CHAPTER ONE
INTRODUCTION AND LITERATURE REVIEW

1.1 General Introduction

In South Africa many of the dissolution caves in the Late Archaean-Early Proterozoic dolomites of the Transvaal Supergroup contain Plio-Pleistocene sedimentary fills. These fills have yielded fossil hominin and other fauna and flora, which have been the focus of much research (Partridge, 2000). However, with the exception of Brain's (1995) work at Swartkrans, the geological aspects of the caves, in particular the sedimentology and stratigraphy of the cave fills, are poorly understood. The previously employed lithostratigrapic approach to cave fills (Partridge, 1978; 1979; 1982; 2000) has proved to be untestable, mainly due to the complex stratigraphy of the breccias, the high degrees of cementation of the breccias and the limited exposures of the deposits (Partridge, 2000).

Gladysvale Cave is situated on a northeast slope of the Gladysvale Valley, in the John Nash Nature Reserve, within the Cradle of Humankind World Heritage Site. The site is located some 13km north-east of the other better known hominin bearing sites of Sterkfontein, Swartkrans and Kromdraai (Figure 1.1).

Gladysvale Cave, apart from being an important hominin locality, is unique among the caves in the Cradle of Humankind World Heritage Site (COH WHS), in three major aspects. First, the cave contains a well-stratified fill, which consists of clastic and chemical sediments (Lacruz, 2002; Lacruz *et al*., 2003; Pickering, 2002). Second, the younger internal deposits at Gladysvale are relatively complete, and the deposit has been trenched and sectioned by lime miners in the early part of last century, providing ideal three-dimensional exposures. Third, the deposits are preliminarily dated to between 250 and 200 kyr by Electron Spin Resonance (ESR) on bovid tooth enamel (Curnoe, 1999; Schmid, 2002), placing them well within the range of Uranium-series, a technique that has been shown to provide a reliable and accurate chronology for the last 500 kyr (Edwards *et al.*, 1984), and one which has been successfully employed in similar cave environments (Ayliffe *et al.*, 1998; Moriarty *et al.*, 2000). These three factors, the unique stratigraphy, the relative completeness of the deposit, and the exceptional exposure and the potential for absolute dating of the deposit, make Gladysvale an ideal location to document the stratigraphic fill and sedimentology and to test previously published models of cave sedimentation (Ayliffe *et al.*, 1998; Partridge, 2000; Moriarty *et al.*, 2000).

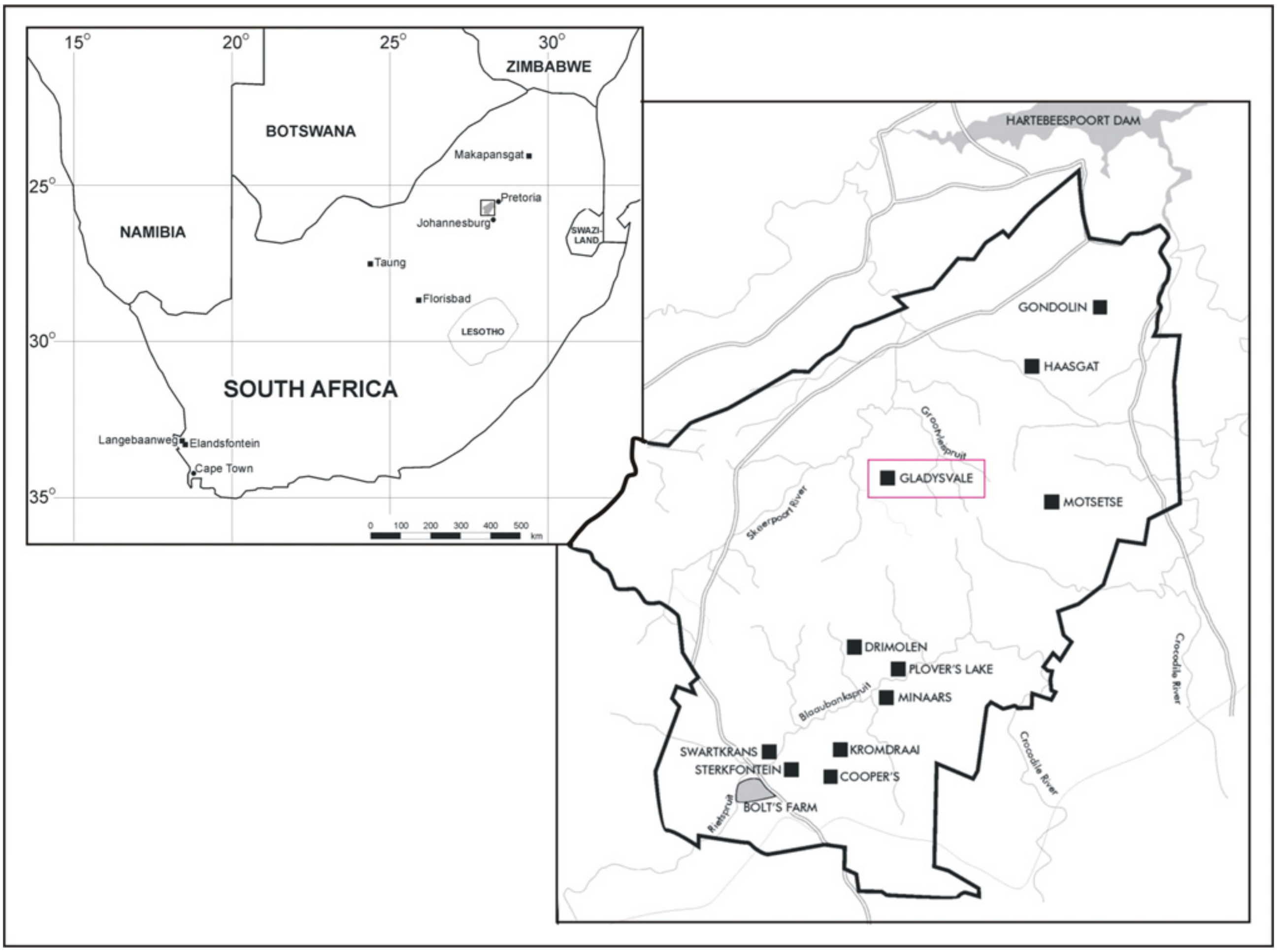

Figure 1.1. Locality map for Gladysvale and other hominin bearing caves sites in South Africa (after Berger & Lacruz, 2003).

As these cave fills are the context in which the hominin and other fossils are found, they provide valuable information about accumulation agents and rates, as well as palaeoclimatic and palaeoenvironmental data (Brain, 1958). Cave sediments are also potentially datable, which together with the use of chronostratigraphic principles could allow for dating of included hominin fossils, as well as aiding in chronological correlation between cave sites. A better understanding of the nature and stratigraphy of these cave fills would therefore greatly enhance our interpretation of the fossil record.

1.1.1 Geological setting

The cave system at Gladysvale was formed in host rocks consisting of the Late Archaean-Early Proterozoic cherty dolomites of the Eccles Formation of the Chuniespoort Group (Malmani Subgroup) of the Transvaal Supergroup (Berger & Tobias, 1994) (Figure 1.2). Transvaal Supergroup sedimentation began with thin, predominantly clastic rocks (the Black Reef Formation), which grade into a thick package of carbonate rocks and banded iron formations (BIF) of the Chuniespoort Group. These lithologies reflect a carbonate-BIF platform sequence, which covered much of the Kaapvaal Craton (Eriksson *et al.*, 1995). Within the Chuniespoort Group, there are six (Eriksson *et al.*, 1995) to eight (Eriksson & Truswell, 1974) carbonate layers, which are grouped together as the Malmani Subgroup dolomites. The Eccles Formation occurs towards the top of the Malmani Subgroup, and is situated stratigraphically between the Lyttelton and Frisco formations (Eriksson *et al.*, 1995). The Eccles Formation at Gladysvale consists of chert-rich dolomitic limestone and reaches a thickness of some 300m, dipping at ~40° to the northwest (Martini & Keyser, 1989).

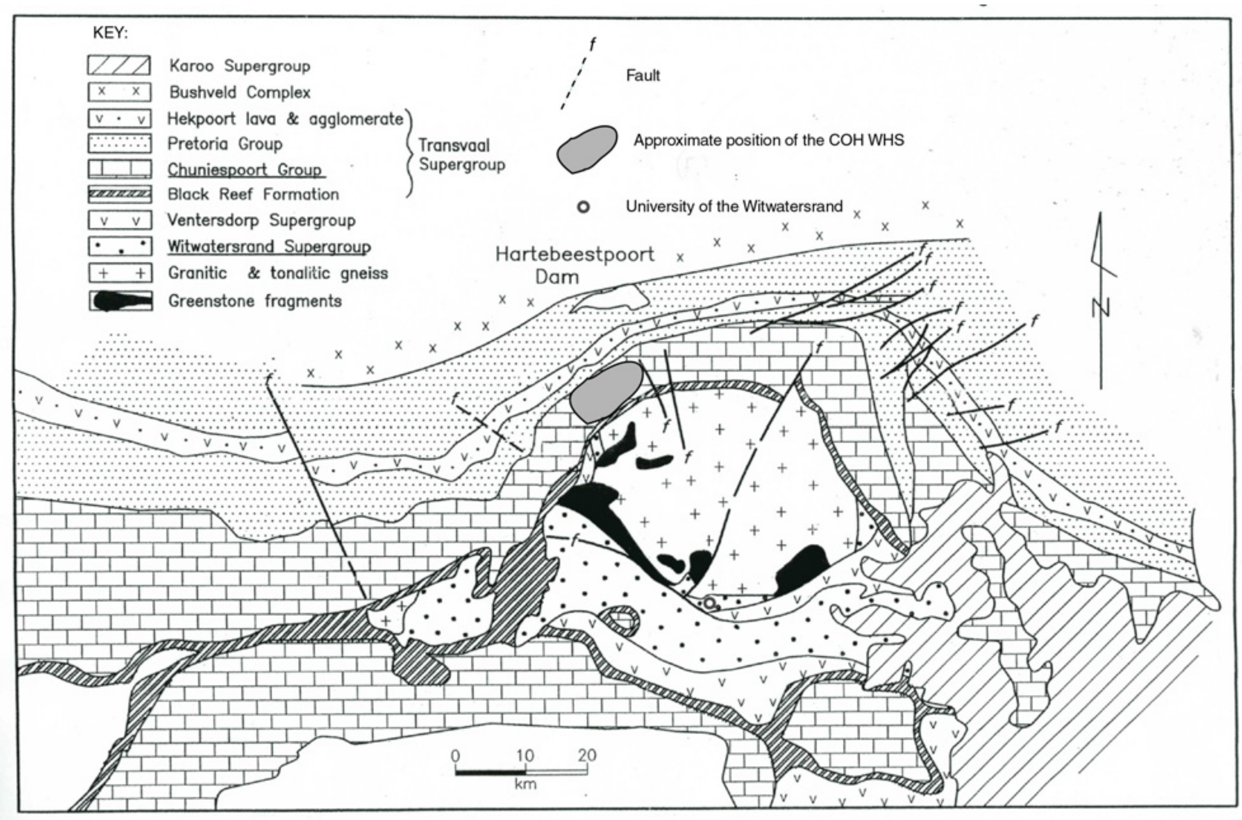

Figure 1.2. Regional geology the Transvaal and Witwatersrand supergroups in the Johannesburg area, showing the approximate position of the Cradle of Humankind World Heritage Site (COH WHS) (After McCarthy *et al.*, 1986).

Gladysvale Cave is situated on the north-east slope of the Gladysvale valley, which is part of a landscape heavily dissected by numerous rivers draining down to the Skeerpoort River (Martini & Keyser, 1989). There are three or more caves in the Gladysvale Cave system, which are situated very close to each other and even overlap. Speleologicallly they consist of a single chamber, which has been separated by rubble chokes (Martini & Keyser, 1989) (Figure 1.3). There was substantial stalagmite, stalactite and flowstone development in both caves, but these have been largely destroyed by lime miners (Martini & Keyser, 1989). The cave morphology is, to an extent explained by the surface topography, as in a steep sided valley, such as the Gladysvale valley, a cave forming inside the slope may open laterally into the side of the valley, by means of breaking through one of the side walls (Brain, 1958) (Figure 1.4).

Based on his observations on a wide variety of South Africa dolomite caves, Brain (1958) divided the processes by which South African caves formed in the Transvaal dolomites into five stages (Figure 1.4). During stage 1 an irregular shaped chamber is dissolved out of the dolomite along a plane of structural weakness by phreatic groundwater, directly below the water table. In Stage 2 the ground water level in the chamber is lowered as a result of surface erosion, deepening of river valleys, or climatic change, and the exposed portion of the chamber is brought into the vadose zone. During Stage 3 water drips down along joint planes into the chamber and causes dissolution of the dolomite, enlarging its seepage plane, and forming an aven. Dissolution is driven by the presence of dissolved carbon dioxide in the water, which allows water to take large quantities of carbonate into solution. Travertines are deposited in abundance, and depending on the drip regime, stalagmites, stalactites and flowstones will grow. Eventually, during stage 4, as the avens continue to become enlarged, the cave opens to the surface. Once open, the cave will start to fill up with surrounding surface derived soil and debris. Stage 5 consists of lime rich water percolating through the cave roof, which slowly calcifies the cave sediments, forming breccias. Surface erosion and acidic rainwater will then progressively destroy the cave and its contents.

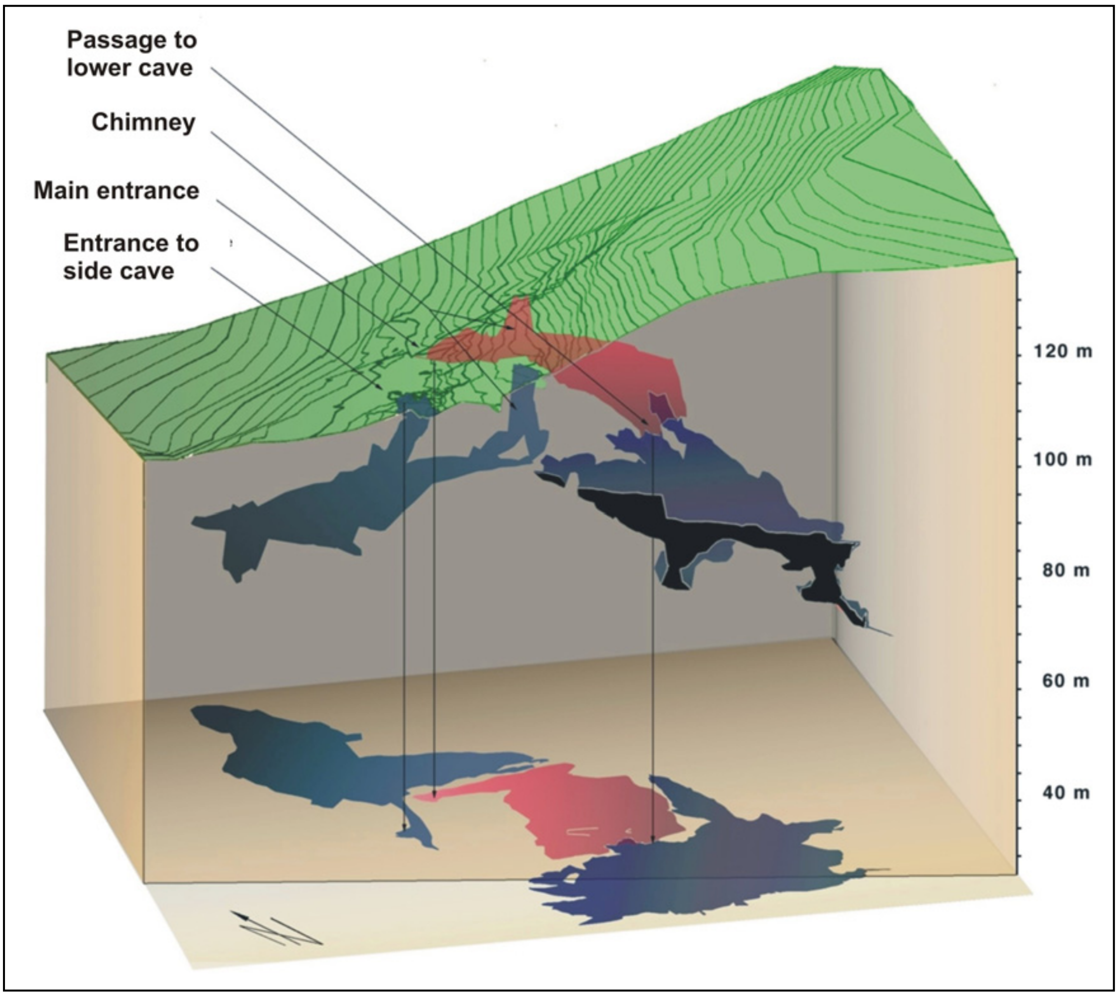

Figure 1.3. Three dimensional GIS generated image of the cave system at Gladysvale, upper chamber shown in red (Courtesy of Schmid, Martin, Berger and others).

