second important landscape feature related to the chert breccia sheets is their weathering pattern: the sheets weather out differentially from surrounding dolomite, to form conspicuous wall-like features that vary in height from 10's of centimeters to 20 m. The walls are generally less than 2 m in width, and width generally increases with wall height. The maximum residual height of the chert-sandstone walls increases from the plateaus along the hill crests at altitudes of >1520 mamsl, where walls are up to 2 m high, to the valley bottom of the Skeerpoort River where walls reach heights of 20 m at ~1350 mamsl altitude (Fig. 5). Chert stockwork veins along faults form similar erosion features, but are less well developed and less wide spread (Fig. 5). Fig. 8 shows a plot for chert breccia walls in the catchment area, in which maximum wall height is plotted against altitude (in mamsl) measured at the base of each wall.

When moving from the plateau into the valleys of the CoH, the observed maximum wall height increases in a linear manner at a rate of about 6 m gain in wall height per 50 m drop in altitude between 1500 and 1350 mamsl (Fig. 8). On the rolling plateau, between altitudes of 1500 and 1560 mamsl, maximum wall heights are ~2 m and appear to decrease slightly at a rate of 0.4 m per 100 m elevation gain. These observations suggest that the flat plateau along the top of the hills is an erosion surface that was lowered gradually at more or less the same rate across the area. Dirks et al. (2010) report a ^{10}Be cosmogenic denudation rate of 3.6 ± 1.1 m/My for a quartz sample at 1524 mamsl from the chert-capped plateau 1 km south of Malapa. Dolomite along the plateau would therefore erode at 5–6 m/Ma to account for the observed differential erosion between chert breccia and dolomite.

The linear decrease in wall height along the slopes of the Grootvleispruit catchment, and measured denudation rates at 1357 mamsl (of 53 + 9 m/My for incision of the Skeerpoort River) and 1524 mamsl (Dirks et al., 2010) can be used to derive a simple relationship between elevation and erosion rate in the CoH; i.e. the maximum wall heights of the chert breccias across the landscape of the CoH can be used to estimate denudation rate. In presenting this relationship, we are aware that detailed reconstructions of denudation rate should take many variables into account including slope, rock type and climate variability (e.g. Portenga and Bierman, 2011), and that the estimates based on wall heights can only be used locally on a dolomite substrate; nevertheless they provide a useful first-order tool in the CoH. Assuming the base of the African erosion surface (with a denudation rate of ~4 m/My) is at 1525 mamsl this relationship is:

$$-0.29[\text{elevation}] + 446.25\ \text{m} = [\text{erosion rate}]$$

providing an estimated denudation rate of ~28 m/My at Malapa. If the base of the African erosion surface (with a denudation rate of ~4 m/My) is set at 1500 mamsl the relationship becomes:

$$-0.34[\text{elevation}] + 514\ \text{m} = [\text{erosion rate}]$$

providing an estimated denudation rate of ~24 m/My at Malapa. Either way, the estimates indicate that the cave would have been ~50 m deep 2 My ago when the *Au. sediba* fossils were trapped. Using this logic, a reconstruction can be made along a schematic

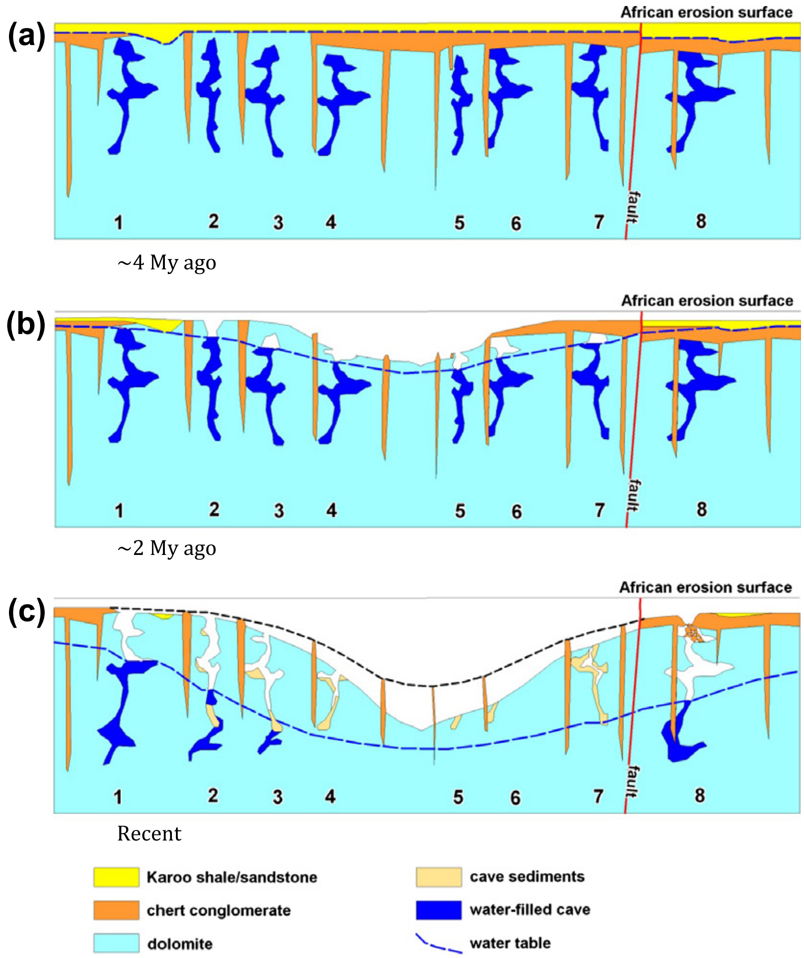

Fig. 9. Cartoons showing a time series for landscape formation and cave opening in the Malapa area, where the base of the cave systems occur at a constant elevation of ~1440 mamsl, with Malapa representing an erosion remnant of the lower part of a cave system. Chert breccia walls form erosion remnants that increase in height in a systematic manner (Fig. 8) when moving from the plateau into the valley. See text for a detailed discussion. (a) situation approximately 4 My ago; (b) situation approximately 2 My ago; (c) current situation.

N–S cross section across the Grootvleispruit valley, to show what the area would have looked like back in time (Fig. 9).

The reconstruction in time in Fig. 9 shows that in the Malapa area, the base of the cave systems appears to occur at a constant elevation of ~1440 mamsl, with Malapa occurring as an erosion remnant of the lower most part of a cave system. Moving up the hill side S of Malapa, caves along the same fracture system gradually deepen at the same rate the land surface rises (Dirks et al., 2010). This observation has been made more generally for caves in the COH (e.g. Martini, 2006) and suggests that the caves formed relatively early with respect to a paleo-water table, presumably at a time that the African erosion surface was still intact in the area. As a consequence, cave systems display different characteristics within the catchment of the Grootvleispruit. On the plateaus at >1500 mamsl, deep cave systems can be found characterized by vertical shafts and sinkholes. Along the valley sides, caves are intersected at a variety of cave levels as horizontal galleries move into the valley sides. Along the valley floors, few cave systems remain; instead erosion remnants of caves can be found (Fig. 9). These features suggest that in the last few million years new cave formation was slow. Instead the caves largely acted as passive holes, that were filled with sediment as they were dissected and progressively eroded.

