A new perspective on the geohydrological and surface processes controlling the depositional environment at the Florisbad archaeozoological site

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DECLARATION

I declare that the thesis hereby submitted by me for the Philosophiae Doctor degree at the University of the Free State is my own independent work and has not previously been submitted by me to another University. Where use has been made of the work, or assistance, of others, this has been duly acknowledged in the text. I further more cede copyright of the thesis in favour of the University of the Free State.

DEDICATION

Dedicated to my parents, Ed and Kath who, from beyond, gave to me an awareness of that of which I was unaware, as well as the insight, inspiration, and guidance to embark upon this project; and to both hominid and beast, who were sustained over the millennia by the spring and its unique formation, and who in their passing, ultimately created the opportunity for this research.

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ABSTRACT

The Florisbad Quaternary Research Station and archaeozoological site is located 45 km north-west of Bloemfontein, Free State Province, South Africa (28° 46` 05.4"S, 26° 04` 10.7"E), and is sited around a series of highly saline, warm water spring vents. The site is partially covered by a large sand dune. The site is significant for three important reasons. Firstly, the discovery of the Florisbad skull (*Homo helmei*) in 1932 by Prof. T. Dreyer, secondly, a collection of faunal fossil remains representing at least 31 taxa, including extinct and extant species, and referred to as the Old Collection and, thirdly, a Middle Stone Age (MSA) human occupation horizon representing a temporary butchering site with evidence of a hearth, butchering tools, and faunal fossil remains.

Spring- and excavation pit water samples were taken and analysed in 1988 during a high rainfall period, and in 1999 during an average rainfall period. In relation to the spring water, the results show that the total dissolved solids (TDS) of the excavation pit water were, in relation to the spring water, higher during the high rainfall period and lower during the average rainfall period. This was contrary to the norm, where it is expected that high rainfall periods should produce a decrease in TDS due to a dilution effect. The TDS of the spring-water remained stable throughout both high and average rainfall periods. Further analysis showed considerable TDS increases between the excavation pit waters, and between the pit waters and the spring-water. It is concluded that the pit waters were not directly related to the spring water and that the two water bodies were separate entities with the pit water being recognized as groundwater. An analysis of rainfall in relation to the TDS of the spring- and groundwater indicated that short-term rainfall affected the quality of the groundwater, but not the quality of the spring-water, while long-term rainfall had little effect on the quality of the spring-water.

The question arose as to why the TDS of the groundwater was so much higher than that of the spring-water, and what factors were causing these differences? Organic-clay (peat) samples from the walls of the excavation pits as well as the walls of the open excavation area were analysed. The results of the analyses, and an examination of the stratigraphy, strongly suggested that minerals had accumulated in the organic-clay layers due to

organic matter having a similar colloidal organization to that of clay, with the ability to adsorb large quantities of minerals on their outer surfaces. A comparison of the groundwater and organic-clay analyses results showed that the TDS of the decomposed Peat II organic-clay layer was considerably higher than that of the groundwater, with the same being true for the far less decomposed Peat IV organic-clay layer. By analysing and combing the water and organic-clay layer results with the many factors, mechanisms, and processes involved, it is concluded that the salinization of the organic-clay layers, and the flushing of ions from the organic-clay layers by percolating water during rainfall periods, is responsible for the increased mineralization of the groundwater. Other factors, mechanisms, and processes, such as rainfall, aeolian deposition, evaporation, capillarity, wind, temperature, matrix-suction, pH, Eh, PCO_2, PO_2, DOC, and biomineralization, all of which support the accumulation of free salts in a semi-arid environment such as Florisbad, were also investigated.

Of primary importance was the question as to whether the spring-water was actually responsible for fossilization of the faunal remains, and could fossilization have taken place within the environs of the spring vents, or in the spring vents themselves? Previous research has suggested that the spring-water was calcium-carbonate rich, with evidence of calcium-carbonate deposition further suggesting that faunal remains of the Old Collection must have been in contact with the spring-water in spring vents for some time. An analysis of the spring-water analysed over the past 84 years indicated that there had never been sufficient Ca (under-saturation) in the spring-water for fossilization to occur, and this is confirmed by the current analyses. The contemporary lack of Ca in the spring-water, combined with other environmental factors within the environs of spring vents, such as the lack of organic matter and clay, combined with a high Eh environment, also strongly indicated that, historically, fossilization could not have taken place within the environs of the springs. Contrary to earlier hypotheses, it is concluded that the spring water and spring flow would directly assist in the de-mineralization of faunal remains.

A detailed investigation of the site, along with an analysis of the stratigraphy and sedimentation, revealed that previous theories on the formation of the site did not sufficiently accommodate the current stratigraphy in the context of the organic-clay layers, the salinization process, and fossilization. From this deduction all the existing and pre-existing evidence was revisited in an attempt to provide a hypothesis which would accommodate the existing morphology of the site, sedimentation, and fossilization. It is hypothesised that the spring site formed around a large drainage-impeded pan which was largely covered by a sand dune that had migrated from the area of the extensive salt pan to the north and north-west (Soutpan). The arms of the dune eventually came to rest up against the windward slope of a dune belt located just south of the spring site, and a dam began to form. High rainfall periods produced organic-clay layers, while sandy layers were produced during drier windy periods. This led to the formation of alternating horizontal layers of organic-clay and sand, eventually building up to almost the top of the sand dune on the leeward face. When the water level in the dam reached the top of the arms of the sand dune, it broke through the eastern arm. The dam water and sediments then evacuated the dam in a flash flood. This flash flood eroded the area to the east of the site to such an extent that the drainage was diverted, and a wide flat-bottomed vlei was formed where much of the dam sediments were deposited. This hypothesis provides an alternative for the formation of the spring site, accommodating all aspects of sedimentation, salinization, and fossilization.

The dating of the Florisbad deposits and fossils has been subject to an ongoing debate since the first ^{14}C dating was carried out in 1954. The ages and depths of recently published profiles did not appear to correspond to the assumption of greater compaction with depth and time. In an attempt to resolve this issue, linear, exponential, and logarithmic mathematical trend lines were then experimentally applied to the published profiles of electron spin resonance (ESR) and optical stimulated luminescence (OSL) dates in order to test the theory of compaction, and to validate the results. The hypothetical effect of manipulating ages on trend lines was also tested. A discussion on some possible shortfalls regarding the dating methods used is undertaken.

A best logarithmic fit to data was obtained by holding the ESR Middle Stone Age Human Occupation Horizon (MSA) age at 127 ka, and advancing the lower deposit age from 250 ka to 420 ka. The next best fit to data occurred by regressing the ESR MSA age from 127 ka to 78 ka, and holding the lower deposit age at 250 ka. The application of exponential and linear trend lines produced poor fits to data. A suggested compaction trend line was also introduced which produced an ESR MSA age of 75 ka and a lower deposit age of 384 ka. In the final analysis, trend line results suggested an MSA age of 92 ±12 kyr and a basal deposit age of 400 ±20 ka. The logarithmic and suggested compaction trend line ages for the lower deposits both produced ages similar to the suggested Florisain – Cornelian faunal boundary of *c*. 400 ka. The exercise confirmed that the ages in the published profiles were disjunct and that this disjunction may be related to a number of different physical forces.

KEY WORDS

Key Words: Florisbad, Archaeozoological site, Spring-water, Groundwater, Organic-clay layers, Salinization, Fossilization, Chemistry, Geohydrology, Geology, Depositional environment, Formation of site, Dating.

OPSOMMING

Die Florisbad Kwaternêre Navorsingstasie en argeosoölogiese terrein is 45 km noordwes van Bloemfontein, Vrystaat Provinsie, Suid-Afrika geleë (28° 46` 05.4"S, 26° 04` 10.7"E) langs 'n reeks uiters sout- en swaelryke warmwater fonteine. Die terrein is gedeeltelik bedek deur 'n groot sandduin. Die terrein is betekenisvol vanwee drie belangrike redes: Eerstens die ontdekking van die Florisbad-skedel deur prof. T. Dreyer in 1932, tweedens `n versameling van dierlike fossieloorblysfsels wat ten minste 31 taksa van uitgestorwe en bestaande spesies verteenwoordig (bekend as die Ou Versameling), en derdens 'n menslike bewoningshorison uit die middel-Steentydperk (MST) wat 'n tydelike slagplaas, tekens van 'n vuurherd, slaggereedskap en dierlike fossieloorblyfsels toon.

Monsters van die fonteinwater en sypelwater in die uitgrawings is in 1988 gedurende 'n tydperk van hoë reënval, en in 1999 gedurende 'n tydperk van gemiddelde reënval, geneem en ontleed. Die resultate dui aan dat die totale opgeloste vastestowwe (TOV) van die sypelwater in die uitgrawings, in vergelyking met dié van die fonteinwater, hoër was gedurende die nat periode en laer gedurende die gemiddelde reënvalperiode. Dit is teenstrydig met die norm wat sou verwag dat hoë reënvalperiodes 'n afname in TOV sal oplewer as gevolg van 'n verdunningseffek. Die TOV van die fonteinwater het stabiel gebly deur beide die hoë en gemiddelde reënvaltydperke. Verdere ontleding het 'n aansienlike TOV toename tussen die sypelwater van verskillende uitgrawings en tussen die sypelwater en fonteinwater getoon. Die gevolgtrekking is dat sypelwater in uitgrawings nie direk verband hou met fonteinwater nie en dat die twee waterliggame aparte entiteite is waarvan die sypelwater as grondwater beskou kan word. 'n Analise van reënval in verhouding tot die TOV van die fontein- en grondwater dui aan dat korttermyn reënval die kwaliteit van die grondwater beïnvloed maar nie dié van die fonteinwater nie. Langtermyn reënval het weinig invloed op die kwaliteit van fonteinwater.

Die vraag het ontstaan waarom die TOV van die grondwater soveel hoër is as dié van die fonteinwater en watter faktore hierdie verskille veroorsaak. Organiese klei (veen) monsters van die kante van uitgrawings asook die kante van die oop uitgrawing is geanaliseer. Die resultate van die analises en 'n ondersoek van die stratigrafie het sterk aanduidings getoon dat minerale in die organiese kleilae versamel het, as gevolg van die feit dat organiese materiaal dieselfde kolloïdale struktuur as klei het en die vermoë besit om groot hoeveelhede minerale in hulle buitenste lae te adsorbeer. 'n Vergelyking tussen die resultate van die grondwater- en organiese klei analise het getoon dat die TOV van die ontbinde Veen II organiese kleilaag aansienlik hoër was as dié van die grondwater, terwyl dieselfde vir die veel minder ontbinde Veen IV organiese kleilaag geld. Deur die resultate van die water en organiese kleilaag te vergelyk met die baie faktore, meganismes en prosesse betrokke, word die gevolgtrekking gemaak dat versouting van die organiese kleilae en die loging van ione uit die organiese kleilae deur sypelwater gedurende reënval periodes, verantwoordelik is vir die toenemende mineralisasie van die grondwater. Ander faktore, meganismes en prosesse soos reënval, aeoliese neersetting, verdamping, kapillariteit, wind, temperatuur, matriks-suiging, pH, Eh, PCO_2, PO_2, DOC en biomineralisasie wat almal bydra tot die opeenhoping van vry soute in 'n semi-ariede omgewing soos Florisbad, is ook ondersoek.

Van primêre belang was die vraag of fonteinwater eintlik verantwoordelik was vir fossilering van die dierlike oorblyfsels en of fossilering in die omgewing van fonteine of in die fonteine self kon plaasvind. Vorige navorsing het daarop gedui dat fonteinwater kalsiumkarbonaatryk was, met aanduidings van kalsiumkarbonaat afsetting wat verder daarop dui dat dierlike oorblyfsels van die Ou Versameling vir 'n geruime tyd in kontak met die fonteinwater in fonteinne moes gewees het. Wateranalise van die fonteinwater oor die afgelope 84 jaar het aangedui dat daar nog nooit voldoende Ca (onderversadiging) in die fonteinwater was vir fossilering om plaas te vind nie en dit word bevestig deur die huidige analise. Die hedendaagse gebrek aan Ca in die fonteinwater, in kombinasie met ander omgewingsfaktore in die omtrek van fonteine, soos die gebrek aan enige organiese materiaal of klei en 'n hoë Eh omgewing, is 'n sterk aanduiding dat fossielisering nie in die verlede in fonteine kon plaasgevind het nie. In teenstelling met vorige hipoteses word die gevolgtrekking gemaak dat fonteinvloei bydraend is tot die demineralisasie van dierlike oorblyfsels.

‘n Gedetailleerde ondersoek van die terrein, saam met ‘n analise van die stratigrafie en sedimentasie, het aan die lig gebring dat vorige teorieë oor die ontstaan van die terrein nie die huidige stratigrafie ten opsigte van die organiese kleilae, die versoutingsproses en fossilisering, genoegsaam in ag geneem het nie. Met hierdie afleiding in gedagte is al die bestaande en vooraf bestaande getuienis weer nagegaan in ‘n poging om met ‘n hipotese voor ‘n dag te kom wat die bestaande morfologie van die terrein, sedimentasie en fossilering sou kon akkommodeer. Daar word gehipoteseer dat die fontein gevorm het in die omgewing van‘n groot pan met beperkte dreinering. Hierdie pan was grootliks bedek deur ‘n sandduin, wat migreer het van die oorspronklike terrein in ‘n noord- en noordwestelike rigting (Soutpan). Die arms van die duin het uiteindelik tot ruste gekom teen die windkanthang van ‘n duingordel wat net suid van die terrein van die fontein geleë is en het ‘n dam gevorm. Hoë reënval periodes het organiese kleilae gevorm, terwyl sanderige lae gedurende droër winderige periodes gevorm is. Dit het gelei tot die vorming van afwisselnde horisontale lae van organiese klei en sand, wat uiteindelik tot amper by die kruin aan die lykant van die sandduin opgebou het. Die stygende watervlak het deur die oostelike arm van die sandduin gebreek en water en sediment in die dam is d.m.v. ‘n blitsvloed gedreineer. Hierdie blitsvloed het die area oos van die terrein tot so ‘n mate geërodeer dat die dreinering herlei is en ‘n wye vlei gevorm het waar baie van die sedimente van die dam gedeponeer is. Hierdie hipotese verskaf ‘n alternatiewe verklaring vir die vorming van die terrein om die fonteine en sluit alle aspekte van sedimentasie, versouting en fossilering in.

Die ouderdom van die Florisbad afsettings en fossiele is sedert die eerste ^{14}C ouderdomsbepaling gedoen in 1954, onderworpe aan ‘n voortgesette debat. Die ouderdomme en dieptes van onlangs gepubliseerde profiele het skynbaar nie ooreengestem met die aanname van hoër kompaksie met diepte en tyd. In ‘n poging om hierdie kwelvraag op te los, is lineêre, eksponensiële en logaritmiese wiskundige tendenskrommes op die gepubliseerde profiele van ESR en opties gestimuleerde

luminessensie (OSL) ouderdomme gebruik om die teorie van kompaksie te toets en die resultate daarvan te bekragtig. Die hipotetiese effek van die manipulering van ouderdomme op tendenskrommes is ook getoets. Die moontlike tekortkominge van die dateringsmetodes wat gebruik is, word ook bespreek.

'n Beste logaritmiese datapassing is verkry deur die ESR Middel Steentydperk Menslike Bewoningshorison (MST) ouderdom van 127 ka konstant te hou en die laer afsettingsouderdom van 250 ka na 420 ka te verander. Die volgende beste datapassing is verkry deur die ESR MST ouderdom van 127 ka na 78 ka terug te skuif en die laer afsettingsouderdom op 250 ka konstant te hou. Die aanwending van eksponensiële en lineêre tendenskrommes het swak datapassings opgelewer. 'n Voorgestelde kompaksie-tendenskromme is ook toegepas. Dit het 'n ESR MST ouderdom van 75 ka en 'n laer afsettingsouderdom van 384 ka opgelewer. In die finale analise het die tendenskromme resultate 'n MST ouderdom van 92 ±12 ka en 'n basale afsettingsouderdom van 400 ±20 ka voorgestel. Die logaritmiese en voorgestelde kompaksie tendenskromme ouderdomme vir die laer afsettings het beide ouderdomme opgelewer soortgelyk aan die voorgestelde Florisiaans – Corneliaanse fauna grens van *c*. 400 ka. Die oefening het bevestig dat die ouderdomme in die gepubliseerde profiele disjunk was en dat hierdie disjunksie verwant kan wees aan verskilende fisiese kragte.

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CHAPTER 1

INTRODUCTION

1.1 A HISTORICAL OVERVIEW OF THE FLORISBAD SITE

Florisbad Quaternary Research Station, 28° 46` 05.4"S, 26° 04' 10.7"E, (spring eyes) is an important archaeozoological site located 45 km north-west of Bloemfontein, Free State Province, South Africa. A large salt pan, Soutpan, located to the north of Florisbad, dominates the local environment, as illustrated in the Google Earth image of the area (Figure 1). The significance of Florisbad is threefold. Firstly, the discovery of the Florisbad hominid cranium (*Homo helmei*) by Prof. T. Dreyer in 1932, its subsequent description in 1935 (Dreyer, 1935), and the recent dating, based on a fragment of a tooth, to 259 ±35 ka by electron spin resonance (ESR) (Grün *et al*., 1996), brought recognition to Florisbad. Secondly, there is the existence of a collection of artefacts and vast number of faunal fossil remains representing 26 species, which is referred to as the Old Collection (Brink, 1987). The Old Collection represents the Florisian Land Mammal Age (Hendey, 1974), or the Florisian-Cornelian Faunal Boundary, with an age of *c*. 400 ka (Klein, 1984). Thirdly, a Middle Stone Age (MSA) human occupation horizon, representing a temporary butchering site, at which butchering tools and faunal fossil remains have been excavated and identified (Brink, 1987; Henderson, 2001a; Brink and Henderson, 2001).

Very little is known about the modern day habitation of Florisbad prior to the settlement of one Hendrik Venter and his family on the farm Rietfontein, and the adjoining farm Jackalsfontein, before the farms were purchased from the Land Commission in 1849 (Henderson, 1995). Rietfontein is the farm on which Florisbad is situated. Because of the interest in, and collection of, early fossils by Martha Venter, Hendrik's wife, many of which formed part of the initial collection donated to the National Museum.

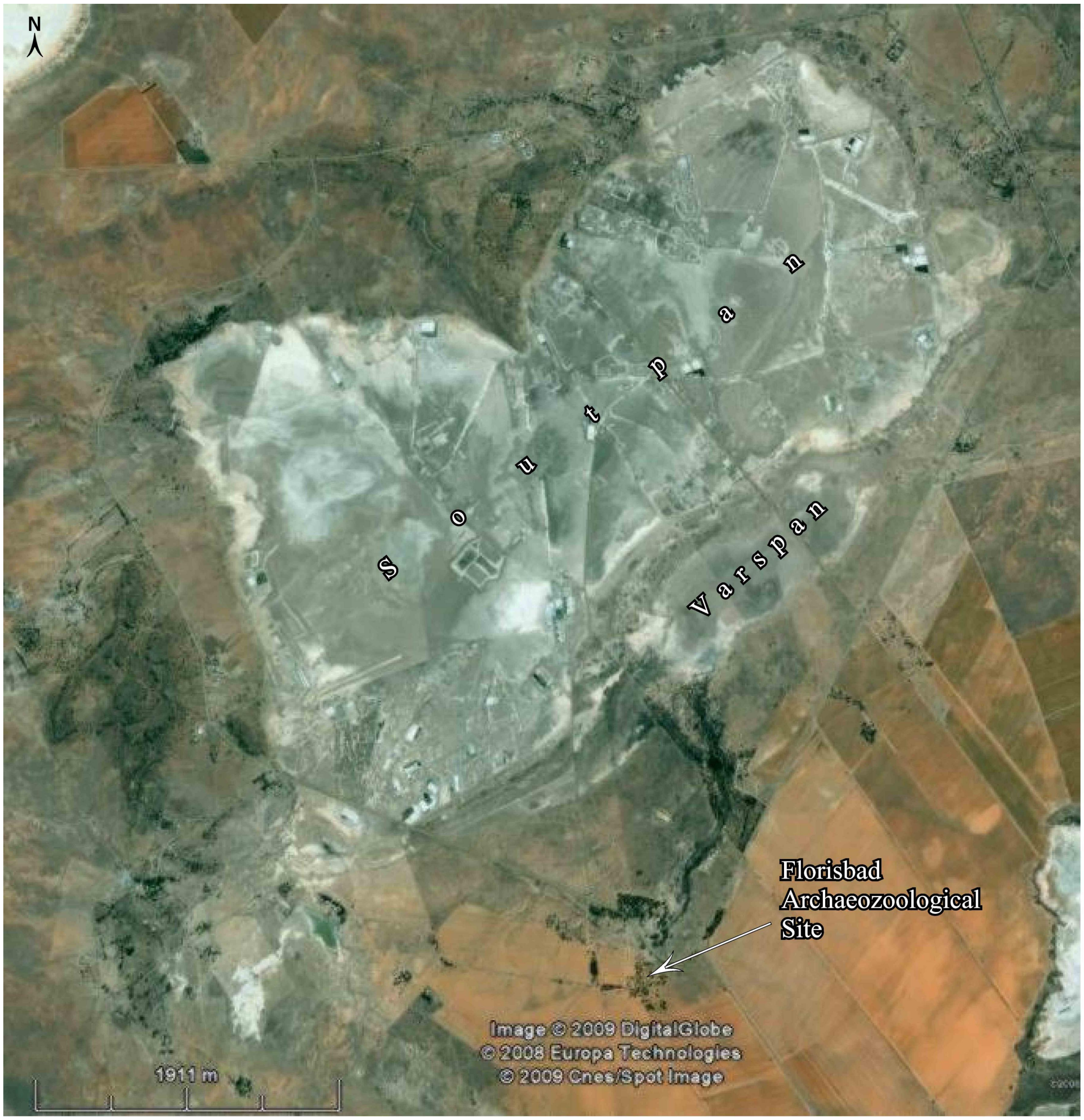


Figure 1. A Google Earth satellite image of the Florisbad area showing Florisbad in relation to Soutpan and the large areas of red aeolian sand deposition in the surrounding areas. Note the evaporative salt industry that has established itself on the pan.

A brief history of the farm, and its owners, is given, based on Henderson, (1995), unless otherwise stipulated. Prior to their settlement at Rietfontein, Hendrik and Martha Venter (née Coetzer) had a son Floris Johannes, born in 1836, who married Renske du Plooy. Renske gave birth to a number of children, including a son born in the Bethulie district in 1869, also named Floris Johannes, after whom Florisbad was later named. Part of Jackalsfontein, was registered in Renske's name in 1886, and all of Rietfontein in 1894. When Hendrik Venter died in 1899, Renske remarried one Gert Abraham Coetzee. In 1917 Renske divided the farms among her children, with the part of Rietfontein, which included the springs, going to Floris Johannes (Hendrik Venter's grandson), This portion was registered as Florisbad, with word of the mineral springs healing properties spreading far and wide, and Floris Johannes earning the reputation of the "healer" (Anon [1], 1980). The springs became known as "Floris se bad", hence the name Florisbad (Anon [1], 1980). Floris married Martha Johanna Breed and they had four children. Later Gert Abraham sold his share in the baths to a niece and her husband, who in turn were bought out by the brothers, Edward and John Sowden. The State bought out 93 ha of the farm, including the springs in 1980, and handed over custody of the site to the National Museum in 1981.

In 1912, due to the increase in the numbers of persons wanting to use the spring, Floris built a small house over the spring eye in order to provide some protection and privacy for bathers. While increasing the size of the swimming bath, a number of fossils were unearthed. The same year, a strong earthquake occurred, with its epicentre near Koffiefontein, 125 km southwest of Florisbad, and having an intensity of VIII on the Mercalli scale (Fernàndez and Guzmàn, 1979). This event has often mistakenly been reported in the literature as having occurred at Fauresmith (Anon [1], 1980). The earthquake reportedly resulted in a new spring eye erupting in the excavations due to a build-up of gasses (Grobler and Loock, 1988a), throwing up many fossils and artefacts (Anon [1], 1980). Martha Venter made a large collection of these fossils, which were later donated to the National Museum by her son, Gert.

1.2 A BRIEF DESCRIPTION OF THE FLORISBAD AREA

In general, the Florisbad area is undulating, broken occasionally by dolerite dykes which have formed low hills. The area is underlain by Ecca Group rocks of the Karoo Supergroup, with no evidence of the rocks outcropping in the immediate area which is partially due to the area being blanketed by red and yellow aeolian sands.. A large salt pan, Soutpan (Figure 1), dominates the area and has had a considerable effect on the formation of the Florisbad spring site in that brackish wind blown material, deflated from the pan, has contributed to the Florisbad sand dune and the site as a whole. A belt of san dunes also occurs south east of Florisbad indicating that aeolian deposition was wide spread. Other smaller salt pans are visible in the top left hand corner and the bottom right hand corner of the image. Figure 2 presents a plan of the Florisbad farm showing the spring and excavation sites in relation to Soutpan as well as other key features.

Figure 3 is an aerial photograph of the Florisbad archaeozoological site indicating a number of key features referred to in the text. The Florisbad archaeozoological site and excavations are centred in the area of a number of warm water spring eyes, around which the swimming pools were built. A number of buildings have been removed over the years in order to expand the excavations. The MSA Human Occupation Horizon excavations are located within, but to the right of, the main excavation area (Figure 3).

Figure 4 is a simplified interpretation of a section of the Florisbad sedimentary deposits as they are today. The original Florisbad spring site was a large pan with water supplied mainly from the spring eyes, which surfaced along a dolerite dyke. It was in this pan that faunal remains were fossilized in association with organic-clay (Peat) sediments which developed at the bottom of the pan. Fossil remains from the lower sediments are referred to as the Old Collection. During the late Middle Pliocene a sand dune developed on the windward side of Soutpan to the north-west, growing in size as it migrated towards the spring site. Eventually its migration was restricted by a previously existing row of largely static sand dunes lying to the south east of Florisbad. This resulted in a dam being

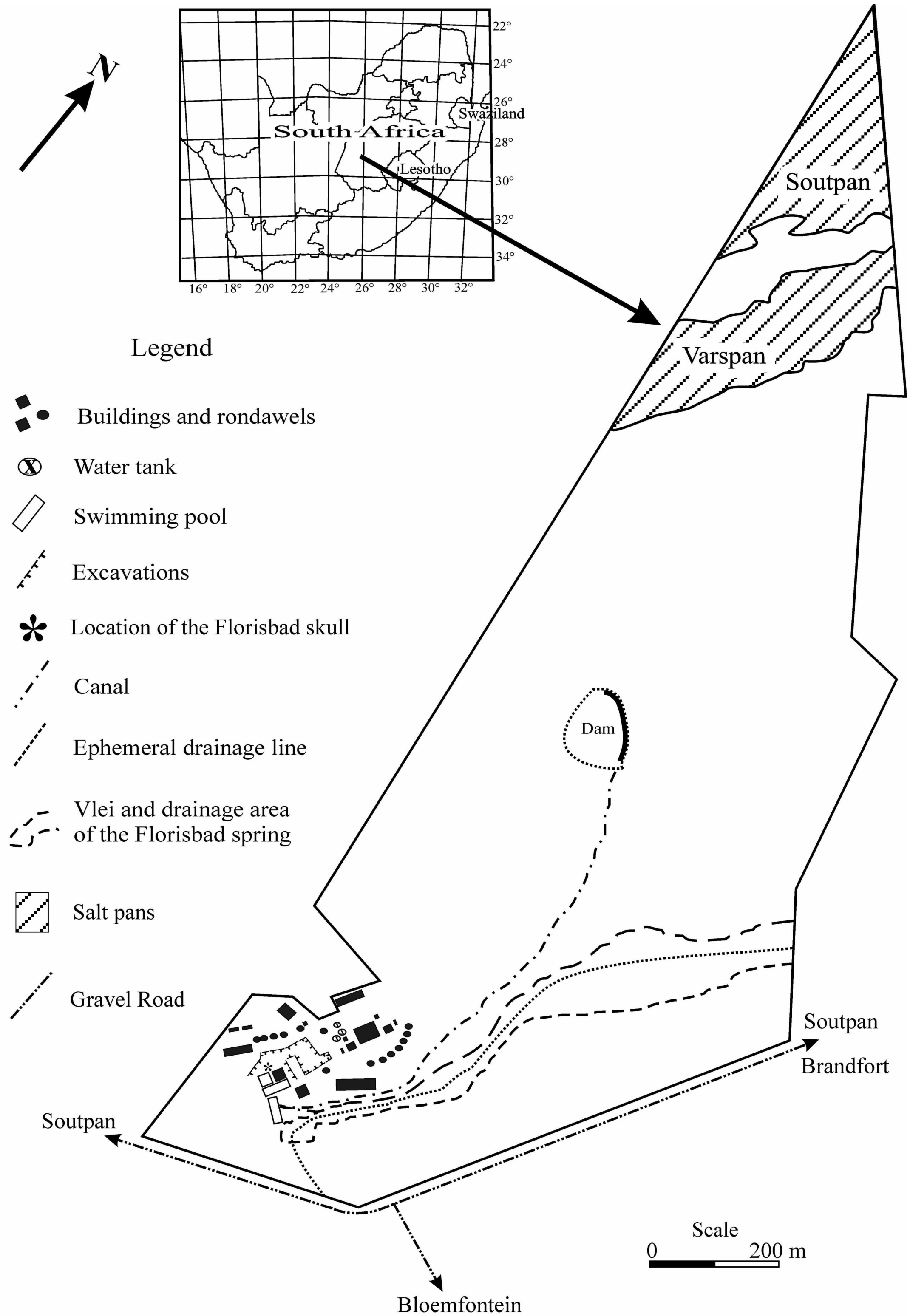


Figure 2. Plan of the Florisbad Farm (after Douglas, 1992).

Figure 3. An aerial photograph of the Florisbad spring site indicating some key features (Photo: National Museum, Bloemfontein)

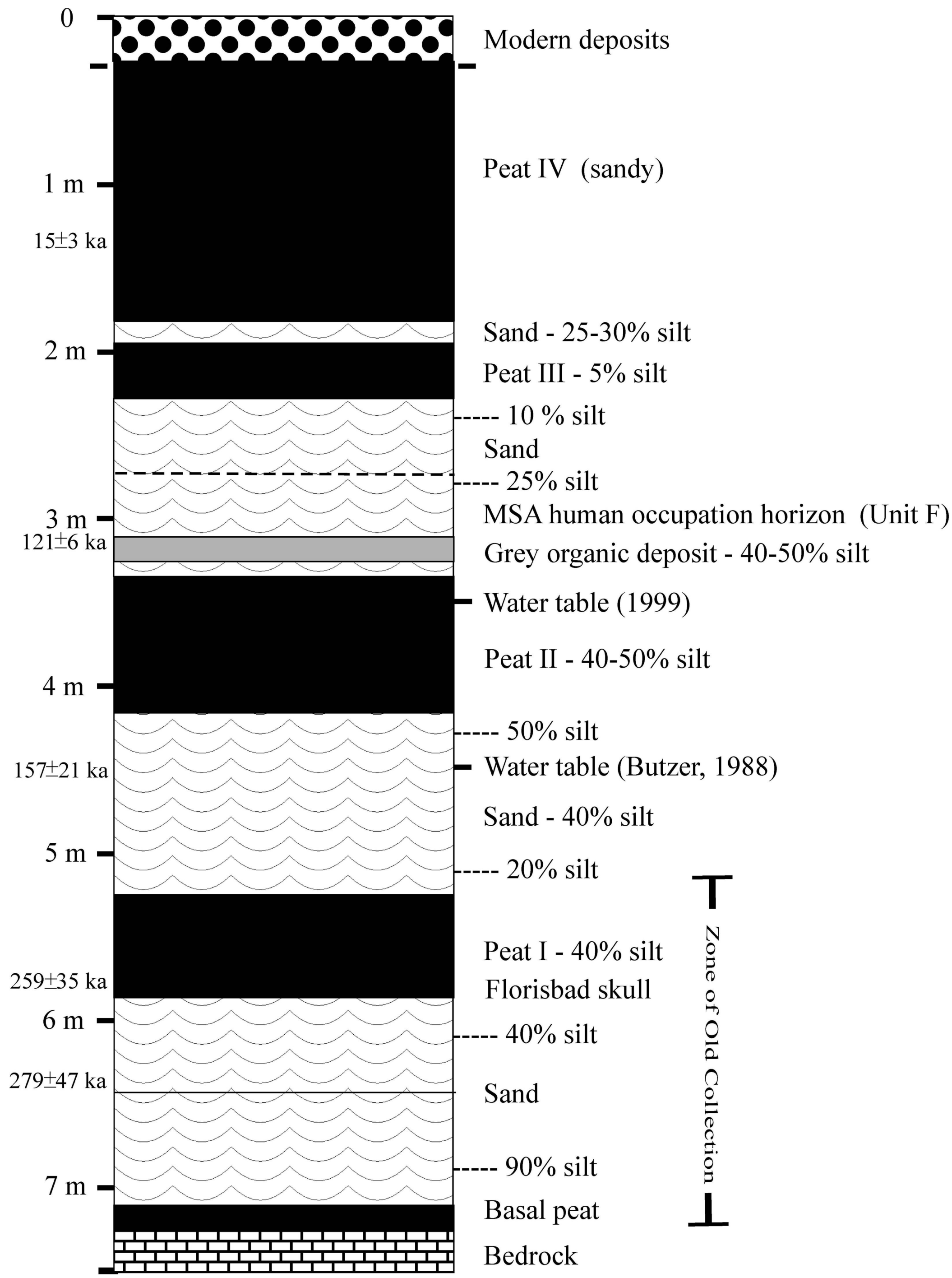


Figure 4. A simplified schematic cross-section of the Florisbad sedimentary deposits, showing the major organic-clay (peat) and sand horizons with their silt content (after Kuman *et al*., 1999 and Appenix IV). Dates in ka after Grün *et al*. (1996)

formed between the leeward face of the Florisbad sand dune and the windward face of the static dunes to the south-east. High rainfall periods produced organic-clay layers while sandy layers were produced during drier windy periods. This then led to the alternating horizontal layers of organic-clay and sand (Figure 4) building up to almost the top of the dune. This stratigraphy is not part of the sand dune formation, but a separate unit. When the water in the dam neared the top of the eastern arm of the sand dune, it broke through the eastern arm, resulting in a flash flood which gouged out a wide vlei to the north-east of the site, where much of the evacuated dam sediments were deposited (Figure 3). The flash flood also left a large eroded area between the spring site and the base of the south-east dune belt, a large part of which was the original dam floor.

This brief description of the Florisbad site is based entirely on the hypotheses presented in this study. More detailed descriptions of the geomorphology and geology are given in Chapter 2, while the geohydrology, sedimentation, and previous hypotheses on the formation of the site are to found in Chapter 5.