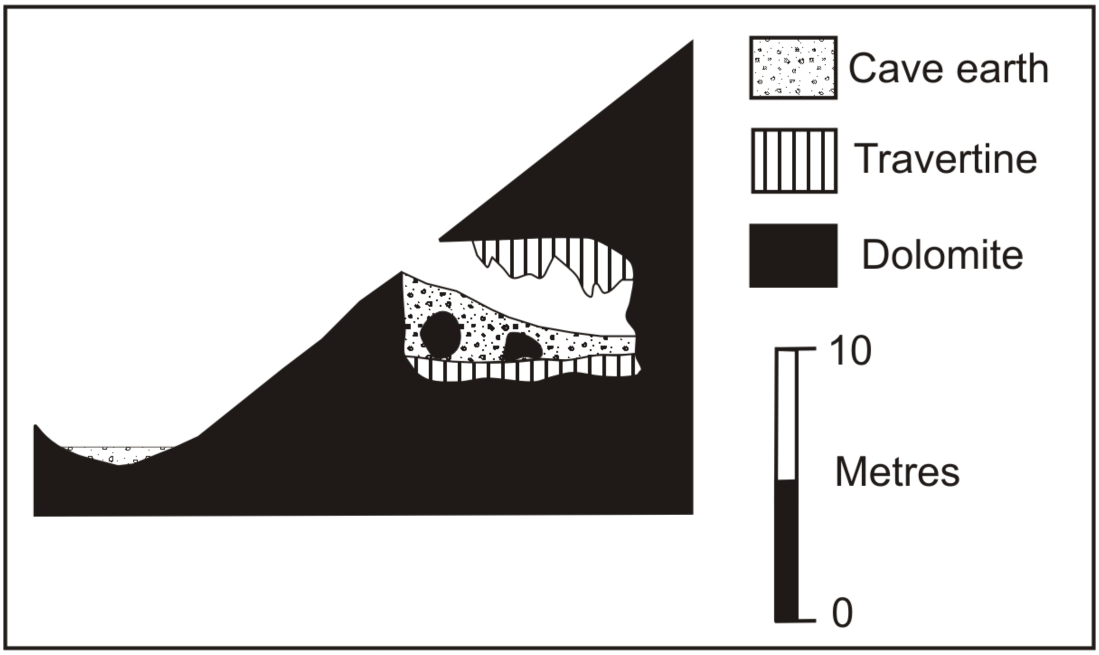

Figure 1.4. Model for caves with valley side openings, such as seen at Gladysvale (after Brain, 1958).

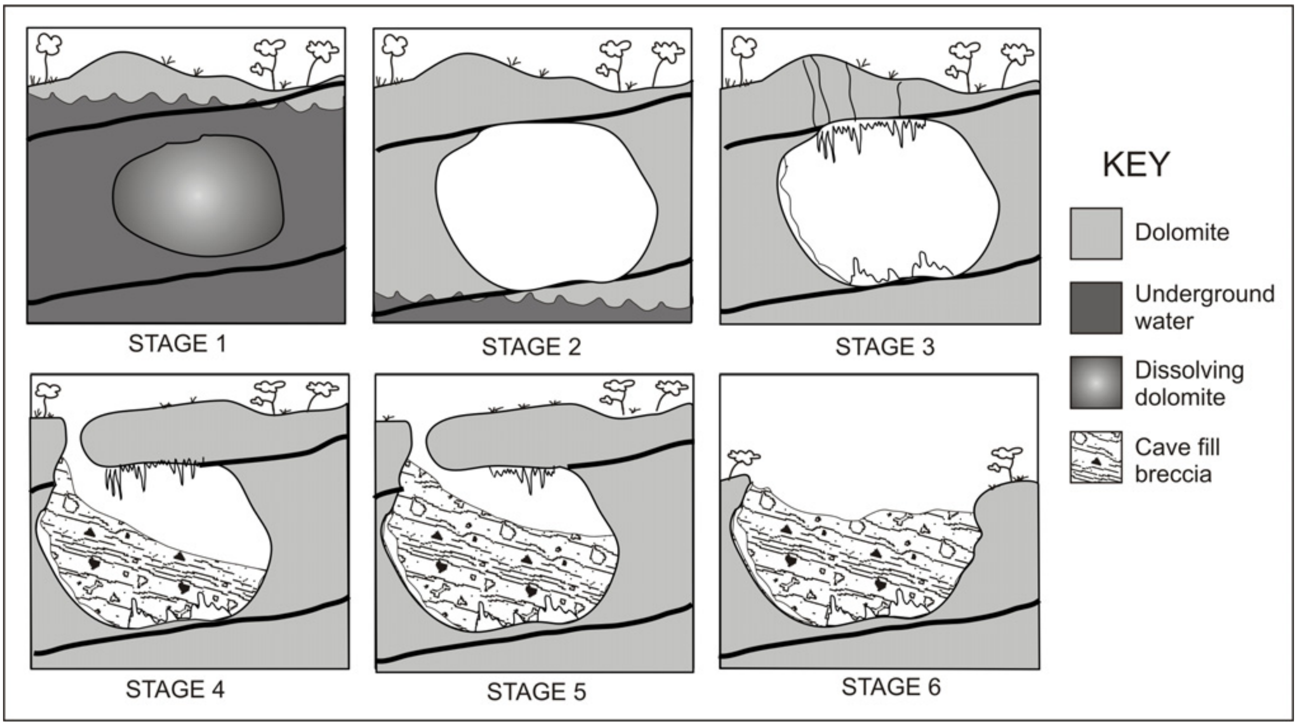

Figure 1.5. Stages in South African dolomitic cave formation (after Brain, 1958).

1.1.2 Previous research at Gladysvale Cave

Gladysvale Cave was mined intermittently from 1902 to 1928 for the speleothem deposits, which were sold as lime. Lime was used as a flux in the gold mining industry, and as an ingredient in fertilizer and toothpaste (Berger & Tobias, 1994). Based on the supposed presence of a hominin skull, Robert Broom made a few short visits to Gladysvale in 1936, collecting a few specimens, but not finding the alleged hominin fossils he was looking for. In 1948 Charles Camp and Frank E. Peabody from the University of California, Berkeley, visited the cave and excavated extensively for over a month (Berger & Tobias, 1994). Peabody was the first to make note of the stratified nature of the small concave chamber at the back of the upper chamber (Peabody, unpublished field notes, 1948), which now bears his name. Recent excavations at Gladysvale began in 1988, and between 1991 and 1992 fragmentary hominin material, consisting of a maxillary third premolar and maxillary second molar (Berger, 1992; Berger *et al.*, 1993) was recovered. At the time this made Gladysvale the first new hominin site found in South Africa in forty-eight years. *In situ* hominin material was recovered in 1997 (Schmid & Berger, 1997). From the hominin material and associated fossil material the younger deposits were given an early Middle Pleistocene age (Berger & Tobias, 1994). A stratigraphy for the exposed calcified strata was proposed by Berger & Tobias (1994), consisting of a lower Stony Breccia and an upper Pink Breccia (see Figure 1.6 A).

Between 1994 and 2002 much of the external cover of the rubble slope was removed, exposing a large de-roofed cave fill, which has been termed the Gladysvale External Deposit (GVED). Lacruz (2002) and Lacruz *et al*. (2003) reviewed the stratigraphy of the GVED using the model of Moriarty *et al.* (2000), in which clastic material is divided into units by flowstone layers, forming flowstone bounded units (FBU) (Figure 1.6 B). In general the GVED deposits are coarse grained, show little microstratigraphy relative to the Peabody Chamber deposits, contain large (up to 15cm), angular clasts, and are believed to represent more proximal sediments within the cave system at Gladysvale. The ESR and palaeomagnetic dates show that the GVED is between 650 ± 63 ka and 779 ± 51 ka (Lacruz, 2002; Lacruz *et al.*, 2003).

Pickering (2002) focused on the stratigraphy of the Peabody Chamber, a small concave chamber in the upper level of the underground cave. The model of Moriarty *et al.* (2000) was used to divide the well-stratified sequences into units bounded by flowstone layers, which in turn were classed into three flowstone bounded intervals – FBI 1, 2 and 3 (Pickering, 2002) (Figure 1.6). According to Moriarty *et al.* (2000), the FBI of the Peabody Chamber should are separated by major temporal unconformities. The ESR dates (Curnoe, 1999; Schmid, 2002) for the Peabody Chamber confirm this, as FBI 1 is between 1300 and1000 ka, FBI 2 is ~800 ka, and FBI 3 is ~250 ka. The cave sediments in the Peabody Chamber consist of the distal members of fan type sedimentation, which are correlatable along stratigraphic sections, and a major flowstone marker horizon, with the GVED (Pickering, 2002). Some 40 meters of cave fill, both inside Gladysvale Cave and in the exposed external deposits, is therefore part of the same fan system, with proximal and distal members represented.

Other recent research at Gladysvale includes the ESR dating of some bovid teeth from the internal and external deposits (Curnoe, 1998; Schmid, 2002; Lacruz *et al.*, 2003). During excavations in 2002, a mid to late Acheulian hand-axe was discovered in the external deposits, showing that these external deposits actually consist of the GVED (as described by Lacruz *et al.,* 2003) and an older cone deposit. The hand-axe is significant as it is not only the first stone tool to be found at Gladysvale, but one of the few *in situ*, reliably located artefacts of this age in the Cradle of Humankind World Heritage Site (Hall *et al*., in press). Research on the geological aspects of Gladysvale Cave and the Peabody Chamber in particular have been presented as papers at a number of conferences (Hancox *et al*, 2002a; Pickering *et al*., 2002; Pickering *et al*., 2003; Pickering *et al*., 2004a, 2004b; Appendix 1).

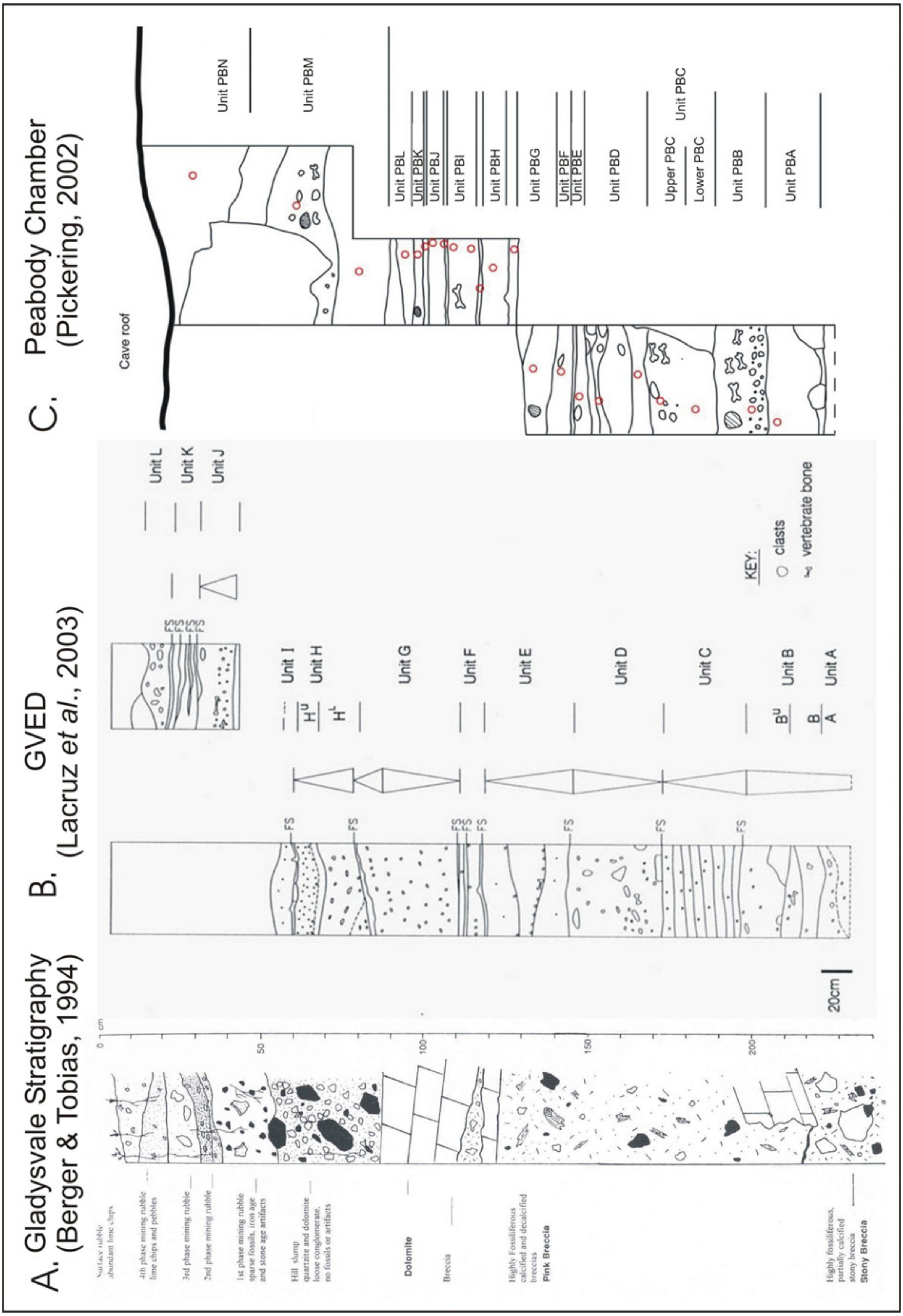

Figure 1.6. Proposed stratigraphies for deposits at Gladysvale Cave: A: Stratigraphy for the external deposits (Berger & Tobias, 1994); B: Revised stratigraphy for the GVED (Lacruz *et al*., 2003; Lacruz, 2002); C: Stratigraphy of the Peabody Chamber (red circles indicate sample locations) (Pickering, 2002).

1.2 Literature Review

1.2.1 Cave stratigraphy

In South Africa, Plio-Pleistocene hominin bearing deposits are found in the caves of the Cradle of Humankind World Heritage Site and at the Makapansgat complex of sites (Figure 1.1).

The Cradle of Humankind World Heritage Site is a 47 000 hectare area some 40km to the northwest of Johannesburg, which encompasses 13 cave sites, including the world famous hominin bearing caves of Sterkfontein, Swartkrans and Kromdraai (Figure 1.1) (Hilton-Barber & Berger, 2002). Several studies have been undertaken on the stratigraphy (Partridge, 2000; Clarke, 1994), sedimentology (Partridge, 1982) and lithostratigraphy (Partridge, 1978; Partridge, 1979; Butzer, 1976) of the sites within the Cradle, and as a result the cave deposits of Sterkfontein, Swartkrans, and Kromdraai have been divided into various members and beds on the basis of colour, grain size and $CaCO_3$ content (Partridge, 1978, 1979, 1982). Correlation of these members between caves has, however, been impossible and the chronostratigraphy and link between sedimentation and environmental change has remained unexplored.