Using the observations above it is possible to make a more general, 1st order reconstruction back in time of what the landscape in the CoH would have looked like (Fig. 10). For a cross-section along the Grootvleispruit valley, 2 My time slices have been drawn in Fig. 10. In reconstructing these sections it has been assumed that: 1. In the Grootvleispruit catchment area the Africa paleo-erosion surface existed at an altitude of ~1520–1550 mamsl, this surface was gradually lowered; 2. The 1520–1550 mamsl level approximately coincided with the contact between the Malmani dolomite and a blanket of overlying Karoo sediment; 3. Shale of the Timeball Hill formation was removed more rapidly than the dolomite resulting in dip slopes along hard grounds (e.g. Rooihoogte formation chert). Fig. 10 illustrates that ~2 My ago, Gladysvale cave would have been uncovered by erosion, whilst the landscape around Malapa was fundamentally different, with narrow valleys incising through Karoo and wad in the top of underlying dolomite; ~3 Ma ago, the land surface across the Grootvleispruit catchment was probably flat and underlain by Karoo and wad, to form a landscape similar to the flat plains that currently occur to the SE of Johannesburg.

4. Controls on cave formation and cave distribution in the CoH

4.1. General setting of caves in Malmani dolomite

About 750 cave systems (~80% of caves in South Africa) are known in the dolomite sequence of the Transvaal Supergroup (Martini, 2006), not counting inaccessible sinkholes, erosion remnants of caves and caves entirely filled with sediment. The highest

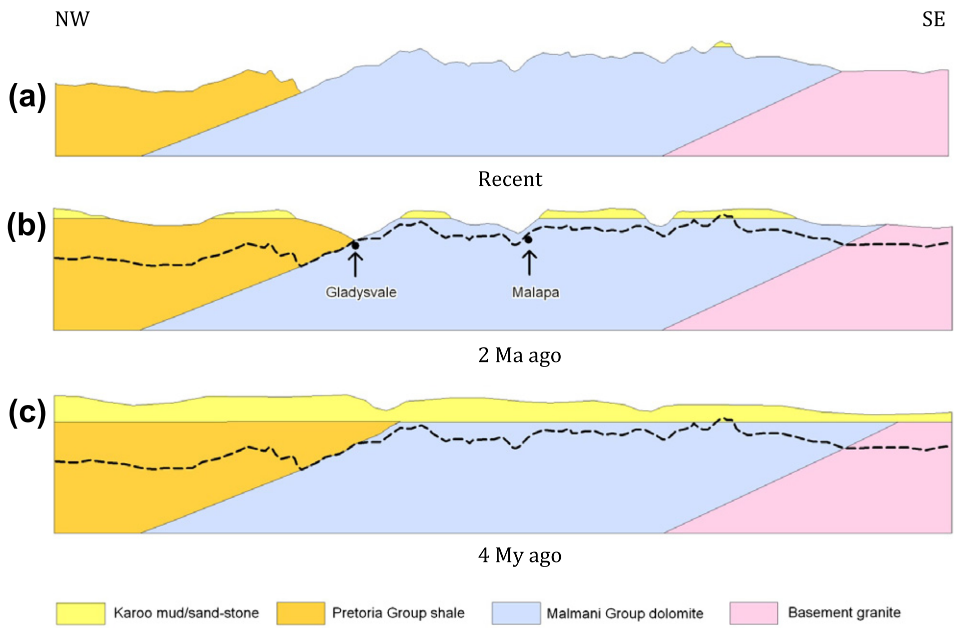

Fig. 10. Reconstruction in time of the landscape in the CoH across the Grootvleispruit catchment. (a) The current situation. (b) 2 My ago and (c) 4 My ago. The topography and geology are based on the geological section in Fig. 4 with a vertical exaggeration of 6:1. In drawing these sections it has been assumed that the Africa paleo-erosion surface existed at an altitude of ~1520–1550 mamsl; that this surface was gradually lowered; that the 1520–1550 mamsl level approximately coincided with the contact between the Malmani dolomite and a blanket of overlying Karoo sediment; and that the shale of the Timeball Hill formation was removed more rapidly than the dolomite resulting in dip slopes along hard grounds near the top of the dolomite. By ~2 My ago, Gladysvale cave would have been uncovered by erosion, whilst the landscape around Malapa was fundamentally different.

concentration of caves occurs in the Malmani dolomite between Pretoria and Potchefstroom; i.e. across the area in which the CoH occurs (Fig. 2). For this study, we compiled a database of 597 caves, sinkholes and cave erosion remnants in and around the CoH, many previously unmapped, through systematic mapping and compilation from geological reports (Enslin, 1973), records from caving clubs and earlier studies (e.g. Partridge, 2000). We know that this data base is not complete as many small cave systems and remains of caves remain unmapped, but for the purpose of this study our database is sufficient, and constitutes the most comprehensive yet analyzed. Of the caves mapped (Fig. 11a), 81 contain evidence of macrofossils and 15 are known to contain hominin or archaeological remains from a variety of ages between 0 and 3 Ma (Partridge, 2000; Tobias, 2000; unpublished data collected by us).

Caves generally developed in a phreatic environment and are characterized by complex mazes of passages along joint systems (e.g. Kavalieris and Martini, 1976; Martini, 2006). The cave channel systems in the CoH are poorly interconnected resulting in slow flow of groundwater. As a result sediment accumulated in the caves, which acted as sediment traps. Sediment accumulation occurred especially in caves exposed between Pretoria and Krugersdorp (i.e. the CoH), whereas little cave fill is reported from the Far West Rand where the largest caves are located (Martini, 2006) and where the African erosion surface is largely intact. This pattern indicates that the accessibility of caves to sediment fill is related to the evolution of the landscape itself; i.e. as the African erosion surface is dissected by rivers, the caves become accessible to animals and are filled with sediment.

Geological studies indicate that most hominin-bearing deposits formed in broadly similar settings involving debris cone accumulations beneath vertical cave openings (Partridge and Watt, 1991; Brain, 1993; Pickering et al., 2007; Pickering and Kramers, 2010). Gravel- to boulder-sized breakdown from ceilings and walls mixed with allochtonous and autochtonous sand- to clay-sized debris, accumulated on talus cones as debris flows and sheet wash. The clastic sediments were cemented by carbonate-rich waters dripping from cave ceilings. Periods of clastic sedimentation alternated with periods of erosion (Brain, 1993) or flowstone formation (Pickering et al., 2007), marking unconformities reflecting episodic cave fill. The fossil-rich Malapa site was previously unrecognized by scientists and differs geologically from other hominin sites in that flow stones are few, and clastic sediments were deposited horizontally in narrow cave chambers formed along fractures that accumulated a rich and highly diverse fauna (Dirks et al., 2010; Kuhn et al., 2011).

4.2. Lithological controls

In the CoH, many fossil-bearing caves occur along stratigraphic contacts within the Malmani Subgroup. For example, Bolts Farm, Swartkrans, Sterkfontein, Coopers, Kromdraai and Minnaars Caves in the valley of the Blaubank River (Fig. 1), are positioned along the contact of the Oak Tree and Monte Christo formations, whilst Malapa and Haasgat (Fig. 1) are located close to the contact between the Lyttelton and Eccles formations. Many of the stratigraphic contacts are sheared (e.g. the strongly foliated, mylonitic tuff horizon central to Sterkfontein Cave; Fig. 6e and f). Mylonitic textures in dolomite are poorly preserved due to post-kinematic annealing, but the deformational nature of the contacts can be deduced from intense folding and dismemberment (i.e. transposition) of chert horizons in which fold asymmetries verge consistently N (Fig. 6f); boudinaging of chert and/or shale marker horizons near contacts, and the preservation of well-preserved mylonitic textures in shale horizons near contact zones (Fig. 6e; Gibson et al., 1999; Alexandre et al., 2006). Caves are common in the immediate vicinity of such layer-parallel shear zones throughout the CoH, which, therefore, provide a strong regional stratiform control on cave formation.