1.3 PREVIOUS SCIENTIFIC RESEARCH AT FLORISBAD

Over the years many and diverse scientific studies have been initiated at Florisbad, many of them unrelated to the fossil or archaeological material. Table 1 provides a summary of a large portion of this work

The first scientific interest shown in the Florisbad fossils was by Dr Robert Broom who published his findings in the *Annals of the South African Museum* (Broom, 1913). A more intensive research programme was initiated by Prof. T.F. Dreyer of the Grey University College, later to become the University of the Free State, at the request of the South African Museum in Cape Town. Prof. Dreyer and Captain R. Egerton Helme (Anon [1], 1980), who assisted Dreyer with funding during the 1920s (Henderson, 1992), made a small excavation at the side of the pool in 1927, and in 1928, Prof Dreyer and Miss A. Lyle carried out further excavations, describing these findings in 1931 (Dreyer and Lyle,

Table 1. Examples of previous research carried out at Florisbad Research Station. Subject fields are arranged alphabetically with sources being referenced at the end of the table.

Subject	Category/field	Description of research	References
Acarology		Fossil oribatid (free living) mites with a description of a new species and genus from Florisbad.	23, 24, 25, 26.
Archaeology	Artefacts	Descriptions of the stone implements found at Florisbad, their occurrence and position within the sediments.	12, 13, 19, 37, 53, 55, 59, 60, 61, 68, 69, 74, 89.
	Middle Stone Age	References particular to the MSA period at Florisbad including the artefacts and the MSA human occupation horizon.	13, 16, 51, 52, 59, 60, 61, 67.
Archaeozoology	Taphonomy	Reconstructing of the fossil history record in determining how animals became part of the fossil record.	13, 14.
	Alceloaphini	Revision and discussions on fossil wildebeest and hartebeest from Florisbad.	13, 54, 80, 85.
	Bovidae	Description and discussions on fossil antelope and buffalo at Florisbad.	13, 18, 86, 87.
	Equus (Ainus**)**	Discovery of an ass from southern Africa with range extension.	15.
	Hominidae	These references represent the stages in the classification of the Florisbad skull and the taxonomic conclusions made by the various authors.	22, 34, 35, 36, 37, 38, 39, 40, 41, 46, 69, 70, 78.
	General	Description and discussion on the faunal fossils of Florisbad.	13, 14, 28, 29, 42.
	Giraffidae	Description and discussions on fossil giraffe found at Florisbad – recorded by Cooke (1964) but not by Brink (1987).	13, 29, 79.

(continued…..)

Table 1. (continued)

Subject	Category/field	Description of research	References
	Hippopotamidae	Revision and discussions on fossil hippopotami found at Florisbad.	13, 50, 55.
	Mustelidae	Revision and discussions on fossil otters found at Florisbad	44.
	Perissodactyla	Revision and discussions on fossil zebras found at Florisbad.	27.
	Suidae	Revision and discussions on fossil pigs found at Florisbad.	43.
Dating		Dating of the sedimentary deposits and fossils. A list of ages and methods employed are presented in Table 4.	7, 11, 12, 17, 30, 49, 60, 62.
Geology and Geomorphology	General	Discussions on the geology of Florisbad including minerals found in the sediments as well as the geology and geomorphology of the surrounding areas.	31, 45, 47, 48, 56, 58, 59, 61, 63.
	Lacustrine deposits	Theories on sediment deposition at Florisbad by inundation (flooding) of the palaeo-lake (Soutpan).	56, 57, 59, 84.
	Liquefacation	Evidence of earthquake induced liquefacation in the sediments.	83.
	Lithostratigraphy and sedimentary deposits	Discussions on the stratigraphy of the sediments as well as various cross-sections of different aspects of the deposits and drilling results.	2, 4, 20, 39, 56, 60, 67, 73, 81, 82.
Entomology	Arachnidae	Composition of surface-active spiders determined by pitfall trapping.	64.
	Coleoptera	Seasonal composition of beetles determined by pitfall trapping.	65
Formation		Various theories on the formation of the Florisbad spring site and the Florisbad sand dune.	2, 3, 13, 19, 20, 39, 39, 59, 60, 61.

(continued…..)

Table 1. (continued)

Subject	Category/field	Description of research	References
Groundwater	Salinization	Description and analysis of the processes of ion enrichment of the clay and organic fractions within the sediments.	2.
Herpetofauna		Composition of reptiles and amphibians at Florisbad and surrounding areas determined by pitfall traps, funnel traps and other methods, including a biogeography.	8, 9, 10, 32, 33.
Ornithology		Diets of various birds relating to reptile and mammal prey.	5, 8, 9, 10.
Palaeoenvironment		Discussions on the palaeoenvironments that existed at Florisbad.	13, 19, 60, 61, 75, 76, 77.
Palynology	Coprolites	Preservation and interpretation of pollen in fossilized hyaena dung.	75, 76, 77.
	Pollen	Discussions on the role of pollen and plants in determining the palaeoenvironments of Florisbad.	75, 76, 77, 81, 82.
	Phytoliths	Discussions on the occurrence of fossilized pollen grains including those from the teeth of antelope species found at Florisbad.	72, 76.
Spring water		Various analyses of the Florisbad spring-water and discussions on various aspects of the springs.	1, 2, 3, 32, 45, 47, 66, 71.
Wood		Discussions and identification of the only significant piece of wood recovered from the sediments.	6, 21.

(continued…..)

Table 1. (continued)

References

(1) Appendix 1; (2) Appendix 2; (3) Appendix 3; (4) Appendix 4; (5) Avenant (2005); (6) Bamford & Henderson (2003); (7) Barendsen *et al*. (1957); (8) Bates (1988a); (9) Bates, (1988b); (10) Bates *et al*. (1992); (11) Beaumont & Vogel (1972); (12) Beaumont *et al*. (1978); (13) Brink (1987); (14) Brink (1988); (15) Brink (1994): (16) Brink & Henderson (2001); (17) Broeker *et al*. (1956); (18) Broom (1913); (19) Butzer (1984); (20) Butzer (1988); (21) Clark (1955); (22) Clarke (1985); (23) Coetzee (2001): (24) Coetzee (2002); (25) Coetzee (2003); (26) Coetzee and Brink (2003); (27) Cooke (1950); (28) Cooke (1952); (29) Cooke (1964); (30) Deacon (1966); (31) De Bruiyn (1971); (32) Douglas (1992); (33) Douglas (1995): (34) Drennan (1935); (35) Drennan (1937); (36) Dreyer (1935); (37) Dreyer (1936a); (38) Dreyer 1936b); (39) Dreyer (1938a); (40) Dreyer (1938b); (41) Dreyer (1947); (42) Dreyer & Lyle (1931); (43) Ewer (1957); (44) Ewer (1962); (45) Fourie (1970); (46) Galloway (1937); (47) Grobler & Loock (1988a); (48) Grobler & Loock (1988b); (49) Grün *et al*. (1996); (50) Henderson (1996); (51) Henderson (2001a); (52) Henderson (2001b); (53) Hoffman (1953); (54) Hoffman (1955); (55) Hooijer (1958); (56) Joubert (1990); (57) Joubert & Visser (1991); (58) Joubert *et al*. (1991); (59) Kuman (1989); (60) Kuman & Clarke (1986); (61) Kuman *et al*. (1999); (62) Libby (1954); (63) Loock & Grobler (1988); (64) Lotz *et al*. (1991); (65) Louw (1987); (66) Mazor & Verhagen (1983); (67) Meiring (1956); (68) Oakley (1954); (69) Protsch (1974); (70) Rightmire (1978); (71) Rindl (1915); (72) Rossouw (1996); (73) Rubidge & Brink (1985); (74) Sampson (1974); (75) Scott & Brink (1992); (76) Scott & Rossouw (2005); (77) Scott *et al*. (2003); (78) Singer (1958); (79) Singer & Boné (1960); (80) Thackeray *et al*. (1996): (81) Van Zinderen Bakker (1957); (82) Van Zinderen Bakker (1989); (83) Visser & Joubert (1990); (84) Visser & Joubert (1991); (85) Wells (1959); (86) Wells (1965); (87) Wells (1967); (88) Wells (1972).

(1931). In 1932, Prof. Dreyer and Gert Venter uncovered parts of a hominid skull, the Florisbad skull (Anon [1] 1980). For the location of the skull see Figure 3. Prof. Dreyer donated his collection to the National Museum in 1936 when, due to a lack of funding, all excavations were halted. With funding from the then Council for Scientific and Industrial Research, Dr A.C. Hoffman, Director of the National Museum, and A.J.D. Meiring, Assistant Director of the National Museum, recommenced excavations in 1952, but this was short-lived as the then owners of the farm, the Sowden brothers, halted all excavations later that year.

In order that the site be preserved, and that further research could be carried out unhindered, the State bought 93 ha of the farm, including the springs, in 1980, and handed this over to the custody of the National Museum. The resort was then permanently closed to the public and new research initiatives were begun by the National Museum in 1981. Recent work at Florisbad has related to extensions to the previous excavations and detailed work on the MSA occupation horizon (Henderson 2001a, 2001b; Brink and Henderson 2001), as well as extensive taphonomic studies (Brink, 1987).

1.4 THE FLORISBAD FAUNAL DEPOSITS

Two distinct archaeozoological and archaeological deposits, which are separated from each other both horizontally and vertically, have been excavated at Florisbad. Figure 4 provides a generalized cross-section of the present day Florisbad sedimentary deposits showing the major organic-clay and sand horizons and their silt content. Further detail on the depositional environment and sedimentation is given in Chapter 5, subsection 5.6.

1.4.1 The Old Collection

The first collection is referred to as the Old Collection (Table 2; Table 3), comprising a basal accumulation of fossilized faunal remains in the areas of spring activity, and representing a death assemblage resulting largely from carnivore killings (Brink, 1987, 1988). Table 2 represents a list of the faunal interpretation by Brink (1987), while Table 3 represents changes made to this faunal list, by Churchill *et al*. (2000). Figure 5 is an example of the extinct giant buffalo, *Pelorovis antiquus*, from the Old Collection. Bones in this assemblage are characterized by evidence of hyaena damage and, to a lesser extent by porcupine gnawing, with no indication of cut marks (Brink, 1987, 1988). Owing to the relatively low incidence of hyaena damage to the bones (21.34%) (Brink, 1987), it is postulated here that the death of animals resulting from various diseases contributed significantly to the Old Collection. Africa has a large number of bovine and equine

Table 2. A comparison of fauna recorded between the MSA Assemblage and the Old Collection based on relative abundance (minimum number of individuals) (after Brink, 1987, 1988). = MSA Assemblage; = Old Collection.

Table 3. Changes and updates to the original taxonomic list of fauna occurring at Florisbad Research Station (Brink, 1987), compiled from Churchill *et al*. (2000).

	Now included in Churchill *et al*. (2000)	Now excluded from Churchill *et al*. (2000)
Amphibia – indet.	X	
Reptilia – indet. X		
Aves – indet.	X	
Mamalia		
Lagomorpha		
Leporidae	X	
Lepus sp.		X
Rodentia		
Murinae – indet.	X	
Pedites capensis	X	
Pedites sp.		X
Carnivora		
Hyaenidae – indet. (coprolites)	X	
Gallerella sanuinea	X	
Herpestes sanguineus		X
Perissodactyla		
Equus spp. (possibly two species, *E*. [*Asinus*] sp. [=*E. lylei*], and a plains zebra similar to *E. quagga*.	X	
Equus burchelli		X
Equus quagga		X
Ceratotherium simum		X
Artiodactyla		
Phacochhoerus sp.	X	
Phacochhoerus aethiopus		X
Phacochhoerus africanus		X
Kobus ellipsipprymnus	X	
Kobus sp.		X
Alcelaphus buselaphus	X	
Raphicerus sp.		X

Figure 5. A graphic reconstruction of the extinct giant buffalo *Pelorovis antiquus* from the Florisbad Old Collection. The horns of this species attained a spread of over 2 metres. (Graphic: E Russouw, National Museum, Bloemfontein).

diseases such as foot and mouth disease, anthrax and African horse sickness, some of which are highly contagious and may reach epidemic proportions. It is suggested here that these, or similar diseases, took a significant toll on herds of animals at various times in the past, with sick animals remaining close to a water source, which may well explanation the large number of faunal remains in the Old Collection. Should this hypothesis be correct, the incidence of bone damage would largely reflect hyaena scavenging, and not carnivore predation. The Old Collection comprises a vast number of faunal fossil remains comprising 26 species, and represents the Florisian Land Mammal Age (Hendey, 1974), or the Florisian-Cornelian Faunal Boundary, with an age of *c*. 400 ka (Klein, 1984).

1.4.2 The MSA Occupation Horizon

The second collection is the MSA human occupation horizon which occurs approximately 3.5 metres above basement, where species diversity is far less than in the Old Collection (Table 2). Figure 6 give an idea as to what the MSA human occupation horizon excavations looked like. This horizon represents a temporary butchering site and has delivered artefacts in the form of butchering tools as well as faunal remains (Brink, 1987; Henderson 2001a; Brink and Henderson, 2001). These were found in horizontal deposits on a sandy substrate, which had been deposited in an aqueous pan type environment with little disturbance (Kuman and Clarke, 1986; Kuman *et al.,* 1999; Henderson, 2001a; Brink and Henderson, 2001). In this assemblage, signature marks on bones indicate slicing, scraping and cutting, as well as bone-breaking, and show no signs of carnivore damage (Brink, 1987). MSA faunal remains are also more friable, in relation to remains from the Old Collection (Kuman and Clarke, 1986; Brink, 1987). It is important to note here that the MSA faunal remains are not directly associated to a peat layer, but lie above the Peat II layer, which may explain their friability.

Figure 6. A section of the Middle Stone Age Human Occupation Horizon excavation site at Florisbad (Photo: National Museum, Bloemfontein).

1.5 THE FLORISBAD SKULL

The position of the Florisbad skull in the evolution of modern man is of considerable importance. Drennan (1937) considered the Florisbad skull (Figure 7) to be the most primitive, and therefore the most important modern lineage human fossil thus far unearthed in South Africa. Dating of the Florisbad skull at 259 ±35 (Grün *et al*., 1996), supports a recent African origin for all modern people, and is considered the single most important find at Florisbad (Brink, 1987). Further to this, the skull is the only relatively complete example of the late archaic phase of modern development in Africa (Brink, 1997). Mr G. Venter discovered a hominid right upper third molar in the debris of one of the western spring eyes in 1932, and shortly thereafter, on 25 July 1932. Only later, did Prof. T.F. Dreyer find a hominid skull-cap and other facial bones in the same debris (Dreyer, 1938a; Clarke, 1985; Henderson, 1992). The remains of the Florisbad skull represent but a portion of the total skull, comprising part of the parietal bone, frontal bone, nasal bone, maxilla, and a tooth. The original reconstruction of the Florisbad skull is given in Figure 7. Dreyer made a brief announcement of this find in The Friend newspaper dated 26 July 1932, with a full description of the skull appearing in The Friend, dated 13 August 1932 (Henderson, 1992). A more detailed description of the skull was published in 1935 (Dreyer, 1935), placing Florisbad firmly on the archaeological and physical anthropology map.

Interpretations on the taxonomic status of the Florisbad skull have been made by Dreyer (1935, 1936a, 1938b, 1947), Ariëns Kappers (1935), Drennan (1935, 1937), Galloway (1937), Singer (1958), Wells (1969, 1972), Rightmire (1976, 1978) and Clarke (1985). With the exception of Ariëns Kappers (1935) and Wells (1969, 1972) who considered the skull to be neoanthropic (relating to more modern forms of humankind), all the other researchers recognized archaic features of the skull and considered it to be palaeoanthropic (relating to earlier forms of humankind). Despite general agreement on this particular point, there was much contention around the morphological interpretation of the skull. Dreyer, (1935) considered the Florisbad skull to be more closely related to

Figure *7*. Reconstruction of the Florisbad skull (*Homo helmei*). Note the hyena tooth impression on the forehead (Photo: National Museum, Bloemfontein).

Homo sapiens than *H. primigenius*. (250 to 20 ka), the latter being a predecessor of Neanderthal Man (130 to 24 ka). It has been hypothesised that *H. primigenius* coexisted with Neanderthal man, possibly interbreeding with Neanderthal Man to evolve into the modern day *H sapiens* (Brodrick, 1960; Pfeiffer, 1969). It was also noted by Dreyer (1935) that the flange of the malar projected outwards and backwards, similar to the condition seen in Bushmen skulls. Dreyer (1935) placed the Florisbad skull in the Hominidae, below *H. sapiens* and *H. primigenius*, giving it the sub-genus status of *H.* (*Africanthropus*) *helmei,* an archaic form of *Homo sapiens*. Both Ariëns Kappers (1935) and Dreyer (1935) agreed that the Florisbad skull differed from Rhodesian Man (*H. rhodesianus*) (*Synonyms*: Broken Hill cranium; Kabwe cranium) from Zambia (300 to 125 ka).

Drennan (1935, 1937) considered the skull to be more characteristic of a "low" type of *Homo sapiens* represented by Rhodesian Man, and concluded that it was an African variant of the Neanderthal race with minor modern features, naming it *H. florisbadensis* (helmei). Both Dreyer (1935) and Drennan (1935) agreed that artefacts from the site were of a Mousterian nature, being representative of a Middle Palaeolithic, or Neanderthal culture. Dreyer (1936a) then proceeded to show similarities between the Florisbad skull and Bushmen skulls, concluding that the Florisbad skull belonged to a pre-historic race of Bushmen. From descriptions of the Steinheim skull, *H. steinheimensis*, (250-350 ka) from Germany by Weinert (1936), Dreyer (1936b) concluded that the Florisbad skull was similar, but a more primitive member of the same race. Drennan (1937) placed the Florisbad skull before the Rhodesian skull in the evolutionary series, while Galloway (1937) placed the Florisbad skull following the Rhodesian skull, seeing it as a link between the Broken Hill and Boskop skulls.

In his interpretation between the Rhodesian skull and the Bothaville skull (a modern European skull), Dreyer (1938b) found them to be so similar that there was no reason to assume that the former did not belong to *H. sapiens*. Dreyer (1938b) concluded that the differences in the frontal lobe of the Florisbad cast were so conspicuous that separate specific rank must be accorded to the Florisbad skull. In comparing the Maatjies River

(MR) skull to that of the Florisbad skull Dreyer (1947) concluded that the Florisbad skull "fits almost ideally" as a predecessor to the MR skull. In his taxonomic classification of South African skulls, Dreyer (1947) further concluded that the race to which he had originally assigned the Florisbad skull, should be maintained as *H.* (*Africanthropus*) *helmi*, that the MR stage should be assigned to *H.* (*Africanthropus*) *dreyeri*, and the Bushmen race to *H.* (*Africanthropus*) *austroafricanus*.

Singer (1958) suggested that the Florisbad skull belonged to the Rhodesian – Saldanha group, while Rightmire (1976) stated that sub-Saharan archaic man could not simply be grouped as an African Neanderthal. As Rightmire (1978) considered the Florisbad skull to be older than Neanderthal, representing a Middle Pleistocene period, he aligned the Florisbad skull to the Broken Hill lineage.

After decades of speculation as to the taxonomic status of the Florisbad skull, Clarke (1985) carried out a further reconstruction of the cranium. During the reconstruction Clarke (1985) came to a number of conclusions, with the most significant being that Dreyer (1935) had incorrectly reconstructed the facial bones in three major areas. As there were no positive joints between any of the facial bones, Clarke (1985) concluded that the plaster reconstruction between the bones was educated conjecture, and therefore any facial measurements would be meaningless (Clarke, 1985). Clarke (1985) further concluded that the Florisbad skull had a much more archaic and larger appearance than that of Dreyer's reconstruction, and considered it to have strong similarities to the Broken Hill cranium. However, Clarke (1985) also noted that subsequent interpretations based on the Dreyer's (1935) reconstruction by authors such as, Drennan (1935), Galloway (1937), and Rightmire (1978), were unreliable.

In 1997 Foley and Lahr (1998) resurrected the name *H. helmei* based on mode 3/Middle Palaeolithic industries and archaeological remains, which resulted in the establishment of the immediate ancestors to the Neanderthals. This classification of the Florisbad skull appears to have been accepted by a number of researchers (Deacon and Deacon, 1999; Kuman *et al*., 1999) as an archaic form of *H. sapiens*. However, from an anatomical point of view there still remain many questions. Lieberman *et al*. (2002) placed *H. helmei* in a group of recent, or modern, human fossil records that has a confusing pattern of variation, with many vaguely defined taxa which are not widely accepted. This confusion, Lieberman *et al*. (2002) stated was born out of a lack of established unique derived features (autapomorphies) between anatomical modern *H. sapiens* (AMHS) and archaic *Homo* species (AH). White *et al*. (2003) felt that in addition to the difficulties in partitioning lineages, many of the available species names were based on inadequate type specimens, such as *H. helmei,* and a lack of substantial and accurately dated hominid fossils between 300 and 100 ka. Greater accuracy and detail is important in light of hypotheses like the one proposed by Lahr and Foley (1998) which propose that. Neanderthal and modern lineages share a common ancestor in an African population between 350 and 250 ka ago. In summary, it would appear that the incomplete cranium and lack of an entire brain case in the *H. helmei* type specimen is a major factor in not being able to pin point its precise taxonomic lineage. Therefore, unless further, or complete, *H. helmei* skulls are found, the *H. helmei* question may never be truly resolved.

1.6 DATING OF THE FLORISBAD DEPOSITS

Dating of the Florisbad deposits has been an ongoing exercise over decades, with the age of the Florisbad skull being a matter of conjecture until it was ESR dated in 1996 at 259 ±35 ka (Grün *et al*., 1996). It will be noted in Table 4 that ^{14}C dating ages of the lower deposit ages are usually greater than the upper limit of the dating method which was used at the time, and therefore these ages are very inaccurate: the true ages being much higher than the greater-than-ages given. Laboratory codes for the ^{14}C dates are given in Table 4. However, the greater-than ages may also have been as a result of there being no measurable radioactivity in the lower Florisbad sediments (Libby, 1954; Oakley, 1955). Greater detail on the modes of fossilization of the faunal remains and details on the dating are discussed in Chapter 5: subsection 5.6.4 and Chapter 6 respectively. Table 4 provides a history of the dating of the Florisbad sediments compiled from Deacon (1966), Kuman and Clarke (1986), and Grün *et al*. (1996). More detail on dating and the influences of the environment on dating is given in Chapter 6.

Table 4. History of the dating of the Florisbad deposits.

Stratigraphic Layer/ Item	Age yr	Sample Ref. number	Method	Reference
Peat IV	3 530 ±80	Pta-1128	^{14}C	Kuman & Clarke 1986
	3 550 ±60	Pta-3617	^{14}C	Kuman & Clarke 1986
	3 580 ±60	Pta-3631	^{14}C	Kuman & Clarke 1986
Sand above Peat III	4 370 ±70	Pta-1127	^{14}C	Kuman & Clarke 1986
	10 000 ±100	Pta-1125	^{14}C	Kuman & Clarke 1986
	11 700 ±110	Pta-3609	^{14}C	Kuman & Clarke 1986
Peat III	6 700 ±500	C-852	^{14}C	Libby 1954
	19 530 ±650	L-271D	^{14}C	Broecker *et al*. 1956
White sand below Peat III	>44 700	Pta-3465	^{14}C	Kuman & Clarke 1986
Organic layer above Peat II	>47 200	Pta-3623	^{14}C	Kuman & Clarke 1986
White sand above Peat II	146 000 ±15 000		OSL	Grün *et al*. 1996
	128 000 ±22 000			
	133 000 ±31 000			
MSA Human Occupation horizon	121 000 ±6 000		ESR	Grün *et al*. 1996
Peat II	9 104 ±420	C-851	^{14}C	Libby 1954
	28 450 ±2 200	L-271C	^{14}C	Broecker *et al*. 1956
	>42 800	Pta-1108	^{14}C	Beaumont & Vogel 1972
Olive green sand above Peat I	157 000 ±21000		OSL	Grün *et al*. 1996
Peat I	>35 000	L-271B	^{14}C	Broecker *et al*. 1956
	>41 000	C-850	^{14}C	Libby 1954
	>44 000	Y-103	^{14}C	Barendsen *et al*. 1957
	>48 100	GrN-4208	^{14}C	Beaumont & Vogel 1972
	281 000 ±73 000		OSL	Grün *et al*. 1996
Hominid tooth	259 000 ±35 000		ESR	Grün *et al*. 1996
Basal brown sand layer	279 000 ± 47 000		OSL	Grün *et al*. 1996

^{14}C = Radiocarbon dating
ESR = Electron Spin Resonance dating
OSL = Optically Stimulated Luminescence

1.7 RATIONALE FOR THE STUDY

The research reported on here originated in 1988 when the ecology and biogeography of the herpetofauna of Florisbad was examined (Douglas, 1992, 1995). Amphibians were found to occur in most of the waters related to the Florisbad spring site. In light of this finding, the water quality of the spring, vlei, an exploration pit that amphibians were inhabiting, the farm dam, and Soutpan (Figure 2), were analysed, and the results examined in light of the water bodies being suitable as habitats and breeding environments for amphibians (Douglas, 1992, 1995). A re-examination of the 1988 water analysis results by the author indicated that the historically low Ca concentration in the spring water did not support theories in the literature (Brink, 1987) that faunal remains at the site had become fossilized in a spring context. Further to this, the TDS (total dissolved solids) of the water in the exploration pit was 27% higher than that of the spring-water, only 22m away, and this disparity was questioned (Douglas, 1992, 1995). This disparity in the water quality was not identified, or questioned, in the original study due to these waters only being examined in relation to their suitability as habitats for amphibians. These two aspects of the water quality led to a number of questions being asked in four main areas, for which no answers could be found:

- Firstly: was there actually a disparity between the spring-water and the excavation pit water, or was this an isolated occurrence peculiar to the sampling period? A fairly accurate assessment of the spring water quality could be made from sporadic water sampling over the past 84 years (Douglas, 1992), but no sampling of the excavation pit waters had been carried out, and therefore no information on the quality of the excavation pit waters was available.

- Secondly: should there be a disparity between the quality of the spring- and excavation pit waters, questions arose as to why there was a difference, what was causing the difference, and what processes and factors were involved. Furthermore, in light of the low spring-water Ca content, which aspect of the site had sufficient mineralization for fossilization of the faunal remains, if it was not the spring-water, from where did these minerals originate?

- Thirdly: in light of the low spring-water Ca content the assumption was made that the spring-water was not responsible for fossilization. If this were the case, where was fossilization taking place, how was it taking place, and what processes and factors were involved?

- Fourthly: Existing hypotheses on the formation of the Florisbad site could not account for a number of unanswered questions. These included the absence of the upper red-brown sand units on the eastern side of the "mound" (Rubidge and Brink, 1985; Brink 1987), the seven metres of clay deposits in the vlei area (Butzer 1984), the size of the vlei in relation to the ephemeral drainage line, the erosion of the east bank of the vlei, the two almost right angle dog legs in the current ephemeral drainage line, and the pinching out and height of the Peat IV layer on the west wall of the excavations.

Dating of the Florisbad deposits were a significant factor in the literature review leading up to these investigations. However, illustration (a) in Grün *et al*. (1996) suggested that when the depth of the sediments was plotted against time, this would reflect equal compaction and compression of the sediments at all depths, which appeared to be illogical. It was hypothesised that the application of mathematical trend lines to depth/age plots would either confirm or disprove this theory. It was further hypothesised that the application of a linear trend line would reflect equal compaction over depth and time, while logarithmic and exponential trend lines would reflect a more logical and gradual degree of compaction over depth and time. It was hoped that from the results it would be possible to verify, or disprove, the dates of depths given in illustration (a), Grün *et al*. (1996), as well as the dates given by Grün *et al*. (1996) for the Middle Stone Age Horizon located in the middle of the profile, and the much older lower sediments, located at the bottom of the profile.

These questions then provided the motivation for further investigations. It should be mentioned that since the publication of the results of the study in Appendices I-IV, certain aspects of the study have been revisited, and the hypotheses updated. Therefore, the hypotheses, illustrations, discussions, and conclusions, presented in this main body of the thesis, should they differ from those in the Appendices, should be considered as being the more correct.

1.8 KEY AREAS TO BE ADDRESSED

- An examination the physical environment in order to determine the many components which may have contributed to the existence and formation of the site, as well as contributing to an environment suitable for fossilization.
- Determine whether there is a difference between the quality of the spring- and subterranean waters away from the spring eyes, and determine to what extent the quality of the two entities may differ.
- Determine the degree of salinization of the organic-clay layers, and relate this to the degree of mineralization of the spring- and groundwater.
- Determine the properties of the organic-clay layers in relation to salinization, as well as possible processes and factors within the environment which would contribute to this salinization.
- Correlate the above information in order to determine whether conditions existed within the organic –clay layers for the initiation of fossilization.
- Determine whether fossilization was capable of taking place within the confines of the spring vents and eyes.
- Develop a theory, based on existing morphological and chemical evidence, for the formation of the spring site that would accommodate the geomorphology of the site, salinization of the organic clay layers, and the fossilization of faunal remains.
- Develop a method of evaluating the variability of the ages and depths of published sedimentary profiles and their fossil components.

1.9 SYNOPSIS

The importance and uniqueness of the Florisbad spring site as an archaeozoological and archaeological site has been briefly illustrated in this chapter. Brouwer (1967) stated that many detailed palaeontological descriptions make no mention at all of the conditions under which fossilisation has taken place. Brink (1987) further stated that, to fully understand the fossil fauna it is important that the palaeoenvironment from which the fossil remains were extracted, is better known. Even today, the priority appears to be largely centred on the collection and description of archaeozoological evidence, with only modest attention being paid to palaeo-, or current, environmental and chemical factors and processes, which resulted in the fossilization. In other words, a more holistic approach needs to be taken when examining a site such as the Florisbad spring site because fossilization, for example, cannot be simply explained, or determined, by its independent component parts. In the following chapters it will be seen how the inter-relationship and behavioural variability of these component parts can affect the system as a whole.

Chapter

2

The Physical Environment

Part I

Geology and Geomorphology

CHAPTER 2

THE PHYSICAL ENVIRONMENT

PART I

GEOLOGY AND GEOMORPHOLOGY

2.1 INTRODUCTION

The geohydrological and surface processes controlling the depositional environment at Florisbad, and the resultant fossilization of faunal remains, fall under the influence of the physical environment, whether it is the influences of geological, geohydrological, depositional, climatic, vegetation, mechanical, or chemical components.

The importance and influence of the physical environment on the existence and formation of the Florisbad spring site cannot be over emphasised. It is therefore important to examine, in some detail, as many aspects of the physical environment as possible in order to present a holistic interpretation of the Florisbad entity. For example, the geology of the area may influence soil types, and in conjunction with climate, will determine vegetation. Further more, geology, along with many factors, will ultimately determine the formation of aquifers and the flow of groundwater. To this end, the physical environment has been divided and grouped into three sections, namely, geology and geomorphology, vegetation and climate, and geohydrology. In Part I of the physical environment the regional geology is examined in order that the local geology, which is an integral part of the greater Karoo Basin, can be put into context. From a geomorphological perspective, the Western Free State panveld, and Soutpan, which lies north-west of the Florisbad spring site, as well as the Florisbad sand dune, which has resulted in the spring sites modern day morphology, have both had major influences on the depositional history of the site.

2.2 REGIONAL GEOLOGY

The Free State lies close to the geographic centre of the southern African subcontinent and has been less strongly influenced and modified by the breaking up of Gondwanaland than the margins (Moon and Dardis, 1988). The Archaen/Precambrian geology, which is comprised of formations of the Witwatersrand (3000-2800 Ma) and Ventersdorp (2800-2650 Ma) Supergroups, and includes the Central Rand and Dominion Groups (3100-3050 Ma), are important in that they are the basis of the Free State gold mining industry (Holmes and Barker, 2006). The Precambrian geology is in turn blanketed almost entirely by rocks of the Middle Jurassic to Late Carboniferous (300-179 Ma) Karoo Supergroup. This Group comprises the Dwyka (300-289 Ma), Ecca (289-255 Ma), Beaufort (255-237 Ma), Stormberg Groups (230-183 Ma), and the Drakensberg Groups (183-279 Ma) (Catuneanu *et al*., 2005), with Florisbad lying within the Ecca Group (Figure 8). Karoo Supergroup rocks cover approximately two-thirds of the South African land surface (see 2.1.13), and are important for their coal bearing seams in the shallower northern parts of the Karoo Basin (McCarthy and Rubidge, 2005). Karoo rocks are also recognised globally for their wealth of fossil tetrapods, which span the stratigraphic record from the Mid-Permian to the Early Jurassic (Rubidge, 2005).

Classification, descriptions, ages, Precambrian geology, and the Karoo Supergroup members have been sourced from Keyser, (1997), Bisschoff, and Mayer (1999), Anon [2] (2001), Vegter, (2001), Chevallier and Woodford (1999), Woodford and Chevallier (2002), Baran (2003), Grandstein and Ogg (2004), Neveling (2004), Catuneanu *et al*. (2005), McCarthy and Rubidge (2005), Rubidge (2005), and Holmes and Barker (2006), unless otherwise referenced.

2.2.1 Precambrian Strata

The Precambrian geology, although poorly represented in the Free State, forms a part of the regional geology. Precambrian strata is largely restricted to the northern Free State at

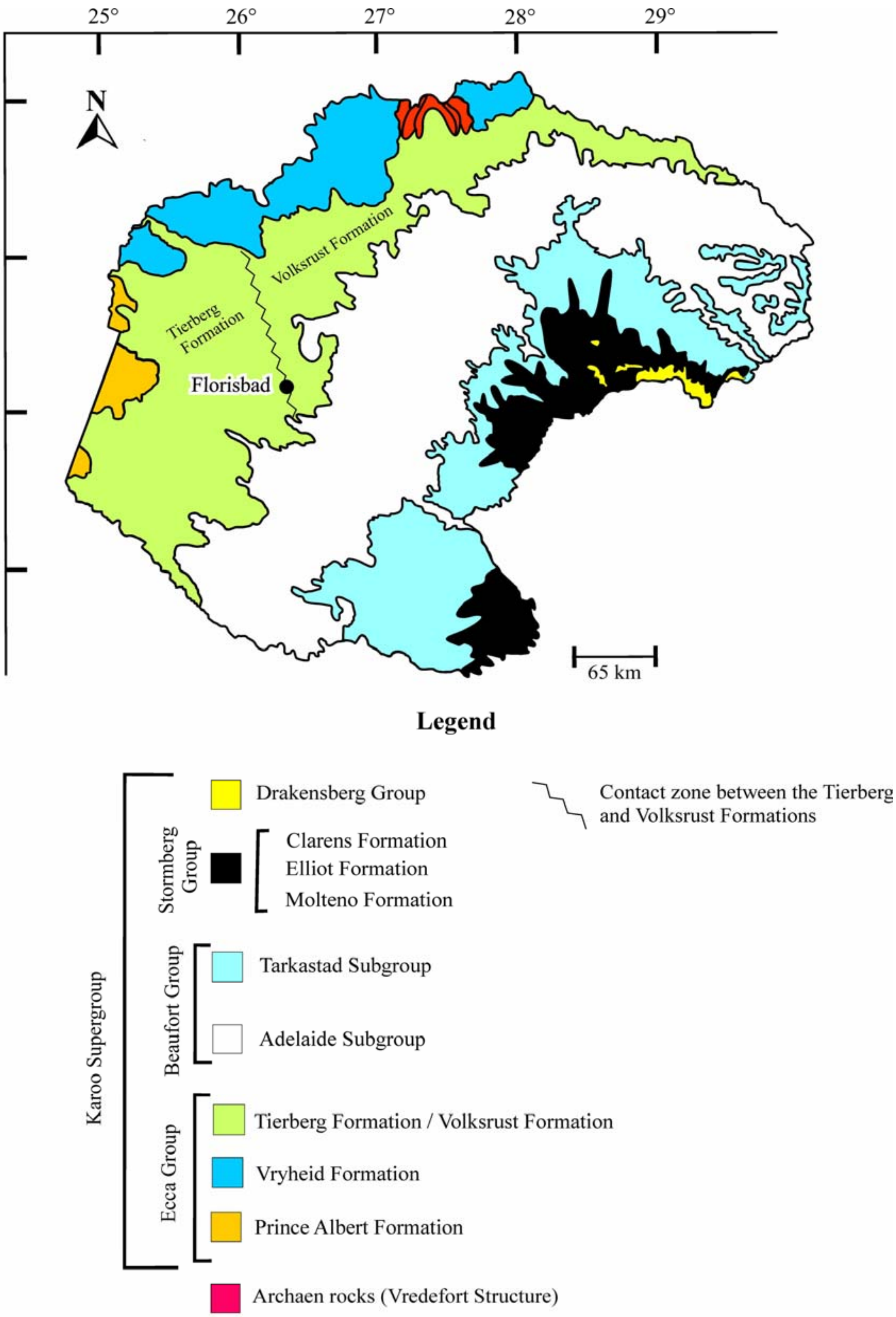


Figure 8. Simplified geology of the Free State showing geological formations mentioned in the text (after Keyser, 1997; Catuneanu *et al.,* 2005; Rubidge*,* 2005).

the site of the Vredefort meteoric impact site with the oldest rocks of the area being Inlandsee gneiss and Parys granite of Swazian age (>3100 Ma). The Vredefort structure (*c*. 2023 Ma) (Figure 8) is the largest verified, and oldest, clearly visible meteorite impact structure on Earth and was declared a World Heritage Site by the United Nations Educational, Scientific and Cultural Organization on 14 July 2005 (Anon [3], 2006; Anon [4], 2007). The Vredefort structure was formed by the impact of a meteorite estimated to have been at least 10 km in diameter (Mayer, 2007). It is estimated that the impact vaporized some 70 cubic kilometres of rock (Anon [3], 2006) at temperatures of up to 2000 °C (Mayer, 2007.), creating a crater ±250 km in diameter (Anon [3), 2006).