The breccias from Sterkfontein Cave system have yielded over 600 hominin fossils, which have highlighted the importance of South Africa in palaeonthropology. Because of the great value and magnitude of the fossils at Sterkfontein, the stratigraphy and sedimentological context of these remains is of great significance, yet it is highly controversial. Robinson (1962) provided the first description and reconstruction of the deposits at Sterkfontein, in which he proposed that the initial infill of material collapsed into the deeper regions of the cave. This was then followed by another infill from a different opening in a less well calcified region of the cave, which in turn was followed by another infill of younger material (Robinson, 1962). Partridge (1978) reviewed this stratigraphy and proposed a six member subdivision for the Sterkfontein deposit, which, with minor reworkings, is sill in use today (e.g. Clarke, 1994).

The dating of the deposits at Sterkfontein has also been extremely contentious. The deposits are believed by some to be beyond the reaches of radiometric methods, such as Uranium-series, and most dating has been done via faunal correlation with better dated sites in

East Africa (Partridge, 2000; Partridge *et al*., 1999a), which provide ages of between 3.0 and 3.5 Ma. Various attempts have been made to date the deposits with palaeomagnetic techniques (Partridge, 2000), but these are plagued with problems (Berger *et al*., 2002), and the comparison of the Sterkfontein magnetostratigraphy with the global polarity time scale by Partridge *et al*. (1999a) is questioned by Berger *et al*. (2002). Partridge *et al.* (1999a) and Partridge (2000) use the patterns of normal and reversed polarity to place the age of the Sterkfontein deposits at 3.22 – 3.58 Ma. Berger *et al*. (2002) argue that the fauna of this deposit suggest a younger age of approximately 2 Ma, and that the magnetostratigraphy can accommodate this younger age. Recent research utilizing experimental techniques, such as cosmogenic nuclides (Partridge *et al*., 2003), suggests an age of approximately 4 Ma, which also can be supported by the magnetostratigraphy. However, once again *a priori* considerations and problems prevail. Current research into TIMS (thermo-ionisation mass spectrometry) U-Pb dating of the speleothems from Sterkfontein is yielding reliable and repeatable ages, which place the deposit containing the “Little Foot” skeleton at ~2.7 Ma (Walker *et al.*, 2004; Walker, *pers. comm.*).

Swartkrans is an equally famous site, and is situated across the valley from Sterkfontein (Figure 1.1). Much of the research undertaken at Swartkrans has been done by Brain (1958, 1981, 1993, 1994, 1995) and includes the development of the discipline of South African cave taphonomy (Brain, 1981) and the first documentation of the controlled use of fire (Brain & Sillen, 1988). The cave stratigraphy and sedimentology of the deposits at Swartkrans have been well documented by Brain (1995), and are discussed later.

Kromdraai is situated about 3km from Sterkfontein, and is one of the earliest known sites, with hominins first being described by Broom (1938). Partridge (1982) hints at a link between depositional hiatuses or unconformities separating members at Kromdraai, arguing that the cyclic development of a vegetational mat outside the cave, which he related to changes in climate, affected the passage of sediment into the cave.

Cooper’s is one of the lesser known hominin bearing sites within the Cradle, although recent research has focused on the palaeontological assemblages and the geology of the Cooper’s D deposit (Berger *et al.*, 2003). The site consists of two spatially distinct fills, which are termed Cooper D East and West (Figure 1.7). The entire sequence rests on a dolomitic floor, with various forms of remnant speleothems, which represent a period of time prior to cave opening, when chemical deposition took place. Two entrances probably fed sediment into the underlying cave, with one in the northeast and the other in the southwest, and the shape of the entire deposit is controlled by the dolomite cave walls. The stratigraphy of the Cooper's D West deposits begins with a coarse pebbly breccia, and in the western-most exposures it consists of very fine sand and micromammal remains. This deposit occurred when the cave roof was still in place, as there is a scarcity of roof blocks, and thin, but well defined flowstones between units. Cooper's D East is a far coarser deposit, with maximum clast sizes of 50cm. The dips on the clasts form a centripetal pattern, suggesting that they formed as a cave fill cone, under a discrete, vertical to subvertical roof entrance. The abundance of roof blocks within the upper breccias suggests that deposition was concurrent with, and followed by, a period of deroofing (Berger *et al*., 2003).

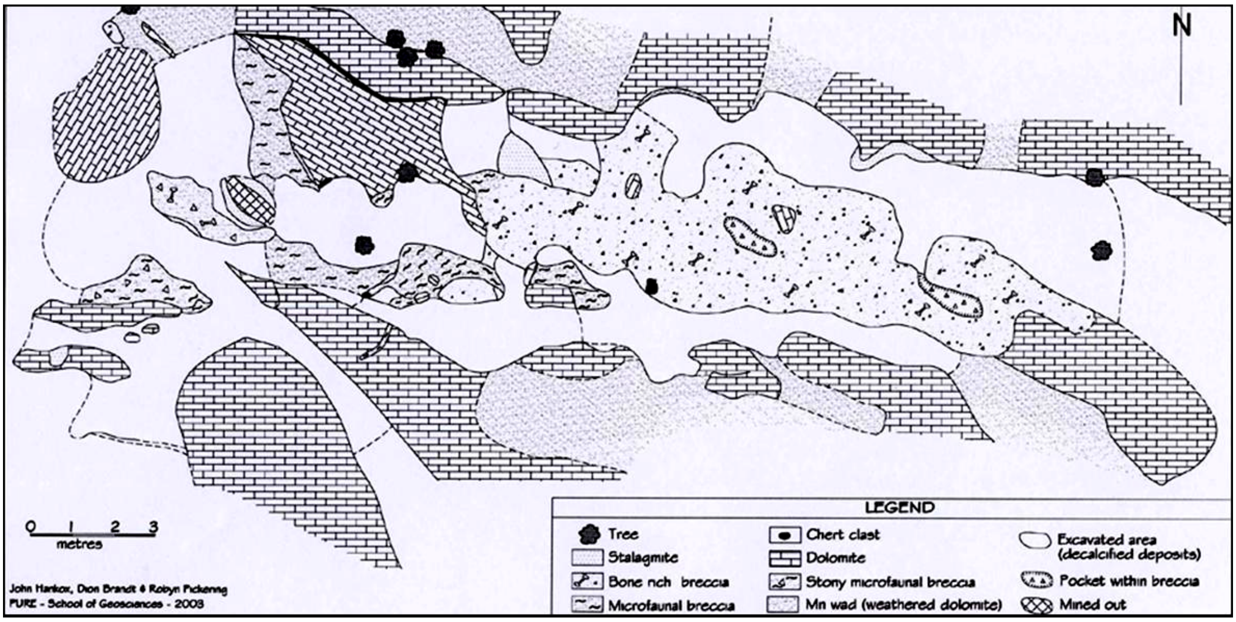

Figure 1.7. Plane table map of the geology of the Cooper's D deposit, showing the east and west breccias (Berger *et al*., 2003).

Drimolen is situated some 7km north of the better known sites of Sterkfontein and Swartkrans, and was only discovered in 1992 (Keyser *et al*., 2000). The site has a rich faunal assemblage, with hominin material attributed to *Paranthropus robustus* (Keyser *et al*., 2000). The cave sediments preserved at Drimolen have been thoroughly investigated and Keyser *et al*. (2000) present a block diagram showing the stratigraphic relationships of the sedimentary fill (Figure 1.8). Various fossils have been recovered from *ex situ* blocks. Of the *in situ* material, two fossil bearing lithologies are recognised: a Blocky Breccia and a Cave Siltstone (Keyser *et al,* 2000). The Blocky Breccia was deposited in a debris cone and is a clast supported deposit with a pinkish brown sandy matrix, with significant dolomite and chert clasts. The unit is highly fossiliferous and all the hominin remains have been recovered from here. The unit is well exposed in the back wall of the main trench of the excavations (Figure 1.8). The Cave Siltstone is a reddish brown, thinly laminated unit, showing evidence of mud cracking along bedding planes and is rich in micromammal fossils. The field relationships of these two units suggest that the Cave Siltstone is the distal equivalent of the cone and comprises the finer sediments washed out from the cone.

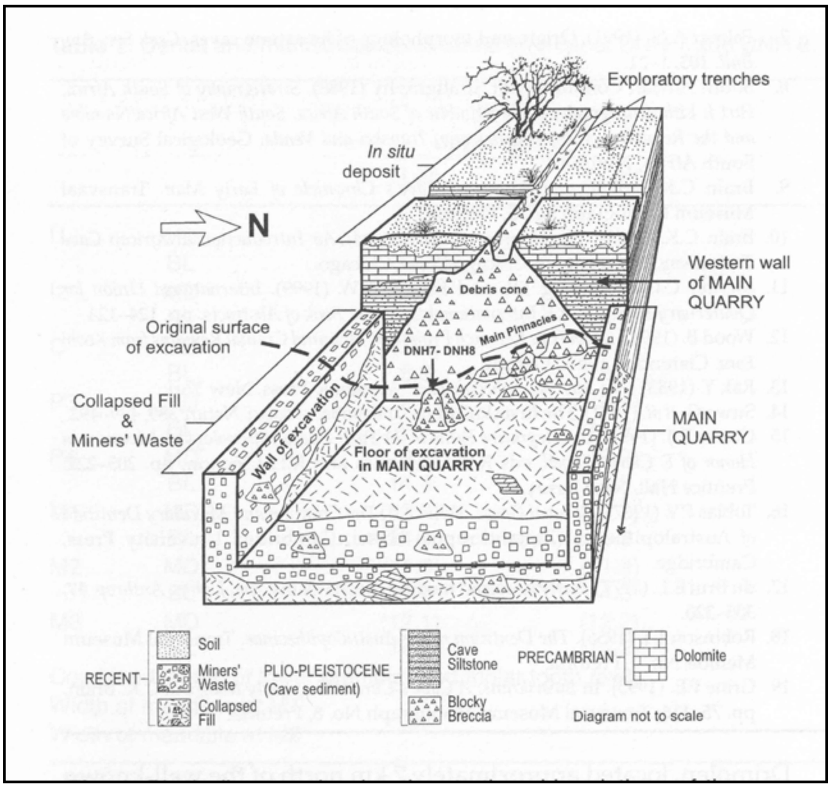

Figure 1.8. Block diagram showing the stratigraphic relations of the breccias at Drimolen (Keyser *et al*., 2000).

Outside the Cradle of Humankind World Heritage Site, the other site where significant hominin fossils have been discovered is the Limeworks Cave at Makapansgat, in Limpopo Province, some 300km north of Johannesburg. The site is currently believed to be between 2 and 4 Ma old, and is well known for the *Australopithecus africanus* fossils found there (see Latham *et al*., 2003 for a recent summary). The cave system is vast and the breccia infillings complex. Latham *et al*. (2003) have reviewed the formation and subsequent sedimentation of the cave, and propose a stratigraphy for the entire deposit, which could be used for future palaeomagnetic work.

1.2.2 Cave sediments

Cave sediments can be divided into clastic and chemical sediments (speleothems). Most research into cave deposits focuses on one or the other, or the overall geomorphology of the entire cave fill deposit, with few integrated models.

Geomorphology of cave sediments

The major work on the geomorphology of the South African cave deposits has been done by Partridge (1978; 1979; 1982) and is summarised in a review paper (Partridge, 2000). In this paper, Partridge (2000) reviews the geomorphology of the sites for which "reasonably complete stratigraphies are now available". He describes the deposits at Makapansgat in a typical section, showing the relationships between various members, which are dominated by a large central debris cone, with other members occurring in the space around this feature. However, Partridge (2000) points out that due to the limited exposures these strata cannot be correlated throughout the cave and that the "vertical relationships between all the members could nowhere be established in a single section". This, together with the lack of dates for the various units, makes it virtually impossible to assess the timing of the deposition of the various members and their relationships at Makapansgat.

Similar problems of poor exposure and lack of dates persist at Sterkfontein, although the deposit is divided into a six-fold stratigraphic subdivision (Partridge, 1978). The same applies to Kromdraai, where again the exposure of the breccias is so limited, and the level of erosion so advanced, that it is difficult to describe the geomorphology of the entire deposit. The deposits at Kromdraai are described as being part of a debris cone, with a vertical entrance shaft (Partridge, 2000), but actual evidence for this is limited. The deposit at Swartkrans is also described as being part of a debris cone (Brain, 1993), but once again the roof of the deposit is missing, and erosion of the members makes it difficult to reconstruct their original geomorphology.

As briefly demonstrated here, the major problems with the geomorphological descriptions of the South African hominin caves sites summarised in Partridge (2000), are the limited exposure of the deposits, and the incomplete nature of these deposits due to erosion. These factors are a function of the assumed Pliocene to early Pleistocene age of the deposits, and the combination of these issues makes the reconstruction of original geomorphology of these deposits exceedingly difficult. As a result, the original nature of the deposits and the processes responsible for their deposition remain poorly understood in the South African context.

These problems are not apparent in younger cave fills, as cave roofs are still intact and entire deposits can be mapped and documented, and processes modelled. Two examples of this are from the Naracoorte Caves in South Australia (Moriarty, *et al*., 2000) and from the caves on the island of Bermuda (Hearty, *et al*., 2004).

Moriarty *et al*. (2000) describe the various deposits in the numerous chambers at the Naracoorte Caves in terms of their "morphostratigraphy" (Figure 1.9). The deposit is described as a cone, which has developed under a vertical shaft, and a fan, which has grown out away from the cone and the entrance shaft, and is divided into a proximal and distal fan region (Moriarty, *et al*., 2000). Speleothems, in the form of stalagmites, stalactites and flowstones, are restricted to areas around drip sources, and are consequently only found around the entrance, in the middle of the proximal region and up against the back wall of the cave. The fan deposit is divided into a lower and upper fan deposit, which together with the U-series ages for the speleothems, suggest at least two periods of cave sedimentation.

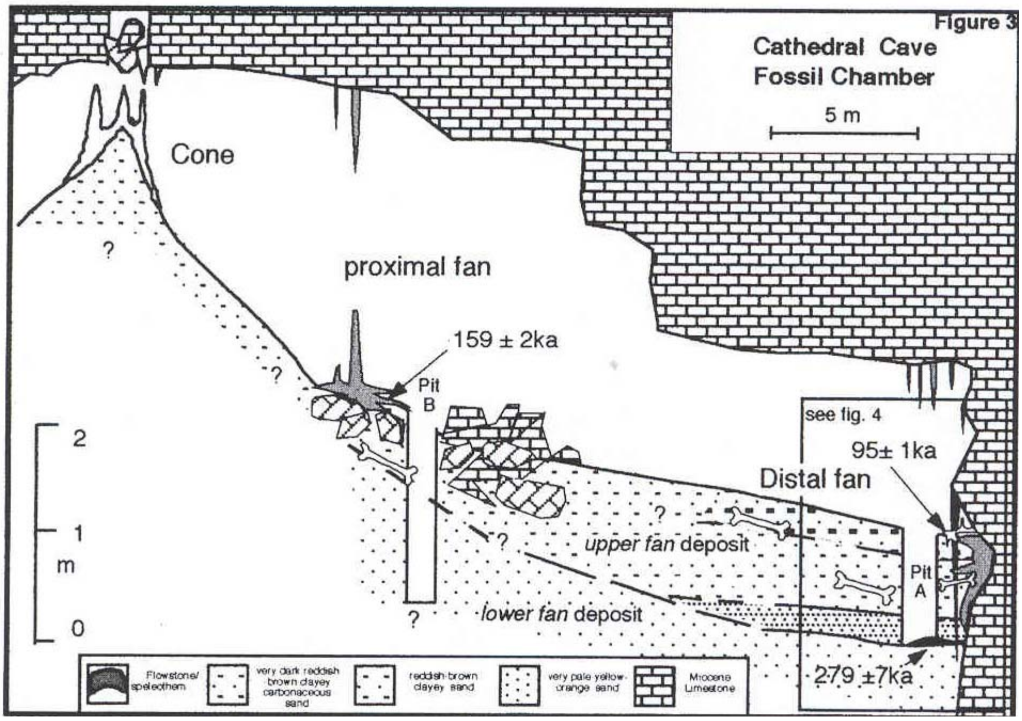

Figure 1.9. Section through Cathedral Fossil Cave, Naracoorte Cave, showing the morphostratigraphy of the deposit (Moriarty, *et al*., 2000).