Locally, cave distribution is controlled by joints and faults (e.g. Kavalieris and Martini, 1976) that truncate layer-parallel shear zones and lithological contacts. Of great importance in this respect is the unconformable lower contact of the Rooihoogte formation, and especially the fault systems and associated breccia–sandstone

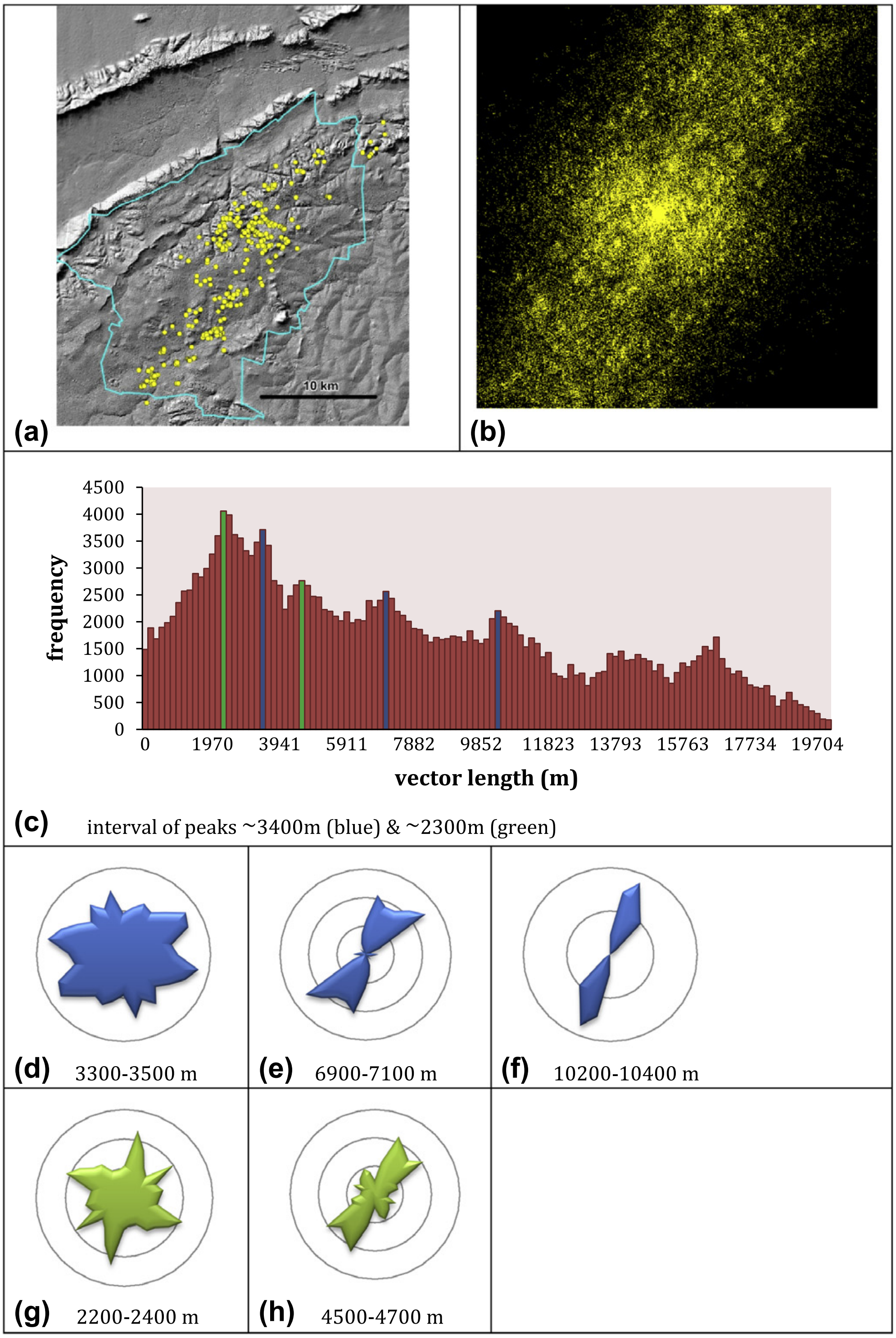

Fig. 11. Fry analysis for all cave sites mapped in the CoH (n = 597). (a) Distribution of all cave sites plotted on a digital terrain model for the CoH. (b) Fry plot for the full cave dataset. (c) Vector-length vs. frequency plot based on the fry plot in figure (b). (d–h) Rose diagrams for vector length domains with the highest frequencies to investigate preferential trends in the datasets. See text for detailed discussions.

filled fractures at the base of the Rooihoogte formation (Fig. 7). The relationship between caves and these fractures is clear in the Malapa area (Fig. 5b), where caves formed as elongated fissure chambers along chert-filled fractures. Caves are most numerous along these fractures near the (irregular) base of the Rooihoogte formation.

Throughout the valley of the Grootvleispruit, Rooihoogte-age fracture systems and caves exhibit a close spatial relationship. This relationship holds equally true elsewhere in the CoH, e.g. Gladysvale cave occurs immediately below the unconformable contact of the Rooihoogte formation, and Sterkfontein is positioned along a NE trending, vertical fracture with breccia that passes through the center of the cave system. Phreatic dissolution of these fractures created the complex mazes of passages that characterize the caves in the CoH (Martini et al., 2003). Thus, on outcrop scale, cave formation is controlled by the intersection of sheared stratigraphic horizons, with some major shears marking stratigraphic