In order to view the Vredefort structure in a global context, the following craters may be considered important. The Sudbury impact structure in Canada rates as the second largest impact site with a diameter of 200 km (Anon [3], 2006). The Morokweng impact crater of South Africa, is >70 km in diameter, and the second largest crater in Africa, estimated to be 145 Ma, which coincided with the Jurassic-Cretaceous boundary (Maier *et al.*, 2006). It is not visible from the surface as it buried beneath a layer of Kalahari sand (Maier *et al.*, 2006). With a diameter of 16 km, the Suavjärvi crater in Russia is recognised as the oldest impact crater at 2400 Ma, but due to regional metamorphism and weathering, it is no longer visibly recognisable as an impact structure (Mashchak and Naumov, 1996).

There appears to be some confusion over the use of the term Vredefort "Dome" (Figure 8). The Vredefort Dome represents the central uplift of a very large impact structure that exposed nearly a complete cross-section through the continental crust of 25-30 km thick (Martini, 1992). This uplift and overturn of Precambrian strata was created when a high-speed primary shock-wave was sent vertically down into the earth, followed almost instantaneously by a rebound shock-wave from inside the earth (Mayer, 2007). This caused a central plug in the floor of the crater to undergo massive uplift and overturn (Mayer, 2007). The overturn at Vredefort has resulted in the deeper Precambrian rock layers of the Witwatersrand Supergroup being exposed almost vertically on the surface, with a dip of 100° to 110°, or 70° to 80° inwards (Hart *et al.*, 1990, 1991, 1995; Tredoux, *et al*., 1999). The core of the structure comprises granite-gneiss, surrounded by a collar of sediments and lavas, which become progressively younger towards the outer perimeter (Hart *et al*., 1990, 1991, 1995; Tredoux, *et al*., 1999; Bisschoff and Mayer, 1999).The Vredefort Dome lies in the central part of the Witwatersrand Basin and is the type locality for pseudotachylite (Eiko, 2001). Pseudotachylite (Shand, 1916) is a dark glassy, grey, granite, formed by frictional heating, and subsequent melting, along oblique impact surfaces during high speed faulting as a result of the meteoric impact (Nakamura, 2003).

Further to the above, the only other occurrence of Precambrian strata is the Allanridge Formation of the Ventersdorp Supergroup which encroaches over the western border of the Free State in a few small areas north-west of Bothaville, north-west of Jacobsdal, west of Hertzogville, and at Allanridge.

2.2.2 The Karoo Basin

Because Florisbad lies within the Karoo Basin, and because the Karoo Basin is the dominant geological structure of the South African landscape, covering approximately two-thirds of the land surface, this is briefly discussed.

While the Karoo Basin (Figures 9a, 9b) is a unique structure in itself, Karoo-age basins, which show clear similarities to the South African Karoo Basin, occur across Africa (Catuneanu *et al.*, 2005). The Karoo Basin is a retro-arc foreland basin. Catuneanu *et al.* (2005) noted that two factors were responsible for the formation of the south-central African Karoo Basins during the Late Paleozoic-Early Mesozoic period when the Pangea supercontinent reached its maximum extent.

Primary control was by tectonic mechanisms, with subsidence mechanisms ranging from flexural in the south, to extensional in the north, and propagating southwards from the divergent Tethyan margin (Catuneanu *et al.*, 2005). These tectonic mechanisms, combined with influences exerted by the inherent structures of the underlying

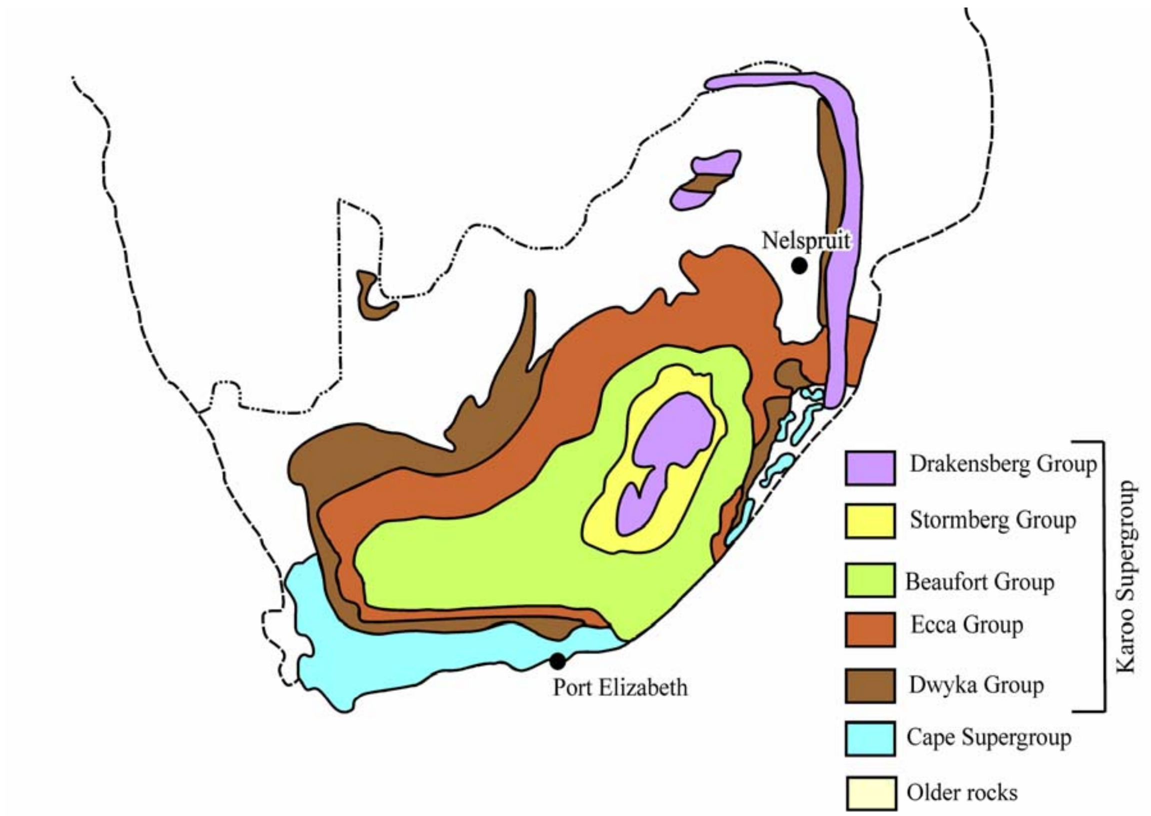


Figure 9a. The extent of the Karoo Supergroup rocks over South Africa (after McCarthy & Rubidge, 2005).

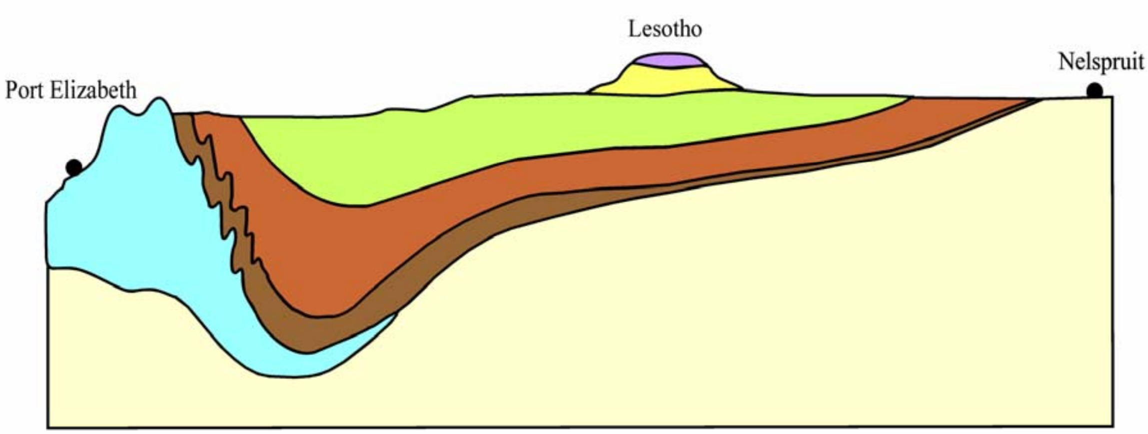


Figure 9b. Schematic cross-section across the Karoo Basin, from Port Elizabeth in the south to Nelspruit in the north-east, showing the stratigraphy of the Karoo Supergroup rocks and the variation in the depth of the Karoo Basin. The Karoo Basin also reflects the asymmetry of the Karoo Sea (after McCarthy & Rubidge, 2005).

Precambrian basement, resulted in discrete depozones, which were further influenced by shifts in climatic regimes. During the Karoo interval, shifts in climatic and tectonic conditions from the northern and southern margins of the African continent resulted in the lithostratigaphic character of the Karoo Supergroup changing significantly across the continent. Karoo basins *sensu stricto* in Africa are restricted to Africa south of the equator, while Karoo-age successions preserved north of the equator are distinctly different (Catuneanu *et al.*, 2005). Woodford and Chevallier (2002) further related geomorphology of the subcontinent to the African surface (±40 ma), which occurred before the African Plate came to rest over the mantle. This is reflected in the topography of the central part of the Karoo Basin when the Great Escarpment, formed during the Jurassic, was still a prominent feature (Woodford and Chevallier, 2002). Topography was further influenced by the post-African I surface (±30 ma) when there was an uplifting centred below Lesotho, resulting in the reactivation of the Great Escarpment (Woodford and Chevallier, 2002). The latter is reflected in the peripheral topography of the Karoo Basin.

At this time, South Africa was located over the South Pole. The subduction zone produced volcanic activity, which helped to increase the size of the developing Cape Fold Mountain range to the extent that the weight of the mountains caused a sagging of the lithosphere to the north. As this basin, created by this sagging, drifted northwards and away from the polar latitudes, the ice sheets melted (Hancox and Rubidge, 2002), giving rise to a large inland sea referred to as the Karoo Sea. The deepest part of the basin lies to the south in the foredeep on the landward side of the Cape Fold Belt, where it is 12 km deep, gradually becoming shallower to the north (Woodford and Chevallier, 2002) (Figure 9b). It was in this basin that sedimentation cycles would deposit the sediments, which were later to form the rocks of the Karoo Supergroup (McCarthy and Rubidge, 2005).

2.2.3 The Karoo Supergroup

The Karoo Supergroup comprises a number of Groups and Formations deposited within the Karoo Basin through progressively changing depositional environments during the Late Carboniferous (310 Ma) through to the Mid Jurassic (185 Ma). In order to understand the position of the Ecca Group, in which Florisbad lies, within the Karoo Supergroup, a brief discussion on the Groups and Formations is presented.

2.2.3.1 *Dwyka Group*

Dwyka Group rocks comprise glacially derived tillites deposited in a marine environment and consists mainly of diamictite that grades upwards into conglomerates, mudstone and shale. Lithological differences occur within this Group over the Basin. The detection of these glacial sediments in India, and South America, provided early evidence in support of the Theory of Continental Drift (McCarthy and Rubidge, 2005). Sedimentation during the Dwyka period was so considerable that in the southern part of the Karoo Sea, the sediments began to form vast tracts of land. The Dwyka Group covers a time span of 300-289 Ma, extending from the early to Late Carboniferous to the early Permian (Catuneanu *et al.*, 2005). Some authors have indicated a swathe of Dwyka Group rocks along the western boundary of the Free State (De Bruiyn, 1971; De Waal, 1978), while others have included the Ecca and Dwyka Groups under a single category (Earlé and Grobler, 1987). By implication, the combining of the Ecca and Dwyka Groups over the Free State implies that Dwyka Group rocks do occur in the Free State. The references used in Figure 8 all indicated that no Dwyka Group rocks occur in the Free State, with the nearest location of these rock types being west of Kimberley, and west of Christiana in the North West Province. This Group forms the base of the Karoo Supergroup with a thickness of up to 330 m, and is of little significance in terms of the geomorphology of the Free State landscape.

2.2.3.2 *Ecca Group*

The Ecca Group comprises 16 formations of which four dominate in the Free State, namely the Volksrust, Vryheid, Tierberg and Prince Albert Formations. These sedimentary rocks consist of muds, silts and other deltatic sediments, accumulated under brackish and fresh water conditions from the drainage of large swampy deltas located along the northern shores of the basin, into the Karoo Sea (McCarthy and Rubidge, 2005). The Ecca Group shales were formed in a shallow intracratonic depression, probably the result of preceding Dwyka glaciation.

The rocks of the Vryheid Formation in the extreme western Free State are of a cyclic deltaic and fluvial origin, and are more arenaceous than the predominantly argillaceous Volksrust formation. This formation comprises thick mudstone and fine- to coarse-sandstone, with secondary shale layers with a thickness of up to 500 m. Being of marine origin, the Volksrust Formation comprises shale, siltstone, mudstone and fine sandstone at the top, and is more argillaceous than the Vryheid Formation. Loock and Grobler (1988) state that Florisbad is underlain by Tierberg Formations The Tierberg Formation largely comprises well laminated, dark grey to black shale, which has been described by Nolte (1995) as a lateral equivalent of the Volksrust Formation mapped further to the east. Catuneanu *et al*. (2005) indicate that the contact lies north of Florisbad, while Villjoen (2005) noted that the vertical division between the two formations occurs in the Boshfof-Hertzogville area, where the Whitehill Formation pinches out. Viljoen (2005) indicates that this contact zone may pass very close to, if not beneath Florisbad (Figure 8).

The Volksrust Formation, with westerly thickness of up to 380 m, extends from the Florisbad area, north, in an arc through to the north-eastern Free State, into Gauteng, Mpumalanga, and down into KwaZulu-Natal. On the other hand, the Tierberg Formation, with a thickness of up to 7700 m, extends from the vicinity of Florisbad, south into, and across, the Northern Cape Province, then southward again into the Western Cape towards Matjiesfontein. The Ecca Group is well represented in the western Free State, and has weathered and eroded to produce a flat to undulating landscape, broken by flat-topped dolerite capped mesas, which will be discussed later. Coal is a major economic deposit within the Ecca Group. The Ecca Group covers a time span of 289-255 Ma, extending from the late to mid Permian period.

2.2.3.3 *Beaufort Group*

These rocks formed from fluvial and deltaic derived sediments which dominate the central and eastern Free State. As deltas to the south gradually built up, the Karoo Sea shrank to become a lake, with this transition to more terrestrial environments signalling the boundary between the Ecca and Beaufort Groups (McCarthy and Rubidge, 2005). Beaufort Group rocks were deposited by north-flowing, meandering rivers, in which sand accumulated, flanked by large floodplains where periodic flooding deposited mud (McCarthy and Rubidge, 2005). This environment provided an ideal habitat for the diversification and evolution of early reptiles, in fact, to such an extent that therapsid fossils have been used to biostratigraphically subdivide the rocks of the Beaufort Group (Kitching, 1977; Keyser and Smith, 1978; Rubidge *et al*. 1995). Within this Group, the older Adelaide Subgroup, comprising shale, siltstone, and fine sandstone, with a thickness of up to 5000 m, dominates in the west. In the east, and the younger Tarkastad Subgroup, comprising mudstone and sandstone, with a thickness of up to 2000 m, dominates in the east. Some areas may record a 70% predominance of sandstone. Weathering and erosion is similar to that of the Ecca Group shale, but due to the higher occurrence of dolerite (see below) in the central and eastern Free State, the relief is not as flat, with a higher occurrence of dolerite capped mesas. Depositional environments indicate a progressive desiccation up the sequence. The Beaufort Group covers a time span of 255-237 Ma, extending from the late Permian to the mid to early Triassic.

2.2.3.4 *Stormberg Group*

Some controversy has surrounded the classification of the Drakensberg and Stormberg Groups over the years, resulting in various classifications. For example, Rubidge (2005) initially referred to the Drakensberg and Stormberg Groups, while later in the paper the Drakensberg Group was referred to as the Drakensberg Formation, and was included under the Stormberg Group. Neveling (2004) classified the Molteno, Elliot and Clarens Formations as independent formations, omitting the encompassing Stormberg Group term, while maintaining the succeeding Drakensberg Group. Catuneanu *et al*. (2005) grouped the Molteno, Elliot and Clarens sedimentary formations under the Stormberg Group, while also maintaining the succeeding Drakensberg Group. The reasoning for this latter decision was that the angular unconformity of the base-Molteno indicated a significant tectonic event across the region, which heralded the Stormberg sedimentation period. The classification by Catuneanu *et al*. (2005) as been adopted in this thesis.

The Stormberg Group comprises three sedimentary formations and covers a time span from 230-183 Ma, extending from the mid Triassic to late Jurassic periods. The Molteno Formation (230-216 Ma), was deposited largely by shallow braided rivers (Hancox, 2000), and hosts an abundance of insect, fish and sea-fern fossils (Anderson *et al*., 1999). As indicated by Catuneanu *et al*. (2005), the Molteno Formation does not lie directly on the Burgersdorp Formation of the upper Beaufort Group, with a ±7 Ma unconformable stratigraphic hiatus between the two. Molteno Formation rocks are of fluvial origin, dominated by sandstone (70%) and mudstone (20 %), with a thickness of up to 70 m.

Rocks of the Elliot Formation (215-203 Ma) were deposited in drier conditions, with loess type aeolian sedimentation of mudstone and siltstone, and fluvial subordinate sandstone (Baran, 2003; Hancox and Rubidge, 2002). This formation, with a thickness of up to 150 m, preserves a large variety of reptile fossils including the oldest known tortoise from Africa (Gaffeny and Kitching, 1994). These deposits also include some of the earliest mammals (Gow, 1986). The Clarens Formation (203-183 Ma) was formed in arid conditions and is of aeolian origin comprising sandstone layers derived from sand dune deposits, which may attain a thickness of up to 230 m. These sandstones (66%) have weathered to produce spectacular vertical cliffs, while differential weathering has produced overhangs and caves. Small quantities of mudstone and siltstone also occur. The Stormberg Group reflects a gradual increase to more arid conditions, sequentially recorded in the rocks, which make up this group (McCarthy and Rubidge, 2005). Towards the end of the deposition of the Elliot Formation, these rocks indicate desert conditions. The rocks of the upper Clarens Formation attests to true desert conditions and a situation similar to that of the Namib Desert (McCarthy and Rubidge, 2005).

2.2.3.5 *Drakensberg Group*

This Group comprises horizontally stratified basaltic lavas, including numerous flows of up to 50 m thick, contributing to a total thickness of 1400 m. When compression of the Karoo Basin was relaxed, due to a drop in sedimentation rate, the Earth's crust erupted along fissures, spreading basaltic lave across the Clarens desert over most of South Africa (McCarthy and Rubidge, 2005). In the eastern Free State, these basalts only remain as remnant capping overlying the Clarens Formation at the highest elevations. This volcanic event signalled the end of Karoo Supergroup sedimentation and the beginning of the fragmentation and dispersal of Godwana into the continents, as we know them today (McCarthy and Rubidge, 2005). Much of this magma was injected into the horizontal sedimentary layers of the Karoo Supergroup, where upon cooling, crystallized to form dolerite sills. The Stormberg Group covers a short time span from 183-179 Ma in the late to mid Jurassic.

2.2.4 Karoo Dolerite Suite

The Karoo dolerite suite represents a post Karoo sedimentation, Jurassic period, of magmatic intrusions, which are younger than the basalts of the Drakensberg Group (Vegter, 2001). This extrusion of basalt took place over the entire African subcontinent, making this event one of the largest flood-basalt outpourings in the world (Chevallier *et al*., 2001). Dolerite dykes and sills are important in the hydrogeology of the Karoo Supergroup as their intrusion resulted in the creation of fracture zones within the host rocks and themselves, forming an important source of underground water (Vivier, 1996; Baran, 2003)). Most of the second-order geomorphological features and drainage systems of the main Karoo Basin are controlled by dolerite dykes, sills and ring-complexes

(Chevallier *et al*., 2001) Dolerite intrusions have had a number of effects on the host rocks. These include the metamorphosis of the host rock due to contact metamorphism, as well as mechanical deformation of the host rocks as in dilation and bending, which in turn may result in fracturing (Vivier, 1996). The suite can be divided into two basic categories.

2.2.4.1 *Dolerite Sills and Ring-Complexes*

These formations are one of the most are the most common type on intrusion in the Karoo Basin, and one of the most prominent features of the Karoo landscape (Woodford and Chevallier, 2002). Their distribution is the same as that of the dolerite dykes, with their emplacement being strongly controlled by the lithology of the country rock (Chevallier *et al*., 2001; Woodford and Chevallier, 2002). Dolerite sills are mostly associated with the argillaceous, lower units of the Karoo Supergroup (Baran, 2003). They re also preferentially associated with the contact zones of the Dwyka- Ecca Group, the Prince Albert- White Hill Formation, the Upper Ecca- Lower Beaufort Group, and other lithological boundaries within the Beaufort Group (Woodford and Chevallier, 2002). Dolerite sills are sheet-like structures exhibiting a saucer-like morphology (Figure 10 and 11) that follows the bedding planes of Karoo structures, with a thickness of between 15 and 300 m, and the outer sill having a diameter of from 10 to 50 km (Vivier, 1996; Baran, 2003; Chevallier *et al*., 2001; Woodford and Chevallier, 2002). Dolerite sills form large coalescing circular, oval, or kidney-shaped structural units, with each unit being composed of several sub-units of smaller size, which are in turn comprise even smaller units (Chevallier *et al*., 2001; Chevallier and Woodford, 2002) (Figure 10). A number of other morpho-tectonic emplacement models have been proposed over the years by researchers such as Rogers and Schwarz (1902), Du Toit (1905), Du Toit (1920), Lombard (1952), Johnson and Pollard (1973), Meyboom and Wallace (1978), Burger *et a*l. (1981), Kattenhorn (1994), Vivier *et al*., (1995), and Vivier, (1996). Structurally dolerite sills are extremely complex with an intricate connectivity, and may be composed of many linear to curvi-linear segments.

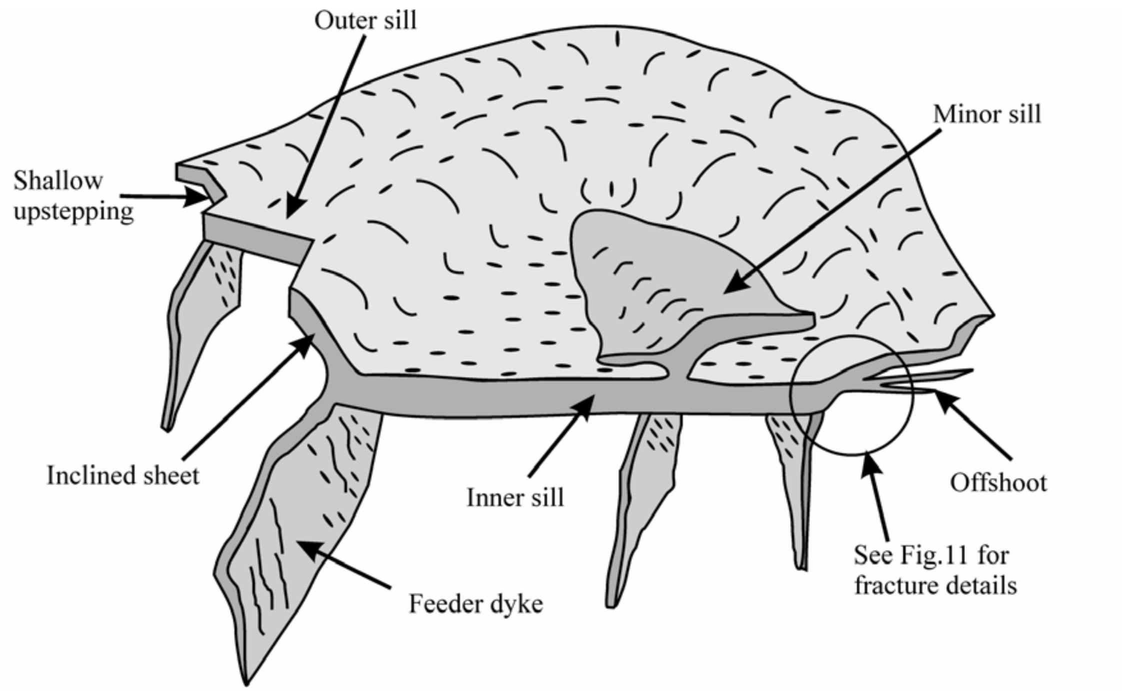


Figure 10. The mechanism of the emplacement of dolerite sill and ring complexes (ring dyke). The saucer shape and ring dyke model after Woodford and Chevallier (2002).

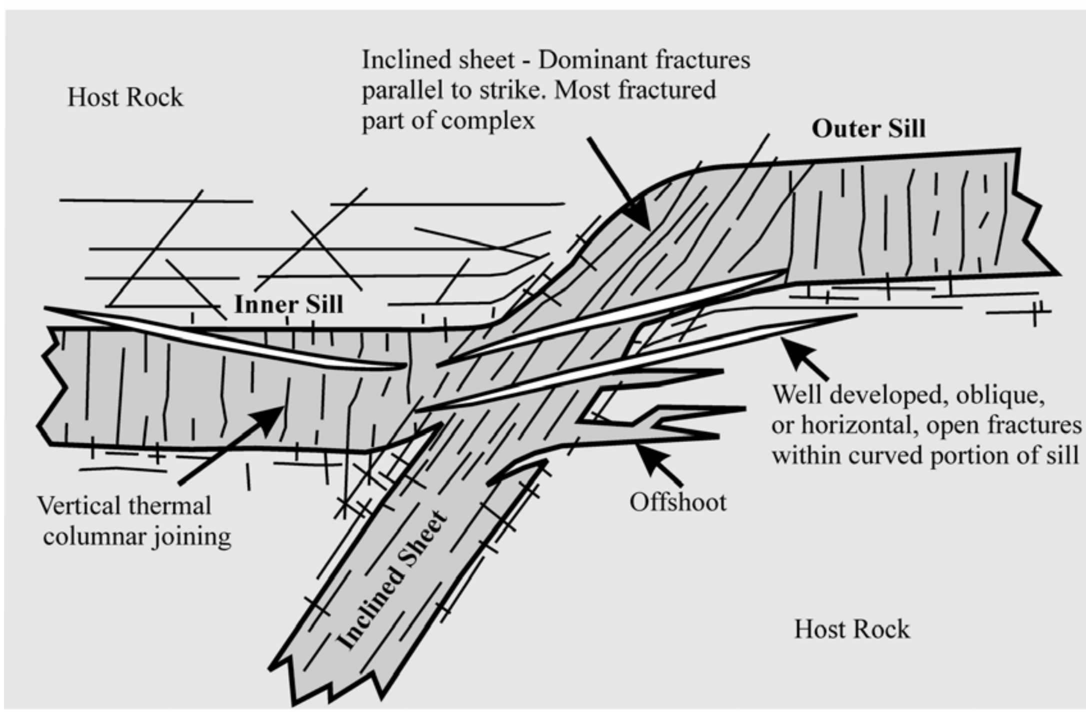


Figure 11. Different types of fractures associated with dolerite sill and ring complexes (after Woodford and Chevallier, 2002).

Chevallier *et al* (2001) proposed three morpho-tectonic models for dolerite sill and ring-complexes, with Woodford and Chevalier (2002) presented two mechanisms for the emplacement of Karoo dolerite sills:

(a) Sills in the northern Karoo have a form similar to that of a laccolith, with the feeder to the laccolith being a central dyke, with a thickening from the outer rim towards the centre of the structure. The rings are seen as peripheral offshoots formed as a result of the warping of the overlying host rock. Vivier (1996) stated that the arms of some laccoliths could be interpreted as sills if the arms of a laccolith were long enough, or, that some sills were extensions of laccoliths.

(b) In the western Karoo it was proposed a feeding system of magma along the inclined sheet, or the ring itself, using a coalescing ring-dyke network. The 60° inward-dipping inclined sheet therefore changes into an upper outer sill, feeding a lower inner sill at the same time.

The geometry of larger Karoo ring-complexes can be expanded on from (b) above in that there are four distinctive morpho-tectonic units associated with ring-complexes (Figure 10). A flat inner sill forming the bottom of the saucer with a thickness of between 30 to 60 m; a flat-lying outer sill exhibiting extensive fracturing and joining with a thickness of between 50 and 100 m, that can extend for hundred of kilometres; a peripheral inclined sheet, where dependant on the area, may dip at angles of between 30° to 80°, and attain a thickness of between 20 and 150 m: and abundant feeder dykes which cut through, into, or out, of the sills and ring-complexes.

In Figure 11**,** the three main types of complex fracturing associated with dolerite sills and ring-complexes are illustrated (Chevallier and Woodford, 1999; Chevallier *et al*., 2001; Woodford and Chevallier, 2002) and are of particular importance in that they form a basis for the storage of ground water.

- Well developed vertical thermal columnar jointing is found within the flat lying inner and outer sills.
- Within the inclined sheet, fractures parallel to the strike of the intrusion are dominant, and this is the most fractured part of the complex.
- Well developed sub-horizontal, or oblique fracture develop within the curved portion of the sill, and may be filled with secondary calcite.

2.2.4.2 *Dolerite Dykes*

Dolerite dykes are associated with the younger Karoo units from the Beaufort Group through to the Clarens Formation, and less frequently in the Drakensberg Group (Baran, 2003). Chevalier *et al*. (2001) proposed three major structural domains based on dyke distribution. There is an extremely complex and intricate relationship between dykes and sills. Dolerite dykes may feed inclined sill sheets and therefore control the shape of ring-complexes as well as uniting with adjacent rings Chevallier *et al*. (2001). Dolerite dykes are vertical to sub-vertical discontinuities that generally represent thin, linear zones of relatively high permeability which act as conduits for groundwater within the aquifer Chevallier *et al*. (2001). Alternatively, they may also act as semi- to impenetrable barriers to groundwater Chevallier *et al*. (2001). Their average thickness varies from 2 to 10 m, with dykes as thick as 300 m having been recorded, and seldom exceed 18.5 m in width, usually being between 2 and 8 m wide (Woodford and Chevallier, 2002).

2.2.5 Other Post-Karoo Intrusions

These post Karoo intrusions are mentioned only briefly as they are not directly related to this study, but are a part of the geology of the Karoo Basin.

2.2.5.1 *Breccia Plugs*

Breccia plugs are mostly restricted to the Ecca Group, occurring in clusters along the western and northern edges of the Karoo, but not restricted to this area (Woodford and

Chevallier, 2002). They occur in clusters of up to >80 plugs, with clusters varying in size from a few hundred metres to 50 km in diameter, usually forming low-relief, circular hills 50 to 80 m in diameter (Woodford and Chevallier, 2002), Alternatively they may form negative-relief depressions characterized by calcrete development with a white alteration halo around them (Woodford and Chevallier, 2002). In the western Karoo they are also associated with swarms of kimberlite fissures in the western Karoo.

It has been proposed by Woodford and Chevallier (2002) that explosive hydrothermal activity often took place when the early dolerite sills intruded into the partially indurated Karoo Supergroup sediments. Two main facies are recognised by Woodford and Chevallier (2002). The first is a molten facies, domed, baked, molten, re-crystallized, and highly contorted sedimentary host rock. These breccias contain xenoliths from the underling strata, with this facies being the most common in the field, possibly due to is resistance to erosion. The second is a breccia facies comprising fractured, broken, shattered displaced, and re-cemented blocks of sedimentary rocks. Breccia plugs may contain extensive mineralization.

2.2.5.2 *Volcanic Vents*

Volcanic vents, or diatrems, represent the first volcanic activity prior to the extrusion of the lava flows at *c*. 180 Ma (Chevallier and Woodford, 1999; Chevallier *et al*., 2001; Woodford and Chevallier, 2002). Largely restricted to the Clarens Formation they occasionally occur in the Drakensberg Mountains (Woodford and Chevallier, 2002). It has been proposed that they were probably formed by pheratic-explosive activity where excessive water and steam pressure overcame that of the magma, resulting in fragmentation, shattering, fluidization, and mobilization of the host rock and/or surrounding basalt by (Woodford and Chevallier, 2002; Svensen, 2006). In diameter, vents may vary in shape and size from a few metres to kilometres, and comprise layered successions of block- and matrix supported breccia and sandstone, as well as tuffs (Woodford and Chevallier, 2002). The only hydrothermal mineral in vents examined by Svensen *et al*. (2006) was zeolite, with cemented sandstone clasts and breccias.

2.2.5.3 *Kimberlites*

Kimberlites within the Karoo Basin occur in clusters of linear swarms of dykes, fissures, and pipes, from Sutherland in the south, through to Kroonstad in the north, as well as in the western Free State, Lesotho and East Griqualand (Woodford and Chevallier, 2002). Kimberlite fracture swarms consist of parallel fissures and associated fractures, or joints, often with an associated upwarping of the surrounding Karoo beds (Woodford and Chevallier, 2002). Kimberlites are largely composed of decomposed igneous material with quantities of crustal, or mantle, xenoliths and megacrysts (Woodford and Chevallier, 2002). Fresh kimberlite may commonly be referred to as blue-ground, weathered kimberlite as green-ground, and decomposed kimberlite as yellow-ground (Woodford and Chevallier, 2002). Deposits of well developed calcrete usually make it easy to identify kimberlites from aerial photographs (Woodford and Chevallier, 2002).