Hearty *et al*. (2004) researched the stratigraphy and geochronology of the pitfall deposits within caves and fissures on the island of Bermuda. In the Admiral's Cave they have documented one of the largest debris cones on the island (Figure 1.10). They describe this deposit as a conical debris accumulation, which formed under a single skylight entrance. A stratigraphic pit through the deposit revealed twelve distinct flowstone, sediment and fossil layers. A layer of flowstone has accumulated on the present surface of the debris cone, fed from overhead drip sources.

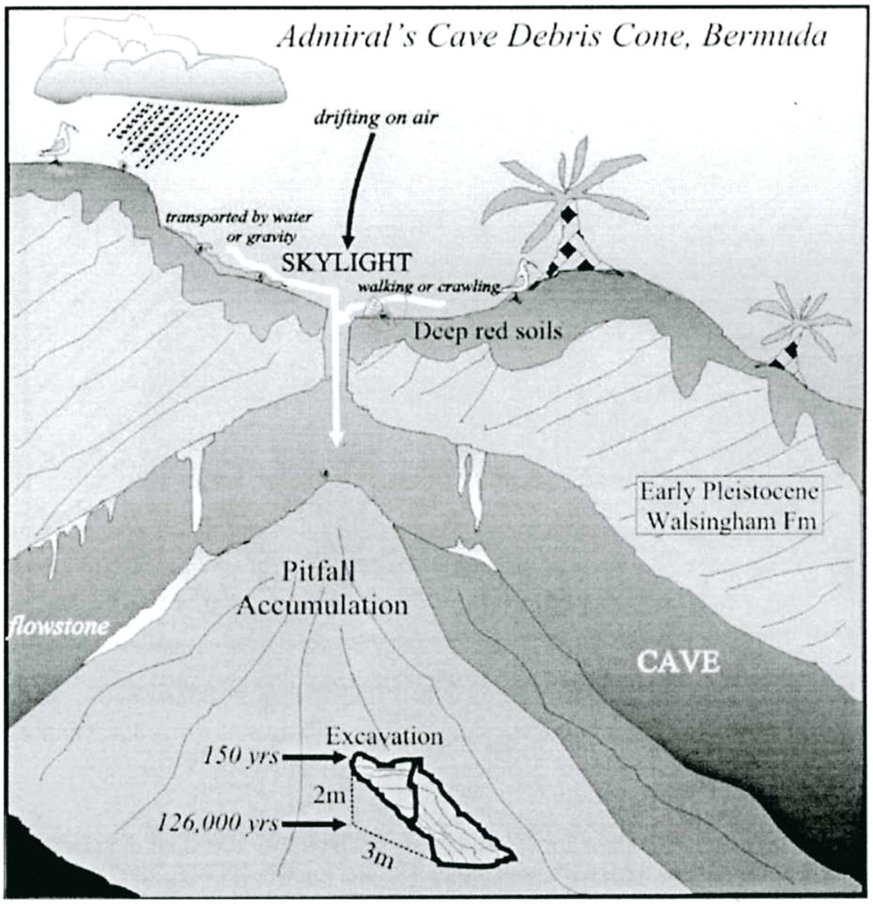

Figure 1.10. A schematic illustration of the Admiral's cave skylight, pitfall accumulation and excavation (Hearty *et al*., 2004).

Clastic sediments

As an introduction to clastic cave sediments, Gillieson (1996, page 143) notes that "a great deal of research has been carried out on the material deposited with the clastic sediments (bones, pollen archaeology) as a means of elucidating environmental or human histories. Less research has been undertaken on the processes by which clastic sediments are produced, transported and deposited within the cave system". Gillieson (1996) argues that this is in part due to the difficulties inherent in observing and measuring flood events in caves. He goes on to argue that there are in fact few differences between the nature of clastic sediments on the surface and those in caves, but that more time transgressive models of sedimentation in caves need to be developed, and the interaction between sedimentation and cave morphology investigated. Gillieson's final observation is that "cave sediments are tricky to understand". With this is mind, approaches to cave sedimentology are reviewed below.

Gillieson (1996) argues that clastic cave sediments form from rock fragments which are broken up by chemical and physical weathering processes while still on the surface. These fragments, in the form of angular boulder debris, cobbles and gravels, sands and silts, then get transported into caves, and are classified as allogenic. Other allogenic cave sediments include organic remains of animals and plants. Authigenic cave sediments also have organic and inorganic components, the latter being predominantly secondary carbonates. These are discussed later in this chapter.

Gillieson (1996) divides the processes of cave sedimentation into gravity fall processes, which occur inside a cave, producing *in situ* sediments, and water-borne processes, which are analogous to those of surface fluvial systems. Water moving through a cave system can deposit sediment as well as erode sediment, and according to Gillieson (1996) is the major control over both the deposition and preservation of cave sediments.

Once emplaced within a cave, sediments will undergo varying degrees of diagenesis. In general, the conditions inside a cave are ones of total darkness, near constant humidity and near constant temperature, which reduces the amount of chemical alteration that can occur (Gillieson, 1996). However, with time some migration of solutes in and out of cave sediments will occur, and depending on local hydrological conditions, sediments are commonly cemented by calcite from cave drip waters (Gillieson, 1996; Moriarty *et al.*, 2000).

Partridge (2000) presents a review of the cave sediments in the South African hominid bearing caves. In general his approach is lithostratigraphic, and he divides cave sediments into various units and members based on their sedimentological characteristics. He argues that in the karst infillings in South Africa there are periods of erosion between periods of sedimentation, and that these are most likely controlled by the external environment coupled to the hydrological conditions within the caves. He further notes that poor exposures of the older, very cemented breccias have hampered the understanding of cave sediments and that research into these older breccias is restricted to bore-core studies.

A slightly different approach is employed by Kos (2001), who uses various sedimentary facies to document and describe the cave sediments in McEachern's Deathtrap Cave, in southeast Australia. These cave sediments span the late Pleistocene to Holocene, and reflect changes in surface processes duirng this time. Kos (2001) divides the sedimentary sequence, into an upper, middle and lower sequence, which in turn consist of eight depositional and erosional facies (Figure 1.11). Facies classification is based on a combination of bedding and textural characteristics that dominate individual sedimentary horizons, taken from Reading (1996). These deposits have been dated by Accelerated Mass Spectrometry (AMS) ^{14}C on charcoal, producing a range of dates from throughout the sequence.

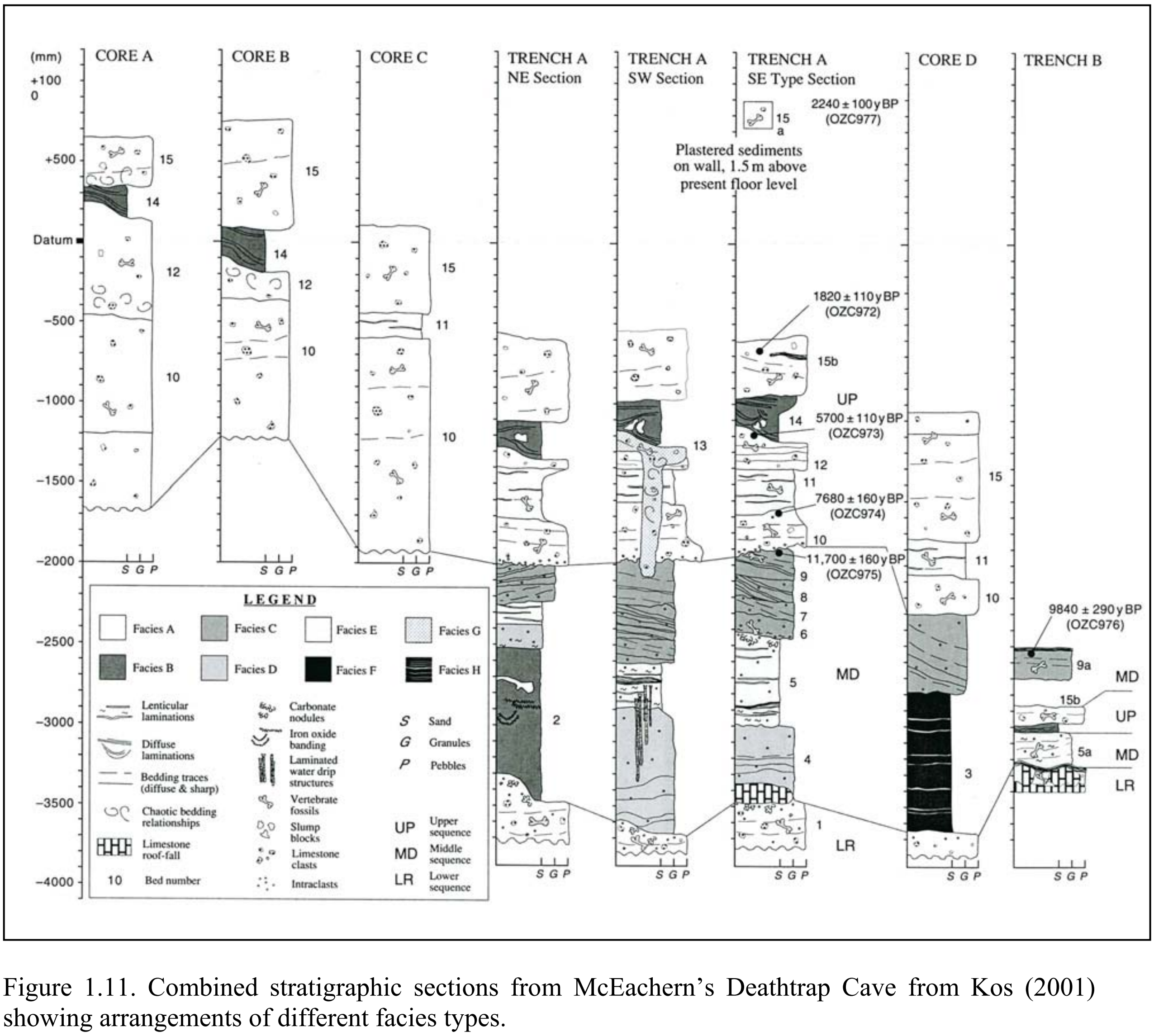

Figure 1.11. Combined stratigraphic sections from McEachern's Deathtrap Cave from Kos (2001) showing arrangements of different facies types.

Kos (2001) reconstructs the palaeoclimatic conditions for the last 120 kyr based on the presence or absence of sedimentation in the sequence, which is a chronostratigraphic approach, coupled with his facies analysis. He relates the cycles of erosion and deposition during the last interglacial to groundwater fluctuations, with erosion occurring during periods of decreased groundwater levels, and visa versa. In Australia the last glacial (around 20 ka) is represented in the cave sediments by hiatuses in the sedimentation and speleothem development, and some periods of erosion, which Kos (2001) equates with drier surface conditions.

The approach used by Kos (2001) is particularly pertinent in that it is one of the very few studies of cave sediments to use a facies approach. A working definition of the term facies is *"an assemblage or association of mineral, rock or fossil features reflecting the environment and conditions of origin of the rock"* (Lapidus & Winstanley, 1990: pg 202). The exact meaning and application of the term has become somewhat blurred and Lapidus & Winstanley (1990) warn that the meaning of the term is, to an extent, determined by its context.

However, the fundamental value and objective of facies studies are that they allow one to make environmental interpretations. Facies are essentially products of their depositional environment and conditions, and can therefore be used to identify that same environment in the rock record (Boggs, 2001). At the time of deposition, different depositional environments occur in a later succession, as a function of their hydrodynamics, with high energy deposits grading into lower energy deposits. Similarly facies corresponding to these environments also change laterally, and thus lateral facies arrangements or facies changes through space reflect the changing hydrodynamics of a depositional system (Miall, 1997).

As laterally contiguous environments in a region shift with time in response to changing geological conditions facies boundaries also shift, so that facies which were initially lateral to each other may be stacked vertically in space. The seemingly simple concept that a direct environmental relationship exits between laterally and vertically stacked facies was first proposed by Walther in 1894, and is known as "Walther's Law" (Boggs, 2001). The driving forces behind the shifting of depositional environments are tectonic up-lift, sea-level change and climate change.

Chemical sediments: speleothems

Speleothems are recognised by growing numbers of scientists as well preserved archives of information about past climate, vegetation, hydrology, sea level, nuclide migration, water-rock interaction, landscape evolution, tectonics and human actions (see Richards & Dorale, 2003 for a full review). Speleothems are particularly useful archives for these sources of information: they occur in many locations around the globe; they can be sampled at high resolution; and may be reliably dated using high-precision uranium-series techniques (Richards & Dorale, 2003).

Speleothems may be defined as bodies of mineral material formed from the chemical precipitation from ground water flowing or dripping in a cave. They generally consist of calcite, formed from the slow degassing of CO_2 from supersaturated ground waters (Richards & Dorale, 2003). Aragonite and gypsum commonly also contribute to speleothem mineralogy, especially near cave entrances where air circulation is increased and evaporative conditions become prevalent (Richards & Dorale, 2003). The mineralogy and morphology of these speleothems is a function of fluid flow, the chemistry of the ground waters and the ambient conditions, including temperature, chemistry and light, of the air or water-filled chamber. A great variety of different types of speleothems are found within caves, and can be divided into subaerial forms (stalagmites, stalactites, flowstones and draperies) and subaqueous forms (rimstone pools, mammillary calcite wall coating and 'dog tooth' spar (Hill & Forti, 1997; Richards & Dorale, 2003).

Speleothems can be used in a number of different ways to investigate past environments, perhaps the most fundamental of which is their presence or absence, because the deposition of speleothems relies on a range of sufficient water supply and soil CO_2 to enable the dissolution and transport of reactants in the vadose zone to underlying caves. The initiation and cessation of speleothem growth is also controlled by factors such as changes in vadose water chemistry, aridity, flooding events, permafrost conditions and random shifts in flow routing (Richards & Dorale, 2003). Once reliable dates are acquired for speleothems, they provide valuable spatial and temporal constraints on environmental factors. The spatial extent of the principal factor controlling speleothem presence or absence may be global (in the form of eustatic sea level change), regional or local (e.g. back flooding of a cave passage from sediment blockage downstream).

The growth frequency of speleothems is another important palaeoclimatic indicator. In general, aridity is not favorable to speleothem growth, as these circumstances are associated with a water deficit, and once arid conditions have halted the supply of drip waters to the vadose region, speleothem growth will cease (Richards & Dorale, 2003). Aridity is often brought on by glacials, as at high altitudes the advance of glaciers dramatically affects karst regions as drip water routes become permanently frozen and the rate of soil CO_2 production is reduced. In some cases soil is stripped from the surface by the ice masses, resulting in a break in speleothem growth (Richards & Dorale, 2003). At temperate to mid latitudes, speleothem growth may continue, but at a reduced frequency under conditions of changing vegetation, soil CO_2 levels, soil moisture and availability.

Richards & Dorale (2003) review global periods of speleothem growth, and show that in general speleothems grow during warm, wet interglacial conditions, with major world wide speleothem growth at ~128 ka during the penultimate interglacial. However, in desert locations speleothem growth is expected during cooler periods, such as at the Naracoorte Caves in Australia, where interglacial periods are characterized by a water deficit, and where speleothem growth occurs during glacials, when there is an increase in effective precipitation (Ayliffe *et al.*, 1998).

Other than the presence or absence or growth rates, speleothems contain important proxies for environmental change in the form of stable light isotopes of oxygen and carbon (Richards & Dorale, 2003). Crucial to both these palaeoclimatic approaches is accurate dating of speleothems (such as U-series dating), as once dated, speleothem records can be compared with independent records from the terrestrial environment, oceans and the cryosphere, so that the amplitude and spatial patterning of the response to different climatic forcing events can be investigated (Richards & Dorale, 2003).