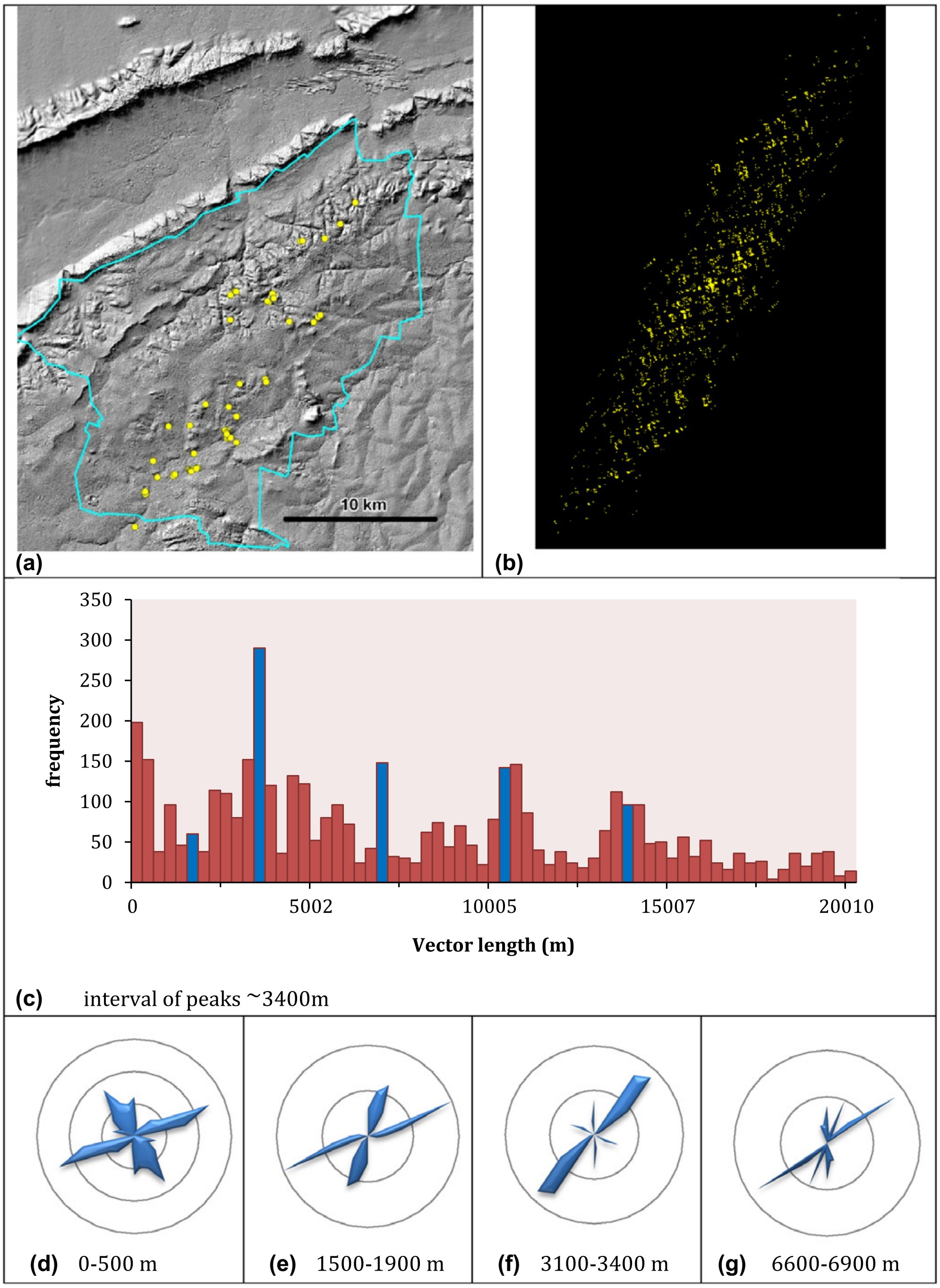

Fig. 12. Fry analysis for fossil-bearing cave sites mapped in the CoH ($n = 81$). (a) Distribution of cave sites plotted on a digital terrain model for the CoH. (b) Fry plot for the fossil-bearing cave dataset. (c) Vector-length vs. frequency plot based on the fry plot in figure (b). (d–g) Rose diagrams for vector length domains with the highest frequencies to investigate preferential trends in the datasets. See text for detailed discussions.

boundaries, and cross-cutting fracture patterns of Paleoproterozoic age, reactivated during the Neogene.

4.3. Cave distribution patterns

Near Malapa, caves formed along chert stockwork fractures (of Rooihoogte formation age) in NNW, NNE and WNW directions, to define a network of apparently interconnected cave openings 500 m long and 100 m wide (Fig. 5b). Post-Pliocene modification of the landscape can be seen in upslope (i.e. S-ward) changes in morphology of karst features in caves that topographically and stratigraphically occur above Malapa. The hominin site at Malapa is positioned at the intersection of a NNE and a NNW fracture. A similar prevalence of N to NNE and E to ESE trending fractures controlling cave systems has been reported for Sterkfontein (Martini et al., 2003) and caves throughout the area (Kavalieris and Martini, 1976; Martini, 2006), indicating the regional significance of this joint pattern.

4.3.1. Fry analysis

Regional trends in distribution of caves and cave deposits can be investigated using Fry analysis (Figs. 11 and 12), which offers a visual approach to semi-quantitatively appraise characteristic spatial trends in point data (Vearncombe and Vearncombe, 1999; Dirks et al., 2009). The caves are potentially aligned along fractures or fracture trends that were conducive to phreatic dissolution at a time when the far field stress allowed water ingress along fracture planes. Fry analysis can be used to search for anisotropies in the distribution of caves, to investigate whether linear trends occur within a given data set at a variety of scales (e.g. Vearncombe and Vearncombe, 1999). For Fry analysis we have made use of the programme DotProc v1.3 (Dirks et al., 2000), which interacts with MapInfo software, and which is freely available on request.

In Fry analysis using DotProc, a 2-D distribution pattern of data points (i.e. caves) is assessed by calculating connecting vectors between all caves within a fixed X–Y reference frame. These vectors have been re-projected from a central point to reproduce a distribution pattern of points or a "Fry plot" that visually enhances any regular patterns that exist in the original data set (Figs. 11b and 12b), e.g. a preferred orientation or spacing of characteristic trends (Vearncombe and Vearncombe, 1999; Dirks et al., 2000). The existence of scale-dependent fabrics or anisotropies in a Fry plot is further investigated with rose diagrams, in which a subset of vectors that fall within a particular length range are plotted and assessed for preferential trends (Figs. 11 and 12). By investigating the vector dataset in a systematic manner, length ranges for vectors displaying the strongest preferential trends in the data set can be selected. Thus, rose diagrams will allow a scaled, directional assessment of cave distributions in the Fry plot to search for dominant trends.

The existence of characteristic spacing between points (i.e. a nearest neighbor analysis) in a Fry plot can be analyzed by investigating vector frequency as a function of vector length. This allows assessment of clustering effects and enables a semi-quantitative determination of the characteristic spacing between clusters of points (i.e. it provides insight in whether groupings of caves are spaced at regular intervals or not). Vector maxima generally coincide with a range of vector lengths, which, when plotted on a rose diagram display strong fabrics indicating that the clusters occur in regularly spaced linear arrays (Dirks et al., 2000).

For analysis of regional trends of caves in the CoH, the cave database has been projected in a UTM WGS84 reference frame. Results of the analysis are shown in Fig. 11. An analysis of the frequency distribution of vector lengths for the full cave dataset shows a non-uniform distribution with clustering at vector lengths of 2.3 km and 3.4 km (Fig. 11c); i.e. a characteristic spacing between clusters of caves occurs at ~2.3 and ~3.4 km and multiples thereof, with each cluster representing a grouping of caves less than 2 km in diameter.

To investigate if the clustered distribution of caves occurs along particular trends, rose diagrams are constructed (Fig. 11d–h). Trends are only well developed at particular vector lengths, again reflecting the non-uniformity of the dataset. At 2.3 km scale the dominant trends are NNE (015°) and ESE (115°) with a minor NE trend. These orientations confirm the information obtained from mapping (e.g. Figs. 3 and 5b), and the known NNE and ESE fracture trends in cave systems (Kavalieris and Martini, 1976), with the NE trend tracing the strike of the lithology. No strong preferential trends in cave alignment occur for the 3.4 km scale clusters. At scales over 4 km, trends are dominated by NNE and NE directions (Fig. 11e, f, h). The NE trend parallels the strike of the lithology and confirms the importance of a regional strata-bound control. The NNE trend reflects the control on cave distribution of a large-scale underlying fracture system. The regional significance of NNE and ESE fracture systems can be observed from a DTM for the CoH (Fig. 11a), which shows that many of the weathering resistant ridges in the area are incised by streams flowing along these trends. Regional (>3 km) controls are likely the result from deeper-seated, larger scale fracture systems, and their incision probably reflects the state of the far field stress at the time caves formed.

Caves with macrofossils are investigated as a sub-set of the total cave dataset (Fig. 12). Fossil-bearing caves show a strongly clustered distribution pattern, with a well-developed characteristic spacing between individual clusters of 1700 m and 3400 m (and multiples thereof). Within individual clusters (at scales of <500 m), fossiliferous caves preserve preferential ENE and NNW trends. At larger scales (>1500 m) trends vary from N–NNE to NE, i.e. similar to the overall cave data set, reflecting lithological (NE) and N and NNE trending, regional fracture controls on cave distribution patterns.

The orientation analysis suggests that the distribution trends for fossiliferous caves differ from the distribution of all caves. It appears that caves with fossils are more strongly clustered, and that they developed along fracture trends that are differently orientated, suggesting that the fossil-bearing caves may have been different or special in some way, when compared to all caves; i.e. they may have developed at a particular time, or formed in a particular way.