A swarm can be divided into sub-swarms of smaller size where fissures are closely spaced approximately 10 to 50 m apart (Woodford and Chevallier, 2002). Hypabyssal kimberlite can form positive-relief hills, or negative-relief, calcrete rich depressions, with a diameter of 10 to 400 m in the western Karoo, and from 200 to 1000 m on the Kaapvaal craton (Woodford and Chevallier, 2002).

Much debate has surrounded the factors controlling the emplacement of kimberlites including hot-spot tracks, crustal tectonics, and dynamics of the mantle (Woodford and Chevallier, 2002). It would appear that tectonics guided the emplacement of kimberlite swarms, with the ability of swarm sub-division indicating a vertical hierarchy of fissures, dykes, parental dykes, and larger bodies, within the fracturing system, at depth (Woodford and Chevallier, 2002).

2.2.6 Cenozoic Deposits and Soils.

The 180 Ma gap between the Drakensberg Group and the Quaternary deposits in the Free State consists of rocks which do not occur in the province, such as those belonging to the

Kalahari Group. Quaternary deposits are normally considered to be <2 Ma, and can be split in to three broad categories, namely, alluvial and colluvial deposits, calcretes, and sands and soils.

2.2.6.1 *Alluvial and Colluvial Deposits*

There is evidence of aggrading conditions along many Free State rivers in the form of river terraces. Grobler *et al*. (1988) proposed that the change from a degrading to an aggrading situation was responsible for the formation of many of the pans in the area (see section 2.3.1 for more detail). Holmes and Barker (2006) refer to a number of Free State rivers which show signs of aggradation, for example, the Orange, Caledon, Vaal, Modder, and Vet-Sand rivers.

2.2.6.2 *Calcretes*

Particularly in the arid to semi-arid western Free State calcretes, in a nodular and hardpan form, are wide spread. Calcretes form from the evaporation of groundwater and the resulting precipitation of calcium carbonate, or may form as a weathering product of dolerite (Woodford and Chevallier, 2002). Because carbonates of different ages occur, they do not appear to be related to any specific events (Myburg, 1977). Calcretes may attain thickness of 30 m, but seldom remain homogeneous over depths exceeding 1 to 5m (Woodford and Chevallier, 2002). Holmes and Barker (2006) state that due to calcretes not being related to any specific event, they serve little purpose as palaeoenvironmental indicators. Calcretes can occur as either surficial deposits, or beneath soil, or sand cover (Myburg, 1977). Nodular calcrete is often associated with unconsolidated sediments and hardpan calcrete with little, or no, surficial cover (Myburg, 1977).

2.2.6.3 *Sands and Soils*

Soil and sand formation is influenced by four factors, namely, parent material, or underlying formations; climate, topography and biological factors, of which parent material and climate have played a dominant role in the Free State (Hensley *et al*., 2006). Unconsolidated sand is a feature of large parts of the western Free State (Figure 12). Holmes and Baker (2006) refer to unconsolidated sands possibly being of aeolian origin, based on morphometric properties. They also mention the presence of lunettes and aeolian sand as a feature of current aeolian processes along fence lines and roads in the western Free State (Holmes and Baker, 2006) (see 2.1.2). Hensley et al (2006) noted that in the western Free State sands are of an aeolian nature, and were derived from the Vaal River and its tributaries over the millennia. The classification of soil and sand types will be dependant on the characteristics of the soil used in the classification, and researchers may use different criteria. Table 5 gives a comparison of soil classification after MacVicar (1973) and MacVicar *et al*. (1977), and Hensley *et .al*. (2006).

Descriptions of the soils of the Free State presented in Figure 12 have been taken from Earlé and Grobler, 1987. Freely drained latosols (#1) with a moderately advanced stage of laterization occur in the high rainfall areas of the east. A red-yellow-grey latosol plinthic catena forms the dominant soil pattern west of the former (#2-#6). The term plinthic indicates that these soils contain a highly weathered mixture of the sesquioxides of iron and aluminium as red mottles, which change to hardpan during alternative wet and dry cycles. These plinthic latosols vary due to drainage and relief. Sub-type #5 is characterized by large tracks of rock outcrop in the Vredefort area, while #6 in the west comprises well drained, sandy, partly aeolian soils.

Solonetzic soils, some with black montmorillontic clay, other with reddish clays, predominate in the central and southern areas, including some areas east of the Vredefort structure (#7-#9). These solonetzic soils have a thin porous upper layer, underlain by a hard clay rich, highly alkaline, columnar horizon. The high alkaline content of mainly sodium and magnesium causes the soil to defloculate and become impervious to water. Usually not suitable for crops, these lands are used for grazing sheep. These poorly drained, clayey, solonetzic soils do not usually support tree growth, which explains the endless grassy plains over the province, broken only by riverine bush along water

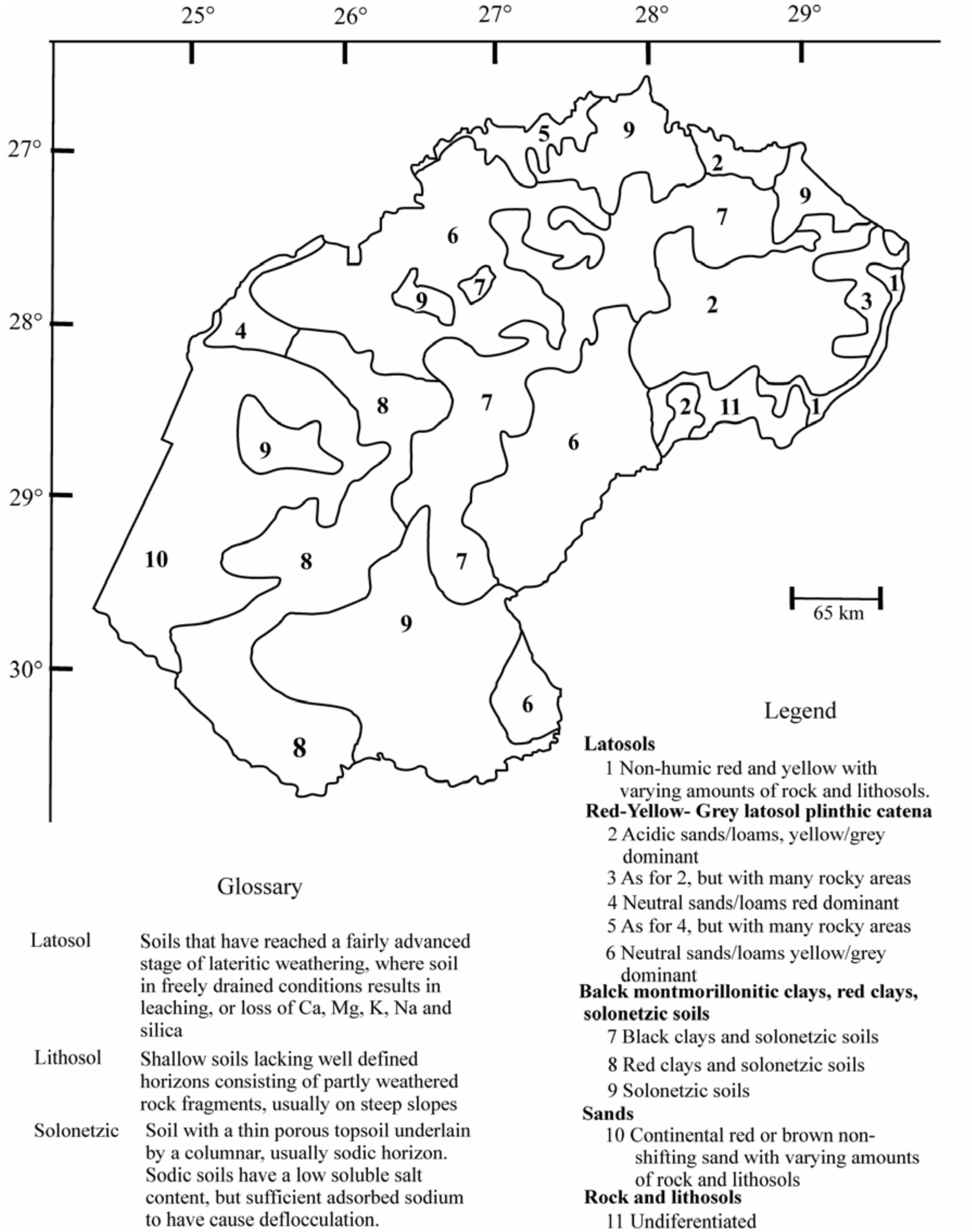


Figure 12. Soil map of the Free State (after MacVicar, 1973; MacVicar *et al*., 1977 and Lynch, 1983).

Table 5. A comparison of Free State soil classifications after MacVicar (1973), MacVicar *et al*. (1977), and Hensley *et al*. (2006).

Soil classification after MacVicar (1973), MacVicar et al. (1977), as per Fig.7.		Predominant soil categories from Hensley et al. (2006) reflected in the soil categories of MacVicar (1973), MacVicar et al. (1977).	Soil clasification after Hensley et al. (2006).	
	Latosols			
1	Non-humic red and yellow with varying amounts of rock and lithosols	A, B, C.	A	Red-yellow, structureless freely drained soils
	Red-yellow-grey latosol plinthic catena			
2	Acidic sands/loams yellow/grey dominant	B, C.	B	Plinthic catena: upland duplex and margalitic soils rare
3	As for 2, but with many rocky areas	B, C.	C	Plinthic catena: upland duplex and magalitic soils common
4	Neutral sands/loams red dominant	A.	D	Duplex soils dominant
5	As for 4, but with many rocky areas	B.	E	Dark coloured margalitic clay soils with marked swell-shrink properties
6	Neutral sands/loams yellow/grey dominant	B, C, D.	F	Shallow soils on rock
	Black montmorillonitic clays, red clays, solonetzic soils			
7	Black clays and solonetzic soils	E, D.	I	Miscellaneous soils
8	Red clays and solonetzic soils	A, C, D.		
9	Solonetzic soils	E, D, I.		
	Sands			
10	Continental red or brown on-shifting sand with varying amounts of rock and lithosols	A.		
	Rocks and lithosols			
11	Undifferentiated	F, I.		

courses. However where the lithosol is sufficiently well drained, such as on the slopes of koppies (mesas), trees and shrubs can be supported.

In the south-west of the province well drained red sandy soils of aeolian origin occur (#10). #10 comprises patches of rock and lithosol of poorly developed morphology on partially weathered rock, with some outcropping. Florisbad lies within this sub-group #10 A analogous situation occurs along the Lesotho border where similar lithosols occur with a high proportion of outcrop (#11).

2.3 GEOLOGY OF THE FLORISBAD AREA

There is no outcropping of Karoo Supergroup bedrock on the Florisbad farm (pers. obs.). Florisbad lies within the Tierberg Formation of the Ecca Group (Loock and Grobler, 1988) (see 2.1.1.3). Loock and Grobler (1988) note that the Ecca beds in the vicinity of Florisbad are sediments of the Tierberg Formation, which comprises well bedded shales and thin siltstones, the deposition of which took place through suspension settling of fine mud and silt under reducing conditions in an inland sea (Loock and Grobler, 1988). As progressively shallower conditions set in, the upper Tierberg beds of coarser material were deposited in a prograding deltaic environment (Loock and Grobler, 1988). Beaufort Group rocks have been recorded 3 km south west of Florisbad, and 14 km south-east of Florisbad on the farm Kalkwal 17, where approximately 4 metres of Beaufort Group rocks were found overlying the Ecca Group rocks (Loock and Grobler, 1988) (Figure 13). eyer and Lyle (1931) thought that the blue basal shale was of the Dwyka Group, but were later described as "Blue Ground", or kimberlite, by Dreyer (1938a). A representative of De Beers reportedly identified ilmenite and corundum (ruby) (Fourie, 1970), but Fourie (1970) noted that, with the exception of some fine ilmenite found throughout the Florisbad sands, no other kimbelitic minerals were identified. The surface area of Florisbad is composed of an unconsolidated covering of red-yellow and pale bleached

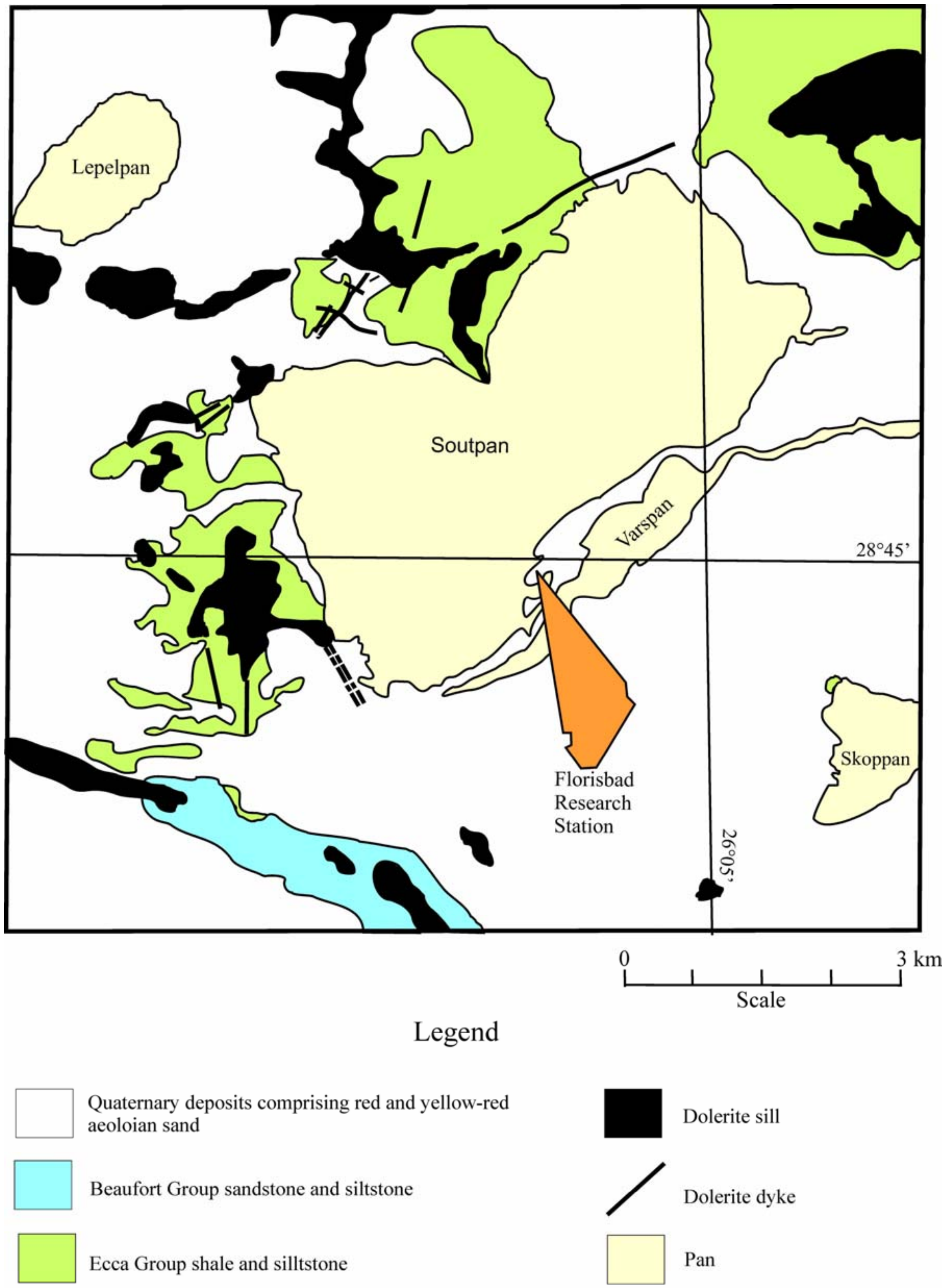


Figure 13. The geology of Florisbad and surrounding areas (after Loock & Grobler, 1988).

aeolian sand of varying depth (Loock and Grobler, 1988). The aeolian nature of the sand at Florisbad is supported by the large sand dune which straddles the Florisbad spring site, the extensive row of sand dunes lying further to the south and south-east of the site (Figure 14), and. by the presence of sand dunes along the southerly margins of most western Free State pans. The aeolian nature of the sand at Florisbad is further supported by researchers such as Grobler and Loock (1988a, 1988b), Van Zinderen Bakker (1989), Visser and Joubert, (1991), and Kuman *et al*., (1999). Calcrete horizons, which are common in the area, have been exposed through erosion that has occurred along the eastern bank of the vlei (low profile drainage system with grasses and semi-aquatic plants growing along the course – often marshy) draining from the spring site. Dolerite intrusions, intermixed with Ecca Group rocks are visible to the west and through to the north of Soutpan (Figure 14).

2.4 GEOMORPHOLOGY

2.4.1 Topography

The extreme planation of the regional landscape of the Highveld, on which Florisbad lies, was hypothesised as being the product of an erosional phase referred to as the 'Great' African planation cycle which occurred over large areas of sub-Saharan Africa (King, 1978). The African planation and denudation cycles occurred over a prolonged geotectonically stable period from the mid-Cretaceous to the mid-Tertiary (80 Ma) (King, 1978), were thought to have manifested themselves in the extensive typical grassland areas that stretch from the interior of southern Africa, through Zimbabwe, Zambia, and East Africa (King, 1978).

The topography of the western to central parts of the Free State province is generally flat to undulating, increasing in altitude from 900 m above sea level (asl.) in the west, to 1265 m asl at Florisbad (Figure 14). Kruger (1983) described the western Free State as slightly irregular plains, the north-western areas as plains and pans, and the central and southern areas as lowlands with hills. From the central to eastern areas of the province the

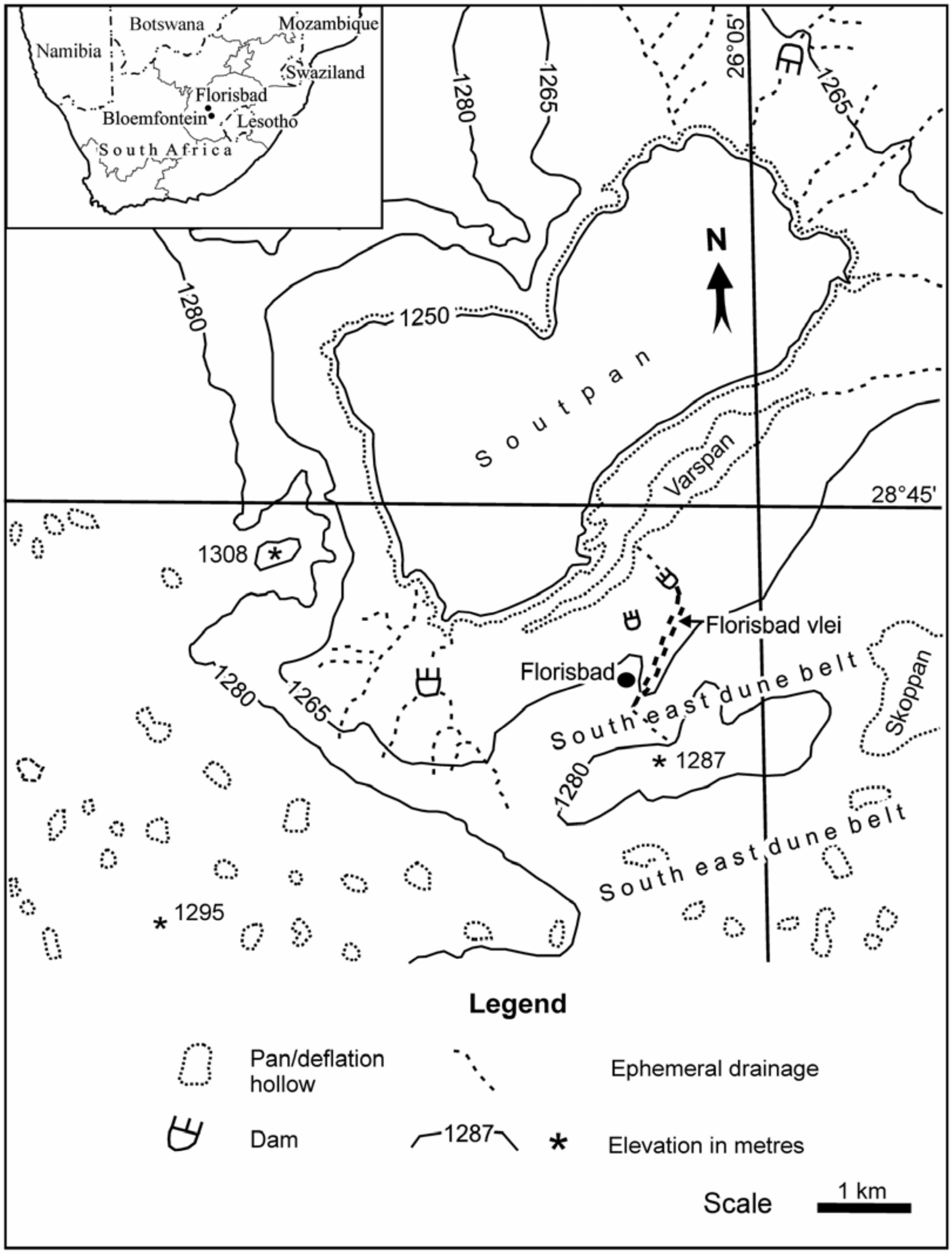


Figure 14. The topography of the Florisbad area showing the location of Florisbad and other features discussed in the text (after Appendix 4).

topography increases considerably to 3282 m asl at the Sentinel, in the Drakensberg mountain range. Kruger (1983) described this area as slightly irregular, undulating plains, with occasional hills rising to mountains. Holmes and Barker (2006) note that the Free State is dominated by areas where slope is <5%. Drainage patterns provide a good reflection of the topography of the province, and this can be clearly seen by the reduction of primary drainage from the higher lying eastern areas to the low lying western areas (Figure 15).

2.4.2 The Western Free State Panveld

Pans are the foci of poorly developed drainage basins that develop within environments of high structural uniformity and low surface relief (Goudie and Wells 1995). Florisbad is located on the eastern boundary of the western Free State panveld, west of the 500 mm isohyet (Figure 15). Verster *et al*. (1992) classed the geomorphic surface of Soutpan as being Post African Phase I. Many researchers have sought to explain the origins and formation of this panveld (Geyser, 1950; De Bruiyn, 1971: Le Roux, 1978; Marshall, 1987a, 1987b; Grobler *et al*. 1988: Marshall and Harmse, 1992). Since the classic study by Hutchinson *et al*. (1932) on the hydrology of pans and other inland waters of South Africa, considerable attention has been given to South African pans. Wellington (1945) described the western Free State panveld as having the greatest density of pans anywhere in South Africa. The floors of the western Free State pans show considerable variation in the soil and rock composition as well as vegetation cover. De Bruiyn (1971) classified pans as follows, with further subdivisions indicating the presence or absence of vegetation. However, based on observations vegetation may vary seasonally, or over the loner term, due to the influence of rainfall.

- Salt pans – occur mainly in the central region, extending to the east of the panveld.
- Calcrete pans – have the widest distribution occurring over the entire panveld with the exception of the extreme north of the panveld.
- Gypsum pans – limited to small areas in the western central part of the panveld.

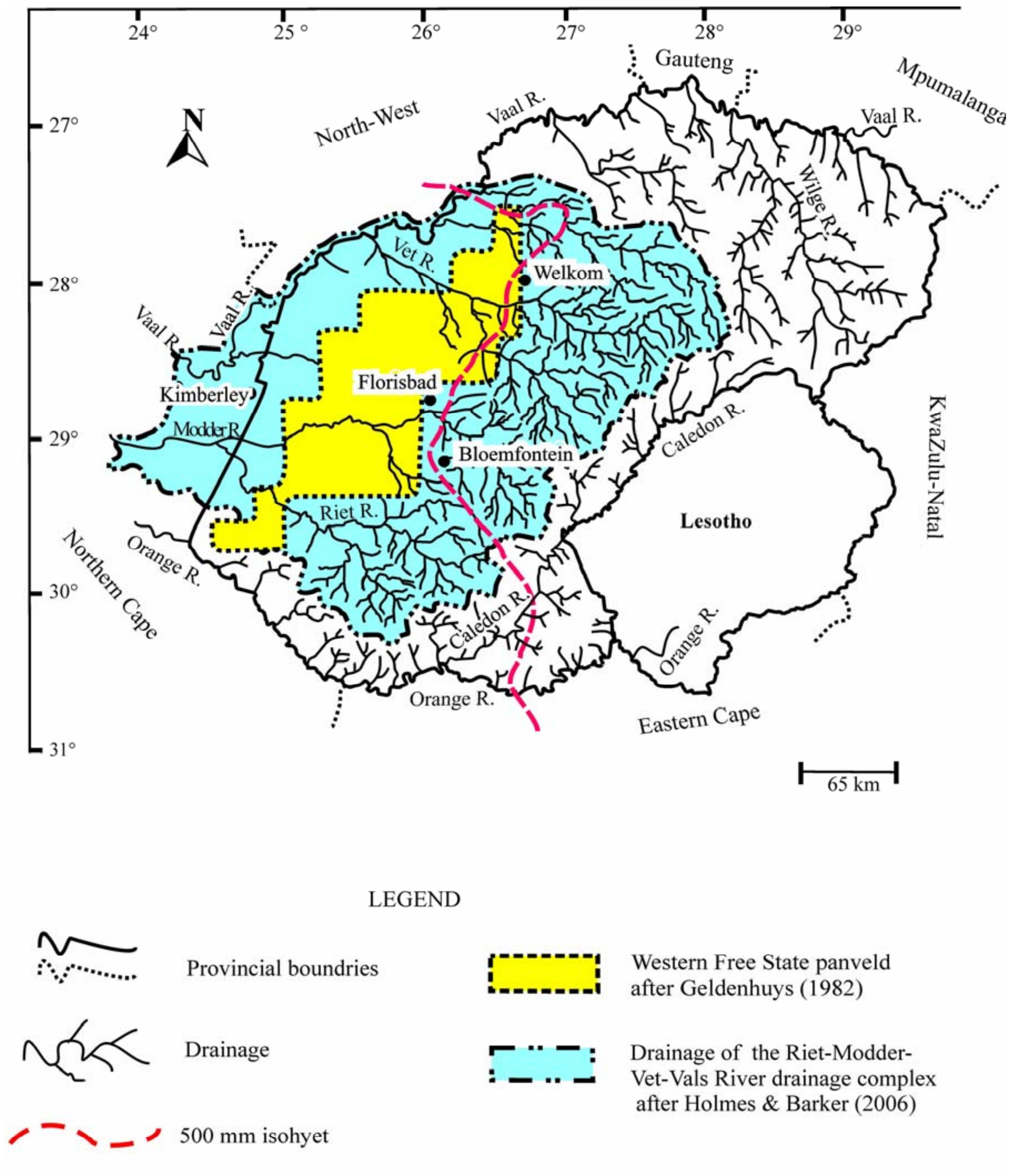


Figure 15. Drainage systems of the Free State Province showing the Western Free State Panveld as defined by Geldenhuys (1982) and Holmes and Barker (2006).

- Clay pans – occur mainly along the central eastern border, and northern areas of the panveld.
- Sand pans – are limited to a belt in the southern central area of the panveld, from the eastern border, westward to the centre of the panveld.
- Water covered pans – permanent water covered pans are limited to one near Viljoenskroon and one near Brandfort. Other pans may be artificially filled for irrigation and evaporative purposes, the latter being related to mining.

The panveld, and the influence of the panveld on faunal distribution would appear to be a somewhat unobtrusive and ignored as a barrier to biogeographic faunal distribution. Zoogeographically the panveld represents a natural western boundary to at least herpetofaunal distribution (Douglas, 1992), and possibly other fauna as well.

Dependant on the criteria used, the classification of a deflation hollow as a pan rests with the particular researcher, and therefore not all deflation hollows may be classed as pans. Therefore, the way in which deflation hollows are defined as pans will determine the boundaries of the panveld as well as the number of pans within the panveld. According to Geldenhuys (1982) the panveld covers an area of 1227 km^2, with some 8 803 pans having been recorded (Figure 15). Douglas (1992) gave a concentration high of 82 pans per km^2. Geldenhuys (1982) defined the panveld as being approximately 400 km long, running from the Vals River in the north to the Orange River in the south. Geldenhuys (1982) indicated the panveld to be an average of 140 km wide (min 50 km, max 200 km), extending eastwards from the western Free State border eastwards towards Welkom, and almost to Bloemfontein. This definition is consistent with definitions given by other researchers (De Bruiyn, 1971; Le Roux, 1978; Earlé and Grobler, 1987; Marshall, 1987a, 1987b; Seaman *et a*l., 1991; Lawson and Thomas, 2002). Although not defining the boundaries of the panveld, Holmes and Barker (2006) showed all pans occurring within the Riet-Modder-Vet-Vals river drainage complex. The highest concentration of pans indicated by Holmes and Baker (2006) clearly fits the boundaries of the western panveld as defined by other researchers, with Le Roux's (1978) delineation of the western panveld being similar to that of Geldenhuys (1982). Le Roux (1978) also indicated large, but very low density, panvelds in the southern, extreme east, and north-east of the province, with minor low density panvelds at, or near, Bethlehem, Harrismith, Kroonstad and Parys.

Data from Geldenhuys (1982) indicates that 54% of all pans occurring in the western Free State, occur in the two degrees, 2825 and 2826, representing 4719 pans (Figure 15). Analysis of further data indicates that size of pans varies considerably with 2561 (54.3%) pans being <2 ha in extent, 1855 (39%) being 2-2.5 ha, 303 (6.4%) being > 25 ha. The southern areas of the panveld, as defined by Geldenhuys (1982), is blanketed by Quaternary red and grey aeolian sand with areas of calcrete and surface limestone. Largely underlying the Quaternary sands, and having more exposure towards the north, are Tierberg and Volksrust Formation rocks of the Ecca Group, with some Adelaide Subgroup rocks of the Beaufort Group occurring along the eastern boundary (Figure 8). Rocks of the Volksrust Formation and Adelaide Subgroup again become covered by Quaternary deposits in the region of the Vet River to the north. Dolerite sills and dykes are common over the entire area.

Holmes and Barker (2006) recorded 16 830 pans occurring in the Riet-Modder-Vet-Vals river drainage complex. If the 10 253 pans recorded by Holmes and Barker (2006) as occurring on the Ecca Group rocks only are considered, this number approximates the 8 803 pans recorded by (Geldenhuys, 1982). Geldenhuys (1982) classified pans based on the breeding of water-fowl, where the criteria for defining pans was based on vegetation growth, determined two months after flooding (see Section 4.3).

The derangement of drainage patterns, or the disturbance of drainage patterns, by tectonic and/or climatic factors, as a possible cause for the pans was suggested by Wellington (1945) and Geyser (1950). Le Roux (1978) disagreed with this hypothesis, stating that it could not account for the wide distribution of pans, and that climatic derangement in the form of wind was the only agent that could have been responsible for the formation of most pans. Grobler *et al*. (1988) saw pan formation being largely as a result of climatic causes and stated that whatever the causes, the hydrodynamic equilibrium of the streams was altered from an actively degrading drainage to an aggrading situation. When dry periods occurred, prevailing winds deflated the sediments, forming hollows, which were the initial sites of the pans (Grobler *et al*., 1988). Lancaster (1978) also stated that deflation was the major factor contributing to the formation of pans, while Goudie and Wells (1995) noted that bedrock weathering was promoted by un-vegetated, dry and saline conditions, with the relationship between pan hydrology, sediment supply, and deflation, being complex. Goudie and Wells (1995) concluded that the result of deflation was the formation of pan-margin lunettes from deflated pan sediments.

Le Roux (1978) stated that there was no evidence of river capture, or backward tilting, as a cause for pan formation in the then Orange Free State. Van Zinderen Bakker (1989) was of the opinion that the Jurassic dolerite intruded the Karoo beds forming sills and dykes, with pans such as Soutpan, which is one of the largest pans at 19.4 km^2 or 1 940, ha, 1.6 km north-west of the spring site, being formed during subsequent erosion cycles through deflation.

Marshal (1987a) examined the panveld within a morpho-tectonic framework and postulated that the pans were remnants of a tectonically disturbed major palaeodrainage system that previously drained the area between the Vaal and Modder Rivers in an east to west direction. This system was referred to by Marshall (1987a) as the palaeo-Kimberley River. Marshall (1987a) envisaged an ancient Modder River flowing directly north-west from near Bloemfontein, to join the Vaal River at Christiana. Through headward erosion and Miocene structural displacements, the palaeo-Kimberley River eroded eastwards and captured the middle reaches of the ancestral Modder River, which then dried out to the north (Marshall, 1987a). At this time the the ancestral Riet-Modder River drained parallel to the palaeo-Kimberley River. Later Pliocene uplift in the region of the ancestral Riet-Modder River also resulted in headward erosion eastwards, which again resulted in the capture of the upper reaches of the ancestral Modder River at the present day elbow at Lombards Drift (Marshall, 1987a), close to Florisbad. Downwarping of the palaeo-Kimberley River during the same period resulted in the disruption of the river, with

Quaternary climates drying out the rivers to form the beginnings of the modern day panveld.

2.4.3 The Florisbad Sand Dune

The Florisbad sand dune, a crescent shaped aeolian deposit, is something of an enigma. Possibly, due to the excavations being concentrated in the area of springs activity, and a preoccupation with the term spring mound (Brink, 1987) no research appears to have been carried out on the Florisbad sand dune, as an entity, in order to determine its actual status. The Florisbad sand dune has been previously referred to as a lunette (Brink, 1987, Loock and Grobler, 1988) and as a spring mound (Brink, 1987). The latter term is disregarded as it does not seem plausible that such large quantities of spring sand would be available in order to replicate a sand dune to such an extent. There also appear to be no other such examples of springs in South Africa producing such large deposits. While it is not questioned that, in some areas the base of the dune does rest on such spring deposits it is felt that these are minimal. Owing to the status of the Florisbad sand dune still being undetermined, and therefore unclassified, due to the lack of information on its composition, sedimentation, and structure, the dune has been referred to in this thesis simply as the Florisbad "sand dune", and no attempt has been made at classification.

Pan fringe lunette dunes within the western Free State panveld, and in the Kalahari, are formed on the lee side of deflation hollows mainly by unconsolidated material being blown from the pan floor by a prevailing north-west wind (De Bruiyn, 1971; Goudie and Thomas, 1986; Loock and Grobler, 1988; Van Zinderen Bakker, 1989; Lawson and Thomas, 2000; Holmes *et al*., 2008). Such dunes are easily recognizable on aerial photographs, and in the field, from their peculiar vegetation cover, colour, shape, erosion degradation, and topographic expression (Loock and Grobler, 1988; Holmes *et al*. (2008). Another characteristic of lunette dues in the southern African region is their often extensive gully erosion on the windward side of the lunettes, with Holmes *et al*. (2008) noting that all lunettes examined by him in the western Free State panveld showed signs of erosion degradation. These gullies allows for the recycling of sediments, where dune sediments are washed back into the pan, and then re-deposited on the windward slope of the dune by aeolian action during drier periods (De Bruiyn, 1971; Telfer and Thomas, 2006; Holmes *et al* 2008).