Speleothems grow under specific conditions which are largely climatically controlled. Consequently in caves with intercalated speleothem and sediment deposits there must be climatic controls over both types of sedimentation. Unfortunately, intercalated speleothems are rare, but models for the climatic control of cave sedimentation have been generated and are discussed in the next section.

1.2.3. Cyclic cave deposition models

Brain (1995) Model

The work of Brain (1958) on the cave fills of Swartkrans, Sterkfontein and Kromdraai, and the long-term and detailed investigation of the Swartkrans by Brain (1981, 1993, 1995) provide the best insights into both cave formation and fill processes in South African hominin bearing caves. Brain (1995) argues that the 'most striking feature of the Swartkrans depositional sequence is the repeated alternation between periods of sediment accumulation and periods of non-deposition, the latter generally accompanied by active erosion within the cave'. Brain (1995) suggests that outside factors, such as changes in climate and the extent of ground-covering vegetation were involved, and that the cave fills at Swartkrans were deposited in climatically controlled cycles.

Brain (1995) uses Member 5, the youngest fossiliferous deposit at Swartkrans, to model the possible climatic changes controlling cave sedimentation. Member 5 consists of a 4 m thick deposit of sediment, which apparently was accumulated rapidly in an irregular solution channel, between 9 and 12 kyr (Brain, 1995). He argues that this period of deposition reflects a time, at the start of the current interglacial period, in which the storm water entering the cave was carrying a large load of soil. This soil was eroded because of the incomplete mat of vegetation in the catchment area. Indications from other sites suggest that the moisture index at this time was low, and rainfall increased in amount and/or intensity, as the last glacial maximum ended (Thackeray & Lee-Thorp, 1992). As a result the hillside soil would have been prone to erosion, as it was unprotected by vegetation, which had died back during the glacial maximum (Brain, 1995). Conversely, he argues that the erosional space, into which Member 5 was deposited, formed under different conditions when the water entering the cave was devoid of sediment load and was acidic. During this glacial interval, temperatures were lower, causing a reduction in evaporation. As a result there was more water passing through the dense vegetation mat, which was rich in humic acid, and would have chemically eroded the carbonate rich cave fills (Brain, 1995).

Brain (1995) therefore implies that all the cave fills (not just Member 5) at Swartkrans were deposited in climatically controlled cycles. During interglacial periods, when there was little vegetational cover and increased rainfall, soils were prone to erosion and cave deposition took place. During the corresponding glacial intervals, vegetational cover was much thicker, resulting in little to no erosion of soil and increased acidity of surface water, which itself was increased, as a result of low evaporation rates. These acid, surface waters entering the cave system caused the erosion of the older cave deposits.

The work of Brain (1995) is not the only instance of a cyclic depositional sequence documented in South African hominin caves. Partridge (2000) presents a model for cave formation and filling in the South African context which is not dissimilar from the Brain (1958) model. He points out that departures from this general scheme are numerous, and relate chiefly to "the timing of erosional episodes between periods of sedimentation" (Partridge, 2000: pg 102).

Ayliffe et al. (1998)/Moriarty et al. (2000) Model

A second model on cave fill processes is proposed by Ayliffe *et al.* (1998) and Moriarty *et al.* (2000) based on work on the limestone caves of the Naracoorte region in South Australia, where extensive deposits of megafauna-rich sediments are intercalated with speleothems and flowstones.

Ayliffe *et al.* (1998) used TIMS U-series to date the speleothems at the Naracoorte Caves and argue that the speleothems document changes in effective precipitation over the past 500 kyr. Ayliffe *et al.* (1998) were able to show that there were four major periods of speleothem growth in the past at: 20-115 kyr, 155-220 kyr, 270-300 kyr and 340-420 kyr, corresponding to the stadial and cool interstadials of the last four cycles. Comparing these growth periods with known climate change records, such as the SPECMAP record and the radiolarian temperature curve from RC 11-120 (Figure 1.12), these authors proposed that the lack of speleothem growth during interglacials indicated that these periods were comparatively arid in South Australia.

No modern speleothem material is found at the Naracoorte Caves, because of the current hydrological balance of the region, in which the potential evaporation exceeds precipitation for most of the year, except for a few months in winter, resulting in a net water deficit. Ayliffe *et al.* (1998) argue that these modern conditions can serve as an analogue for the interglacial conditions, during which the increased regional temperatures affected the potential evaporation, leading to a net water deficit and no speleothem growth. Conversely, during the glacials and cool interstadial regional temperatures were lowered, depressing evaporation and increasing the effective precipitation, allowing for speleothem growth to take place. Ayliffe *et al.* (1998) suggest that the lower regional temperatures and increased precipitation were driven by changes in atmospheric circulation in the form of stronger Walker circulation in the region during glacials.

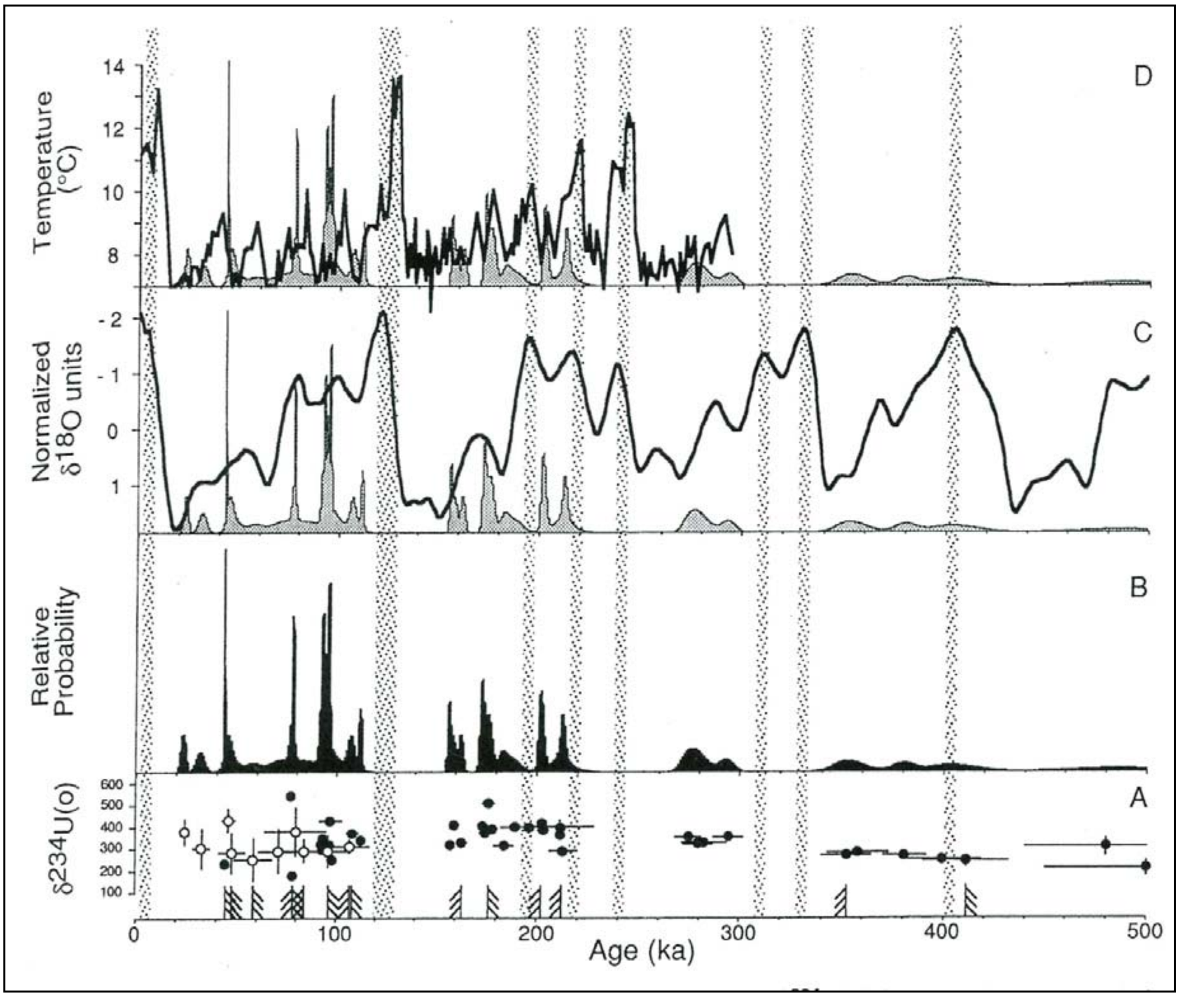

Figure 1.12. Composite diagram from Ayliffe *et al* (1998) showing from bottom to top: A: uranium-series ages for speleothems at Naracoorte; B: relative probability of speleothem growth; C: relative probability of speleothem growth with SPECMAP climatic record (Imbrie, *et al*., 1984), stippled bars are warm periods; D: temperature curve from RC-11-120. Speleothems at Naracoorte grow during glacial periods.

Building on the Ayliffe *et al.* (1998) model, Moriarty *et al.* (2000) interpreted development of flowstone at Naracoorte as representing a wet environment with good vegetational cover outside the cave, and a restricted or closed cave entrance, i.e. that they represent a clastic hiatus. The sediments are interpreted as representing a drier environment, with less vegetational cover and better run off and sediment supply into an open cave.

Moriarty *et al.* (2000) have shown with high precision TIMS U-series dating of flowstones, that these layers do indeed represent temporal breaks in clastic sedimentation, and that there is a distinct cyclicity in the timing of the sediment and flowstone depositional events in the Naracoorte caves.

Moriarty *et al.* (2000) go on to make more specific palaeoenvironmental interpretations from flowstone form and growth rates, as well as the clastic material and fossil bone. These interpretations are summarised in Table 1. In terms of the flowstones themselves, Moriarty *et al.* (2000) argue that a single uncontaminated, massive flowstone layer within a sedimentary sequence indicates that the entrance of the cave was sealed for a substantial period. During this time minimal clastic material and run-off water could have entered the cave. Conversely, a flowstone rich interval with inclusions of detrital material implies more open conditions during a wet surface episode.

Moriarty *et al.* (2000) also correlate the degree of calcification of the clastic layers with climatic changes. Layers with no calcification correlate to arid periods or to very fast sedimentation in a wetter period. Calcification of sediments under drip waters occurs very rapidly, even in a climate too dry for flowstone formation, so clastic sediments with minor calcification may indicate drier conditions (Moriarty *et al.,* 2000). Clastic layers with intercalated layers of contaminated flowstone and well-calcified sediments imply a wetter climate and an open cave entrance, through which sediments and water could enter.

	Palaeoenvironmental interpretation			
Surface climate	Wet		Arid	
Cave entrance	Open	Closed	Open	Closed
Clastic input	Low	Nil	High	Nil
Bone input	High	Nil	Low	Nil
Flowstone growth rate	Fast	Moderate	Nil	Nil
Flowstone form	Detrital contamination, inclusion rich	Clean, few inclusions	n/a	n/a
Stratigraphy	Contaminated flowstones with inter-laminated clastics and bone	Clean flowstones, minor reworked clastic grains	Laminated clastics, sparse bone	Hiatus surface, no bone, no clastics, no flowstone

Table 1. Summary of the palaeoenvironmental interpretations of Moriarty *et al.* (2000) of the stratigraphic successions of flowstones and sediments in the Naracoorte caves, Australia.

Summary

Model	Palaeoclimatic conditions		
	Glacial (and cool interstadials)	Glacial maxima	Interglacial
Brain 1995	Erosion of cave sediments from increased volumes of more acidic water (lower temperatures, therefore reduced evaporation + dense vegetation, therefore acidic water)		Cave sediment deposited by storm water entering cave, reduced vegetation cover outside cave
Ayliffe *et al.* 1998	Speleothem growth: temperatures and evaporation decreased, therefore increased effective precipitation.	No speleothem growth: relative aridity	No speleothem growth: increased temperatures and evaporation, therefore net water deficit
Moriarty *et al.* 2000	Flowstone growth: wet phases. Some sediment input if cave entrance is open	Breaks in speleothem growth: dry	Sedimentary input, with minor flowstone growth: water deficit, increased input of fauna

Table 2. Summary of models of climatic control of cave deposits.

Brain (1995) does not include speleothems in his model, making his interpretation fundamentally different to the models of Ayliffe *et al.* (1998) and Moriarty *et al.* (2000). This aside, the models of Brain (1995) and Moriarty *et al.* (2000) both predict that sediments will be washed into caves during relatively arid interglacials. The key factor in controlling the hydrological balance in all three models is the effect of regional temperatures on evaporation, and the knock-on effect of evaporation on effective precipitation. It is a relatively well established fact that speleothems grow during wet times, with a net increase in effective precipitation (Richards & Dorale, 2003), but these wet conditions are not synonymous with interglacials, as Ayliffe *et al.* (1998) illustrate. The only way to establish if these wet conditions occur during glacial or interglacial times is to date the speleothems present in the cave, and the most suitable dating technique to do this is Uranium-series.

1.2.4 Uranium-series Dating

Uranium-series dating has revolutionised Quaternary research as it can provide an accurate and precise chronology back to 500 ka (Edwards *et al*., 1984), which is well beyond the 50 kyr cut off point for ^{14}C dating. The radio-active decay of the uranium minerals in two decay chains led to the discovery of radioactivity itself (Becquerel, 1896a, b). In 1907

Rutherford was the first to recognise that the decay chain of uranium and thorium could be used to estimate the age of the earth. The decay series of Uranium (^{238}U and ^{235}U) follows a multi-branch decay chain of radioactive isotopes of a number of elements ending in stable Lead (^{206}Pb) (Faure, 1986) (Figure 1.13). Uranium-series dating can be applied to a wide range of materials, including bone and shell, but the most accurate results are from chemically precipitated calcium carbonates, which occur as speleothems, travertines, calcretes and biogenic carbonates (Schwarcz, 1992).

Uranium-series dating of speleothems is made possible by the varying solubility of U and Th in natural ground waters. Thorium has a very low solubility while U has a relatively high solubility, which results in the separation of Th and U in the precipitation of speleothem calcite from drip waters (Schwarcz, 1989; Blackwell & Schwarcz, 1995). Freshly precipitated calcite therefore contains traces of U, but is free of Th, meaning that any measured Th must be the decay product of the *in situ* U (Schwarcz & Latham, 1989) and ages for the sedimentary material are therefore derived from the ratio of $^{230}Th/^{234}U$, as the ^{230}Th is produced from the decay of the *in situ* ^{234}U (Blackwell & Schwarcz, 1995; Schwarcz, 1992).

Mass spectrometric methods are used to measure the amounts of the parent and daughter isotope (Blackwell & Schwarcz, 1995) and this has greatly improved the precision of the dates (Edwards *et al.*, 1984). The use of TIMS has been particularly successful in acquiring precise dates of material as old as 500 kyr (Stirling, *et al.*, 1995; Ayliffe *et al.*, 1998; Kaufman *et al.*, 1998; Moriarty, *et al*., 2000). Recent research, however, has shown that Multi Collector ICP-MS is a more reliable and accurate method of measuring uranium isotopes (e.g., Stirling, *et al*., 2000; Hellstrom, 2003). ICP-MS is a preferred method because the U and Th yield is higher, as aspiration of dilute solutions into high temperature (~7000 K) plasma produces almost 100% atomisation and ionisation of Th and U. Th and U isotopes are also measured at the same time, removing the need for chemical separation of U and Th. These factors, together with increased machine capabilities and technological advances, generate precisions of 0.1% of 1σ, even with small sample size requirements (10-100ng Th or U), which are equivalent to the highest quality TIMS results (Stirling *et al*., 2000; Hellstrom, 2003).