4.3.2. Fractal analysis

The difference in distribution of all caves and fossil-bearing caves can be further investigated with fractal analysis of the distribution pattern (Fig. 13). A point distribution is fractal if it exhibits self-similarity; i.e. the distribution pattern is composed of copies of itself when scaled down, making the distribution scale invariant. The distribution of caves resulted from the favorable interaction of a number of controlling parameters including fracture patterns and the far field stress, which allowed fractures in certain orientations to be dilational resulting in groundwater ingress and preferential dissolution of dolomite (i.e. cave formation). There is no guarantee that such controls acted in the same way on all scales or originated as a result of the same tectonic process at the same time. However, if controls are related, e.g. caves formed along one generation of fractures, a fractal distribution of cave occurrences is expected within a certain scale range, if only because fracture distributions that are mimicked by the caves are fractal (Turcotte, 1986, 1997; Mandelbrot, 1983).

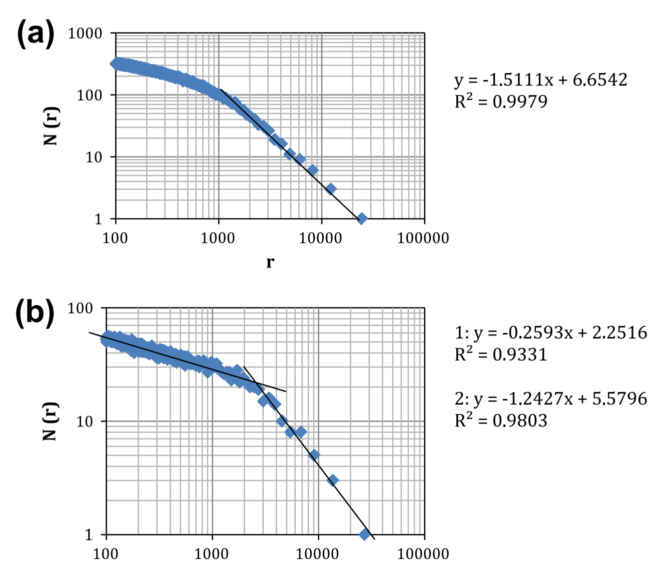

Fig. 13. Fractal plots for the spatial distribution of all caves (a) and fossil-bearing caves (b) in the CoH. Plots are generated using a block counting method (see text for details). (a) For the full cave dataset, a fractal distribution is apparent at scales of >2 km with a fractal dimension $D = -1.51$ and roll-off (i.e. under-representation of the data set) at smaller scales. (b) For fossil-bearing caves a well-developed, bimodal fractal distribution pattern is apparent. At scales of 0–2.3 km the fractal dimension $D = -0.26$ reflecting a highly clustered distribution at local scales. At scales of >2.3 km the fractal dimension $D = -1.24$. See text for a full discussion.

To test the fractal nature of a point distribution (caves are represented as points on the map; e.g. Fig. 11a), a box counting method is applied. This is done by placing a regular grid of squares on the point distribution that is analyzed, followed by a count of the number of squares containing part of the point distribution. The number of squares (N) required to cover the entire point distribution, will depend on the size of the squares, measured as side length r; N is therefore a function of r. In the box counting method, the size of the grid, i.e. r, is reduced repeatedly, and the variables, $N(r)$ and r are recorded. The relationship between $N(r)$ and r is fractal if a plot of $\log[N(r)]$ versus $\log[r]$ is linear. The slope of this linear relationship is called the fractal dimension (D). Thus, the fractal relationship can be expressed as follows:

$$N(r) = Cr^{-D}$$

with 'C' equaling a fractal constant.

Fig. 13a shows the fractal distribution pattern of all caves mapped in the CoH, next to the fractal distribution of fossil-bearing caves (Fig. 13b). All caves and sinkholes are distributed along a fractal slope of -1.51 at scales >2500 m, with roll-off at smaller scales (suggesting under sampling of caves at this scale range; Turcotte, 1997), indicating that cave distribution is controlled by a regional fracture system that exerts a systematic control on the distribution of caves at scales >2.5 km. The result also suggests that at scales >2.5 km, our mapping of caves has resulted in a representative sample of the overall distribution of caves in the area. However at local scales (<2.5 km) the distribution of caves may be underrepresented; i.e. not all caves within clusters of caves have been mapped.

The fossil-bearing caves display a well-developed bimodal fractal distribution pattern that is distinct from caves in general, with a well-developed fractal distribution at 0–2.5 km scale (fractal slope of -0.26), and a different fractal distribution at scales over 3 km, (fractal slope = -1.24). At local scales (<2.5 km), the low fractal dimension is indicative of a highly clustered distribution of fossil-bearing caves (Turcotte, 1997), whereas the regional distribution of cave clusters corresponds to a two dimensional distribution similar to what would be expected for caves along fracture patterns (e.g. Turcotte, 1986, 1997). Thus, clusters of fossil-bearing caves are distributed differently relative to one another when compared to the distribution of fossil-bearing caves within individual clusters, i.e. consistent with the observation made in Fry analysis (Fig. 12). This strongly suggests that local controls on the distribution of fossil-bearing caves (or fossils in caves) are distinct from regional controls. It is noted that highly clustered distribution patterns with $D < -1$ are uncommon in geological processes that involve fracturing (Turcotte, 1997).

5. Discussion

We will start the discussion by briefly summarizing the main points made earlier. The landscape in the CoH over the past 4 My has been dynamic in the sense that it has undergone major changes in its physical appearance, reflected in a general denudation of the geology, and incision of the upper reaches of the Crocodile River into the African erosion surface. Geomorphological features such as chert breccia dykes, which weather differentially from surrounding dolomite, can be calibrated with ^{10}Be exposure dates (Dirks et al., 2010) to estimate erosion rates for the landscape across the CoH (Fig. 8). In this way it can be shown that 2 My ago Malapa cave was probably ~50 m deep, when the landscape was less deeply incised, and remnants of regolith of the African erosion surface and Karoo cover rocks were more extensive. Landscape reconstructions provide estimates for the time of opening of cave systems to allow access to animals.

Caves are distributed across the changing landscape as a function of geological features that controlled their formation: firstly layer-parallel controls in dolomite such as lithological boundaries, layer-parallel shear zones (Gibson et al., 1999), and shale horizons (paleo-erosion surfaces of Paleoproterozoic age); secondly fracture systems that formed in the Paleoproterozoic during deposition of the Rooihoogte formation (plus some later fractures); and thirdly an extensional far field stress which allowed water ingress along selected fracture orientations at the time dolomite was dissolved to form cavities. The latter happened before the landscape was incised, and related to a water table that no longer reflects the current situation (e.g. Martini et al., 1977). It resulted in a systematic distribution of cave geometries across the landscape in the CoH with sinkholes and deep cave systems occurring on plateaus above 1500 mamsl, partly eroded cave systems like Sterkfontein and Wonder caves occurring at elevations between 1500 and 1420 mamsl, and shallow cave systems and erosion remnants of caves occurring at lower elevations.

Cave distribution patterns on a regional scale are controlled by NE trending dolomitic lithologies and NNE and ESE trending, basement fractures, reflecting the state of the far-field stress at the time of cave formation consistent with NW extension (Fig. 11; Bird et al., 2006; Viola et al., 2012). The distribution of all caves and sinkholes in the CoH is significantly different from the distribution of fossil-bearing caves, in terms of directional controls, degree of clustering and spacing, and fractal distribution (Figs. 11–13). Fossil-bearing caves are distributed in a highly clustered, fractal manner, with clusters spaced at regular intervals of 1700 m and 3400 m (Fig. 12c), reflecting a fundamental control on fossil distribution that is not random.