In many instances, pan-margin sand dunes are inextricably related to pan formation in the drier areas of southern Africa, as indicated by Lancaster (1978, 1986, 1989) in the southern Kalahari of Botswana, by Lawson and Thomas (2002) west of the Molopo River valley in the south-west Kalahari, and De Bruiyn (1971) in the Free State. Of the ten pans examined by De Bruiyn (1971), with the exception of one, all had sand dunes lying to the south, south-east, and east. De Bruiyn (1971) also noted that these sand dunes had been eroded, in some cases rather severely, by small courses on the windward side. Pan margin related sand dunes therefore form an integral part of the pan system in the drier regions of southern Africa. Telfer and Thomas (2006) used optically simulated luminescence dating to show that lunette features are spatially complex and that their formation may be relatively rapid. A 5 m accumulation of sediment over 570±40 years, and a 6 m accumulation over 660±40 were recorded by Telfer and Thomas (2006) at Witpan. Studies on the Witpan lunette showed that formation may have taken place within 2 ka, and that primary deflation from the pan had not contributed significantly to the formation of these lunettes (Telfer and Thomas, 2006). In fact, Telfer and Thomas (2006) noted that most of the lunette material was derived from the recycling of older lunette sediments from neighbouring dunes, with some contribution from the linear dunes.

It is also recognized that the southern African Kalahari dune system, and that of the Sahel of West Africa, were both established during multiple arid phases since the last interglacial (Thomas *et al*. 2005). Therefore, climate plays a critical role in dunefield dynamics and the interplay between dune surface erodibility and atmospheric erosivity (Bullard *et a*l. 1997; Thomas *et al*. 2005) and thus the mobility of sand dunes. Lawson and Thomas (2002) state that elsewhere lunettes have been derived from deflated sediments transported downwind to the pan margin by wave action. Thomas *et al*. (1993) state that the relatively small size of most southern African pans mitigates against inundation related lunette development, particularly in the Kalahari. In the case of

Florisbad, the considerable size of the adjacent Soutpan does not place it in this category, and its extensive surface area would possibly have made it more conducive to deflation factors and the formation of the Florisbad sand dune.

Lawson and Thomas (2002) saw pan-margin lunettes as palaeoenvironmental indicators, and their view supports the theory put forward in this thesis that the Florisbad sand dune may hold the key to the history of the site. Pans with more than one lunette could possibly preserve evidence of episodic, or multiple, episodes of aridity with dune orientation indicating variations in the prevailing wind direction and thus the direction of aeolian deposition (Lancaster, 1978). Young and Evans (1986) note that a large portion of pan deflated sediments are deposited within a short distance of the pan on the downwind side. Pan deflation occurs only to a limited degree today in the Kalahari and significant lunette construction is not evident, therefore pan margin lunettes may be regarded as palaeomorphs, representing periods which indicated favourable deflation conditions (Lawson and Thomas, 2002).

Lawson and Thomas (2002) note that the Kalahari pans are the most westerly expressions of closed deflated basins and display characteristics of pan basins operating under relatively arid conditions. Therefore, pan and lunette processes which operate in the Kalahari may be similar to those which operate to the east, and consequently may have implications, and shed greater insight, into contemporary and recent environmental processes over much of southern Africa and other dryland regions.

The Florisbad dune rises 27 m above the pan floor and is located 1.65 km from the pan margin. South and south-east of Florisbad aeolian deposition rises to 37 m above the pan floor on the crest of the south-east dune belt, while south of the Modder River, this aeolian deposition continues to rise to 93 m above the pan floor. It would appear that the lands on the farm directly west of Florisbad were once used for agricultural purposes and reached almost to the pan shore, indicating fairly deep aeolian deposits.

It is evident from aerial photographs and satellite images that vast areas of the western Free State were blanketed with a thick layer of aeolian deposits. It is suggested that, if Florisbad and surrounding areas were blanketed with a thick layer of aeolian sand during the Holocene, the pans would also have been covered. Considering the extensive area of Soutpan, this accumulation on the pan floor would have been quite considerable. Therefore, it is postulated that at various times phases of heavy aeolian deposition covered Soutpan with sand from outside the area. Subsequent dry and windy periods then deflated the recently deposited aeolian sands from the pan floor, until the harder pan floor was again exposed.

It is suggested that it was these more recent aeolian deposits that were responsible for the formation of the sand dune which, as it grew, migrated towards the Florisbad spring site. These deflated deposits then further contributed to the formation of the dunes south and east of Florisbad. Alternatively, it is possible that an outer lunette may have formed near the Florisbad spring site, to be later covered by aeolian deposition. This would imply the presence of a fossil lunette beneath the more recent aeolian deposits. The four metres of red sand recorded by Rubidge and Brink (1985) on the western flank of the site would tend to support both these hypotheses. De Briuyn (1971) has recorded individual sand dunes forming on the floor of the Kalgat (421) pan. Holmes *et al*. (2008) noted that the upper levels of the lunettes at Morgenzon and Sunnyside Pan both contained from 80-95% sand. This could possibly be related to periods of Holocene aeolian sand deposition and deflation.

What is significant about Soutpan is that, although it is not entirely devoid of fringing lunettes, such lunettes are far less common and pronounced than at other pans in the area (Figure 16), particularly when considering the vast expanse of the pan. Fringe lunette formation at most pans in the area, including pans such as Morgenzon, Sunnyside, and Geluk Pan, are very distinct, well defined, and clearly merge with the pan margins (Figure 16). As mentioned in subsection 2.4.2, De Bruiyn (1971) recorded nine out of ten pans examined by him as having distinctive lunette margins.

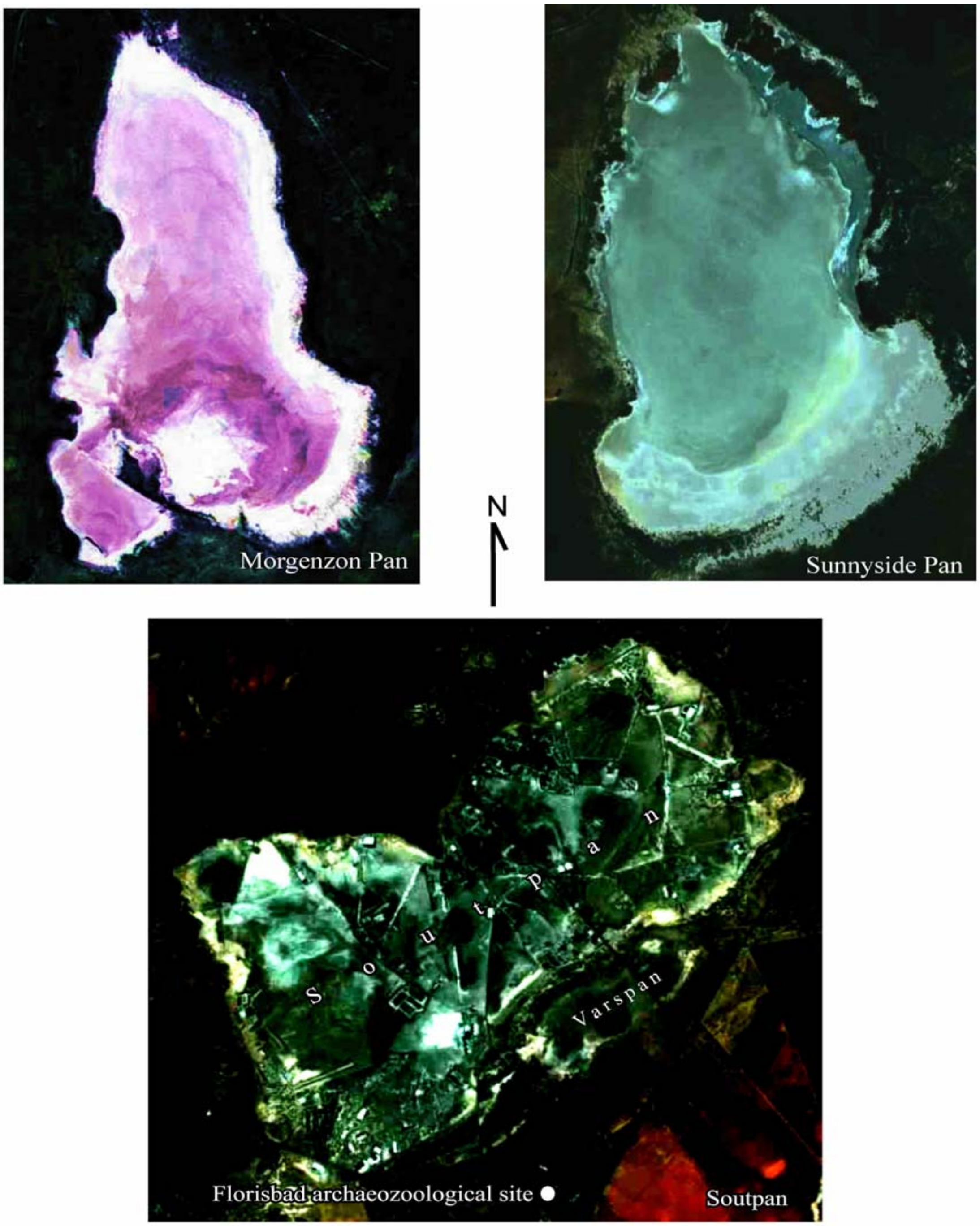


Figure 16. Google Earth images giving a comparison of the south and east pan fringe lunette development between Morgenzon Pan, Sunnyside Pan and Soutpan. Note the low degree of pan fringe lunette development at Soutpan and Varspan and the orientation of the pans. Images have been adjusted to define the pan fringe lunettes in more detail (Not to scale).

De Bruiyn (1971) also noted that pan fringe lunettes did not attain great heights. Holmes *et al*. (2008) recorded a prominent fringing lunette with a height of 5 m, 60 m from the edge of Morgenzon Pan, while similar sized lunettes were recorded at Sunnyside Pan, Deelpan and Salpeterpan.. Lawson and Thomas (2002) and Holmes *et al*., (2008) mention outer lunettes which can occur at far greater distances from the pan margins than pan fringing lunettes themselves. Lawson and Thomas (2002) recorded the height of an inner lunette at Koopan Suid at 9 m, while an outer lunette was recorded 1,25 km from the pan with a height of 80 m. This would beg the question as to whether or not outer lunettes occur at the base of the Florisbad and south east dune belt dunes. Holmes *et al*., (2008) recorded an outer lunette 660 m from the primary lunette at Morgenzon Pan.

Rubidge and Brink (1985) drilled 31 auger holes at the site, and although the thickness of individual lithostratigraphic units, sand size and colour were recorded, little information was produced regarding the composition and status of the dune. That up to 4 metres of red to brown sand were recorded on the western side of the springs and towards the top of the dune (Rubidge and Brink, 1985), would seen to confirm a thick layers of more recent aeolian deposition. By examining sand size of the top 200 mm of the dune over a four km distance, from the Soutpan shore in a south-easterly direction, Loock and Grobler (1988) conclude that the true aeolian nature and origin of the dune had been established. As previously mentioned, Mucina and Rutherford (2006) noted that aeolian dust may be transported several thousand metres into the air by strong winds.

It is put forward here that many of the questions relating to inconsistencies in the stratigraphy of the site, particularly through auger drilling, (see subsections 2.3.3 and 2.3.4), are directly related to the formation of the dune and its internal stratification. In this instance, a combination of factors such as variations and t Rubidge and Brink (1985) drilled 31 auger holes at the site, and although the thickness of individual lithostratigraphic units, sand size and colour were recorded, little information was produced regarding the composition and status of the dune. That up to 4 metres of red to brown sand were recorded on the western side of the springs and towards the top of the dune (Rubidge and Brink, 1985), would seen to confirm a thick layers of more recent

aeolian deposition. By examining sand size of the top 200 mm of the dune over a four km distance, from the Soutpan shore in a south-easterly direction, Loock and Grobler (1988) conclude that the true aeolian nature and origin of the dune had been established. As previously mentioned, Mucina and Rutherford (2006) noted that aeolian dust may be transported several thousand metres into the air by strong winds.he composition of aeolian deposition, the aquatic environment which existed when the dune initially moved over the site, and most importantly, the angle of internal cross-stratification of the dune, where the angle of repose on the leeward slope is usually between 30 and 35 degrees, but as the dune migrates, the angle of internal stratification decreased down the windward slope (Appendix IV), are responsible.

It is suggested that the vegetation cover and the low angle of repose of the Florisbad dune, as well as the dunes making up the south-east dune belt, indicate that the transportation of aeolian material has been inactive for some considerable time. According to Loock and Grobler (1988), the Florisbad site varies from other dunes in the area in that it has a higher brackish soil content which has been derived from Soutpan.

In summary, the Florisbad sand dune shares no similarities or characteristics with other lunette dunes referenced in this subsection, nor to those observed from aerial photographs and satellite images. Further to this, characteristics such as size, height above the pan floor, distance from the pan, vegetation, extensive gully erosion and possibly shape, all preclude the Florisbad sand dune from being classified as a lunette in the sense that the term has previously been used. It is however acknowledged that an as yet undetected fossil outer lunette may exist under the more recent aeolian deposition. As no detailed study of the Florisbad san dune has been undertaken in order to establish its true status, it is proposed that the Florisbad sand dune is more closely related to a barchanoid dune, rather than a lunette. Only further analysis and dating at greater depths will determine the Florisbad sand dunes true nature and status.

In that the Florisbad dune is considered here as being more barchanoid related, having being formed from aeolian sand deposits originally from outside the area, rather than a lunette comprising deflated pan floor and recycled lunette material, periods of dune activity in the south-west Kalahari may be pertinent. In data compiled by O'Conner and Thomas (1999), from Thomas *et al*. (1997), Stokes *et al*. (1997) and Eifel and Blumel (1998), it is shown that primary phases of linear dune activity, which is thought to have had an influence on the Florisbad sand dune, occurred at 30-23 ka and 16-10 ka, with the probability of the 16-10 ka high activity phase being extended to 17-8 ka. Holmes *et al*. (2008) recorded lunette building periods in the western Free State panveld at 12-10 ka, 5.5-3 ka, 2-1 ka, and 0.3-0.07 ka, indicating the more recent nature of lunette development.

2.5 SYNOPSIS

By relating the regional geology of the Free State to the Karoo Basin, a foundation has been established for understanding the geology of the Florisbad area, as well as the many aspects that will emanate from the geology in future chapters. The importance of geomorphological features, such as the western Free State panveld and the Florisbad sand dune, that have played such an important role in the morphology, formation, sedimentation, and chemistry of the Florisbad spring site, have also been brought to the fore as background, and as a prelude to their contributions towards the site. A new hypothesis has been proposed regarding the formation of the Florisbad sand dune.

Two other critical components, namely, climate and vegetation, which have also played major contributing roles in the depositional environment at the Florisbad spring site, will be dealt with in the following chapter.

Chapter

3

The Physical Environment

Part II

Climate and Vegetation

CHAPTER 3

PART II

CLIMATE AND VEGETATION

3.1 INTRODUCTION

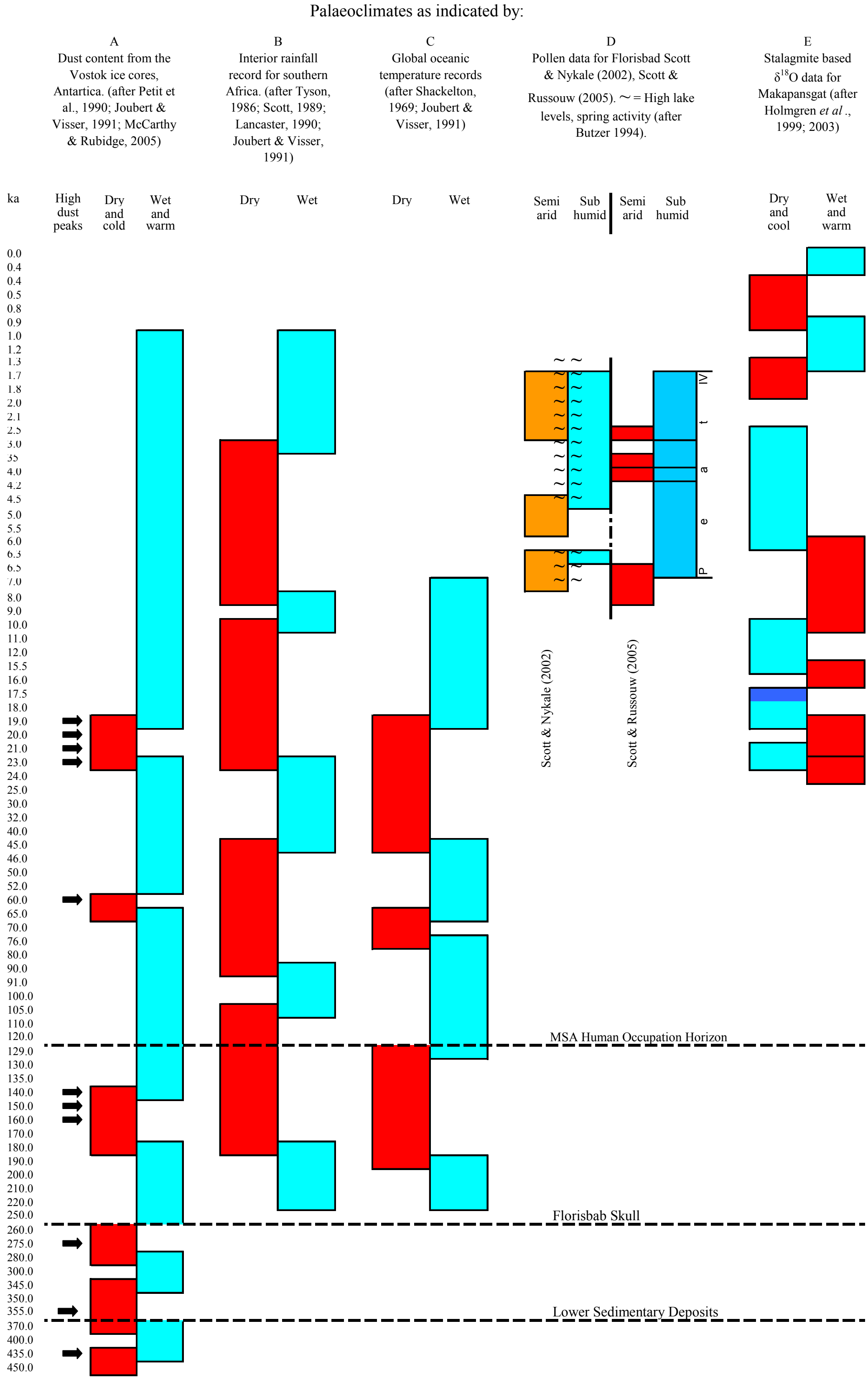
The various facets of climate are responsible for a number of processes that have contributed to the depositional and geohydrological environment at Florisbad. These would include weathering, diagenesis, chemical reactions, the formation of aquifers, including related aspects such as recharge and flow, aeolian deposition, and vegetation. Climate would have influenced the type and production of vegetation, which in turn would have determined the species composition of the area. Vegetation would also have influenced the morphology of the area by varying the degree of sand dune mobility during long-term wet and dry periods.

The climate of the Free State is partly influenced by the relief of the province and is associated with a west to east increase in altitude from 900 to 3282 metres asl. Florisbad Research Station is located on a plateau almost in the centre of the southern African region (Brink, 1987). This effectively prevents the free circulation of air from the costal regions from reaching the inland plateau, resulting in Florisbad having a climate of extremes (Brink, 1987). Aspects of climate are discussed below.

3.2 PALAEOCLIMATE

Palaeoclimate is important in that it would have determined aspects such as temperature, wet and dry cycles, vegetation, and thus the environment in which the Florisbad spring site was formed, and faunal remains fossilized. Many different factors have been used in attempts to project palaeoclimate locally and on a world wide scale. A few of these factors such as pollen, stalagmites, dust, rainfall and temperature, are presented in Table 6. In many instances, the projection of

Table 6. An interpretation of examples of global and local palaeoclimates as determined by various authors (Log scale).



palaeoclimate is uncertain and speculative. Table 6 gives examples of palaeoclimatic interpretations based on a various factors and reflect the differences in the results between broader surveys covering extended periods, and the more detailed results obtained from shorter term individual site climatic predictions.. It will be noted in Table 6 that there is often little correlation of the broader interpretation of palaeoclimatic conditions between studies. It is also apparent that as more intense and detailed research is carried out for a specific region or site, although be it over shorter periods, the accuracy of palaeoclimatic projections may become more reliable. For example, on a broader scale the hydrogen isotope ratio, temperature, oxygen isotope ratio, and sea surface temperature of the Vostok ice core analysis all follow the highs and lows of the core dust content closely (McCarthy and Rubidge, 2005). This would indicate that as these highs and lows are comparable, they are also fairly accurate in supporting the occurrence of events at a particular time but reveal little regarding short-term palaeoclimate. Variability at a regional level may be judge from data compiled by O'Conner and Thomas (1999) which shows the differences in phases of linear dune activity (arid periods) for western Zambia, western Zimbabwe and the south-west Kalahari, but all within the Kalahari region.

Even although palaeoclimate is an all encompassing term covering millennia, palaeoclimate as such must be seen in a very site specific context. This is largely because of the many interrelated climatic factors, as well as factors such as the geology, vegetation, deposition and erosion, all on both a micro- and macro-scale, will differ considerably between sites, even within a region. This is illustrated in studies such as those by Scott and Nykale (2002) and Scott and Russouw (2005) on the Florisbad pollen, and Holmgren *et al*. (1999; 2003) on the Makapansgat stalagmite. It will be noted from columns "D" and "E" in Table 6, that there is little correlation between the two sites, and attempts to correlate different sites in different regions are unlikely to correspond due to climatic variability and local conditions.

It can be deduced from evidence presented by previous researchers that, unlike the present day relatively dry climate, the Florisbad area previously experienced more humid periods (Butzer, 1984; Brink and Lee-Thorpe, 1992; Scott and Nyakale, 2002). Joubert and Visser (1991) suggested that lacustrine deposits had formed at the Florisbad site due to the water level from the nearby Soutpan rising and expanding to an area of twice its current size during cyclical humid periods. Scott and Brink (1992) stated that the pollen levels could not be precisely linked to layers cited in the sedimentological study by Visser and Joubert (1991). Scott and Russouw (2005) noted that the high water levels proposed by Visser and Joubert (1991) were in conflict with the pollen evidence where Chenopodiaceae pollen was an indication of dry conditions. Scott and Brink (1992) also contended that the findings of Van Zinderen Bakker (1957), who interpreted the ratio between Compositeae pollen and grass pollen as an index of dry periods, was contradictory to the rises of the palaeolake as proposed by Visser and Joubert (1991). Rises and falls in the lake complex were correlated by Joubert and Visser (1991) to past southern African and global climatic data. Scott and Brink (1992) also rejected Van Zinderen Bakker's (1957) interpretation and proposed that an increase in the ratio between Chenopoiaceae, grasses, and other pollens signified moist periods, while a decrease in the ratio signified dry periods.

The presence of fresh water gastropods in the clay facies of the Florisbad site also indicates the previous presence of large bodies of fresh water (Fourie, 1970; Visser and Joubert, 1991). Other indications of a more humid past is the fossil evidence of species which might rely on an aquatic environment, such as hippopotamus (*Hippopotamus amphibius*), lechwe (*Kobus leche*), clawless otter (*Aonyx capensis*) and water mongoose (*Atilax paludinosus*) as well as the possibility of waterbuck (*Kobus ellipsiprymnus*) (Brink, 1987; Henderson, 2001a; Appendix II).

Brink and Lee-Thorpe (1992) suggested that the area was grassland with a grazing succession similar to that of the Serengeti in east Africa. This suggestion was based on the presence and grazing habits of *Antidorcus bondi*, an extinct springbok, and a relative of the extant springbok *A. marsupialis* (Brink and Lee-Thorpe, 1992). *A. bondi*, unlike *A. marsupialis,* was a specialized grazer and is thought to have fed almost entirely on newly sprouted grass shoots, which could only be ensured through regular mowing by larger herbivores and sufficient soil moisture (Brink and Lee-Thorpe, 1992). This specialized diet meant that *A. bondi* could not shift its diet in winter, and therefore the veld must have had a high primary production to allow for year-round production of new shoots, which could only have been attained with year-round high soil moisture content (Brink and Lee-Thorpe, 1992), and a frost-free climate.

It must therefore be assumed that localized sedimentary deposits, such as the peat layers, were formed during wet periods, and that the sand layers were accumulated during dry periods. This would be accordance with the stratigraphy given in Figure 4 where major wet and dry periods are reflected in the cross-section. There does seem to be consensus between the hypothesis proposed in this thesis that the water level in the Florisbad dam rose to the top of the arms of the san dune, and palaeoclimate. Both Scott and Nykale (2002) and Scott and Russouw (2005) concur that the development of Peat IV occurred during a predominantly wet period. Scott and Russouw (2005) also concur that Peat IV almost reached to the top of the site during this wet period, with its width indicating that the wet period must have been a considerably extended one. From the results of Scott and Nykale (2002) and Scott and Russouw (2005) who base their analysis on pollen derived from the surrounding area, it can also be deduced that the Florisbad dam was not filled entirely by spring-water during this wet period.

As the stratigraphy of the sand dune and sedimentary deposits vary considerably, this would reflect a multitude of different environmental influences, with each section having to be considered independently. Greater detail on the environmental history of the site is given under section 5.6.1 and many of the other sections. Another drawback with the current state of palaeoclimate analysis and interpretation is that it is very much limited to the near surface sediments. As will be seen from Table 6, these interpretations come no where near to resolving issues in important layers such as the MSA Human Occupation Horizon, the Florisbad skull, or the lower sedimentary deposits where the Old Collection is to be found. However, much more research is required in order to associate the sedimentary developmental stages at Florisbad directly to the palaeoclimate.

3.3 PRESENT CLIMATE

3.3.1 Rainfall

Evidence presented in the appendices (Appendix I, II, III, IV) indicates that rainfall was a critical factor in the mobilization and transport of ions for the salinization of the ground water, the salinization of the clay and organic material, and the fossilization of faunal remains. Rainfall was also a critical factor to the formation of the Florisbad spring site, with alternating layers of sand and organic clay, reflecting alternating dry and wet periods (Figure 4).

The South African climate is almost entirely under the influence of circulation to the west of the country, with changes being dominated by disturbances in the southern hemispheres westerly circulation, which in turn appear as cyclones or anticyclones, moving across, or around the coast. (Schulze, 1994). Climatic conditions are further influenced by factors such as latitude and solar radiation, altitude, position relative to land and sea, and ocean currents and temperature (Schulze, 1994). Rainfall in the Free State can be brought about by three basic weather systems (Louw, 1979; Schulze, 1994; Kruger, 2007).

3.3.1.1 *Convergent Rainfall*

Convergent rainfall is the primary cause of rainfall over the interior of South Africa, including the Free State. This rainfall is largely as a result of a cut off low pressure cell forming over southern Africa, and drawing in a broad stream of warm, moist, equatorial air, from the north and north-west. This influx of warm moist air, and concentration of air in the lower atmosphere, causes a simultaneous outflow, or divergence, in the upper layers. A well developed cell of low pressure over southern Africa favours strong convergence over the Free State, resulting in convective storms and heavy rainfall.

3.3.1.2 *Orographic Rainfall*

A low pressure system over southern Africa is usually followed by an anticyclone moving from the west, to the north east along the coast, gathering moisture from above the warm Auglhus and Mozambique ocean currents. Easterly and south-easterly air is then forced against the Drakensberg escarpment, resulting in orographic rain along the eastern side of the country, which affects the eastern and north-eastern Free State. At the same time this anticyclonic system also promotes uplifting and convergence of the equatorial air flowing southwards. Orographic rainfall is also important over the south-western and southern Cape.

3.3.1.3 *Frontal Systems*

These cold systems, which originate in the higher latitudes of the Atlantic Ocean to the west of South Africa, where warm air is forced to rise, cool, and moisture to condense. Frontal systems play an important role in both summer and winter rainfall, but particularly in winter rainfall over the south-western Cape and along the Cape coast, while strong frontal systems also bring some winter rainfall to the Free State. Contrary to the above three systems, the development of a high pressure cell over southern Africa results in dry periods.

3.3.1.4 *Rainfall of the Florisbad Area*

Rainfall in the Free State province increases from west to east, as illustrated by the mean annual rainfall for Jacobsdal (west) of 349 mm, and the mean annual rainfall for Witzieshoek (east), 1016 mm (Douglas, 1992). Florisbad Research Station falls within a summer rainfall region where most rainfall occurs between October and March, with approximately 78% of rainfall occurring in summer (Earle and Grobler, 1987).

Florisbad lies just west of the 500 mm isohyet within the 400 mm to 500 mm mean annual rainfall zone of Van der Wal (1977). The 500 mm isohyet is of particular importance because, as a meteorological feature, areas receiving rainfall above 500 mm are regarded as being suitable for dryland agriculture. (Petja *et al*., 2004). As with all isohyets, the position of the 500 mm isohyet is variable, with its position being determined by fluctuating seasonal and annual rainfall. The mean annual rainfall for the 78 year period 1922 to 1998 was 496 mm, with a minimum rainfall of 271 mm in 1965 and a maximum rainfall of 957 in 1988. When expressed as a percentage of the mean annual rainfall, between 1961 and 1999, Florisbad experienced a maximum of 175% of the mean, and a minimum of 46% of the mean. Long dry periods are characteristic of the area with mean annual rainfall for the period 1964 to 1970 being 406 mm and for the period 1982 to 1986 being 379 mm (Douglas, 1992). Figure 17 provides a 23 year rainfall record for Florisbad showing the considerable annual variation in rainfall. More details on the rainfall at Florisbad can be found in Appendix I.

3.3.2 Temperature

Figure 18 gives an example of mean monthly maximum and minimum temperatures for the Florisbad area. Temperatures in the Free State province decrease from west to east. This is illustrated by the mean January (summer) temperature for Boshof (west) of 24 °C and the mean January temperature for Harrismith (east), 18 °C (Douglas, 1992). The mean July (winter) temperature for Boshof is in the order of 10 °C and the mean July temperature for Harrismith, 7 °C (Douglas, 1992). In the Florisbad area, temperatures can vary considerably between summer and winter, with a maximum of 38.5 °C (January) and a minimum of –7 °C (June) having been recorded at Glen Agricultural College, 32 km east of Florisbad.

3.3.3 Evaporation

The most important requirements for the deposition of salts in a semi-arid environment such as Florisbad, and the salinization of sands, soils, and organic material, are evaporation and capillarity. Capillarity is reliant on the conductivity of the soil in relation to the surface evaporation rate (Hillel, 1971), and will increase in fine textured soils of high clay and humus content. Bohn *et al.* (1985) illustrated that where the water table was 900 mm below the surface, the effects of salt deposition began at about 400 mm, where the electrical conductivity (EC) was 0.2 ms/m, with EC increasing to EC 70 ms/m at 50 mm. In order to illustrate the effect of evaporation

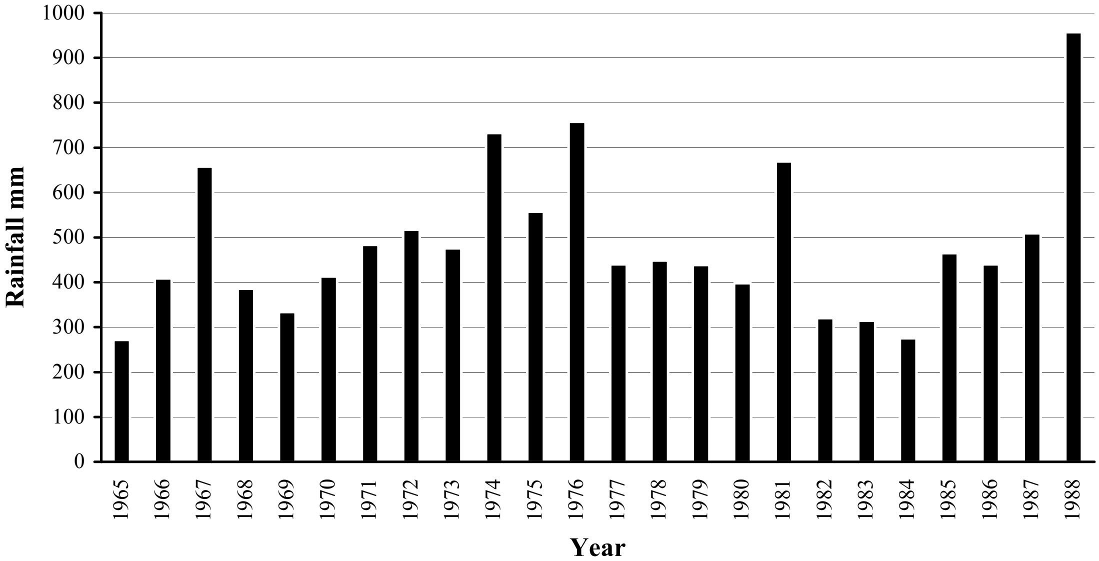


Figure 17. Twenty-three year annual rainfall at Florisbad showing the variation in annual rainfall between the lowest (1965) and highest (1988) recorded rainfall periods.

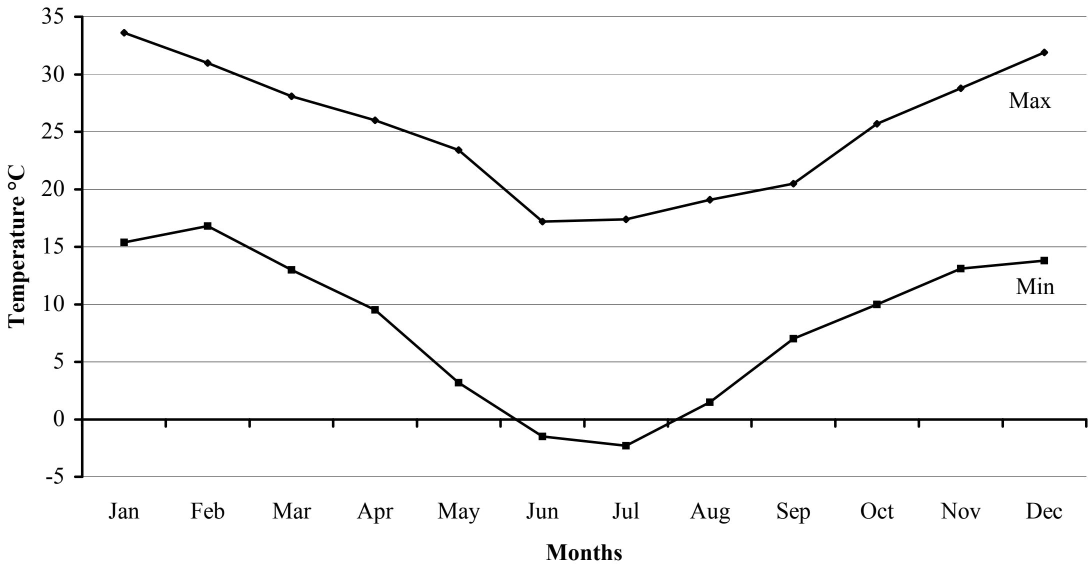


Figure 18. An example of the monthly minimum and maximum temperatures for the Florisbad area (taken at Glen Agricultural College 1987).

and matrix suction at Florisbad, rainfall and evaporation figures were taken from Glen Agricultural College, 32 km east of Florisbad. Rainfall for the period January, February and March 1992 was 66.5 mm, while the mean evaporation rate for the same period was 315.0 mm. This resulted in a rainfall deficit of 248.5 mm for the three-month period, which is an indication of the amount of moisture and salts being brought to the surface by capillarity and matrix suction.