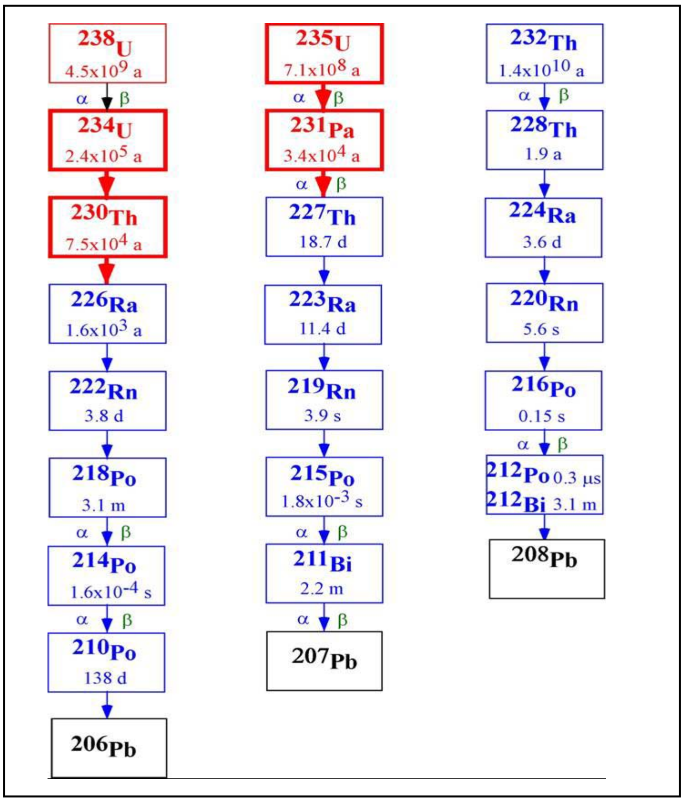

Figure 1.13. Uranium-series decay chain, from Uranium to stable Lead. Species involved in Uranium-series dating are marked in red (from Eggins, 2003).

The major problem with U-series is the contamination of carbonates by detritus, such as clay minerals containing U-series decay minerals in equilibrium (Grün, 2003; Schwarcz, 1992). During the chemical preparation of samples, Th and U are released from the detrital minerals as well as the calcite, meaning that the assumption that the sample does not contain any ^{230}Th at the time of formation is invalid. Also present in the detrital minerals is ^{232}Th, which does not occur in calcite, and as it behaves identically to ^{230}Th can be used to estimate the degree of ^{230}Th contamination, by the use of isochrons (Schwarcz & Latham, 1989; Bischoff & Fitzpatrick, 1990).

Age errors are asymmetrical due to the exponential nature of the age function. Finite ages can be produced as long as the measured ^{230}Th/^{234}U ratio is statistically distinct from equilibrium. Error calculations can be made on an Isoplot program (Ludwig, 2003) or Monte Carlo strategies (e.g. Hellstrom, 1998). Monte Carlo strategies are preferred for age calculation close to equilibrium.

1.2.5 Stable light isotopes of oxygen and carbon as records of palaeoclimate

The Quaternary is the latest period of the Earth's history and spans about the last 1.8 Ma (Aguirre & Pasini, 1985). It is subdivided into two geological epochs, the Pleistocene (from 1.8 Ma to about 10 ka) and the Holocene (from 10 ka to the present). The Quaternary is characterised by the rapid cycles of climatic change, with successions of warm (interglacial) and cold (glacial) climatic phases. Glacial phases are characterised by the occurrence of large ice sheets in the northern and southern hemispheres. Oxygen isotopes have played a major role in our understanding of these climate changes.

Oxygen ($\delta^{18}O$)

Oxygen occurs as three stable isotopes, ^{16}O, ^{17}O and ^{18}O. The fractionation of these isotopes between sea water and air, or fresh water and precipitating calcium carbonate is temperature dependent. The values for $\delta\ ^{18}O$‰ may therefore be used as signals of climate change, with positive $\delta\ ^{18}O$ values indicating a colder climate, and negative values signalling a warmer climate (Faure, 1986). This phenomenon has been applied to ice cores, deep sea cores and speleothems, which coupled with various dating techniques, has provided palaeoclimatic data for the Quaternary (Faure, 1986).

The current state of knowledge about the precise sequencing of the cold and warm phases during the Quaternary has been deduced from oxygen isotope analyses of deep-sea cores (Emiliani 1955, Shackleton and Opdyke, 1973). During cold climatic phases, a large amount of the atmospheric water is precipitated in polar regions and causes the build up of polar ice caps (and terrestrial ice masses). The ice of these polar regions is depleted in ^{18}O relative to sea water, i.e. it is "isotopically lighter". During the coldest climatic phases the sea-level is about 130m lower than today; current conditions are typical of an interglacial phase. This water is bound in the ice masses and therefore the remaining sea water is "isotopically heavy". When the climate changes, most of the ice is melted, and therefore the seawater becomes "isotopically lighter". These changes are documented in the shells of small marine organisms, namely foraminifera, which are continuously deposited on the ocean floors (Chappell & Shackelton, 1986). Oxygen isotope analysis of foraminifera from deep sea cores gives the major frame-work for Quaternary chronology. Figure 1.14 shows a generalised oxygen isotope curve for the last 900 kyr (Bassinot *et al*., 1994), indicating at least ten warm-cold cycles in this period, and numerous smaller oscillations.

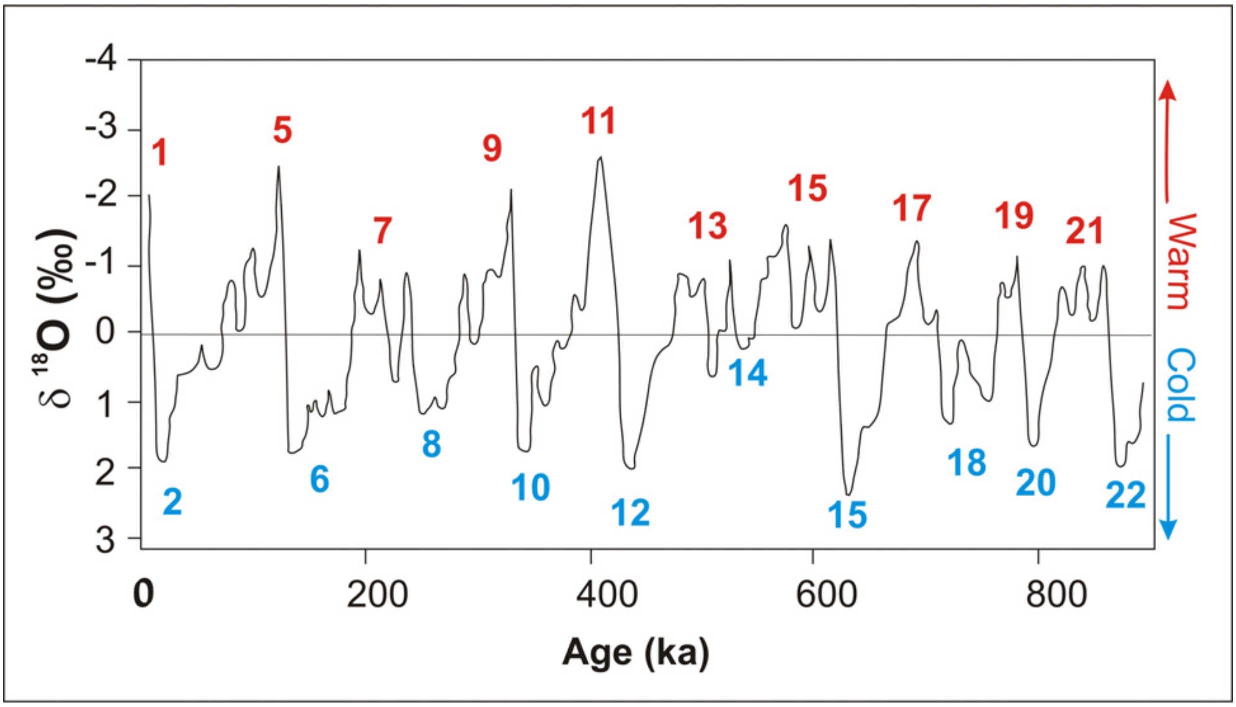

Figure 1.14. Generalised oxygen isotope curve for the last 900 kyr. Negative oxygen isotope ratios indicate warmer climatic periods and positive ratios indicate generally colder climates. The odd numbers label interglacial periods, and the even numbers indicate glacial periods (Bassinot *et al.,* 1994).

The driving force for the climatic oscillations are changes in the orbital parameters of the Earth, eccentricity of the Earth's orbit around the sun (on a 100 kyr cycle), obliquity (the tilt of the rotational axis of the Earth with a 41 kyr cycle) and precession (rotation of the axis itself with a 22 kyr cycle). These orbital parameters cause a change in the relative distribution of energy flux from the sun between the northern and southern hemispheres, which in turn triggers either the build-up or melting of ice masses. The relationship between orbital parameters and energy received from the sun has been calculated by Milankovitch in the 1920s and 1930s (Milankovitch, 1941), and these periodicities are thus known as 'Milankovitch cycles'.

There are several high resolution records of Quaternary climate change, which are used as references for other records to be compared to. The two most commonly used records, which pertain to southern hemisphere studies and are therefore pertinent in this study, are the SPECMAP record of sea level change, and the Vostok ice core record of temperature change.

The SPECMAP Archive No.1, which is freely available from the U.S. National Geophysical Data Centre (NGDC), contains climate time series of the past 700 kyr for 17 sediment cores from the Atlantic Ocean, and the basic downcore and core-top data from which these time series were derived. The records include quantitative data on the species and assemblages of planktonic fauna and flora that reflect changing surface conditions of the Atlantic Ocean. Measurements of $\delta^{18}O$ and $\delta^{13}C$ on the benthic and planktonic assemblages provide further palaeoclimatic data (Imbrie *et al.*, 1984). An age model is applied to each downcore record, to transform in into a time series by correlating the $\delta^{18}O$ records of each core with the published $\delta^{18}O$ chronology of Imbrie *et al.* (1984). The final result is a smoothed curve, which represents global sea level changes as a function of changing ice volumes for the last 780 kyr (Figure 1.15).

The Vostok ice core record is from the ice sheet in east Antarctica, and records global climate change for the last 420 kyr (Figure 1.14), with four glacial-interglacial cycles recorded (Petit *et al.*, 1999). The ice cores provide a range of palaeoclimatic data that includes local temperatures (analysed via the deuterium content of the ice) and precipitation rates, moisture source conditions, wind strength and aerosol fluxes of marine, terrestrial, cosmogenic, volcanic and anthropogenic origin (Petit *et al.*, 1999). Changes in global ice volume and in the hydrological cycle are assessed via the $\delta^{18}O$ content of the ice. The Vostok record of temperature change for the southern hemisphere is a smoothed record of the deuterium content of the cores (Petit *et al.*, 2001), and is also available online from U.S. National Geophysical Data Centre (NGDC).

Richards and Dorale (2003) review the application of oxygen isotopes ($\delta^{18}O$) on speleothems. The distribution of ^{18}O between calcite and water during the precipitation of speleothems is temperature dependent only if the system remains in isotopic equilibrium, such as at the back of deep caves, where ventilation in minimal and humidity high. Experimental studies on modern, inorganic calcite have shown that a 10°C change in temperature results in a 0.02‰ shift in fractionation (O'Niel et al., 1969; Friedman & O'Niel, 1977). Deviations in $\delta^{18}O$ in ancient calcite cannot be attributed to temperature shifts alone, and are a function of: changes in the $\delta^{18}O$ of the sea water, which fluctuates with changing ice volumes; changes in the path of the water from the source to the site of precipitation; and the isotopic composition and amount of rain and percolating vadose waters.

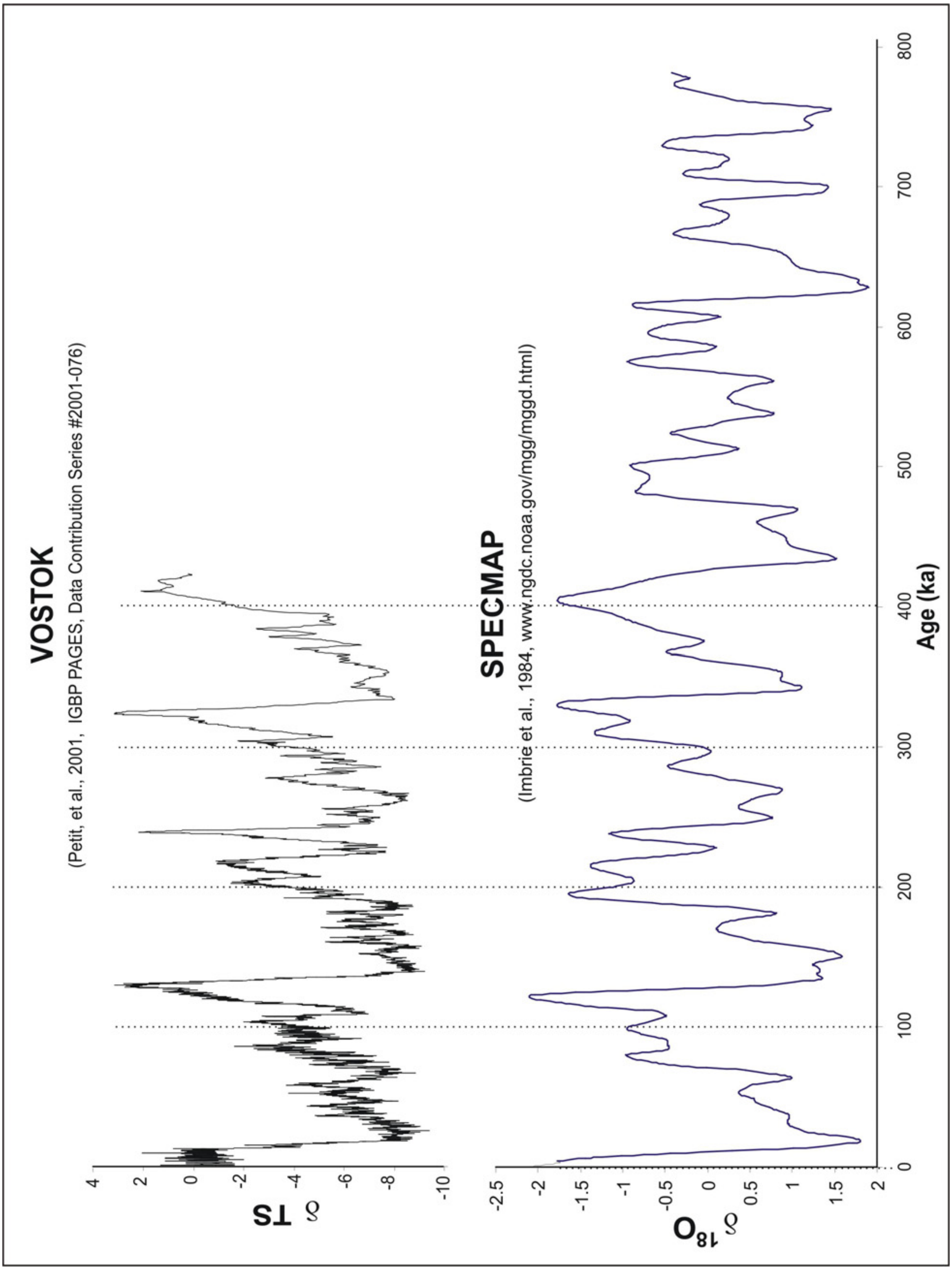

Figure 1.15. Graphs of climate change: Vostok ice core temperature change records for the last 400 kyr (top) in Antarctica (Petit *et al*., 1999) and the smoothed, global SPECMAP record (Imbrie *et al*., 1984).

In some cases calcite deposited during interglacial periods has a less negative $\delta^{18}O$ signature (Gascoyne *et al.*, 1981; Goede *et al.*, 1986; Dorale *et al.*, 1992; Lauritzen, 1995). In general negative shifts in $\delta^{18}O$ values are attributed to a combination of factors including glacial-interglacial changes in ocean $\delta^{18}O$, and intensification of rainfall and seasonal patterns of rainfall, which may result in shifts too great to be attributed to temperature alone (Richards & Dorale, 2003).