The observations from the CoH presented in this paper, indicate that the landscape in the Cradle was dynamic, that the change in landscape were systematic and proceeded in a predictable manner, which allows for reconstructions of the landscape back in time, and that the distribution of fossils in caves developed in close relationship with the evolving landscape.

As pointed out before, when considering the distribution of caves and fossils within the CoH and their relationship to the geology and evolving landscape, many pertinent questions remain. Based on the observations presented here we would like to discuss some of these questions in more detail:

1. How dynamic was the Pliocene–Pleistocene landscape; what were the tectonic drivers, and how did they influence the distribution of caves and fossils within them? Are there geomorphological and geological patterns that determine how the landscape was shaped over time; how caves formed, opened up to be filled with sediments and then became exposed?
2. When did caves become accessible for occupation and burial of fossil remains? Is it possible to predict where caves with fossils can be found, and how old fossils within caves are likely to be?
3. Why do fossils occur in certain caves and not in others; do all caves have the potential to contain hominin fossils and associated macrofauna, or are some more amenable then others, and if so why? Are the vectors of bone accumulation biological (predation, habitation) or geological (geomorphological traps, taphonomic reasons linked to accumulation rates)?

5.1. How dynamic was the Pliocene–Pleistocene landscape?

The observations presented in this paper illustrate that the physical characteristics of the landscape in the CoH changed significantly in the past 4 My. The change in the physical environment was largely driven by knick point retreat and valley widening (Partridge, 1973) as creeks in the headwaters of the Crocodile River were incising into the plateau landscape. In so doing caves opened up. These caves provided shelter and security and possibly access to water for animals that lived at the time (e.g. Dirks et al., 2010). They could also be dangerous traps, with unseen vertical shafts and complex passageways hiding pitfalls and predators (e.g. De Ruiter and Berger, 2000).

Bailey et al. (2011) and Reynolds et al. (2011) present the dynamic change in landscape in the CoH as if driven by large, active fault lines that repeatedly moved to rejuvenate the land surface, resulting in fault scarps, knick points and valley incision along faults. They developed this picture for the CoH based on satellite interpretations of geomorphological features at resolutions of 10m or more, with little or no ground truthing, with their interpretations relying on comparisons with the tectonic geomorphology of plate boundary zones such as the African Rift system and the Middle East (e.g. King et al., 1994). Whilst they recognize that the South African landscape is significantly different from East African rifts, Bailey et al. (2011) assume that active faulting did occur; e.g. along the valley of the Blaubank River below Sterkfontein.

However this picture appears to be incorrect. Although significant intra-cratonic earthquake activity occurs in areas around the CoH (e.g. Singh et al., 2009), there are no single, large fault lines that accommodated repeated fault motion with the same kinematic sense as encountered in plate boundary zones, such as the normal faults of the East African Rift with many kilometers of offset marginal to deep rift basins, or the strike-slip faults in the Middle East and Turkey that accommodated hundreds of kilometers of displacement (e.g. King et al., 1994; King and Bailey, 2006).

Fault scarps identified by Bailey et al. (2011) in the CoH did not result from faults with significant displacements of marker horizons in the Malmani dolomite and Black Reef formation; e.g. Bailey et al. (2011) equate a linear feature visible on DTM, SPOT and Landsat along the Blaubank River (Figs. 1 and 11) with a large active fault line, but field relationships indicate that the total displacement along this structure since the Archaean is a few 100's m at most. Similarly, Bailey et al. (2011) point at a nick point in the Blaubank River where it passes through Swartkops gorge, and suggest that the river profile reflects active faulting. Outcrop in the gorge is excellent and no evidence for a major fault exists. Instead the river incised along a mafic dyke that intruded weathering resistant quartzite of the Witwatersrand Super Group (Fig. 2), with the knick point in the river occurring where the river crosses the quartzite horizon. In other words, Bailey et al. (2011) and Reynolds et al. (2011) misinterpret the tectonic geomorphology of the the CoH by comparing erosion features too closely with fault scarp morphologies in young, active plate boundary terrains.

This is not to say that there are no active faults near the CoH, and that faulting does not play a role in the dynamic changes occurring in the landscape. There are many faults in the CoH and surrounding areas, but almost all of them are small, narrow structures, with relatively short strike lengths (<10 km; Enslin, 1973) and minor displacements (<50 m of total displacement). Many of the larger fractures were intruded by mafic dykes, creating distinct lithological breaks in the landscape. Collectively, the faults and dykes crisscross the area of the Johannesburg Dome (Fig. 2) and control the landscape dynamics through a subtle interplay between: 1. Multiple smaller fractures and joint sets accommodating minor displacements; 2. Differential erosion rates between different lithologies; and 3. Uplift of the land surface and concomitant plateau incision. This process is guided by a radial extensional far field stress, that may have changed over time, with the current-day, principal horizontal extension direction trending NE in the CoH area (Bird et al., 2006). This interplay of geomorphologic drivers has rendered a tectonic texture to the landscape (Fig. 11a) reflected in the way rivers incised along minor fractures and joints (with preferential incision along NNE and ESE trends), in the way springs and caves are distributed along similar orientations (Kavalieris and Martini, 1976), and the way nick points have developed above weathering resistant lithologies, mainly cherts (e.g. the chert breccia units if the Rooihoogte formation) and silicified quartzite. The principle tectonic control on river incision and landscape development is therefore not repeated movement along large faults (e.g. Bailey et al., 2011), but large-scale uplift of the landscape itself, and especially uplift since the Pliocene (e.g. Partridge, 2010).

It has been suggested that the land surface in the CoH rose by hundreds of meters (e.g. Partridge, 2010) since the mid-Pliocene resulting in incision of the African erosion surface and exposure of caves (e.g. Partridge, 1973). Apatite fission track data for South Africa suggest that denudation rates in the Cretaceous were an order of magnitude higher than the Cenozoic (Brown et al., 2002; Tinker, 2005; Doucouré and de Wit, 2003; Tinker et al., 2008a,b), whilst U-Th/He dating indicates that unroofing along the eastern escarpment has been less than 850 m in the Cenozoic (Flowers and Schoene, 2010); i.e. erosion of the plateau has been slow since the Cretaceous, a fact consistent with limited Cenozoic sediment accumulation along the South African passive margin (McMillan, 2003; Tinker et al., 2008a).

Cosmogenic isotope studies indicate that in the last few hundred thousand years denudation rates for outcrops on the stable, flat areas along the southern African escarpment and inland plateau are generally between 1 and 3 m/Ma. (Kounov et al., 2007 for ^{3}He and ^{21}Ne; Decker et al., 2011 for ^{3}He). Denudation rates obtained from basalt (1.5–4.0 m/Ma; Kounov et al., 2007; Decker et al., 2011) are generally higher than those obtained from weathering resistant quartzite (1.0–2.1 m/Ma; Kounov et al., 2007), although higher rates have been recorded (e.g. 3.9 ± 1.1 m/My for ^{10}Be, Dirks et al., 2010), indicating lithological control on erosion rates. Denudation rates along the dry western escarpment of southern Africa are significantly lower (0.4 m/Ma for ^{10}Be and ^{26}Al, Cockburn et al., 2000; 1.5–3.0 m/Ma; Kounov et al., 2007) than rates along the wet eastern escarpment (6.6 m/Ma for ^{36}Cl, Fleming et al., 1999) suggesting that climate also exerts a strong control on denudation processes with a further control being the morphological position of the sampling site (e.g. Portenga and Bierman, 2011).