Van Zinderen Bakkker (1989) gave an average annual rainfall deficit for eight stations, with records varying from 18 to 47 years, of 1260 mm, while the mean annual evaporation rate given by Baran (2003) for the area is 1 600 – 1 700 mm. Evaporation in the Free State is converse to rainfall decreasing from west to east, with higher rainfall areas experiencing less evaporation, and lower rainfall areas experiencing higher evaporation. This is largely due to the west to east increase in altitude and cooler temperatures. Particularly in semi-arid environments where evaporation exceeds precipitation, calcium carbonate forms as a result of the high evaporation rate (Hillel, 1971). The effects of capillarity and salinization have both played a major role in the high mineralization of the upper Peat IV horizon (Appendix II).

3.3.4 Wind

Winds at Florisbad (Figure 19), as with rainfall, are regulated by disturbances in the circulation systems and patterns to the west and south-west of the country, and their influence on the intensity of low and high pressure cells over the land. The latter factors will determine the intensity with which warm, moist, tropical air is drawn in from the tropics during summer, and the intensity with which cold fronts reach inland during winter. Wind has played a major role in the formation of the Florisbad spring site (Appendix IV). Wind is critically important in the Florisbad context in that it not only determines rainfall, but also the aeolian transport of sand, and the resultant formation of sand dunes. Mucina and Rutherford (2006) have noted that aeolian dust may be carried several thousand metres into the air by strong winds. Florisbad may never have existed as such an important archaeozoological and archaeological site had

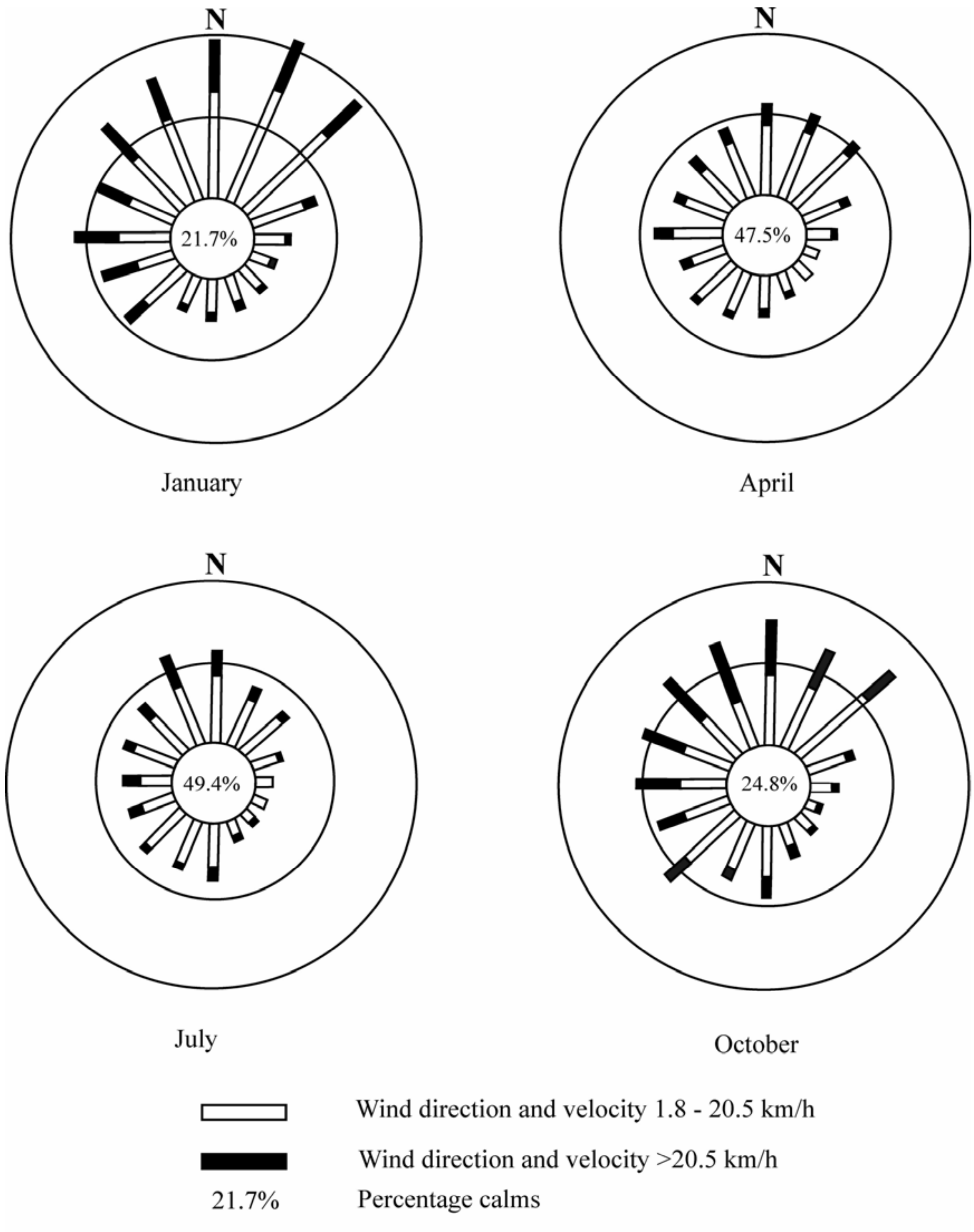


Figure 19. Wind roses for Bloemfontein, 45 km south-east of Florisbad, showing wind direction and velocities for the months of January, April, July and October (after Kruger, 2002).

it not been for the formation of the sand dunes (Appendix IV). In Figure 19, wind roses show the predominantly north-easterly, through to south-westerly, and even southerly (anticlockwise), wind flow throughout most of the year.

Despite some authors (Loock and Grobler, 1988; Van Zinderen Bakker, 1989) referring to a prevailing north-west wind, it is evident from Figure 19, that currently, it is difficult to assign any prevailing direction to local winds, as was confirmed by Kruger (2002). However, data on wind direction by Schulze (1994) does show a north-west predominance for both January and July. Current wind directions in Figure 19 may therefore not be a reflection of historical wind direction. Judging from the extent of the south-east dune belt (Figure 14), which continues for kilometres south and south-east from the southern banks of the Modder River, it is apparent that historically, there must have been extended periods with a very predominant prevailing north to north-west wind in order to have moved such large quantities of sand.

3.4 VEGETATION

The Grassland Biome of Rutherford and Westfall (1986) is defined as having two categories of grasses. The first category is the sweet grasses, which have a low fibre content and maintain their nutrients in their leaves throughout winter. The second category is the sour grasses, which have a high fibre content and tend to withdraw their nutrients from the leaves during winter. Sour grasses prevail with a higher rainfall and more acidic soil, with 625 mm being the level at which sour grasses prevail. This then places Florisbad in the sweet grass category. Acocks (1988) veld type (AVT) 50 (Figure 20) has since been classified by Low and Rebelo (1996) as Dry Sandy Highveld Grassland. Mucina and Rutherford (2006) describe the vegetation unit as Dry Highveld Grassland and the Florisbad sub-units as Western Free State Clay Grassland (Gh 9), partly surrounded by Vaal-Vet Sandy Grassland (Gh 10), which extends from the north-west of Florisbad, around to the south east (anticlockwise).

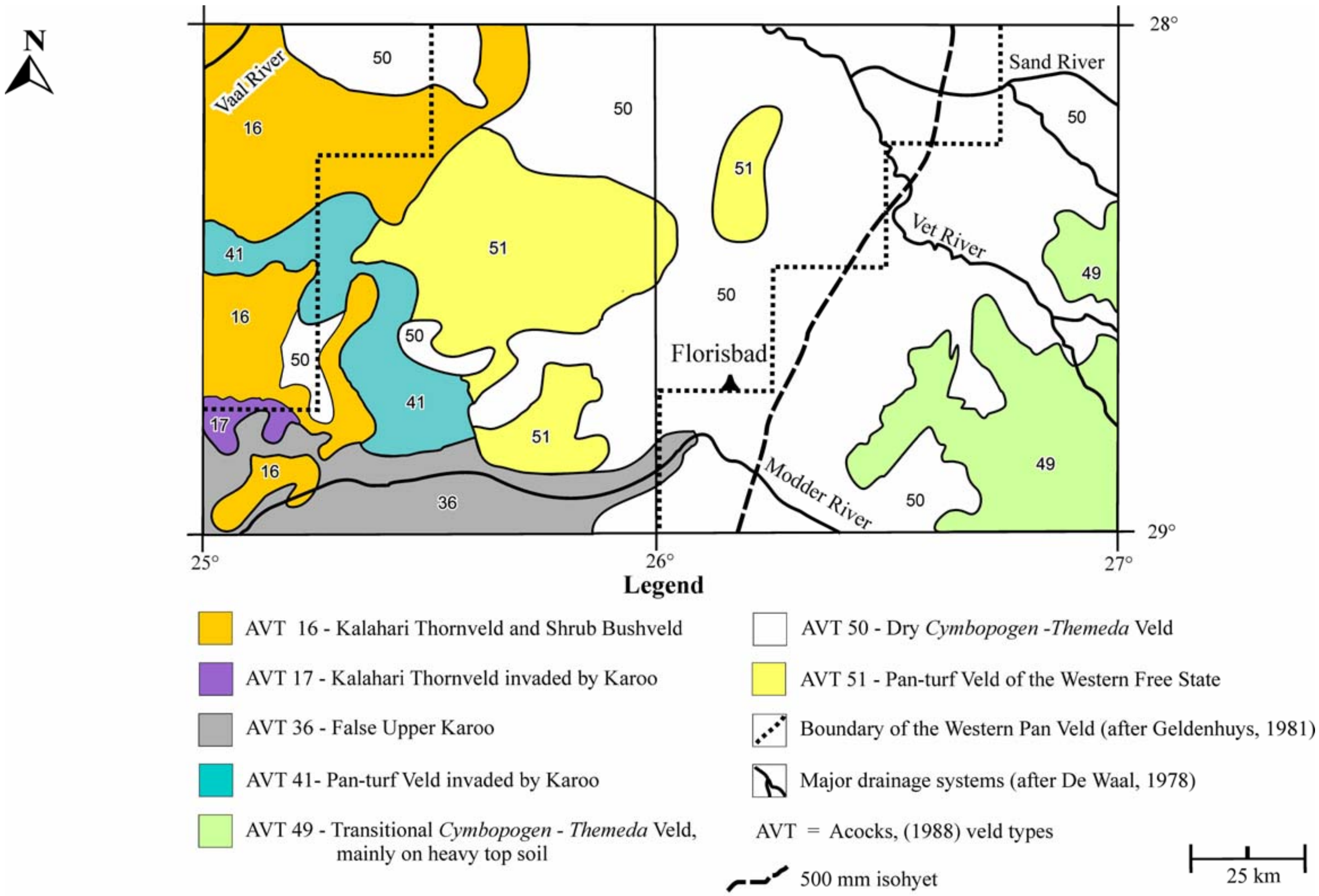


Figure 20. Vegetation of the central-western Free State showing veld types after Acocks (1988), panveld boundaries after Geldenhuys (1982), and the main drainage systems.

The Florisbad spring site is located in Dry *Cymbopogen- Themeda* Grassveld, Acocks Veld Type 50 (Acocks, 1988) (Figure 20), and in close proximity to an easterly protrusion of False Upper Karoo (AVT 36). The site also lies within, and close too, the western boundary of the Grassland Biome of Rutherford and Westfall (1986), with the Nama-Karoo Biome lying directly to the west. Karoo and Desert vegetation types which are encroaching from the south-west (Acocks, 1988), along with the close proximity of the False Upper Karoo and Nama-Karoo Biomes, may well see the area falling into a considerably drier vegetation zone. Vegetation is a reflection, and integral aspect, of past and present climates and environments. Vegetation is not only a reflection and record of present climatic and soil conditions, but also a record of vegetation and climatic changes over the millennia. This history is recorded in the palynology (study of pollens) of the area, in the form of pollen (organic) analysis from the sediments and hyaena coprolites. The study of phytoliths (inorganic biogenetic silica formed in the intercellular and intracellular spaces in plants), particularly from grasses, is another valuable analytical tool in recording the vegetation and climatic history of an area. Phytoloths also occur in hyaena coprolites (Scott and Russouw, 2005), and phytoliths from the dental cavities of Middle Stone Age land surface bovid species were examined by Russouw (1996), providing an indication of the species diet. Another illustration of the importance of vegetation is that the only piece of wood recovered from the sediments was identified as the tree *Xanthoxylim chalybeum*, which could imply a major climatic shift to warmer, wetter inter glacial conditions (Bamford and Henderson, 2003; Scott and Russouw, 2005).

Highveld salt pans comprise a large variety of low shrubs, succulent shrubs, megagraminoids, graminoids, herbs, and succulent herbs (Mucina and Rutherford, 2006). Douglas (1992) recorded the following dominant vegetation types for the study area. The vegetation of the site largely comprises taller *Themeda triandra*, *Eragrostis lehmanniana* and *Hetrodon contortus* grasses, with low growing grass being represented by *Tragus koeleroides*. Bushes are scattered, and represented by *Salsola glabrescens* and *Filicia muricata*. Along the banks of the vlei area *Cynodon dactylon* grass is predominant, with dense stands of *Salsola glabrescens* bushes and *Conyza bonariensis* annuals. In the wetter regions of the vlei some areas are densely vegetated with *Schoenoplectus* (*Scirpus*) *triquete*r and *Typha latifolius* with *Mesembryanthemum* sp. growing along the drier edges.

The western Free State panveld vegetation unit is classified as Inland Azonal Vegetation, with the sub-unit of the pan vegetation being classified as Highveld Salt Pans (Azi 10) (Mucina and Rutherford, 2006). Based on vegetation growth, determined two months after flooding, Geldenhuys (1982) classified pans as bare pans, sedge pans, scrub pans, mixed grass pans, closed *Diplachne* (*Leptochloa*) pans, and open *Diplachne* (*Leptochloa*) pans. Highveld salt pan vegetation differs considerably from pans in other areas due to their predominance of cyperoids (Mucina and Rutherford, 2006).

The Florisbad area is an area of natural transitional zones for a number of factors, and these factors have been grouped together here because vegetation and climatic boundaries play a dominant role (Table 7). The actual influences of these transitional zones on the Florisbad area were not examined in this study. It is however proposed that, in conjunction with morphological features such as the Western Free State pan veld, they contribute in making the Florisbad area distinctive in its own right.

3.5 SYNOPSIS

The introduction of the regional climate of the Free State province establishes a further background for understanding the local climate of the Florisbad area. It will be noted from this chapter how various aspects of the climate are interrelated, and how in future chapters, they will contribute and influence, in their own individual way, to the depositional and fossilization environment at Florisbad mentioned in the Introduction to this chapter.

In the following chapter, yet another feature having a direct influence on the Florisbad spring site, geohydrology, is examined. As with geology in Chapter 1, geohydrology is discussed on a regional basis in order to bring perspective and understanding to the geohydrology of the Florisbad area.

Table 7. Natural zones which have their boundaries passing through the Florisbad area, as defined by various authors.

Zone	Detail of natural zone	Reference
Grasslands	Boundary between the Karoo and Sweet Grassveld.	Acocks 1953
Grasslands	Boundary between the Karoo (South-West Aarid) and Grassveld (Southern Savanna Grassland).	Acocks 1975
Biotic zones	Boundary between the South-West Arid and Southern Savanna Grassland Biotic Zones.	Keay, 1959; Davis, 1962; Meester 1965; Rautenbach 1978
Biotic zones	Boundary between the South-West Arid and Southern Grassland Biotic Zone based on the rainfall map of Wellington (1955).	Wellington, 1955; Davis 1962
Biotic zones	Boundary between the Grassland and Nama-Karoo Biome based on vegetation.	Rutherford & Westfall 1986
Vegetation	Boundary between the Highveld Grassland (58) and the Karoo Grassy Shrubland Transition from Karoo Shrubland to Highveld (57b).	White 1981
Vegetation	Boundary between the Eastern Mixed Nama Karoo (52) and the Dry Sandy Highveld Grassland (37)	Rebelo & Low, 1996
Rainfall	The 500 mm isohyet.	Poynton, 1964; Van Der Wal 1977
Humidity	Boundary between the Sub-Humid Moisture Region (C) [moisture index –20 to 0] and the Semi arid Moisture Region (D) [moisture index –40 to –20 based on the silvicultural map of Poynton (1971).	Poynton, 1971
Frost	Boundary of the Cooler –Temperate (Mesothermal) Thermal Region (5) [Thermal Efficiency Index – severe frost] and the Cooler-Temperate (Mesothermal) Thermal Region (6) [Thermal Efficiency Index – moderate frost].	Poynton, 1971
Panveld	The eastern boundary of the western Free State panveld.	Geldenhuys, 1982
Herpetofauna	The eastern boundary of the T2 Squamata Transitional Zone and B2 Squamata Province for the Free State.	De Waal, 1978

Chapter

4

The Physical Environment

Part III

Geohydrology

CHAPTER 4

THE PHYSICAL ENVIRONMENT

PART III

GEOHYDROLOGY

4.1 INTRODUCTION

Water is a scarce resource in South Africa, with Vivier (1996) and Chevallier *et al*. (2001) stating that groundwater contributes to only 10% to South Africa's water needs, while surface waters have been virtually exploited to their limit. However, Free State towns such as Ficksberg, Jagersfontein, and Fauresmith, may towns derive from 50 to 100% of their water requirements from groundwater (Vivier, 1996). Vegter (1995) lists 74 municipal supply schemes with populations >2500, whose water supply is based solely on groundwater. By virtue of secondary openings, over 90% of South Africa's rocks are water bearing (Vegter, 2001) This means that groundwater is becoming a more important and valuable resource, with the Karoo Supergroup formations, which underlie approximately 50% of South Africa, potentially becoming South Africa's largest and most important source of water (Vivier, 1996). Chevallier *et al*. (2001) note that, within the Karoo Basin, dolerite dykes represent the most commonly exploited targets due to their well known water yielding capacity. Although having considerable water bearing potential, it was noted by Chevallier *et al*. (2001) that dolerite ring and sill complexes, were on the other hand, overlooked as a source of water, and extensive research is currently being carried out in this area.

That the Florisbad springs have a perennial-, as opposed to a sporadic or cyclical flow, requires greater emphasis to be placed on understanding the geohydrological properties and processes of the area. With Florisbad lying within the Karoo Basin, and the Florisbad springs being controlled by dolerite intrusions, a background to the regional geohydrology, and factors and processes affecting this hydrology, are presented in this chapter.

The geohydrological chapter has been compiled from Vivier (1996), Vegter (1995), Chevallier and Woodford (1999), Chevallier *et al*. (2001), Vegter, (2001), Woodford and Chevalier (2002), and Baran (2003)

4.2 BACKGROUND

South Africa is divided into 64 groundwater regions based on primary, or secondary pores, physiography, and climate, while conformity between lithostratigraphic units and phsiographic features strengthens the case for lithostratigraphy as a primary basis for subdivision. Vegter (2001) suggests that the following criteria govern the availability and occurrence of ground water:

- the storage and transmissive properties of the geological formation;
- topographic relief, which in turn determines the hydraulic gradient
- the volume and frequency of discharge;
- the rate of groundwater movement to discharge points/areas;
- the rate of groundwater discharge as springs and effluent seepage in streams; and
- loss through evapotranspiration.

The transmissive properties of a formation may be restricted by:

- fractures being under compressional stress and consequently tight;
- recrystalization in fractures by hydrothermal fluids or a drop in the groundwater level;
- fractures being filled with swelling, or non-swelling clays as a result of weathering, or past hydrothermal activity.

As opposed to the availability and occurrence of groundwater, Vegter (2001) proposes the following criteria for recharge:

- factors in relation to rainfall such as, volume, intensity, frequency, and temporal distribution,
- availability of surface water,
- land surface configuration,
- soil and vegetation cover; and
- subsurface moisture retention and evapotranspiration.

Lithification occurs in the form of, compaction and cementation desiccation, crystallization, recrystallization, and compression, or a combination of processes. Compression of the Karoo sediments, caused largely by the Drakensberg lavas covering the area, caused a considerable decrease of porosity in the sediments, particularly the porosity and elasticity of the clays (Vivier, 1996). A relatively rapid erosion of the Drakensberg lavas then removed the pressures being exerted on the underlying Karoo sediments, resulting in an isostatic uplifting of the sediments, and the formation of fractures and interstices, which in turn improved the permeability and porosity of the strata (Vivier, 1996). Subsequent weathering and diagenesis of the strata further improved permeability and porosity (Vivier, 1996).

In categorizing different interstices, Vegter (2001) differentiated between four types of saturated interstices:

- pores in disintegrated/decomposed, partly decomposed rock and fractures restricted primarily to a zone below groundwater level;
- fractures restricted primarily to a zone directly below groundwater level;
- openings varying in size from fissures to caves; also pores in dissolution residuum and collapsed unconsolidated deposits; and
- pores in semi- and unconsolidated deposits.

- Vegter (2001) expanded on this by examining interrelated factors contributing to the complex interactions of weathered thickness, porosity, permeability and groundwater flow (Table 8).

Vegter (2001) classified groundwater regions into two subdivisions:

- Regions of primarily water-bearing formations comprising hard-rock formations underlying primary water-bearing formations which may act as aquifers.
- Regions of secondary water-bearing formations, including fluvial deposits, while being further classified on principal rock type and erathem/system:
  - crystalline metamorphic and igneous;
  - intrusive
  - extrusive
  - sedimentary
  - composite.

  Composite regions were further classified as:
    - having no dominant major rock type;
    - having several major lithostraphic units involved; and
    - having more extensive primary aquifers than alluvial deposits present.

Two basic types of flow are present in Karoo aquifers, namely, matrix flow and mesofractural flow. Fractures in Karoo aquifers have a limited storage capacity, and therefore storage must be in the matrix, which is usually composed of fine grained mudstones, siltstones and shale, interbedded with coarser grained sandstone (Vivier, 1996). Karoo rocks originally had a very high porosity, but as previously mentioned, through compaction and cementation due to the overburden, the primary porosity of the rocks was severely reduced, resulting in pores and microfractures being very small. Thus, the modern hydraulic conductivity of these rocks is also very low. The flow of water in a fractured medium is largely controlled by the dimension of the fractures, connectivity, and orientation, with the former two being either limited, weak, or non-existent in Karoo aquifers. Vivier (1966) concluded that borehole yields in Karoo aquifers was largely controlled by the horizontal fractures which may deform considerably due to pumping and extraction.

Table 8. Interrelated factors contributing to the complex interactions of weathered thickness, porosity, permeability, and groundwater flow (after Vegter, 2001).

The degree of tectonic deformation and fracturing.

The degree of non-tectonic fracturing such as thermal shrinkage and sheet formation.

The climate, as in past and present rainfall and temperature.

The age of the land surface.

The relief.

The mineralogical composition and texture of the host rock

Chemical weathering is the dominant process in the development of a weathering profile on crystalline basement rocks.

Groundwater is the principal weathering agent in the saturated lower part of the weathering profile.

The amount of dissolved oxygen and carbon dioxide in the percolating water, where groundwater flow rates control the rate of chemical reaction by the extent to which it supplies hydrogen ions and dissolved oxygen.

The flow rate through the weathering system is dependant on the availability of recharge, the permeability of the weathered material, and the hydraulic gradient between the recharge and discharge areas.

The hydraulic conductivity of the weathered material is therefore a function of the extent of chemical weathering, and is inherently linked to the history of the groundwater flow through the system.

4.2.1 Aquifers and Aquitards

Vegter (1995) defined an aquifer as a stratum which contains intergranular interstices, or a fissure/fracture (as such), or a system of interconnected fissures/fractures (as such) capable of transmitting groundwater rapidly enough to directly supply a borehole or spring". Groundwater fills intergranular open spaces in the unconsolidated sediment and in weathered previously consolidated rock, as well as fractures in the hard rock. Saturation alone does not imply the existence of an aquifer, as a precondition in identifying the formation of an aquifer is the gravitational mobility of water in a saturated rock formation as a result of its permeability.

An aquitard was defined by Vegter (1995) as "a body of poorly permeable rock that is capable of slowly absorbing water from and releasing water to an aquifer. Aquitards do not transmit groundwater rapidly enough by themselves to directly supply a borehole or spring as they usually comprise layers of clay or non-porous rock with a low hydraulic conductivity. A formation that has no interconnected openings, and therefore cannot absorb or transmit water (impenetrable), is referred to as an aquiclude, or an aquifuge.

Baran (2003) recognized four basic types of aquifer.

4.2.1.1 *Intergranular Aquifers*

Intergranular aquifers are comprised of unconsolidated sediment, such as sand and gravel, where water is stored in the intergranular pores. In the Free State, intergranular aquifers are poorly represented and only occur as narrow strips of alluvium along some major river valleys such as the Vaal, Wilge, Sand, Vet Modder and Vals Rivers. Some towns, such as Ficksberg in the eastern Free Sate, obtain up to 50 % of their water from alluvial beds along the Caledon River (Vivier, 1996).

4.2.1.2 *Fractured Aquifers*

Fractured aquifers occur in hard, mainly quartzitic, rock formations where water occurs in fissures, joints, fractures, and faults. Tectonic forces, and to a much lesser degree, weathering processes, have produced these fractures, resulting in a limited and lower storage capacity than other aquifers.

4.2.1.3 *Karstic Aquifers*

Karstic aquifers are associated with carbonate rocks such as dolomite and limestone, where water is stored, and transmitted through solution cavities, channels, and fractures. A characteristic of these aquifers are their solution chambers, formed by the enlargement of cracks and fissures through circulating groundwater containing carbonic acid, which in turn dissolves the carbonate rock. These aquifers are characterized by a high storage capacity and high yields.

4.2.1.4 *Intergranular and Fractured Aquifers*

Intergranular and fractured aquifers occur in tectonically altered and subsequently weathered and fractured rock. They are found in Precambrian rocks and Karoo Basin sediments. Water occurs in a vertical profile where the upper weathered and decomposed rock zone is in hydraulic contact with fractured zone down to the solid and fresh, rock formations below. These aquifers occur throughout most of the Free State with the Florisbad spring aquifer falling into this category. It will be noted from Figure 21 that there is little difference between the yield of boreholes located in Ecca Group rocks and those located in dolerite sills. Borehole yields from the Ecca and other Groups, as well as the dolerites, area are considered low.

4.3 GEOHYDROLOGY OF THE KAROO SUPERGROUP

Usher *et al*. (2006) state that the behaviour of Karoo fractured aquifers is ultimately determined by their unusual geometry, particularly where horizontal, bedding-parallel fractures are present. These features provide not only the conduits for water to boreholes in Karoo aquifers, but also play a prominent role in the interactions responsible for the behaviour of these aquifers. One very important consequence of these interactions is that flow in Karoo aquifers is not only radial and horizontal, but also linear and vertical. This property differs so much from that of the theoretical media usually presented in the literature on aquifer mechanics that the existing conceptual models are useless for the analysis of hydraulic tests performed on Karoo aquifers.

4.3.1 Dwyka Group

Dwyka Group rocks have a very low hydraulic conductivity and virtually no primary voids, with the Group constituting of a low-yielding, fractured, aquifers where water

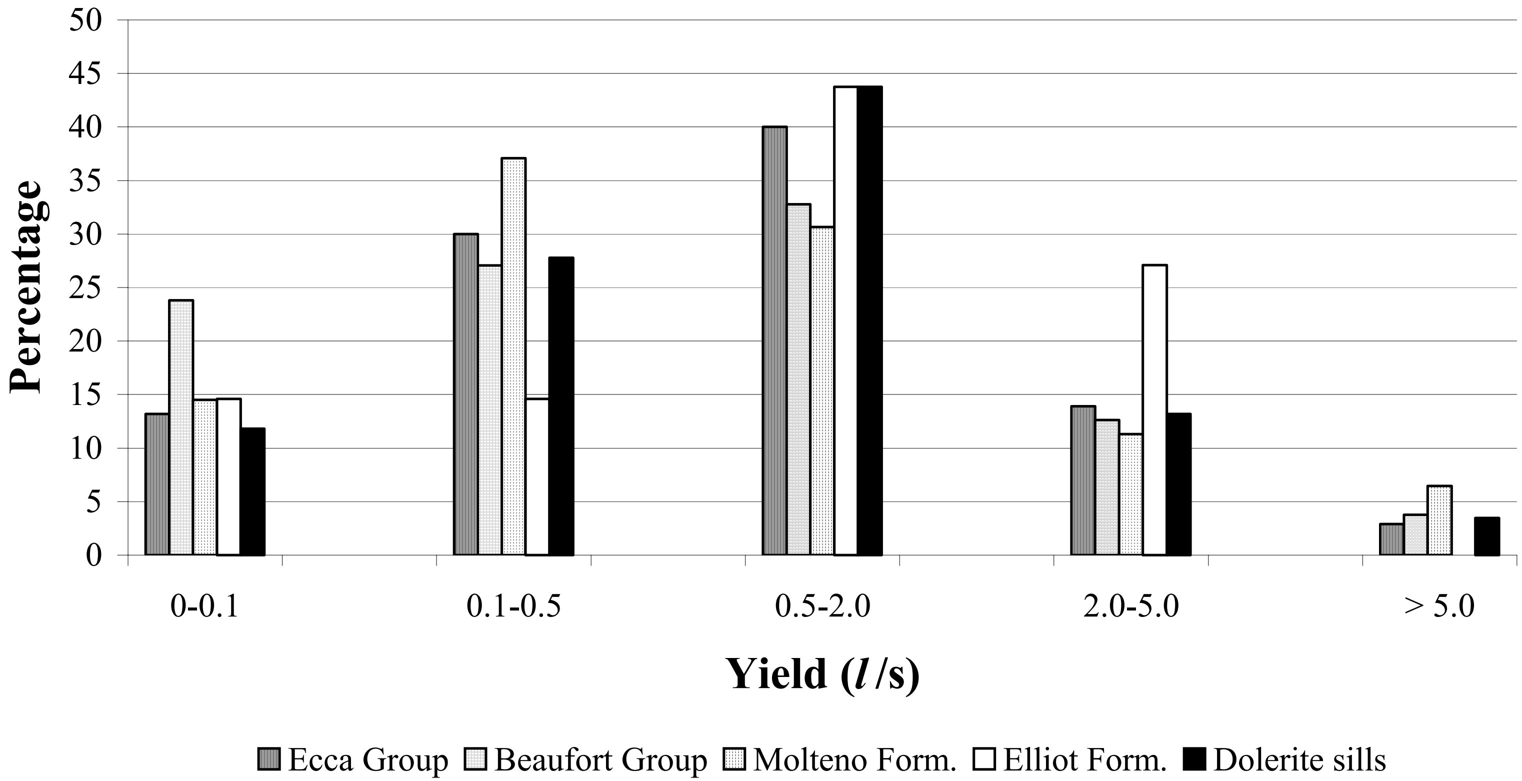


Figure 21. Some comparative examples of borehole yields between the Karoo Group Formations and dolerite sills (after Baran. 2003).

is confined to narrow discontinuities like jointing and fracturing (Woodford and Chevallier, 2002). This has resulted in the formation of aquitards, rather than aquifers (Woodford and Chevallier, 2002). Because of the marine environment under which the Dwyka sediments were deposited water from these aquifers tends to be saline. Exploitable aquifers therefore only occur where sand and gravel were deposited on beaches, or where the Dwyka group was significantly fractured (Woodford and Chevallier, 2002). Little information is available in the geohydrology of the Group. Despite this, the artesian borehole at Glen, produced a yield of 2.53 l/s in Dwyka tillites at ± 760 m (Baran, 2003).

4.3.2 Ecca Group

Ecca Group rocks do not form high yielding aquifers due to the prevalence of dense shales, with the Volksrust shales being considered totally impenetrable (Woodford and Chevallier, 2002); Baran, 2003). Shale thickness varies from 1500 m in the south to 600 m in the north (Woodford and Chevallier, 2002). Ecca sandstones have a low permeability as a result of them having been poorly sorted, with porosity being further decreased through diagenesis (Woodford and Chevallier, 2002). Porosity decreases from 0.10% in the north to >0.02% in the south (Woodford and Chevallier, 2002). Being formed in a fluvial environment, sand-grains tend to align their longest axis parallel to the flow direction to form an anisotropic matrix (Vivier, 1996). This suggests that the hydraulic conductivity of the matrix will be greater in the direction of the alignment of sand grains, as opposed to perpendicular to the alignment (Vivier, 1996). Baran (2003) expected the weathering and fracturing of the arenaceous Vryheid Formation to be more developed than that of the argillaceous Volksrust Formation. Ecca Group waters are of a sodium bicarbonate nature, with 45% of boreholes having an EC value above the recommended limit for drinking water (70 mS/m), 2% above 300 mS/m, with an average of 78.7 mS/m (Baran, 2003). Yields from boreholes are generally considered as low, with comparative borehole yields being given in Figure 21.

4.3.3 Beaufort Group

The older Adelaide Subgroup consists predominantly of argillaceous sediments, while the younger Tarkastad Subgroup is more arenaceous and described as being argillaceous and arenaceous (Baran, 2003). Since the depositional environment of the Beaufort Group has many similarities to that of the Ecca Group, it can be expected that Beaufort aquifers will also be of an anisotropic nature (Vivier, 1996). The geometry of Beaufort Group aquifers is further complicated by the migration of braided and meandering streams over a flood plain, resulting in multi-layered, multi-porous, aquifers of variable thickness (Vivier, 1996; Woodford and Chevallier, 2002). The dominance of mudstone, shale and fine-grained sandstone render the sedimentary units of the Beaufort Group, low in permeability. The complexity of Beaufort Group aquifers is enhanced by many of the coarser and more permeable sediments being lens shaped (Woodford and Chevallier, 2002). Borehole yields of the entire Beaufort Group are considered low (Figure 21).

4.3.4 Stormberg Group

4.3.4.1 *Molteno Formation*

Molteno Formation rocks were formed by braided and meandering streams with basal layers being comprised of conglomerates and coarse sandstone followed by fine-grained sandstones and shales (Vivier, 1996; Woodford and Chevallier, 2002). A characteristic feature of the Molteno Formation is that the sediments are laterally uniform over large areas (Vivier, 1996). However, despite borehole yields being considered low to moderate yielding (Baran, 2003), the persistence of the sedimentary bodies makes for a more favourable aquifer geometry in so far as groundwater storage is concerned (Woodford and Chevallier, 2002). Pebble conglomerates and coarse- grained sandstone at the base of the formation should also allow for ideal aquifers (Woodford and Chevallier, 2002). There appears to be little information on the geohydrology of this formation. (Figure 21).

4.3.4.2 *Elliot Formation*

Rocks of the Elliot Formation were deposited by meandering streams under highly oxidizing conditions, and comprise red mudstone, succeeded by a red fine-grained sandstone (Vivier, 1996; Woodford and Chevallier, 2002). The red mudstone is more conducive to aquitards than aquifers (Woodford and Chevallier, 2002). Boreholes are considered moderate to high yielding (Figure 21).