Long, high resolution speleothem $\delta^{18}O$ records of climate change, such as the Devil's Hole Calcite, Nevada, USA, can be correlated with other records, namely marine sediment cores and ice cores. Richards & Dorale (2003) argue that to match these, and any other palaeoclimatic records meaningfully, one must "identify common causal mechanisms that manifest themselves synchronously". Richards and Dorale (2003) also point out that there will be leads and lags between speleothem records and other proxies for climatic change, but that these are most likely due to either inaccuracies in the chronologies involved (only speleothems are at present directly datable) or in the correlation between the records.

Carbon ($\delta^{13}C$)

Carbon is one of the most abundant elements in the universe and is the basis for the existence of life on Earth. Carbon has two stable isotopes ^{12}C (98.99%) and ^{13}C (1.11%), and one unstable isotope ^{14}C (Faure, 1986). The stable isotopes of carbon are fractionated by a variety of natural processes. The processes pertinent to this study are photosynthesis and isotopic exchange reactions among carbon compounds. Photosynthesis results in an enrichment of ^{12}C, while the isotope reactions between CO_2 gas and aqueous carbonates leads to enrichment in ^{13}C. Consequently, the isotopic abundance of ^{13}C in terrestrial carbonates varies by about 10% (Faure, 1986). Carbon isotope values are expressed as $\delta^{13}C$ values. As differences caused from fractionation are very slight, and to make data more user friendly values are quoted as permil (‰) and not percent.

Photosynthesis centres on the incorporation of atmospheric carbon, in the form of CO_2 gas, by green plants, and the synthesis of this carbon into simple sugars through a complex set of biochemical steps. Atmospheric carbon has a $\delta^{13}C$ value of -7‰ and this value is significantly enriched during the fixing of carbon by plant tissue during photosynthesis (Faure, 1986). Photosynthetic plants can be divided into three major groups, based on a particular chemical pathway used during one stage of photosynthesis: C_3 plants, C_4 plants and CAM plants.

C_3 plants use a photosynthetic pathway which involves the enzyme rubsico. Only three carbons are used in this step of the reaction, hence the appellation C_3. C_4 plants use a slightly different pathway, without rubisco, which has an extra step, involving an extra carbon. C_4 plants have a different cellular arrangement to C_3 plants, and are specialised to fix carbon in low CO_2 conditions (O'Leary, 1981, 1993). The C_4 pathway is a modification of the C_3 pathway, and is relatively modern and occurs in more derived plant families, particularly monocotyleadonous plants, such as grasses and sedges (O'Leary, 1981, 1993). These different pathways result in different degrees of fractionation of carbon within the plants tissues, causing a significant difference in the $\delta^{13}C$ values of C_3 and C_4 plants (Vogel, 1993).

Vogel (1993) relates the geographical distribution of C_3 and C_4 plants in southern Africa to different climatic zones. In general, winter rainfall regions are dominated by C_3 plants (trees and shrubs), while warm, arid summer rainfall areas are 95–100% C_4. Vogel (1993) argues, therefore, that C_4 grasses are adapted to more arid conditions and high temperatures during the growing season. These climatic characteristics of C_3 and C_4 plants, together with the different isotopic signatures of the two groups, can be extended back into the past and are a proxy for climate change. Care must be taken when undertaking such research, as C_4 plants are not pure indicators for arid conditions, and systems need to be assessed holistically, with all other palaeoclimatic proxies.

Our understanding of the relationship between the $\delta^{13}C$ signal and changing vegetation patterns is predominantly derived from the study of secondary pedogenic carbonates. In terms of speleothem studies, the dissolved carbon in cave seepage waters is from three major sources: atmospheric C; decaying organic matter and root respiration in soil above the cave; and dissolved limestone (or dolomite) along the flow route to the cave (Richards & Dorale, 2003). The net isotopic composition of the dissolved HCO_3^- and the precipitated $CaCO_3$ is dependent on the $\delta^{13}C$ reactant in the system; the kinetic fractionation factors in the H_2O–$CaCO_3$–CO_2 system; the saturation state of $CaCO_3$; the rate of exchange with the gaseous phase and the rate of precipitation (Richards & Dorale, 2003). The interplay between these factors produces a complex system, and while changes in rainfall and temperature can cause variation in $\delta^{13}C$ records, the most useful information derived from speleothem $\delta^{13}C$ records relates to vegetation changes.

As already discussed, the photosynthetic pathway of different plant groups defines their $\delta^{13}C$ signature. In secondary carbonates a C_3 plant signature would produce $\delta^{13}C$ values in the range of -14 to -6‰ (Dreybrodt, 1980) to -30 to -24‰ (Vogel, 1993). C_4 plants produce $\delta^{13}C$ values in the range of -6 to +2 ‰ (Dreybrodt, 1980) to -16 to +10‰ (Vogel, 1993). Richards and Dorale (2003) point out that caution must be exercised when analysing $\delta^{13}C$ values, as other factors, such as vegetation density and soil respiration, can also play a major role in the creation of the $\delta^{13}C$ signature in secondary carbonates.

1.2.6 Quaternary climate change records in South Africa

The study of Quaternary climate changes in southern Africa is plagued by the relatively arid and erosive conditions which prevailed over this region, which limited the accumulation of continuous stratified deposits capable of yielding the kind and quality of information needed to access climate variability, change and periodicity (Lee Thorp, 2003). The lack of reliable, high resolution dating techniques have further limited this field.

However, with the advent of Uranium-series dating of stalagmites and speleothems, well-dated, continuous, high resolution records of climate variability are becoming available (Lee-Thorp, 2003; Holmgren *et a*l., 1999; Holmgren *et al*., 2003). High resolution carbon and oxygen stable light isotope data ($\delta^{13}C$ and $\delta^{18}O$) from stalagmites in the Makapansgat Valley document the regional climate change in southern Africa. New TIMS U-series dates of stalagmite T8, from Makapansgat, indicate that this record spans the Late Pleistocene and Holocene, with speleothem growth taking place from 24.4-12.7 kyr and from 10.2-0 kyr, interrupted by a 2.5 kyr hiatus (Holmgren *et al*., 2003) (Figure 1.16).

Another record of climatic change is from a speleothem from the Cango Caves, in the Western Cape (Talma & Vogel, 1992), which provides a $\delta^{18}O$ record for a large part of the last 30 kyr. The $\delta^{18}O$ data is translated to direct temperature change data, by using the isotopic composition of the confined groundwater in the Uitenhage aquifer, which is believed to be of a similar age (Figure 1.17). The stalagmite was dated using ^{14}C and ionium (^{230}Th) techniques, and in general the two techniques are in agreement; however, in the older sections of the stalagmite discrepancies do occur (Talma & Vogel, 1992). The dating shows that there is a major break in the Cango Cave stalagmite between 13,8 kyr and 5 kyr (Talma & Vogel, 1992), which may most likely be correlated with the break seen in the other records of this period already discussed. Talma & Vogel (1992) also present data on changing vegetation patterns for the last 30 kyr, based on the $\delta^{13}C$ values of the speleothem (Figure 1.17).

The Wonderkrater spring sediments in the Limpopo Province of South Africa provide a valuable source for pollen analysis, and by proxy, environmental changes in the Savanna Biome during the late Quaternary. The sequence has recently been re-dated using radiocarbon (Scott *et al.*, 2003), allowing for correlation with other local and global climate changes for this period. The pollen data has produced a temperature and moisture index for the last 20 kyr, with cold conditions at ~17 ka, and increased temperatures at ~13 ka (Figure 1.18).

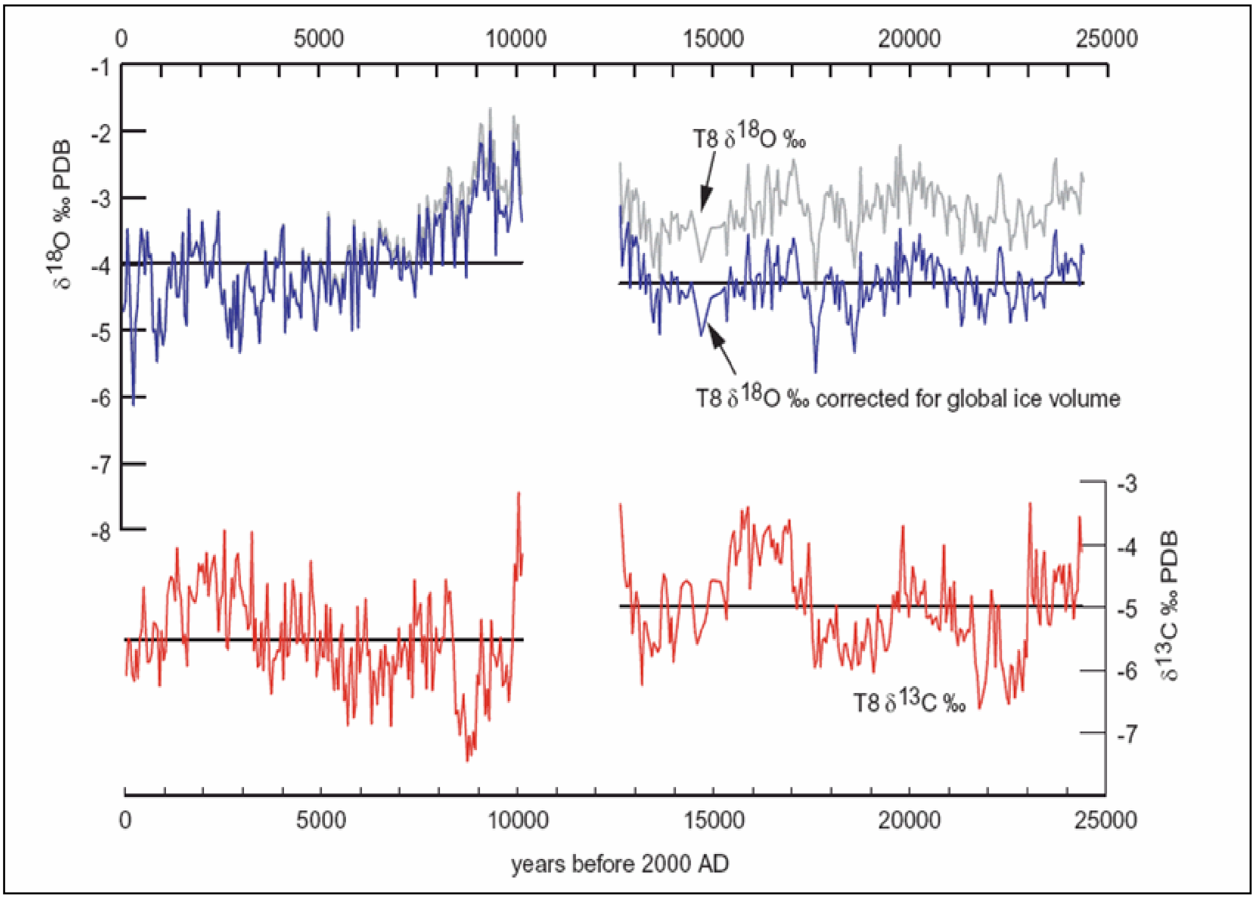

Figure 1.16. Oxygen (blue) and carbon (red) isotope data from the T8 stalagmite from Makapansgat, showing a major break in the sequence between 12.7–0.2 ka (Holmgren *et al.*, 2003).

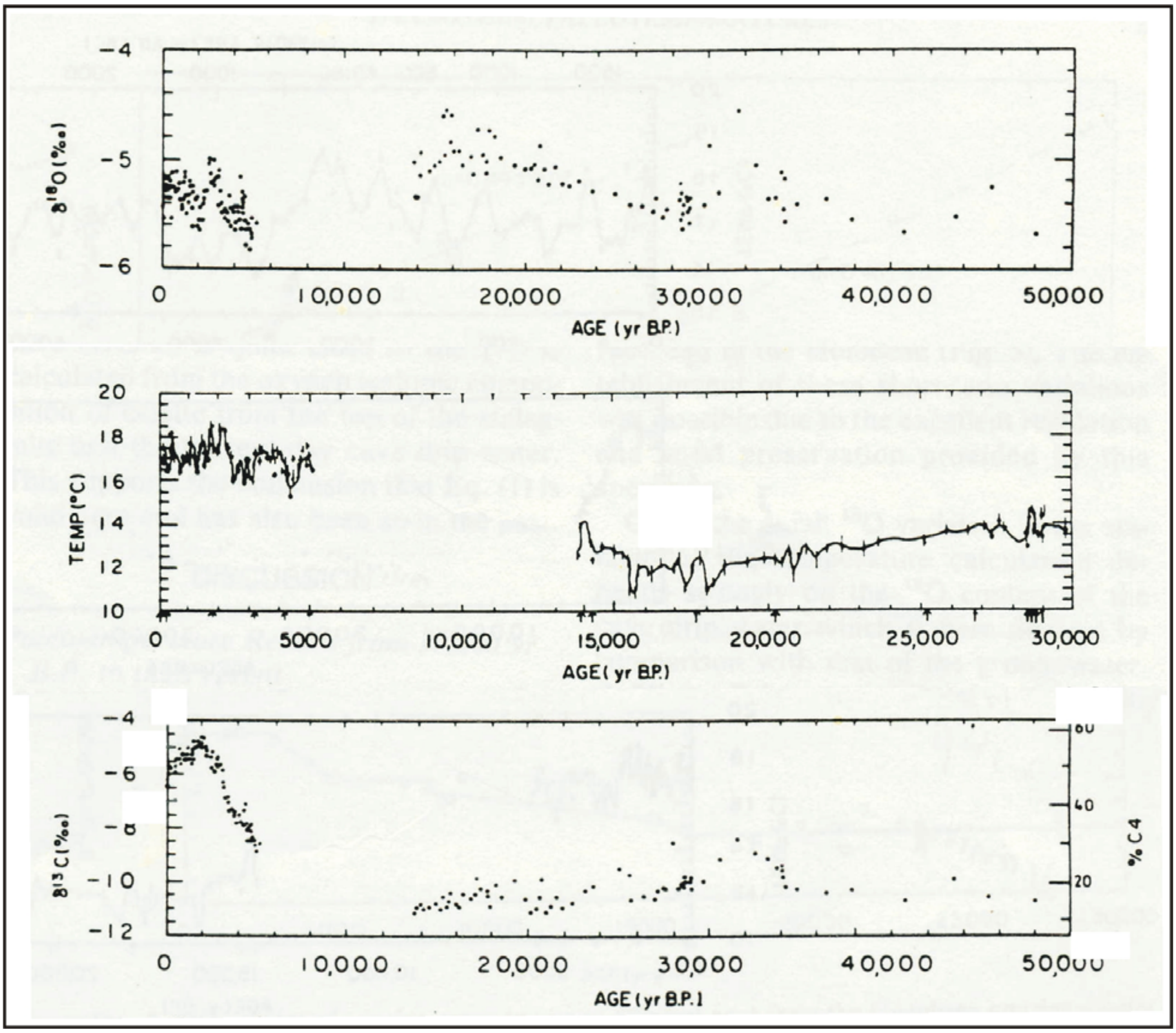

Figure 1.17. Palaeotemperature calculations, carbon and oxygen isotopes from the Cango stalagmite (Talma & Vogel, 1992).