Most available estimates for erosion rates have come from hard grounds and positive erosion features, which experienced lower denudation rates than the surrounding landscape, i.e. valleys erode faster than plateau areas (Portenga and Bierman, 2011); e.g. Dirks et al. (2010) reported an erosion rate of 53 m/My (^{10}Be) for valley incision in the CoH. The exact amount of uplift in the past 4 My is unclear and remains contested (e.g. Doucouré and de Wit, 2003; Partridge, 2010; Erlanger, 2011), but denudation of the flat plateau areas in central southern Africa has been limited. It is clear, however, that areas near the margins of this plateau, such as the CoH, underwent significant erosion and denudation, and that this landscape changed from the days *Au. sediba* roamed there (Dirks et al., 2010).

The fact that denudation since the Cretaceous has been limited on the South Africa high paleau (e.g. Tinker et al., 2008b), and that remnants of Cretaceous and Miocene regolith (and wad) formed on the African erosion surface remain in and around the CoH (e.g. Partridge and Maud, 1987; van Niekerk et al., 1999; Beukes et al., 1999), strongly suggest that much of the current high relief of southern Africa is indeed relatively recent (i.e. <4 Ma as suggested by Partridge, 2010). An important implication of significant uplift since the Pliocene is that the change in climate in the CoH is not just a function of global climate change patterns, but is also a function of uplift of the Southern African plateau creating the typical southern African 'high veld' with its cool winters, violent summer thunderstorms and grasslands.

Another important aspect of uplift accommodated by multiple smaller interacting faults is the position and orientation of the African and Post-African erosion surfaces. There are several lines of evidence that suggest that the relatively flat, landscape surfaces have been affected and rearranged as a result of block faulting, and do not represent continuous planar surfaces as suggested by e.g. Partridge and Maud (1987) and many subsequent papers based on that work. Around the Johannesburg Dome the base level of the Karoo sediments varies from 1495 mams (in the Pretoria area) to 1670 mamsl (in the Johannesburg area). It has been suggested that this reflects glacial paleo-topography (e.g. Johnson et al., 1997) or regional epeirogenetic warping (Partridge and Maud, 2000), but deformed and displaced Karoo along fault zones indicate that there is also an element of down-faulting to preserve inliers at different base heights similar to what can be observed near Malapa. Beukes et al. (1999) have described a paleo-regolith near the West Wits gold mine, ~10 km SE of the CoH, incised by fluvial channel deposits ascribed to a weathering horizon belonging to a Post-African I surface dated at 12–15 Ma (cryptomelane in pedogenic, Mn crust; van Niekerk et al., 1999). Active pedogenesis along this surface indicates that until the mid-Miocene the landscape was hot, flat and humid (and probably not all that high). These outcrops occur on top of the Witwatersrand plateau at ~1700 m, which according to Partridge and Maud (1987) was mountainous terrain standing above the African surface in post-Gondwana (Cretaceous) times. It suggests that the African surface has been structurally dismembered since the Miocene and possibly earlier, along numerous small faults, and it is not a continuous surface as generally assumed (e.g. Partridge, 2010); an important point when attempting landscape reconstructions.

5.2. When did the caves became accessible?

The distribution of sinkholes and deep caves on the plateau, and eroded cave systems in the valleys of the CoH, suggest that active cave formation and deepening of caves was probably not an active process as the landscape of the CoH was exhumed. In many places around the Johannesburg Dome and further west, caves developed down to a constant elevation, with networks of passages cutting through the gently dipping stratigraphy (e.g. Wondercave and Sterkfontein: Martini, 2006; Malapa; Dirks et al., 2010) indicating a strong control by a former water table. Martini et al. (1977) report that in the Far West Rand (west of Johannesburg) almost all caves occur at a level between the current water table and 40 m above it; i.e. caves formed when the water table was 40 m higher. They drained and opened after the African erosion surface had degraded and had been lowered by post-African incision (Martini, 2006).

Thus, caves had largely formed before rivers started incising the plateau landscape that had characterized the CoH in earlier times. Caves probably formed below one of the African Erosion surfaces (Fig. 9) when climates were wetter, and water tables more stationary and stable. By the time the caves were exposed in the CoH, they were largely passive holes in the ground that were not significantly enlarged through dissolution as they were exhumed. Many caves in the CoH display evidence of re-solution of cave sediment, but this occurred in a vadose rather than phreatic environment (e.g. Martini et al., 2003) as mildly acidic surface waters were washed down caves during periods when caves opened to surface as the climate changed (Brain, 1995; Pickering et al., 2007), but this process did not enlarge cave chambers at depth (e.g. Wilkinson, 1983).

The reasons why active cave formation stopped in the CoH before caves were exhumed can be linked to a combination of factors including a general drying out of the climate (e.g. Vrba et al., 1995; Bamford et al., 2010) and rapid uplift of the area since the mid-Pliocene (i.e. ~4 Ma; e.g. Partridge, 2010), which in combination with valley incision would have lowered water tables at a rate faster than cave formation through dissolution could proceed (Martini, 2006).

Thus, when caves entrances opened, caves largely acted as passive openings, some providing access to the water table (Martini et al., 2003; Dirks et al., 2010), all acting as sediment traps. Due to the poor interconnectivity of cave systems in the area (Martini, 2006), sediments washed into the caves were trapped and could not be removed, e.g. through flow of subterranean rivers. On the flat African erosion surface west of the CoH, caves do not commonly contain cave sediment (Martini, 2006).

Because caves are largely static features in the landscape, the age of cave fill is a proxy for the age of river incision and landscape evolution in the CoH. If the age of the oldest cave sediments is known, the age of river incision and first cave exposure in the CoH will be known as well, and vice versa, if erosion rates across the landscape are known, an estimate for the maximum age of sediments and fossils within them can be made; e.g. erosion remnants of chert along the Scheerpoort River and Grootvleispruit suggest that Gladysvale cave first opened ~2 Ma ago (Fig. 10) making it unlikely to find older fossils in that cave.

More, precise absolute estimates for erosion rates are needed in the CoH, but using erosion remnants like the chert-breccia dykes (Figs. 6d and 8), it is possible to predict with a reasonable degree of accuracy when caves became first exposed, and hence what the maximum age would be for fossils within caves. In general it would appear that first exposure of caves systems in the CoH would have occurred earlier towards the N, where the oldest cave sediments may be found, although they could have been removed from the landscape due to advanced erosion. Further S, the average age of cave sediments is likely to be younger. This general picture is complicated by the fact that caves underlying the dip slope of Rooihoogte chert (e.g. Gladysvale, Haasgat, Fig. 1), exposed when shale of the Timeball Hill formation was removed as a result of down-cutting of the Scheerpoort River, may be much younger, and this down-cutting is most pronounced towards the N of the CoH.

Bearing the various constraints in mind, reconstructions of the physical landscape, back in time can be made as shown in Figs. 9 and 10. Using estimates for uplift rates, the position of the African erosion surface, the lack of major fault-lines and the presence of Karoo erosion remnants, it appears that 2 My ago at the time *Au. sediba* entered Malapa cave, the landscape in the CoH would have been mostly flat undulating planes underlain by a still largely intact African erosion surface. Access to caves would have been limited to those broad valleys that had started to incise through the cover of Karoo and wad into underlying dolomite to expose deep cave systems, some providing access to water. The belt of dolomite exposed in the CoH would have been narrower by up to several kilometers as shale of the Timeball Hill formation was still covering large parts of the now exposed dip slopes of Rooihoogte formation. At 4 My ago, the landscape would have been largely flat and covered by wad and Karoo, there may have been some sinkholes providing access to caves below, but they would have been few and access to dolomite would have been limited. Copeland et al. (2011) argue that male and female australopiths 2–3 My ago, occupied aerially distinct parts of the landscape underlain by different geology, with some (presumed) males occupying relatively small territories on dolomite and (presumed) females growing up in areas remote to dolomite. In their analysis the distribution of geological units was taken as near-static, which is clearly not correct as a geologically much more complex picture existed in the area 2–4 My ago.