4.3.4.3 *Clarens Formation*

The sediments of the Clarens Formation are related to aeolian sand, playa lakes, and sheet-flood deposits (Woodford and Chevallier, 2002). Owing to the aeolian nature of the deposits, the rocks comprise homogeneous fine to very fine-grained sandstone, with a high clay content being derived from silt deposited in the playa lakes (Woodford and Chevallier, 2002). The Clarens Formation is dominated by red mudstones, which again are more conducive to aquitards, rather than aquifers. This Formation is considered to have the most homogeneous geometry in the Karoo Supergroup, and is poorly fractured with a low permeability, but with a relatively uniform and a relatively high porosity average of 8.5% (Woodford and Chevallier, 2002). Due to the formation being poorly fractured, hydraulic conductivity is low. Boreholes are considered moderate to low yielding

4.4 GEOHYDROLOGY OF THE KAROO DOLERITE SUITE

4.4.1 Dolerite Dykes

The intrusion of dolerite bodies in the host rock formations created zones of fracturing in both the host rock and the dolerite itself, with these fracture zones becoming natural underground drainage systems for groundwater (refer to previous discussion in section 2.2.4.2). Fracturing usually occurs on both sides of the dolerite intrusion, with subsequent weathering enhancing their permeability, but at greater depths (>30-40 m) fractures become fewer, and borehole yields decrease considerably. Dolerite dykes are, to a large extent, impenetrable and restrict the flow of groundwater (Vivier, 1996; Woodford and Chevallier, 2002). Vivier (1996) states that a solid dolerite dyke's state of weathering is an important indication as to yields that can be expected from boreholes drilled near to dolerite dykes, with highly weathered dykes producing higher yielding boreholes.

Dolerite dykes are the preferred drilling targets for groundwater in the Karoo Basin, with 25% of boreholes in the Victoria West district having being sited into, or along side, dolerite dykes (Woodford and Chevallier, 2002). In northern Kwazulu-Natal, 80% of successful boreholes are directly, or indirectly, related to dolerite intrusions (Woodford and Chevallier, 2002). The state of a dyke may be critical in the success of a borehole. For example, weathered dykes may themselves be fractured, while unweathered dykes may have few, or no, fractures. The hydrogeology of intrusive dykes is complex. It has been stated that the high permeability of dyke contact zones was not only as a result of the fracturing of the host rock, but also as a result of shrinkage joints developed during the cooling of the intrusion (Woodford and Chevallier, 2002). The degree to which the host rock is metamorphosed and fractured away from the dyke is also critical, with the zone of metamorphosis commonly being less-than, or equal to the width/ thickness of the intrusive body (Woodford and Chevallier, 2002). This distance varies in the literature from 1 to 20 m, depending on the area, while the dip of the dyke will also play an important role in determinations (Woodford and Chevallier, 2002). Where younger dolerite dykes cut across older sills, particularly in valley-bottom situations where sill material is highly weathered, these provide good locations for groundwater (Woodford and Chevallier, 2002). Transgressive fracturing is often well developed in such areas (Woodford and Chevallier, 2002).

Horizontal and oblique transgressive fractures within the dolerite and country rock are the dominant water bearing features in the Queenstown area, with fractures in some areas extending for up to 90 m away from the dyke contact (Woodford and Chevallier, 2002). The dip of a dyke *per se* does not appear to have an influence on the borehole yield, but is important in the sitting of the borehole. Conversely, porosity of the host rock may decline in the host rock contact area with the metamorphic aureole intrusive, as a result of re-crystallisation of the host rock silicates and infiltration/cementation by magmatic silica (Woodford and Chevallier, 2002). It has also been found that in the eastern Free State the best borehole yields originated from dykes which were from 7 to 11 m wide (Woodford and Chevallier, 2002), but again, this will vary from area to area. Although groundwater is often associated with dolerite dykes, this is not always the case. In the drilling of the Fish River tunnel between the Gariep Dam and the upper reaches of the Fish River, a distance of 83 km, 55 dykes were intersected, of which 49 were dry (Chevallier *et al*. (2001).

4.4.2. Dolerite Sills and Ring-Complexes

Because of factors such as size, hardness, thickness, and structural complexity, dolerite sill and ring-complexes have largely been overlooked in a hydrological and water exploration sense (Vivier, 1996; Woodford and Chevallier, 2002). This is often due to the easier way in which the multitude of linear dykes and fractures can be detected and located (Woodford and Chevallier, 2002).

Baran (2003) provided almost no information on the geohydrological characteristics of dolerite sills and ring-complexes, while due to the above mentioned factors, Woodford and Chevallier (2002) cite specific site investigations. At their Victoria West dolerite sill and ring-complex site Woodford and Chevallier (2002) noted that very few water bearing fractures were encountered between the top of the inner sill and the host rock. High yields were encountered in horizontal water bearing dolerite offshoots, or fractures, found in the host-rock on the foot wall contact of the inclined sheet, as well as at the base of the of the sill. Moderate to high yields were encountered in the sediments above the inner sill, most probably due to weathering and the near surface release of residual compressive stress developed inside the saucer during intrusion. It has been shown that water bearing open fractures occur at specific locations within the dolerite and surrounding host-rock (Woodford and Chevallier, 2002). These are to be found in the sediment above an up-stepping sill or at the base of an inner-sill, and at the junction between a feeder dyke /

inclined sheet and sill (Woodford and Chevallier, 2002). Generally dolerite sill borehole yields are considered low to moderate.

4.5 GEOHYDROLOGY OF OTHER POST KAROO INTRUSIONS

4.5.1 Breccia Plugs and Volcanic Vents

These structures are both pipe-like and filled with brecciated fractured material, providing highly permeable targets for groundwater (Woodford and Chevallier, 2002). This is due to their geometry and large vertical extent of uniform material. Due to their limited size, it has been proposed that they are only of significance as conduits for rapidly extracting water from the subsurface, and therefore their yield sustainability is dependant on recharge and storativity of the host-rock (Woodford and Chevallier, 2002). The high success rate of boreholes (70% yield in excess of 3 *l*/s) at Calvinia, Luiperdkop, Doorn-Laagte, Carnarvon, and Bitter Poort, can be considered excellent when compared to yields encountered in Ecca shale of the western Karoo (Woodford and Chevallier, 2002). Yields of >8 l/s are almost always encountered in intensely brecciated sections of a plug (Woodford and Chevallier, 2002).

Volcanic vents have a similar occurrence of groundwater to breccia plugs, due them having similar shapes, sizes, and degree of brecciation (Woodford and Chevallier, 2002). However yields may be higher due to their eastern location in a higher rainfall and recharge area. Boreholes in Ladybrand and Matatiele produced yields of 38 *l*/s and 20 *l*/s respectively (Woodford and Chevallier, 2002).

4.5.2 Kimberlites

Unlike the dolerites and volcanic plugs, kimberlite intrusion did not result in an intense thermal metamorphism of the Karoo sediments, and therefore did not significantly alter the hydrological properties of the host sediments (Woodford and Chevallier, 2002). On a larger scale strong regional jointing and reactivation of existing structures that accompanies the emplacement of kimberlite swarms may be important for the occurrence of groundwater (Woodford and Chevallier, 2002). Larger kimberlite blows may represent more permeable zones as they are more heterogeneous, brecciated, and more highly weathered. Jagersfontein and Fauresmith in the Free State both obtain water from the now abandoned Jagersfontein diamond mine (Vivier, 1996; Woodford and Chevallier, 2002). Overall, kimberlite fissures are seen to yield small amounts of water due to the clogging of near surface joints by clays derived from the weathering of the kimberlite (Woodford and Chevallier, 2002). Transgressive water bearing fractures seen in dykes do not appear to have been developed along kimberlite fissures (Woodford and Chevallier, 2002).

4.6 GEOHYDROLOGY OF RECENT DEPOSITS

Intergranular aquifers in alluvium are discussed in Section 4.2.1.1. Alluvial beds, although being limited to narrow strips along main river courses, supply large volumes of groundwater to towns in the Karoo Basin. The Caledon River, for example supplies Ficksburg with approximately 50% of its water (Vivier, 1996). Calcrete deposits are yet another source of underground water: Vivier (2003) and Woodford and Chevallier (2002) note that recharge to these aquifers is 2 to 5% higher than in average Karoo aquifers. Farmers in the Petrusburg district of the central Free State tap into calcrete aquifers, up to 30 m thick, for irrigation purposes (Vivier, 2003; Woodford and Chevallier, 2002).

4.7 NON-INTRUSIVE TECTONIC FEATURES

A number of non-intrusive tectonic features may also influence the hydrology of areas of the Karoo Basin, but these are not within the scope of this study, and are not dealt with detail here. These include regional lineaments, folding, vertical jointing and faulting, bedding-plane fracturing, and sesmotectonic / neotectonic / uploading features (Woodford and Chevallier, 2002).

4.8 SYNOPSIS

As has been emphasised in the foregoing sections, the rocks of the Karoo Super Group and Karoo dolerite suite are extremely complex, while research on the geohydrology of dolerite ring and sill complexes can be considered to be in its infancy. It is only in recent years that hydrocensuses have been carried out on previously existing data. However researchers have had to be extremely careful in compiling this information as often no proper records were kept, or the records are not of a suitable standard for such an exercise.

The geohydrology of the main Karoo Basin is as complex as the formations it examines and, interested parties wishing to obtain more comprehensive information on the subject, are advised to obtain this from any of the references used here, but in particular, from Woodford and Chevallier (2002)

Chapter

5

The Florisbad Springs

CHAPTER 5

THE FLORISBAD SPRINGS

5.1 INTRODUCTION

The previous chapters have provided information on the many components which have contributed to the formation of the Florisbad spring site and the fossilization of faunal remains. Had the merger of all of these components, in conjunction with their interrelated processes, not occurred in the way, and at the times that they did, the Florisbad spring site, as we know it today, may very well never have existed. It is therefore imperative that as many components, and sub-components, pertaining to the formation and sedimentation of the Florisbad spring site are examined in order to obtain a holistic representation of how the site developed into a unique and important archaeozoological site. This chapter synthesises these components in developing and presenting a comprehensive picture of the geohydrological and surface processes controlling the depositional and fossilization environment at Florisbad.

5.2 CLASSIFICATION OF SOUTH AFRICAN SPRINGS

As the Florisbad springs have played such an important role in the history of the site it is prudent to briefly examine the classification of South African springs (Table 9) in order to be able to compare the Florisbad springs to the various types of springs that occur in South Africa.

Springs often form at the site of faults and fissures, but they may also form at dykes, when such a dyke impedes the subterranean water flow. A number of springs, such as those at Aliwal North, Cradock, Rooiwal, Badsfontein, and Knegha Drift, have been recorded as forming at dolerite dykes (Kent, 1949). It is at such a dolerite intrusion

Table 9. Classification of South African spring waters and the criteria used in their classification.

Reference	Basis of Classification	Classification	Criteria for Classification
Rindl (1916)	Chemical composition	Indifferent springs	<1000 mg/*l* dissolved solids; <1000 mg/*l* CO_2. Water temperature well above ambient temperature
		Sulphur springs	Based on an odour of sulphurated hydrogen. Water temperature usually above ambient temperature
		Chalybeate springs	>700 mg/*l* of ferrous iron. South African chalybeate springs thermal, other countries cold
		Saline springs*	High sodium chloride content and appreciable potassium (values not given) (K at Florisbad low)
		Alkaline springs	>1000 mg/*l* of solids with carbonate or hydrocarbonate ions predominating
		Earthy springs	>1000 mg/*l* of solids with bicarbonates of calcium and magnesium predominating
		Table water springs	Form a heterogeneous group comprising alkaline waters
Bond (1946)	Industrial and power production use	Highly mineralised chloride-sulphate waters	Total solids >1 000 mg/*l*:- Cl >27%; SO_4 >5%
		Slightly saline chloride waters	Total solids >300 - <500 mg/*l*:- Cl >27%; SO_4 >3%
		Temporary hard carbonate waters	Total solids <800 mg/*l*:- pH 7.6
		Alkaline soda carbonate waters	Total solids <1 000 mg/*l*:- Na_2Co_3 or $NaHCO_3$ >15%. Permanent hardness nil

(continued…..)

Table 9 (continued).

Reference	Basis of Classification	Classification	Criteria for Classification
Bond (1946) (cont.)		"Pure" waters	Total solids <150 mg/*l*:- pH <7.1
Kent (1949)	Allocated temperature ranges to spring waters based on the air temperature of the region	Warm*	25°-37°C
		Hot or hyperthermal	37°-50°C
		Scalding	>50°C
Kent (1949)	Hydrogen sulphide content as a measure of therapeutic value	Sulphuretted waters	Over 10 mg/*l* dissolved H_2S (including HS\`)
		Moderately sulphuretted waters	Between 5 and 10 mg/*l* dissolved H_2S (including HS\`)
		Slightly suphuretted waters	Between 1 and 5 mg/*l* dissolved H_2S (including HS\`)
		Sulphurous waters	Springs containing dissolved sulphur dioxide – not known from South Africa
Mazor and Verhagen (1983)	Total dissolved ions	Fresher water springs	Total dissolved ions 90-432 mg/*l* (no intermediate group classifications are given)
		More saline springs*	Total dissolved ions 936-2364 mg/*l*

that the Florisbad springs have formed (see Section 5.4.1). Most springs, within the Karoo Supergroup rocks, rise along dolerite dykes of an early Jurassic (Karoo) age and can mostly be classified as warm water springs, as opposed to hot water springs which originate from older and deeper archaean rocks (Kent, 1949). Vegter (1995) lists 59 thermal springs and a further 57 cold water springs yielding >1000 cm^3 per day.

South African spring-waters have been classified by a number of researchers, dependant on their particular interest in spring-waters. From the categorization of spring-waters in Table 9, and marked with an apteryx, the Florisbad spring can be described a highly salinized, highly saline, predominantly methane gas, warm water spring.

5.3 GEOHYDROLOGY OF THE FLORISBAD AREA

Florisbad falls into a region composed of mid Palaeozoic to early Mesoic strata, namely groundwater region 31, the Central Pan Belt (Vegter, 2001). The principal rock types are shales of the Tierberg Formation (Ecca Group) and dolerite dykes (see Sections 2.2.4 and 4.4). The surface lithology is predominantly argillaceous rocks comprising shale, mudstone and subordinate sandstone. Florisbad also lies in a zone of inter-granular and fractured aquifers in to which dolerite dykes and sills have intruded (see Section 5.4.1). With a groundwater yield of 0.5 –2.0 l/s (Vegter, 2001), the region could be described as having a very low, to low, development potential (pers. com. Bertram, E. Department of Water Affairs and Forestry, 20 May 2008).

From information supplied by Vegter (1995, 2001) the hydrology of the Florisbad area can be described as follows:

- Florisbad falls into the sedimentary groundwater region.
- The depth to groundwater level is 10-20 m with a range depth of 8-15 m.

- The storage coefficient for the compact (i.e. lacking significant primary porosity) argillaceous strata is <0.001. This is represents 1 cm^3 of rock containing less than one thousandth part of a cubic metre of water i.e. one litre
- The recommended drilling depth below groundwater level is <20 m.
- The mean annual recharge in mm per annum is 15-25 mm.
- The total dissolved solids (mg/*l*) range in terms of geometric standard deviation (it is noted that fluoride concentrations exceed 1 mg/*l* in more than 20% of the analysed samples) is 1000-1500 mg/l.
- The ground water quality of the area will have an EC range of from 70 to >1000 mS/m.
- There are two principal hydrochemical facies, based on a Piper diagram, with each comprising between 30% and 40% of analysed samples: (a) calcium/magnesium chloride/sulphate facies, and (b) calcium/magnesium bicarbonate/carbonate facies.
- The contribution of groundwater to river flow is negligible, or non-existent.
- The probability of a successful borehole yielding greater than 2.0 l/s is 20-30%.
- The probability of drilling a successful borehole yielding more than 0.1 l/s is 40-60%.

5.4 THE FLORISBAD SPRING AQUIFER

The Florisbad spring aquifer, as related to the consistency of the spring flow is of considerable importance in that it has never been reported that the springs have ceased to flow. It is therefore surprising that no specific research has been carried out on the Florisbad spring aquifer, and therefore information regarding its location, size, recharge, storage capacity, flow rates, travel times, and abstraction, is either unavailable, or unreliable. The research reported on in Appendix I determined that short-term rainfall had a considerable effect on the chemistry of the groundwater. It was further concluded that neither short-term, nor longer-term rainfall had any significant effect on the properties of the spring-water over the study period. It is proposed that any probable fluctuations in the quality of the spring water, due to recharge from long-term rainfall, will be smoothed out by the suggested large size of the aquifer, its probably considerable distance from the spring eyes, and its depth. (Appendix I). On this basis it was concluded that if long-term rainfall did have any affect on the spring-water, this would be over a considerably longer period than the 10 year period examined in this study, or the 84 year period over which the spring-water had been analysed. This, however, does not alter the opinion put forward in this thesis that the spring-water was never historically responsible for the fossilization of faunal remains (see Section 5.6.4).

Grobler and Loock (1988a) postulated that the intake of the Florisbad aquifer was located 30 km north of Florisbad at Basberg where permeable Beaufort Group sandstone occurred at an elevation 150 m above Florisbad. The possibility that the intake area of the spring may lie equidistant to the south-east of Florisbad, in the hills north of Bloemfontein, was also suggested by Grobler and Loock (1988a). Further more, there appears to be uncertainty as to the travel time that recharge water could take to travel from the intake area, through the aquifer, to the spring eyes. Grobler and Loock (1988a) calculated a travel time of anything from 160 to 16 000 years, with a probable 1 600 years.

It has been shown that warmer and deeper circulating water has a higher fluoride content (up to 5 mg/*l*) than cold-water springs and shallow aquifers (0.2 mg/*l*), indicating that a high fluoride content may be an indication of deep-seated thermal groundwater (Woodford and Chevallier, 2002). This would also tend to give support to Grobler and Loock's (1988a) estimates of the Florisbad aquifer being 500 metres below ground level (m. bgl), with water issuing at 29 °C with a fluoride content of 5.5 mg/l. In an attempt to locate oil within the Karoo sediments, a deep exploration borehole (1 495m) was sunk at Glen railway station, 32 km east of Florisbad. This borehole encountered warm artesian water in Dwyka tillite at between 747 and 775 m. bgl, with water issuing at 35 °C, and a fluoride content of 5.3 mg/*l* (Baran, 2003). The properties of the Glen borehole (Table 10) are in many ways similar to those of the Florisbad springs. Fluoride concentrations at

Table 10. A comparison of water quality between the Glen artesian borehole and the Florisbad spring (after Baran, 2003 [B03]); Kent, 1971 [K71]); Fourie, 1970 [F70]; App. I [Appendix 1].

Property	Glen Borehole (B03)	Florisbad spring (App. I) 1988	1999	Florisbad Other analyses
Temperature °C	35 °C issuing from between 747 and 775 m	-	29 °C issuing from 500 m	-
Flow rate *l*/s	2.5	-	-	2.5 (K71)
TDS mg/*l*	2 034	2202.9	2362.7	2277 (F70)
EC mS/m	325	388	398	-
pH	7.9	8.91	9.43	8.3 (F70)
Flouride mg/*l*	5.3	-	5.5	6 (F70)
NaCl mg/*l*	1 895	2086	2226	2099 (F70)
Ca mg/*l*	6 (K71)	98.4	100	87 (F70)
Bicarbonate (HCO_3) mg/*l*	43	-	-	49 (F70))
Temporary hardness ($CaCO_3$) mg/*l*	35	-	-	177 (F70)
Permanent hardness ($CaCO_3$) mg/*l*	215	-	-	40 (F70)

other warm water springs in the Karoo basin range from 4.8 mg/*l* at Aliwal North (36.9 °C) to 13.2 mg/*l* at Fort Beaufort (27.0-29.0 °C) (Woodford and Chevallier, 2002).

5.4.1 Control of the Springs by Geological Features

The intrusion of dolerite bodies into host rock formations created zones of fracturing in the host rock and in the dolerite itself, with the fracture zones becoming a natural underground drainage system for groundwater stored in the fractured and weathered, rock (Baran, 2003). In dolerite related zones weathering is important as it enhances the permeability of the rock, and thus the degree and depth of weathering could be critical in water yields (Baran, 2003). No specific data is available for the yield of water from dolerite dykes in the area due to the poor quality of borehole logs over the years, by, in many instances, unqualified persons (Baran, 2003).

Dreyer (1938a) uncovered a section of a dolerite dyke in the excavations at Florisbad, which led him to divide the site into eastern and western sections (Figure 22). It was suggested that the unconsolidated deposits on the eastern side of the dyke were older than those on the western side of the dyke (Dreyer, 1938a). This was based on the difference in the way in which the artefacts he had found had been prepared: The western artefacts were described as being parallel-sided flakes, while the eastern artefacts exhibited convergent flakes and points (Dreyer, 1938a; Oakley, 1954; Sampson 1974).

The spring eyes rise along an east to west strike and in an attempt to establish what was causing the spring eyes to surface along this strike, Fourie (1970) carried out a magnetometer survey. The results showed no anomalies outside of the excavations, possibly due to the thickness of the sand (Fourie, 1970). In the excavation, where there was minimal sand cover, anomalies were recorded, but these were attributed to noise from nearby buildings and water tanks, and were considered unreliable (Fourie, 1970). Grobler and Loock (1988b) also carried out a magnetometer survey of the site, stating that all the data, including borehole results, defined a dolerite sill with a plunge to the

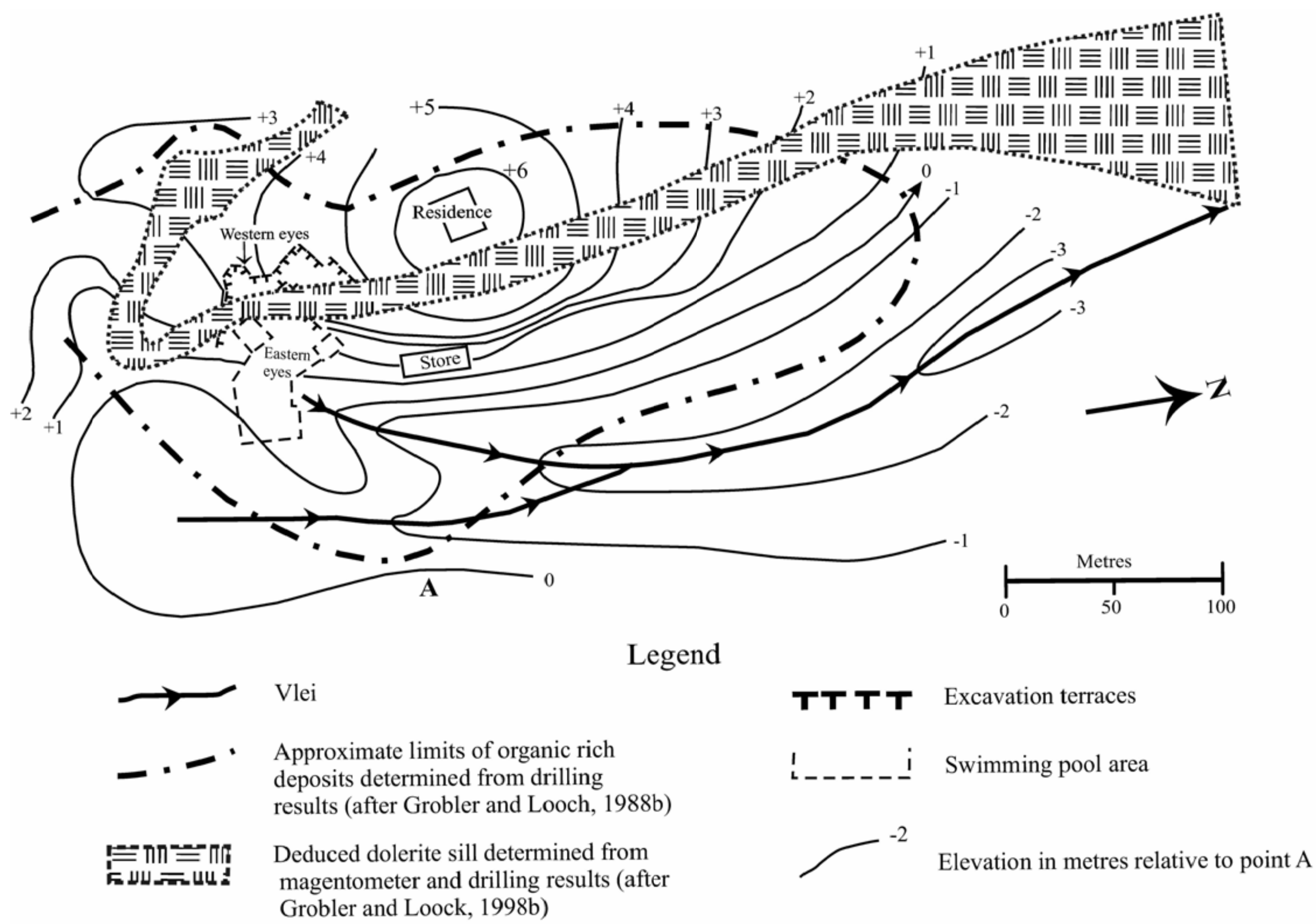


Figure 22. Plan of Florisbad showing the deduced dolerite sill and the extent of the organic deposits (after Rubidge and Brink, 1985; Grobler and Loock, 19988b).

north-west (Figure 22).

In the excavations, just west of the dyke, Dreyer (1938a) unearthed what he described as, a boulder layer in a deep blue clay matrix. A 1.2 m excavation into this layer revealed a zone of homogeneous, highly weathered material. The dyke delineated by Grobler and Loock (1988b) was thought by them to be an appropriate structure for the channelling and emission of underground water in the form of springs at Florisbad.

There is no information for groundwater yields of boreholes located on dolerite dykes in the immediate Florisbad area. However, considerable success has been recorded for borehole yields in dolerite dykes located in other areas of the Karoo Supergroup. At Phuthaditjhaba in the eastern Free State, for example, 87% of boreholes sunk into dolerite dykes were successful with an average yield of 3.2 *l*/s (Kruger and Kok, 1976), where 0.125 *l*/s is considered successful (Baran, 2003). Where boreholes were sunk into the fractured sediments along the dykes, only 40% of boreholes were successful, with an average yield of 1.4 *l*/s (Kruger and Kok, 1976). These results can be compared to boreholes sunk into Karoo sediments not affected by dolerite dykes, which were 16% successful, with a yield of 0.5 *l*/s (Baran, 2003).

5.4.2 Temperature

The temperature of the Florisbad spring-water has remained constant at ± 29 °C over the past 84 years, having only fluctuated by 0.17 °C over this period, with a maximum of 29.05 °C and a minimum of 28.88 °C (Appendix I). Beside the temperatures mentioned in Appendix I, Kent (1964) measured 29.8 °C, while Kent (1948) recorded a temperature of 34.4 °C for the main eye, which is considered here to be in error. It was calculated by Grobler and Loock (1988a) that, if the water intake recharge area were located at Basberg, north of Florisbad, water would have to descend to the contact zone between the Karoo Supergroup sediments and Upper Ventersdorp Group basement rocks at approximately 500 m in order to reach a temperature of 32 °C to 33 °C, so as to issue at 29 °C.

It is interesting to note that of the many boreholes drilled in the immediate Florisbad area, none encountered warm water above 25 °C (Kent, 1964) (see Kent, 1949, Table 9). At Vlakkraal, 5 km south of Florisbad, two springs existed in a vlei which was later covered by the waters of the Krugersdrift Dam. These springs occur in shales and sandstones of the Ecca Group, and according to Kent (1971), their chemical composition was similar to that of Florisbad springs, although the temperature of the water was only 22 °C, implying a much shallower aquifer.

The temperature of spring water is largely determined by the depth at which it originates. It is apparent that if the hypothesis put forward by Grobler and Loock (1988a) is correct, and water must descend to approximately 500 m in order to reach a temperature of 32 °C to 33 °C in order to issue at 29 °C, then the constant temperatures measured at the Florisbad springs would suggest that the temperature of the spring water must have remained fairly constant throughout its history, with no evidence suggesting that it ever had deeper origins, or higher temperatures. Thermal springs with higher temperatures usually exhibit characteristics reflecting these higher temperatures, such as deposits of travertine, sinters, silicified or mineralised micro-organisms, stromatolites, and other chemical, or biochemical precipitation. (Kuman *et al*., 1999; Farmer, 2000; Allen *et al*. 2000). At Florisbad there is no evidence of these, supporting a hypotheses for low palaeotemperatures. Allen *et al*. (2000) noted that organic material is rare in carbonate sinters deposited at above 30 °C, due to the high decomposition rates in high thermal environments. Therefore, the large quantities of organic material in the Florisbad sediments would also strongly contradict a higher palaeotemperature. Kent (1949) noted that, possibly due to their low temperatures, very few South African thermal springs have given rise to mineral deposits. Kuman *et al*., (1999), and research reported on in Appendix III, concluded that the palaeotemperature of the spring-water was similar to the present.

5.4.3 Discharge Rate

Information on the discharge rate of the Florisbad springs, such as the methods used and location of the samples, appear to be mostly vague and unreliable. The sampling locations could have a considerable influence on the discharge rate measured, and it appears that discharge rates in the literature are confined to the area of the swimming baths. Fourie (1970) recorded 21 spring eyes of varying sizes on the floors of the three swimming baths, and another five eyes mainly to the west-north-west of the indoor pool. Due to the lack of detail in the literature, these latter springs appear to have been largely disregarded in any discharge evaluations, which would mean that any flow rate given for the springs as an entirety, would be a minimum flow rate.

For example, Rindl (1915) gave a flow rate of 31.25 *l*/s (112.5 m^3/h), while Kent (1948) has provided the highest flow rate at Florisbad of 44.72 *l*/s (161 m^3/h). However, neither of the aforementioned provided information regarding where the flow test was taken, how the flow rate was determined, or who carried out the determination. In 1961, Mr N. Viljoen of the Provincial Administration measured flow rates of between 1.01 and 3.28 *l*/s, with an average of 2.27 *l*/s (Kent, 1971). On a visit to Florisbad in 1971, Kent (1971) estimated a flow rate of 2.53 *l*/s. Fourie (1970), who himself carried out flow determinations in 1953 on two of the pools (Kent, 1971), gave a combined flow rate of 13.86 *l*/s (49.4) – 15.53 *l*/s (55.9 m^3/h) with an average yield of 14.65 *l*/s. Grobler and Loock (1988a) gave an increase in the flow rate from 1.25 *l*/s (4.5 m^3/h) to 5.22 *l*/s (18.8 m^3/h), after a light earthquake was felt by the residents at the spring site in 1912 (Anon [1], 1980) (see Section 1.1). The accuracy of this latter flow rate may be in question.

Why there should be such a discrepancy between the values given by different researchers is not known. However, Kent (1971), strongly suggests that pumping and dewatering by the mines of the Free State goldfields, centred on Welkom, 90 km north-north-east of Florisbad, and 60 km north of Basberg, the assumed recharge area for Florisbad, has affected the discharge rate of not only the Florisbad springs, but also of other springs and boreholes in the area. Table 11 gives the chronological order of flow rate determinations at Florisbad and lends strong support for Kent's (1971) hypothesis. This would strongly suggest an extensive system of inter-linked aquifers, of varying depths, underlying the region. Fourie (1970) also recorded a daily increase in flow rate and gas discharge between 09h00 and 11h00, and again at 15h00, with a corresponding slight increase in atmospheric pressure, although this was not considered significant. The flow rates for Florisbad (Table 11) can be compared to the Brandvlei spring, which is the strongest recorded spring in South Africa, at 128.1 *l*/s (461 m^3/h), the Aliwal North spring at 28.61 \/s (103 m^3/h), and the Warmbaths spring at 8.33 *l*/s (30 m^3/h) (Kent, 1949).

It has been mentioned (section 5.3) that Vegter (1995, 2001) gave a recharge value of 15-25 mm per annum for the Florisbad area. Kok (1992) estimated that Karoo springs recharged, on average, 8% of the annual rainfall. This would suggest a 40 mm recharge

Table 11. The chronological order of flow rate determinations at the Florisbad spring site showing the decrease in flow rate as a possible result of the dewatering of the Free State goldfields.

Year	Flow Rate *l*/s	Average flow rate *l*/s	Reference
1912	1.25 – 5.22	-	Grobler and Loock (1988a)
1915	31.25	-	Rindl (1915)
1946	First deep mine shaft sunk	-	Kent (1971)
1948	44.72	-	Kent (1948)
1953	13.86 – 15.53	14.65	Fourie (1970)
1961	1.01 – 3.28	2.27	Kent (1971)
1971	2.53	-	Kent (1971)

rate per annum for the Florisbad area, with an average rainfall of slightly less than 500 mm per annum . Kok (1992) noted that areas of mean low rainfall could also experience extreme events, which would increase the recharge rate Kok (1992) also noted that thermal springs, with temperature >25 °C may well be related to deeper regional groundwater systems that extend beyond the localized rainfall area. This extension might imply that groundwater systems may extend into areas with a much higher recharge rate due to a combination of factors such as higher rainfall, a more porous and/or fractured local geology, and the effects of geomorphology on the catchment area. In particular, this may be relevant to permanent springs.

5.5 CHEMISTRY OF THE SPRING- AND GROUNDWATER

As previously mentioned in Section 1.7, the 1988 water sampling was in light of the water bodies being suitable as habitats and breeding environments for amphibians (Douglas, 1992, 1995). Water samples were taken from the spring, vlei, an exploration pit that amphibians were inhabiting, the farm dam, and Soutpan (Figure 2). Water sampling was carried out under the supervision of the Institute of Groundwater Studies, University of the Free State. The Institute of Groundwater Studies analysed the water samples using Inductively Coupled Plasma Optical Emission Spectrometry (ICP-OES) analysis. Samples from the vlei and the farm dam were excluded from this study as the water originated directly from the swimming pools, and thus the springs. The Soutpan water sample was also excluded, as it was the first time in living memory that the pan had filled with water, and the piles of evaporated salt would have had an artificial influence on the results.