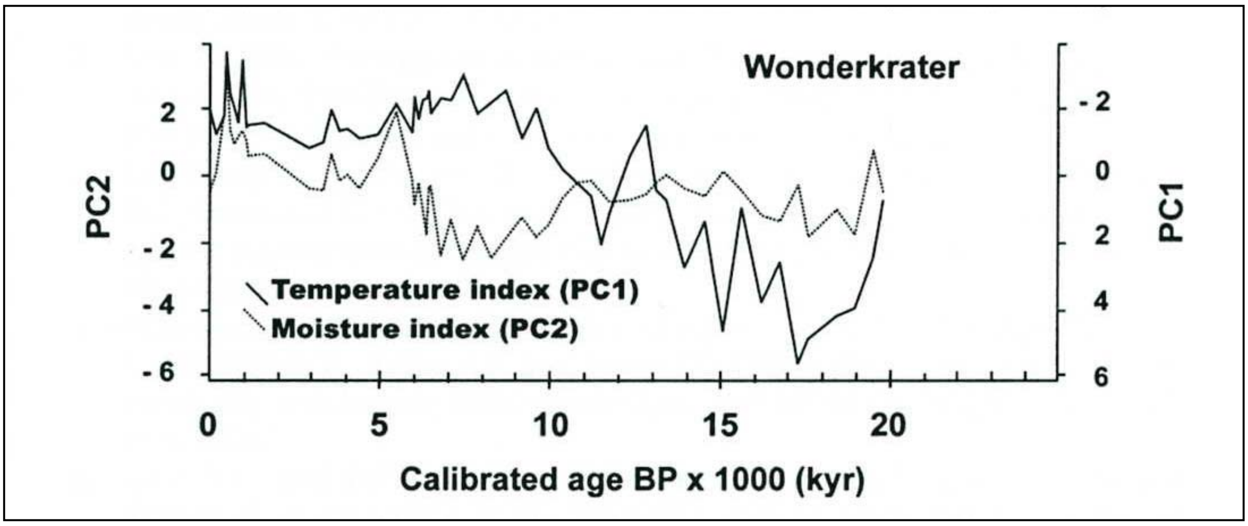

Figure 1.18. Wonderkrater temperature and moisture indices (Scott *et* al., 2003).

The crater lake sediments of the Tswaing Impact Crater provide a number of proxy records, including stable light isotopes, pollen, diatoms and phytoliths (Partridge *et al.*, 1999a). Reimold, Koeberl & Brandt (1999) conclude the Tswaing crater is of meteoritic impact origin. The impact is dated via fission track to 220 ± 52 kyr, and since about 200 kyr sediments have been accumulating in the lake in the middle of the impact crater (Partridge *et al.*, 1999b).

The dating of the sediments of the Tswaing Impact crater lake has been somewhat problematic. Radiocarbon ages were obtained for specific carbonate minerals and the organic remains (mainly dead algae and plants) found within the sediments (Partridge *et al.*, 1999b). The radiocarbon ages are, however, restricted to the upper 20m of the crater sediments, yet these dates, together with estimates of erosion rates have been used to create an age model for the entire profile.

The carbon and oxygen isotope work undertaken on the crater lake sediments was focused on the authigenic carbonate minerals of the sequence (Horstmann & Bühman, 1999). Only twelve samples were run and the resultant $\delta^{18}O$ and $\delta^{13}C$ profiles were not combined with the radiocarbon dates and age model to provide a time series profile of climatic change. The isotopes do however indicate that carbonate precipitation took place under arid to semi-arid conditions, and that in the upper section of the sequence conditions were more evaporative. The enriched $\delta^{18}O$ values suggest that the lake remained a closed hydrological system, in which evaporative conditions were dominant (Horstmann & Bühman, 1999).

More direct palaeoclimatic inferences are made by Partridge (1999b), who developed a facies analysis of the crater sediments, and related changes in facies with changes in the amount of runoff from the sides of the crater, which in turn he related to changes in the amount of precipitation. Principal component analysis statistics were used to determine the link between the properties of the lake sediments and mean annual precipitation. From the combination of the vertical facies changes of soil sediment texture, and the age-depth model, Partridge *et al.* (1997) presents a rainfall time series for the last 200 kyr in the Pretoria region (Figure 1.19). Partridge *et al*. (1997) argues that the main periodicity of this record is around 23 kyr and that nine cycles of orbital precession are reflected in the record (Figure 1.19).

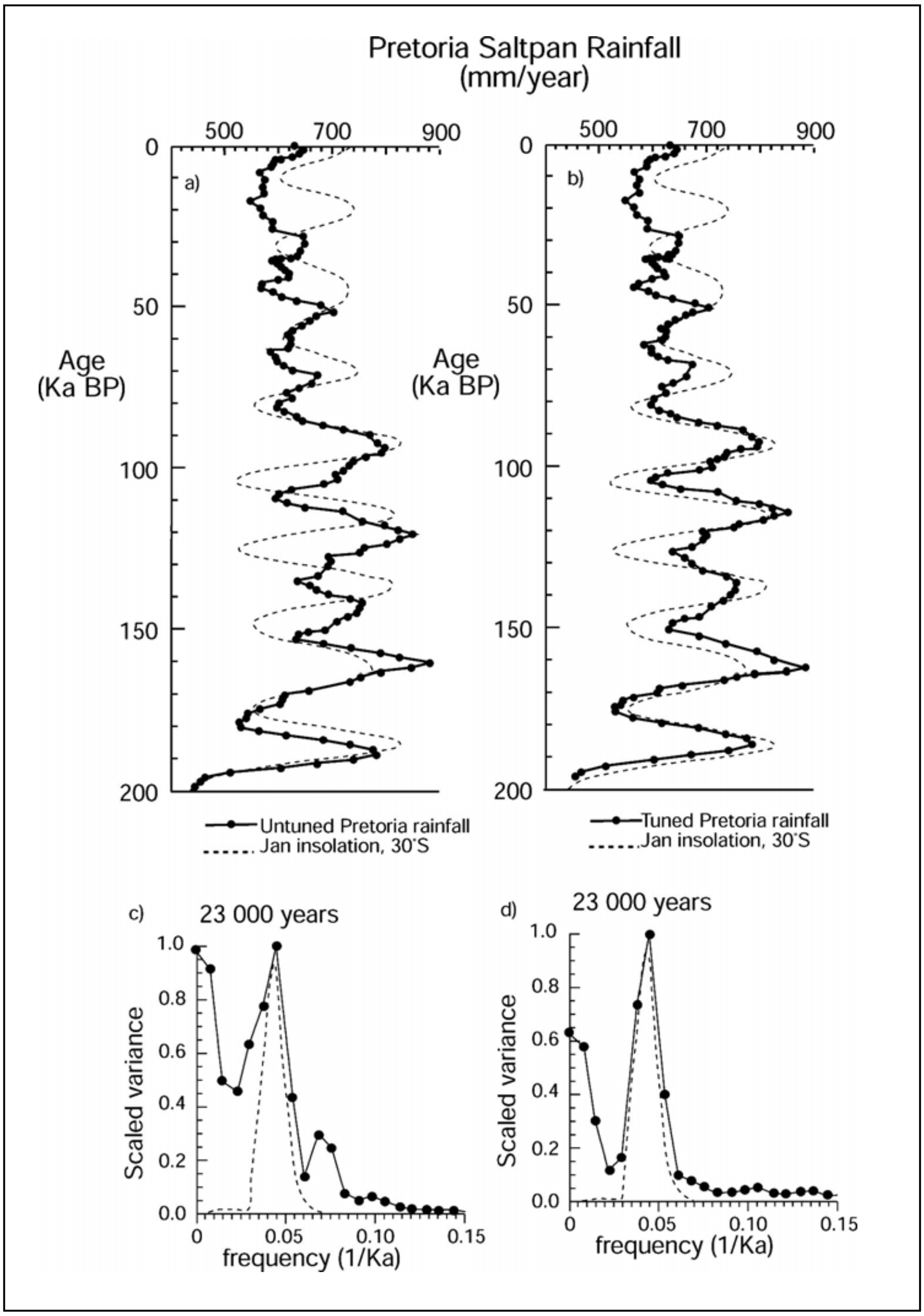

Figure 1.19. Untuned (left) and tuned (right) rainfall records from the Tswaing (Pretoria) Impact Crater (top), with frequency histograms showing 23 kyr cyclicity (Partridge *et al*., 1997).

Resent research in the colluvial badlands, at a site called Voordrag, in northern KwaZulu-Natal, by Clarke *et al*. (2003) provides a 100 kyr record of hillslope evolution. A sequence of sandy colluvium, interbedded with organic rich paleosols has been dated via Infrared Stimulated Luminescence (IRSL) and ^{14}C respectively. The two dating techniques show good concordance, and provide dates ranging from ~96 to ~13 kyr. Clarke *et al*. (2003) argue that, when compared to other palaeoenvironmental records (namely the Tswaing Crater), the sandy colluvial deposits appear to correspond to arid phases, while the paleosol layers repeatedly occur at intervals of hillslope stability, which may reflect periods of increased humidity (Figure 1.20). Both paleosols and colluvium deposits are recorded during the Last Glacial Maximum (LGM) and the warm Early Holocene, which Clarke *et al*. (2003) use to argue that the Voordrag sequence is not controlled by temperature. They suggest, instead, that the sequence is controlled by changes in regional precipitation, which may also reflect enhanced seasonality. The mixed signal at the LGM (OIS 2) suggests climatic fluctuations during this time, while the paleosols P4 and P5 (figure 1.20) may be related to periods of increased precipitation and vegetational cover during late OIS 4 and in OIS 3 (Clarke *et al*., 2003).

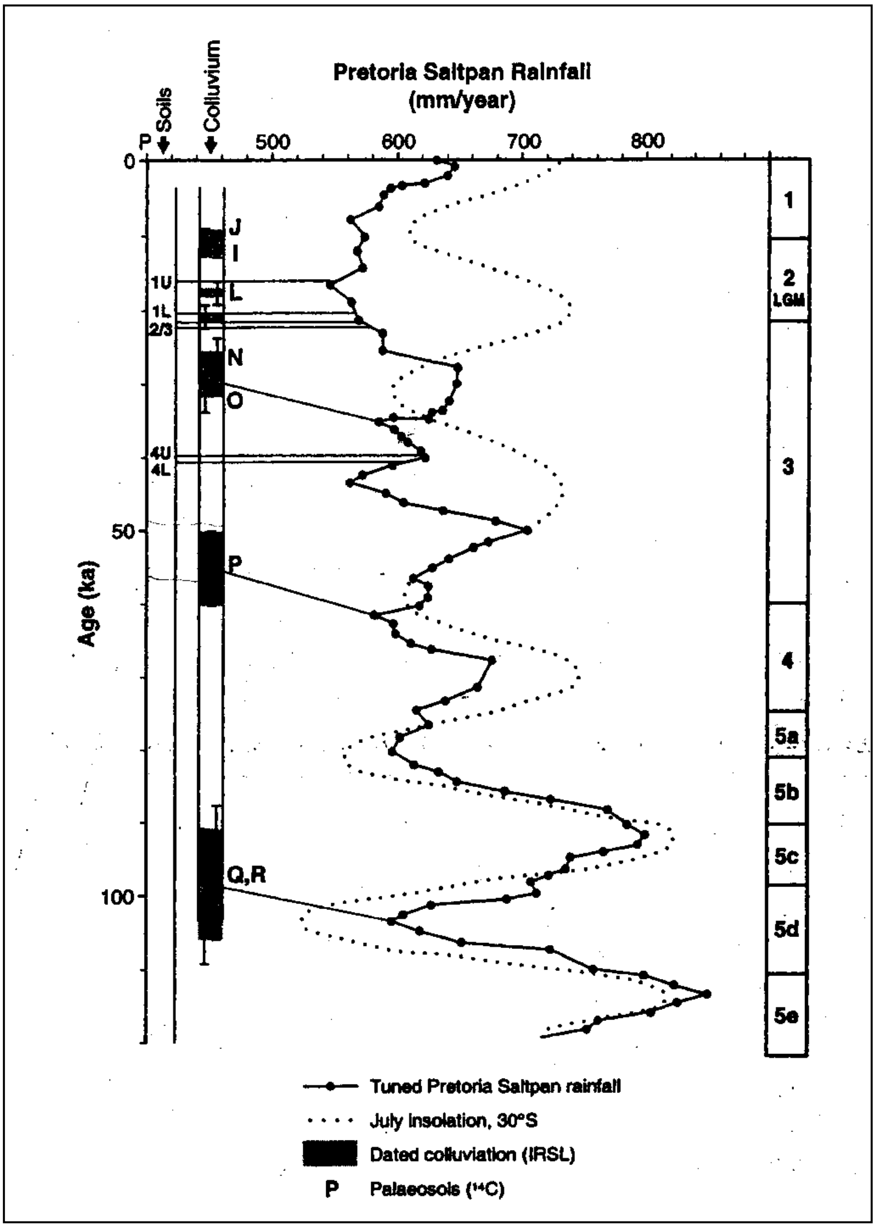

Figure 1.20. Dated colluvium and paleosols from the Voordrag sequence compared to the precipitation record from the Tswaing Impact Crater (From: Clarke *et al*., 2003).

Another potential source of palaeoclimatic data from alluvial deposits is from the ongoing research in the semi-arid Steelpoort region, Mpumalanga (Hancox *et al*., 2004). Here alluvial fans cover many of the mountain footslopes, and grade downslope into the terraces of major rivers, and are extensively dissected by 'dongas' (gullies), revealing successions of clay-rich paleosols intercalated with, and sometimes cross-cut by, gravelly sands. There is no regional evidence for tectonic control of these alluvial fans, suggesting instead that these deposits were climatically controlled. As with the Voordag sequence, the sandy units are thought to have accumulated during more arid climate regimes, with paleosols forming during more humid periods. The sandy units are preliminarily dated via TL (Thermal Luminescence), showing that three main episodes of sedimentation have occurred over the last 125 kyr (at ~ 13 ka, ~40 ka and ~ 125 ka) (Hancox *et al*., 2004).

The north-eastern KwaZulu-Natal coast of South Africa is host to a series of dune cordons, which have formed during the Quaternary, and are another potential source of palaeoclimatic data. Sudan *et al*. (2004) have described the Quaternary evolution of these coastal dunes between Lake Hlabane and Cape St Lucia, using the mineralogy, geochemistry and textural properties of the dunes, their spatial distribution, and luminescence dating. Sudan *et al*. (2004) recognise three generations of dunes, making up the dune complex, which range in age from at least 200 kyr to the present. Periods of dune accretion are correlated with high sea-levels. The oldest sediment package is correlated with the Kosi Bay Formation and may be of Mid Pleistocene age, or older. The middle package of dunes is attributed both to the penultimate glacial, OIS 7 or older, and the last interglacial, OIS 5. The youngest dune formation took place during the Holocene sea-level high (Sudan *et al*., 2004).

Other African records of more recent Holocene climate change are preserved in the ice caps of Mount Kenya and Mount Kilimanjaro in East Africa. These ice caps have been cored and analysed for oxygen isotopes, dust and iron content, as proxies for climatic change (Thompson *et al.*, 2002). The six cores from Mt Kilimanjaro provide a record of environmental variability for the equatorial region of Africa from ~11.7 kyr to the present. Within the record are three major periods of abrupt climatic change documented at ~8.3 ka, ~5.2 ka and ~4 ka.

Long, continuous, high resolution, well dated terrestrial climate change records for the interior of South Africa are rare, as this review has shown. However, with the emergence and wider application of dating techniques such as U-series and luminescence, well dated records for the last ~50 kyr are becoming available. There is good concordance between the various climate change records, especially during the last 20 kyr and in the Holocene. Records back to 200 kyr and beyond are still rare and suffer from poor dating control. The younger fill at Gladysvale is preliminarily dated to 200- 250 kyr via ESR (Curnoe, 1999; Schmid, 2002) and it is hoped that, through the U-series dating and stable light isotope analysis of these strata, this study may add to our understanding of the palaeoclimate in the interior of South Africa.

1.3 Aims

This study is structured around the following three aims:

1. To fully document the three-dimensional sedimentological architecture of the upper cave fill fan at Gladysvale, in terms of its surface morphology and internal architecture, including the proximal to distal changes within the sedimentary body, and the relationship between the fan and the available accommodation space. The deposit in the upper chamber at Gladysvale Cave is ideal for this purpose as it is well exposed from the blasting and trenching activity of lime miners in early part of last century.

2. To date the intercalated flowstones layers of the ~250 ka fill at Gladysvale via Uranium-series dating, and thereby test the chronostratigraphic models of Ayliffe *et al*. (1998) and Moriarty *et al.* (2000) in a South African cave context.

3. To investigate the palaeoclimatic controls on the nature and rate of sedimentation through Stable Light Isotope ($\delta^{18}O$ and $\delta^{13}C$) analysis of the flowstones and sediments, to compare these data, and the newly acquired U-series ages for the flowstone layers, with known records of climate change on a global and regional scale.