To get a sense of what the physical landscape would have looked like in the CoH in the past 4 Ma, time travel is possible. When going due east from Johannesburg, after about 50 km, an undulating landscape occurs N of the town of Springs, largely underlain by a thin blanket of Karoo, incised by broad valleys locally exposing the top of the Malmani dolomite (Fig. 2); this could be a good proxy for the CoH 2 My ago. Continue in a SE direction and the landscape soon becomes completely flat and underlain by horizontally layered Karoo rocks, possibly reflecting what the CoH may have looked like 4 Ma ago.

5.3. Why do fossils occur in certain caves and not in others?

The distribution of fossil-bearing caves in the CoH is different and distinct from the distribution of all caves in the area (Figs. 11–13), for reasons that are not immediately obvious. Presumably caves with fossils in them formed in a particular way or at a particular time to make them somehow conducive to the accumulation of faunal remains. They must have been accessible or attractive to animals in ways other caves were not. Fossil-bearing caves contain fossils from a wide variety of ages ranging from recent to ~3 My old (Partridge, 2000; De Ruiter and Berger, 2000; Tobias, 2000; Berger et al., 2010; Pickering et al., 2011b). Thus, the fossil-bearing cave systems collected these fossils in a preferential manner for a long period of time. This happened irrespective of whether the remains were those of hominins or other species of fauna occupying the South African, Plio–Pleistocene landscape. It also happened along particular trends (Fig. 12e–g), and in a characteristically clustered manner (Figs. 12c and 13b).

The fact that complex fossil assemblages with a variety of ages occur in fossil-bearing cave sediments (e.g. Vrba, 1995; Dirks et al., 2010), would suggest a long-lasting geological control on the taphonomy of these cave deposits, which acted continuously for the past 3 My to collect fossils. If this control involved a particular pattern of fractures, it would not readily explain the highly clustered nature of the fossil-bearing caves, nor does it explain the regular spacing of the clusters (Fig. 12.b), because fracture pattern and hence structures like caves that mimic these patterns, typically display fractal distributions with fractal dimensions (-D) of >1.3 (Turcotte, 1986, 1997). Field mapping indicates that fossil-bearing caves are no different from non-fossil bearing caves in the way they occur within the landscape or in terms of geological controls on their formation. Fossil-bearing caves occur along the edge of the plateaus, and near valley floors (e.g. Malapa), they do not appear to occupy a particular part of the landscape, although they occur in accentuated terrain in which the topography displays a higher degree of roughness (Bailey et al., 2011); i.e. in areas where the African plateau has been dissected by rivers to allow access to cave systems. This relationship is easily explained in that river incision makes caves accessible to sedimentation, occupation and burial alike; but whilst many caves in the CoH contain sediment fill only few collected fossils, and they did so in an apparently systematic way.

If geological processes cannot readily explain the presence of fossils in regularly spaced clusters of caves, it is likely that a biological process was at play. Highly clustered, fractal distribution patterns have been described for optimal foraging behavior in animals, resulting in patterns referred to as Lévy flight (Mandelbrot, 1983). Lévy flight is a kind of random walk in which the measured step lengths of foraging animals follows a Lévy probability density function characterized by a long power-law tail (Mandelbrot, 1983). A number of animals including non-human primates (spider monkeys), are reported to forage in Lévy flight patterns (Ramos-Fernandez et al., 2003; Brown et al., 2005), which includes highly clustered localized motion alternating with long-range steps in seemingly random orientations. These foragers may be moving around in Lévy flight patterns because the resources they are seeking (food, water, prey), or the landscape in which they are moving (shelter) are distributed in a fractal manner. Fractal distributions for Lévy flight patterns of foraging animals can be bi-fractal with similar fractal dimensions as the distribution observed for fossil-bearing caves.

Could it be that scavengers like hyenas, or hunters like large cats collected carcasses in groups of caves (Brain, 1981; Pickering et al., 2004), positioned towards the centers of their territories, with the characteristic cluster spacing reflecting a territorial control? Or could it be hominins themselves that occupied certain caves and not others, with clustering reflecting territorial behavior in early hominins (Copeland et al., 2011). If so, this type of territorial behavior would have persisted for 3 million years, as different species evolved and others became extinct, but always using the same groups of caves; a fact hard to imagine.

One control that may be realistic to consider is that certain cave systems were closely associated with springs that made such caves water-bearing for long periods of time as the landscape changed. Sterkfontein cave, which is easily accessible and has a large lake at its base, may be an example of what such caves looked like. Malapa cave too, shows clear evidence that the lower chambers were water-logged at the time *Au. sediba* was buried, and it has been suggested that access to water was the reason why a highly diverse fauna, of seemingly incompatible bedfellows like hominins and false sabre-toothed cats frequented the site (Dirks et al., 2010; Kuhn et al., 2011). The water-bearing qualities of certain caves could have resulted from the fact that they were orientated favorably with respect the far field stress in the past 3 My to allow water ingress in these caves; a fact that would explain the characteristic orientation of fossil-bearing caves at local scales (Fig. 12d). Alternatively these caves simply occurred close to permanent spring sites, near the margins of hydrological compartments in the CoH. The presence of a nearby water source would explain the highly clustered nature of fossil bearing caves, especially if the fractal distribution of this water source is combined with Lévy flight behavior of animals occupying the caves near the spring. Permanent spring sites also dramatically affect floral communities and thus the micro-environment. The characteristic spacing of clusters could represent a landscape dimension controlling groundwater flow. Whilst it is interesting and useful to speculate in this manner, further work is clearly required.

6. Conclusion

Our understanding of hominin evolution is critically dependent on understanding the sites where fossils are found in relation to the landscape in which the fossil sites occur. In this context no question is more important than whether fossil sites merely represent convenient trapping sites with superior taphonomic characteristics, or whether the fossil sites are a reflection of habitation and land-use patterns by animals that occasionally got trapped within them.

With a detailed description of the evolving landscape in the CoH, focusing on the catchment of the Grootvleispruit, we have attempted to show that this landscape has changed considerably in the past 4 Ma. These changes were largely driven by erosional processes that resulted from broad uplift of the landscape. Erosion patterns did not involve active faulting along major fault lines, but instead resulted from the subtle rearrangement of the landscape through the interactions of multiple fractures in an evolving, extensional, far-field stress. Erosion exposed caves in the CoH, some of which attracted large numbers of different animals for long periods of time. The preferential attraction of certain caves over others, as displayed in the analysis of cave distribution patterns, probably reflects the presence of a stable water source either inside a cave or nearby. Therefore, it appears that the landscape of the CoH, with its caves and stable water sources, and incised valleys with variable vegetation patterns at localized scales, did provide a preferred environment that attracted not only hominins, but also many other creatures, i.e. it is not just taphonomic coincidence that the fossils ended up in the caves.

Acknowledgements

This research has been the result of 4 years of mapping in the Cradle of Humankind. The project started in early 2008 when we decided to map the distribution of caves in the CoH in an attempt to better understand the geological controls on cave formation and their distribution in the landscape. Early discussions benefited immensely from interactions with Geoffrey King. In August 2008 our mapping resulted in the discovery of the Malapa site and the fossils of *Australopithicus sediba*, and the subsequent involvement of a large number of students and collaborators, many of whom have contributed to this research.

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