Of the water samples taken during the 1988 herpetofaunal study only the results of those from the spring eye and one of the exploration pits (Figure 23, E88, P188) were deemed relevant to this study. In 1999, the 1988 localities were re-sampled (E99, P199) and sampling expanded to the three other excavation pits (P299, P399, P499). Water sampling was carried out under the supervision of the Institute of Groundwater Studies, University of the Free State. The Institute of Groundwater Studies analysed the water samples using Inductively Coupled Plasma Optical Emission Spectrometry (ICP-OES) analysis. All

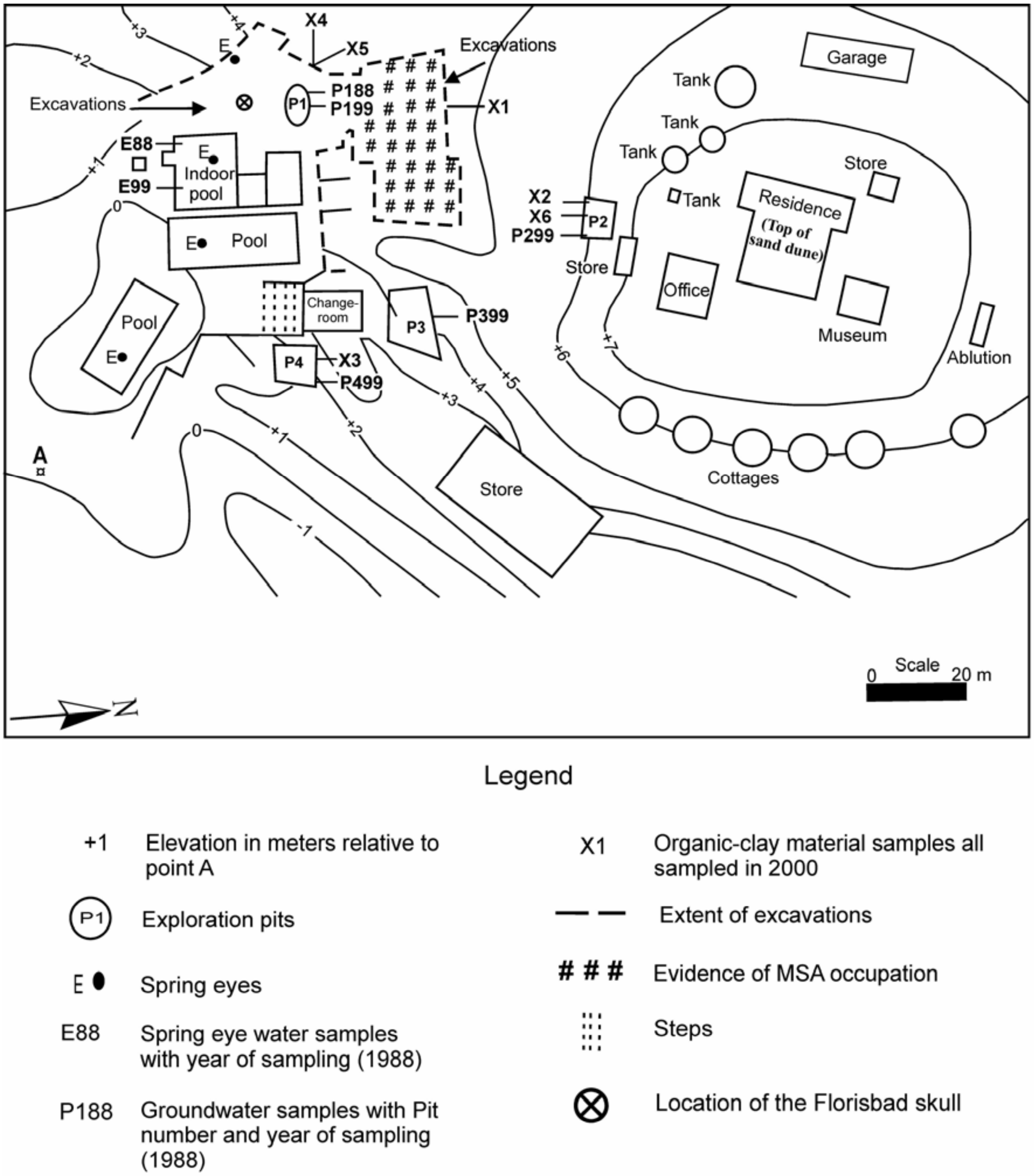


Figure 23. Detailed plan of the Florisbad spring site showing spring water, groundwater and organic-clay material samples (after Douglas, 1992; Appendix I; Appendix 2).

previous water sampling records (Rindl, 1915; Fourie, 1970; Mazor and Verhagen, 1983) were incorporated as part of the data. Rainfall for the year preceding the 1988 water sampling was exceptionally high at 957 mm, while rainfall for the year prior to the 1999 water and peat sampling was slightly above average at 545 mm. This reflected a 42% drop in rainfall between the two periods. It was hypothesised that if the spring-water was the source of the water in the exploration pit, then the quality of the water at the two localities should be comparable.

Rainfall for the year prior to the 1988 sampling period was nearly double the mean annual rainfall of 500 mm, at 957 mm, while rainfall for the year prior to the 1999 sampling period was slightly higher than the mean annual rainfall at 545 mm. The 1998/9 rainfall could therefore be seen as being much lower than the 1987/8 rainfall, but was in fact average rainfall relative to the mean annual rainfall. Mean annual rainfall over the 10 year period prior to the 1988 sampling period was 459 mm, while the mean annual rainfall over the 10 year period prior to the 1999 sampling period was 514 mm, despite the high 1987/8 rainfall. Therefore, despite short-term punctuated high and low extremes in rainfall a mean annual rainfall of around 500 mm appears to have been maintained over the short to medium term. This resulted in a 42% decrease in rainfall between the 1988 and 1999 sampling periods. When comparing the 1988 spring-and exploration pit water sample results to the 1999 results, it was found that the TDS of the exploration pit water was now 58% lower than that of the spring-water (Table 12). This was contrary to the hypothesis mentioned above that, if the spring-water was responsible for the water in the exploration pit, then the quality of the water at the two localities should be comparable. This was also contrary to evidence in the literature, which stated that if concentrations of dissolved substances are high, then the rate of groundwater renewal is low (concentration factor), whereas low concentrations indicate regular recharge (dilution factor) (Bredenkamp, 2000).

Although there was only a slight increase between the 1988 spring-water results (2203 mg/*l* TDS, high rainfall period) and 1999 results (2 363 mg/*l* TDS (average rainfall period) (Appendix I, Table 2), these results conformed with Bredenkamp's (2000) statement. However, results from Appendix I (Table 1, 2) also showed that, not only was the TDS of the pit waters generally higher than that of the spring-water, but that the TDS of the water in pit 1 was higher (2799 mg/*l* TDS) during a high rainfall period (1988) and lower (1174 mg/*l* TDS) during an average rainfall period (1999) (Table 12), this being contrary to Bredenkamp's (2000) statement.

Results between exploration pits 1 and 4 (Appendix 1, Figure 1) showed a west to east TDS increased ion concentration of 666%, while the TDS increase between the spring-water and pit 4 was 287% (Appendix I, Table 1). Groundwater, *per se,* has not been mentioned in the literature, with the general assumption appearing to be that all

Table 12. The difference in water quality (TDS, mg/*l*) between the spring-water and exploration pit 1 water during high and average rainfall periods (after Appendix I).

Locality	TDS mg/*l* 1988 wet period	TDS mg/*l* 1999 average period	% increase/ decrease
Spring-water	2203	2364	+7
Exploration pit 1	2799	1174	-58
Spring-water/Exploration pit 1	2203/2799	-	+27
Spring-water/Exploration pit 1	-	2364/1174	-50

subterranean water at Florisbad originated from the spring eyes. This strongly supported the argument for two separate hydrological entities. Table 13 provides some examples of the variations in ion concentrations, between the spring-water and exploration pit 4. It will be noted from Table 13 that higher ion concentrations occurred at all individual levels of concentration, over a wide range of ions. Further to this, 73% of ions in

Table 13. Some examples of variations in the increase in anion and cation concentrations between the spring-water (1999) and the water of exploration pit 4 (1999) (after Appendix 1).

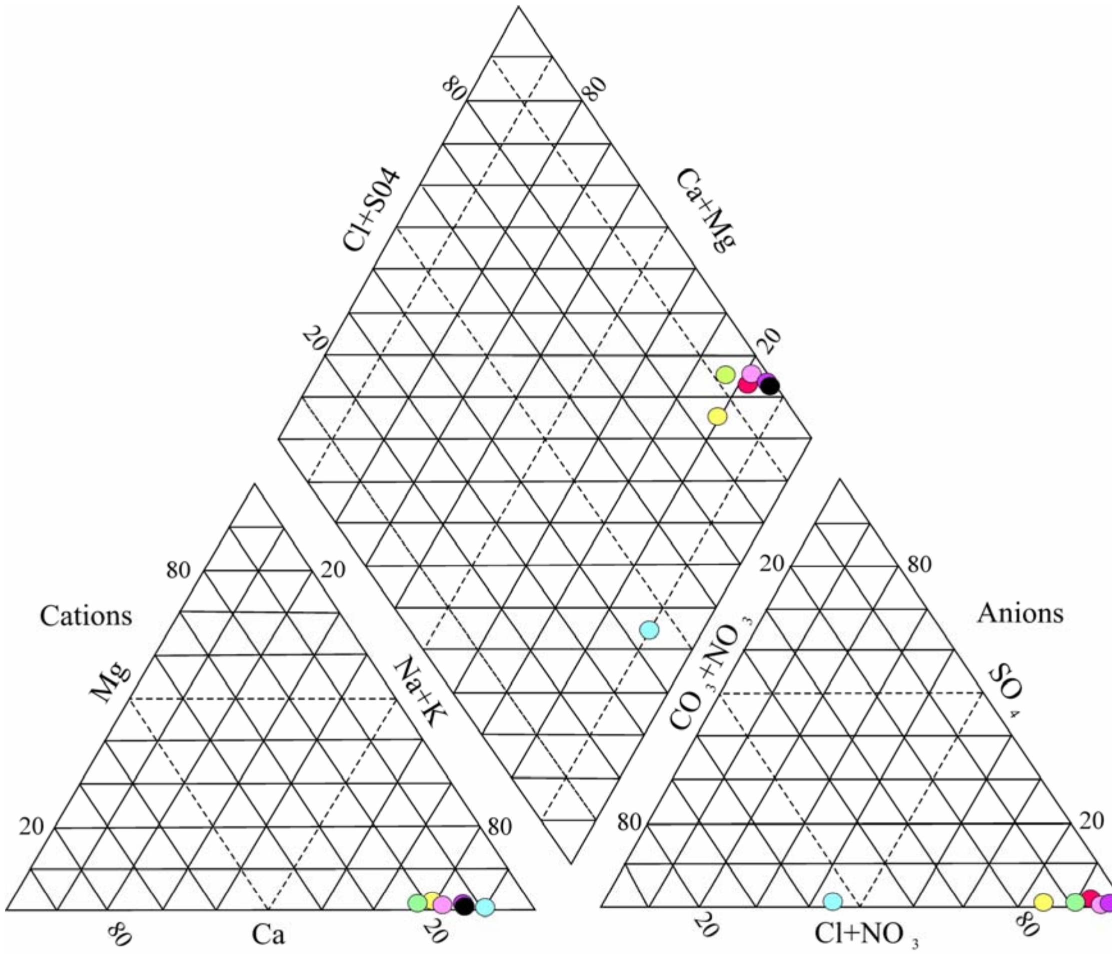
Anion/Cation	Spring-water mg/l	Exploration pit 4 water mg/*l*	Relative ion increase in Pit 4 as a %
Na	784.00	2 920.00	272.45
Bromide	19.31	71.64	273.00
Cl	1 442.00	5 648.00	291.68
B	2.59	10.16	292.28
Ca	100.00	400.00	300.00
Sr	2.97	11.20	301.43
K	10.10	42.54	321.19
Cu	0.004	0.017	325.00
Nitrite	0.73	4.81	566.21
Li	0.35	2.31	560.00
Mg	0.44	9.04	1 954.56
Sulphate	1.50	81.60	5 340.00

exploration Pit 2 showed higher concentrations over the spring-water, while ions in exploration Pits 3 and 4 were 86% higher. This evidence strongly indicated that the exploration pit water could not be directly related to the spring-water, and should be seen as a separate entity, namely groundwater. The organic-clay layers, and in particular Peat II, has contributed to an elevated, or perched, water table within areas of the spring site. Greater detail of the chemistry of the spring- and groundwater is given in Appendices I and II.

The ground water is not thought to be of meteoric origin, as with the TDS of the groundwater already being that much higher than the spring-water, a higher undiluted groundwater TDS would only provide even stronger support for the groundwater hypothesis. The level of the water in the excavation pits did not drop from evaporation or seepage even during hot dry spell, which further supported the hypothesis of a perched water table.

Figure 24 is a Piper diagram plot of all the Florisbad water samples in this study. It should be noted that some of the results were so similar in Figure 24, that a number of the points have been concealed behind other points. The Piper plot confirms the high Na, Cl content of both the spring- and groundwater, representing a strong sodium chloride facies. This does not agree with the deduction made by Vegter (1995, 2001) that there are two principal groundwater hydrochemical facies for the Florisbad area, namely, a calcium/magnesium chloride/sulphate facies, and a calcium/magnesium bicarbonate/carbonate facies. The Florisbad waters analysed reflect a very low Ca, Mg and $S0_4$ content. A Piper diagram presented by Grobler and Loock (1988a) supports these results, with Grobler and Loock (1988a) drawing the conclusion that the Florisbad spring water is relatively old.

For purposes of comparing the TDS of the Florisbad water, Figure 25 represents Stiff diagrams of the water samples from Figure 24, based on the same scale. Samples E88 and E99 reflect the chemical stability of the spring eye water over the high (E88) and average (E99) rainfall periods respectively, while samples P188 and P199 show the TDS variation of the water in exploration pit 1 over the high (P188) and average (P199) rainfall periods. Samples P299, P399, and P499 show the increase in salinity in a west to east direction between the exploration pits, and in particular the increase in Ca in pit 4 (P499). Stiff diagram, R15, based on Rindl's (1915) spring water analysis, reflects the stability of the spring-water over a longer period, while the rainwater sample of Litthauer (2007) provides further comparison.



Legend

For the localities of the samples see Figure 22.

E Sample from indoor spring eye

88 1988 sampling period (High rainfall)

99 1999 sampling period (Low rainfall)

P Sample from exploration pit

R15 Spring eye sample after Rindl 1915

E88 E99 P188 P199 P299

P399 P499 R15 Rain Bloemfontein (After Litthauer 2007)

Figure 24. Piper diagram of the spring eye and exploration pit waters at Florisbad. For comparative purposes the spring eye water analysis of Rindl (1915) and Bloemfontein rainwater (after Litthauer, 2007) have been ploted.

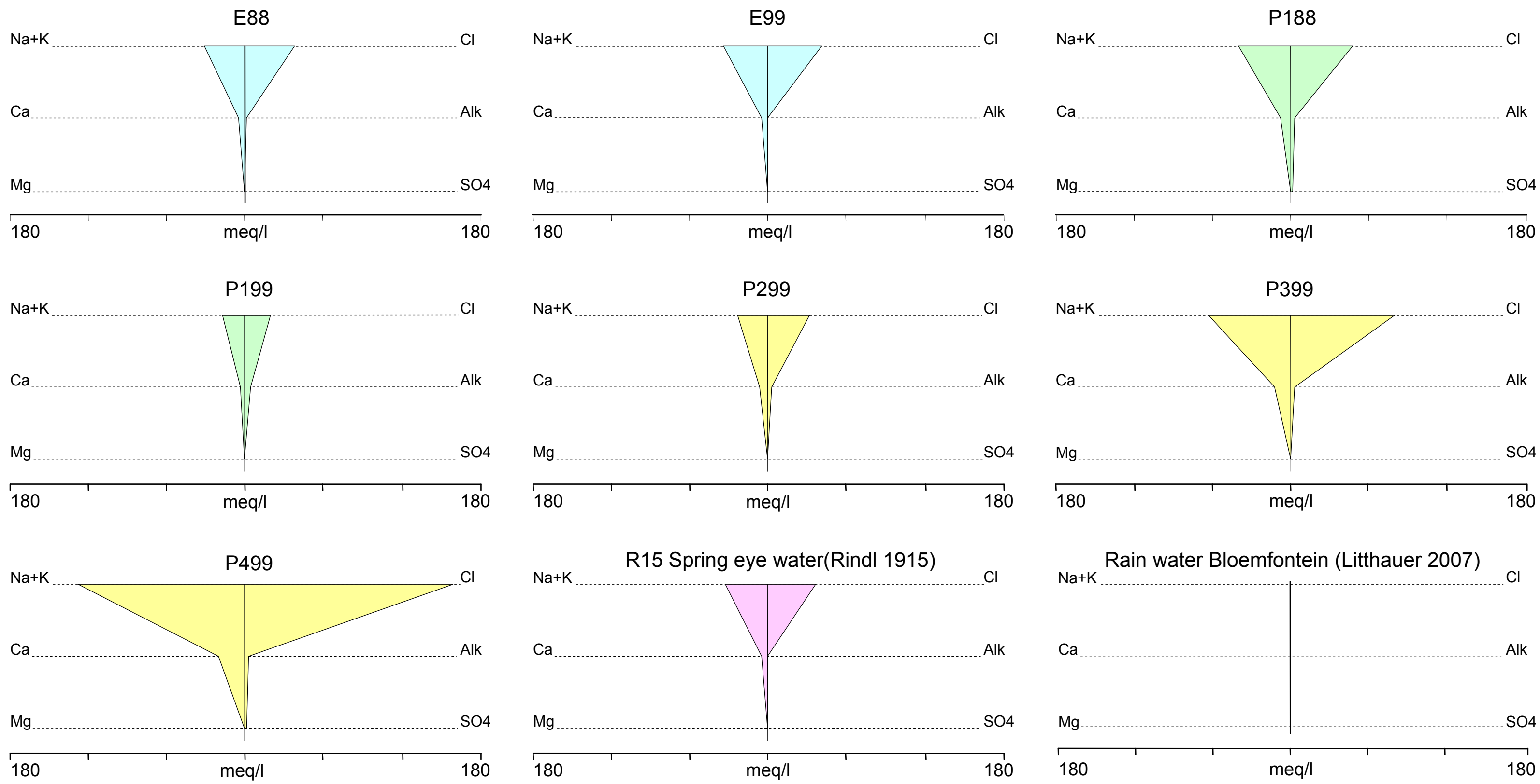


Figure 25. Stiff diagrams for the Florisbad spring eye water (E88, E99), exploration pit 1 waters (P188, P199), and the waters of exploration pits P299, P399, P499. The spring eye water analysis of Rindl 1915 and Bloemfontein rainwater (Litthauer, 2007) are presented for comparison. For further comparison purposes all samples have been plotted on the maximum scale. For further detail on the sample codes see Figure 23.

As the surface elevation of exploration pit mouths (P299 to P499) fell from west to east, so did the depth of the pits to groundwater level, with P299 being 6.70 m deep, P399 being 4.20 m deep and P499 being 2.95 m deep, suggesting that the top of the water table was more or less at the same elevation between pits. The walls of P399 and P499 had both slumped to some extent, with P499 having slumped significantly more than P399, providing a larger groundwater surface for evaporation. The depth of P299 helped reduce the effects of evaporation, as the groundwater was shaded from the sun for most of the day and was also relatively sheltered from the effects of wind. P399, although being shallower, was heavily shaded by large eucalyptus trees, and was later partially covered with iron sheeting. P499, on the other hand, was totally exposed to the sun and wind. In examining these factors it is suggested that this increase in TDS from west to east is largely as a result of evaporation resulting in a concentration of salts as suggested in Appendix I. The lower TDS of samples P188 and P199 (450 mm deep), in relation to the other exploration pits, is thought to be due to it being located in the open excavation area where the elevation of the excavation was below the level and influence of Peat II, with the dilution effect of the nearby spring eyes having a significant effect.

Loock and Grobler (1988) and Grobler and Loock (1988a) suggested that the chemistry of the spring-water was directly related to the organic content of the Tierberg Formation sediments of the Ecca Group. The Tierberg Formation sediments were deposited in a marine environment containing well defined zones of carbonate concretions and fossiliferous zones, which also contained carbonate concretions (Loock and Grobler, 1988). Kent (1949) felt that such a high NaCl content was uncharacteristic of the Ecca Group rocks, and was more related to the underlying Dwyka Group rocks, through which he suggested the water had permeated. Loock and Grobler (1988) credited salinization of the spring water to the Ecca Group rocks, but mentioned that the water would have to have descended to the contact zone between the Karoo Supergroup sediments and Upper Ventersdorp Group basement rocks, upon which the Dwyka Group rocks rests. With the possible exception of Na, and Cl, and work reported on in Appendices II and III, it is concluded that, due to the low water temperature and alkalinity of the spring water, dissolution in the sub-surface geology had contributed little in the form of salinization to the spring water. These conclusions are further supported by the low Ca content of the spring water, particularly in light of the carbonate concretions mentioned by Loock and Grobler (1988).

The fluoride content of the Florisbad spring-water is high at 5.5 to 6.0 mg/*l*, this being in relation to the recommended limit for drinking water quality of 1.0 mg/*l*, and the maximum allowable limit for drinking water quality of 1.5 mg/*l*. This fluoride level compares to that of the artesian borehole at Glen Agricultural College of 5.3 mg/*l* (see section 4.4). Flourine is the most abundant halogen in sedimentary rocks and must be considered as a major source of fluoride in groundwater (Woodford and Chevallier, 2002). Woodford and Chevallier (2002) note that fluoride concentrations cannot only be controlled by near surface lithology, because this does not explain the marked fluoride variability found, on a local scale, in shallow aquifers in the upper 150 m bgl, where the host-rocks are similar.

Analysis of borehole water indicates that the chemical composition of groundwater varies considerably within Ecca Group rocks. Generally, the salinization of borehole water within Ecca Group rocks are considerably lower than that of the Florisbad spring water. Chemical values taken from Baran (2003 for Ecca Group rocks (Florisbad spring-water value are given in brackets) show that TDS ranged from 130.0 to 778 mg/*l* (2363 mg/*l*), NaCl 11 to 208.9 mg/*l* (2226 mg/*l*), and Ca 5.9 to 15.5 mg/*l* (100 mg/*l*). Woodford and Chevallier (2002) give the following groundwater ranges for Karoo sediments: TDS 450 –1000 mg/*l*; pH 8.0-8.5, dropping to 7.5 in the east and north; Ca 30- 80 mg/*l*, dropping to 10-30 mg/*l* in the east; Na <100 mg/*l*, Cl >1000 mg/*l*; K 10-20 mg/*l*; total alkalinity 200-300 mg/*l*; Si 15-25 mg/.

While there is a faint hydrogen sulphide (H_2S) smell emanating from the springs, the quantity of dissolved H_2S is very low at 0.003 mg/l (Fourie, 1970). The main constituent of the free gas component is methane (CH_4) at 70%, with the balance being comprised of nitrogen (N_2) (27.4%) and carbon dioxide (CO_2) (3.6%) (Rindl, 1916; Fourie, 1970).

With total dissolved ions of 2363 mg/*l*, Mazor and Verhagen (1983) noted that the Florisbad spring was the most saline of all South African springs. While the Florisbad water may be the most saline of South African spring-waters, even at current day levels, it is still relatively sweet and potable for most modern animals, and was most probably just as potable for the palaeofauna, (Douglas, 1992). Smit (1977) noted that Kalahari animals could tolerate water with a TDS of 6 000 mg/*l*. Kruger and Lubczenko (1994) gave the salinity tolerances for pigs and milking cows as <3 200 mg/*l*, dry dairy cows and horses as <4 500 mg/*l*, beef cattle as <5 760 mg/*l*, and sheep as 6 400 mg/*l*. Kempster *et al*. (1980) set an upper limit for livestock water quality at TDS 14 000 mg/*l*. This would also imply that, even if spring-water salinity levels had risen in the past, for whatever reasons, the substantial presence of the fossil fauna indicates that the spring-water was potable to the vast herds of game inhabiting the area over time. Based on total dissolved solids of 2201 mg/*l,* Florisbad has a salinity 82 % higher than that of the Aliwal North spring, and 886 % higher than the Warmbaths spring (Douglas, 1992). Based on saline content, Florisbad has a salinity 112% higher than the Aliwal North spring and 2617% higher than the Warmbaths spring (Douglas, 1992).

Not only are the Florisbad TDS levels within the potable drinking water range for mammals, but also well within the tolerance levels for habitation by a number of amphibian species, which are used as modern water quality bio-indicators. With TDS levels of up to 2798 mg/*l*, the outdoor swimming baths, vlei area and exploration pits were inhabited at times by permanent water breeding species, such as *Xenopus laevis laevis, Amietia fuscigula*, and *Amietia angolensis,* although these species were never observed breeding (Douglas, 1992). Munsey (1972) gave an NaCl tolerance level for *X. l. laevis* of 7582 mg/l, while the pH of 9.43 for the spring-water (Douglas, 1992) was higher than the maximum natural pH of 8.86 given by Picker (1985) for this species.

Temporary pond breeders, on the other hand, where breeding sites are used only in conjunction with seasonal rainfall, and where the mineral content of the water is much lower, appeared to avoid waters associated with the spring and groundwater i.e. swimming pools, the vlei, and exploration pits. These species included *Pyxicephalus* *adspersus*, *Cacosternum boettgeri*, *Kassina senegalensis* and *Tomopterna cryptotis*, who were all observed breeding at the only fresh water site on the farm, namely, the area above the dam site (Figure 2) (Douglas, 1992). The use of amphibians as bio-indicators of water quality was particularly evident at this dam where the actual dam was avoided as a breeding and habitation site due to water being pumped from the swimming baths into the dam. The saline Florisbad spring- and groundwater therefore appear to be totally unsuitable for the permanent habitation of temporary pond breeders. The terrapin, *Pelomedusa subrufa*, was also often observed in the outdoor swimming pools (Douglas, 1992).

5.6 THE DEPOSITIONAL ENVIRONMENT AND SEDIMENTATION

5.6.1 Theories on the Deposition of the Florisbad Sediments

Numerous researchers have put forward theories on the formation of the spring site. Gardner (1932) and Deacon (1970) considered spring mounds to be derived from underlying bedrock through elutriation. Brink (1987) saw the spring site developing through sand being brought to the surface by spring vents, which became enlarged due to mechanical and chemical actions, with vegetation developing around the margins of the spring pools. The vents then became blocked, with the size of the mound increasing due to a combination of factors such as choking vegetation, the deposition of windblown sediments, and a diminishing supply of groundwater (spring-water) (Brink, 1987). Brink (1987) further hypothesised that after this closure, the spring eye moved laterally along a bedrock fissure to form a new passage, which cut through the existing strata to form another sand unit, partly overlapping the mound created by the original vent. It was contended that these processes would have repeated themselves to form a build-up of sediments, which includes alternative layers of organic and non-organic deposits (Brink, 1987).

There appears to be little contention in the literature that "peat" (Figure 4 and 23) has been used as a convenient term for describing the four organic rich horizons at Florisbad.

These horizons were described by Van Zinderen Bakker (1989) as so-called ‘peat’ layers of carbonaceous clay and silt, and by Butzer (1988) as organic horizons where the ‘peat’ lacked definition in terms of inter-woven vegetation structure. The quantities of sand deposited within these layers would also tend to make the term “organic rich” more appropriate. The lack of inter-woven vegetation structure, as noted by Butzer (1988), was particularly noticeable in Peat II, but vegetation structure was very high in Peat IV. The author supports the point of view that the “peat” layers should rather be seen as layers of organic rich material, clay, silt, and sand, and these layers are therefore referred to here as organic-clay layers, based on their principal components.

Butzer (1988) described the site as a 7 m mound of spring beds inter-bedded with organic layers, and only partly covered by aeolian material. The site was seen as having developed through spring flow, which was determined by a deep-seated regional aquifer and fluctuations in recharge, with the bulk of the quartz grains comprised of detrial sand released from the underlying Ecca, through which the spring-waters had passed (Butzer, 1988). Sandy pools then developed at the spring site, with peaty organic horizons developing as vegetation encroached on less active springs during periods of low discharge (Butzer, 1988). Vegetation was later submerged by periods of more active spring discharge, and subsequently buried by spring sediments (Butzer, 1988). Dreyer (1938a) also suggested that the Florisbad deposits represented sand output from the spring and considered the Florisbad “mound” as being formed by the sand from a huge (unknown) eye beneath the highest point of the “mound”. This would suggest that the spring pan extended well to the north of the current eyes.

It has been contended that a close relationship and correlation existed between the spring sedimentation and the shoreline position of the adjoining palaeolake complex (Soutpan) and that this relationship played a significant role in the modification of the spring mound sediments (Joubert and Visser, 1991; Visser and Joubert, 1991; Kuman *et al*. 1999; Henderson, 2001a). It was proposed that deposition at Florisbad was directly related to the palaeolake levels (Soutpan), which reflected climatic conditions at the time (Joubert and Visser, 1991; Visser and Joubert, 1991). This involved cyclic sedimentation with soil horizons forming during arid stages when palaeolake levels were low, while deposition of palaeolake bottom silts occurred during wet periods when the spring area was flooded by the palaeolake (Joubert and Visser, 1991; Visser and Joubert, 1991). These cyclic transgression and regression sequences of the palaeolake shore line were translated into four low water-level phases and three high phases (Joubert and Visser, 1991). It was noted that waterlogged conditions had existed at levels higher than the existing water table, and that load structures had been identified as high as the top of Peat II (Brink, 1987). These load structures, as well as faunal evidence, led to the suggestion of the existence of a large water body in the past, but whether or not this was related to the water levels in the palaeopan still needed to be established (Brink, 1987).

Sedimentary deposits to the east of the spring site have been referred to as the "lacustrine sequence" (Joubert and Visser, 1991; Visser and Joubert, 1991) because of their presumed association with the Soutpan complex. These sediments were interpreted as being palaeolake bottom deposits, directly related to Soutpan, being based largely on the presence of freshwater gastropods in the clay facies, (Joubert and Visser, 1991). It was also suggested that during humid phases the Soutpan palaeolake complex enlarged to more than twice its present size (Visser and Joubert, 1991). Despite their "flooding of the palaeopan" hypothesis, Visser and Joubert (1991) noted that the distribution and lateral variation of the organic-rich deposits at the spring site reflected waterlogged, or bog conditions, on a poorly drained flood plain, reflecting the local influence of constant freshwater discharge at the spring. Fourie (1970), who originally described these deposits and the freshwater gastropods, did not see them as being related to the palaeolake, but rather as an integral part of the spring site itself. As mentioned in section 3.2, Scott and Rusouw (2005) noted that the high water levels proposed by Visser and Joubert (1991) were in conflict with the pollen evidence where Chenopodiaceae pollen was an indication of dry conditions.

Butzer (1984) postulated that sedimentation had occurred in a flood plain environment and suggested that Peat I had a similar semi-aquatic accumulation as Peat II. Van Zinderen Bakker (1989) agreed with this, stating that the intervals in the organic-clay layers were due to rises in the level of water in the pan (Soutpan) and water table, and that during periods of stable spring flow, vegetation around the spring produced organic deposits.

It was noted by Grobler *et al*. (1988) that prior to the formation of the pans, the area was already in a state of low topographic relief. In effect, there was most probably not much difference between the elevation of the Soutpan floor, which came under the influence of deflation, and the spring site, which came under the influence of aeolian deposition. As the processes of deflation and deposition continued, the surrounding area deflated, lowering the elevation, while aeolian deposition increased the elevation to the south and south-east of Soutpan. The top of the Florisbad dune is presently 25 m (approximately) above the Soutpan floor (Appendix IV) while there is only a 19 m decrease in elevation across the panveld from Florisbad westward, over a distance of some 120 km (Douglas, 1992). Should the flooding of the palaeopan hypothesis be a consideration (Joubert and Visser, 1991; Visser and Joubert, 1991), including the suggestion that the palaeolake may have enlarged to twice its size (Visser and Joubert, 1991), this would have implied that the water in Soutpan would have had to have risen to at least the level of the top of Peat IV (±21 m) to ensure its formation.

Any such significant increase in the relatively fresh water level in Soutpan would therefore have resulted in a sheet of water spreading far beyond the Florisbad spring site, covering a large part of the western Free State. There appears to be no evidence of this, and with so many pans in the area, any such flooding would have filled these pans and other low lying areas with relatively fresh water as well. Judging by the large number of fossil remains recovered from such a small area at Florisbad (Brink, 1987), and with the availability of so much standing fresh water over such a large area, it may be asked why game would have concentrated to the extent they did, at the Florisbad spring site. There also appears to be no evidence of sedimentary deposits similar to those at Florisbad occurring at any other localities in the area, particularly the presence of the organic-clay layers. The pollen content of the argillaceous green sands between Peats I and II, which corresponded to one of the proposed high lake levels by Visser and Joubert (1991), was seen as a horizon dominated by Chenopodiaceae-type pollen, indicating a period of dry conditions (Van Zinderen Bakker, 1989; Scott and Brink, 1992).

Kuman *et al*. (1999) suggested that a number of micro-environments existed in response to the waxing and waning of the spring, as well as the expansion and contraction of the nearby palaeolake. Based on grain sample size and distribution, it was concluded that most of the sediments not only accumulated under uniform, low-energy, subaqueous environmental conditions, but also under several composite geomorphic regimes, and had not been reworked after deposition (Kuman *et al*., 1999). It was further noted that the sand layers were usually more than 10 cm thick, with an absence of thin layers, which would denote annual, or shorter, cyclical deposition systems (Kuman *et al*., 1999). The sharp bedding contacts of the Pleistocene levels were seen as indicating sudden changes in environmental conditions due to the possible increase and decrease in the spring-water discharge, with the stratigraphy of the layers reflecting changes in the environmental water regime, from open dam conditions to vegetated marshland (Kuman *et al*., 1999).

5.6.2 Current State of the Florisbad Sediments

The sediments and stratigraphy of the Florisbad spring site are important in that they hold the key to the history of the site. Lawson and Thomas (2002) noted that geomorphic features can be seen as an environmental archives. This archival record holds the key to aspects such as the formation of the site, palaeoclimate, and the preservation and fossilization of both the Old Collection and the MSA human occupation horizon.

Rubidge and Brink (1985) described investigations into the lithostratigraphy and depositional history of the Florisbad sediments as still being in their initial stages, and suggested that several models could be proposed. Kuman and Clarke (1986) noted that the spring cycles, which are recorded in a repeated alternation of sands, silts and organic deposits, are poorly understood and controversial. The stratigraphy and sedimentology for a relatively small part of the Florisbad spring site has been documented, and it is hoped that the hypotheses presented in this thesis will go some way to resolving some of the issues. The stratigraphy of the Florisbad site is complex (Rubidge and Brink, 1985), which has been compounded by a yet unresolved depositional history, internal stratification of the sand dune, and erosional forces on the leeward and eastern side of the sand dune. These aspects are addressed in Section 5.6.5 and Appendix IV. The stratigraphy on the face of the excavations may vary considerably from one section to the next, over relatively short distances, while being unbroken and horizontal on other faces. The generalized stratigraphy is reflected in Figure 26.

The sedimentary deposits underlying Florisbad are made up of alternating layers of sand, and what are generally referred to as peat layers (Figure 26) (Dreyer, 1938a; Kuman and Clarke, 1986; Brink, 1987; Henderson, 2001a). Four of these organic clay layers have been identified at Florisbad, but not all are in evidence on the face of current excavations, with only Peat II and Peat IV being clearly discernable during the study. Various authors have described the state of the present day organic-clay layers. Fourie (1970) mentioned a thin basal peat layer (Peat I) resting directly on the dolerite. Peat I, which is located below the water table was described by Dreyer (1938a) as being bituminous, and by Meiring (1956) as being waxy and free of modern tree roots. Fourie (1970) described Peat I as being waxy with scattered pieces of decomposed wood, where the fibrous structure could still be identified.

Peat II is currently at the same level as the water table in pits 2, 3 and 4. Meiring (1956) described Peat II and III as being heavily contaminated with modern tree roots. The Peat II sampled in this study did not however reflect any heavy contamination of modern tree roots, vegetation or fibrous structure. Butzer (1984, 1988) saw Peat II as an organic, very dark grey sandy loam layer, inter-bedded with dark grey loam and pockets of light grey loam. Organic material comprised what Butzer (1984, 1988) referred to as vegetation structures, carbonised with fine vertical roots, and also noted the lack of inter-woven vegetation structure. The Peat II examined in this study was of a grey sandy/clay nature with a very fine black clay fraction which had the consistency of light grease. The effect

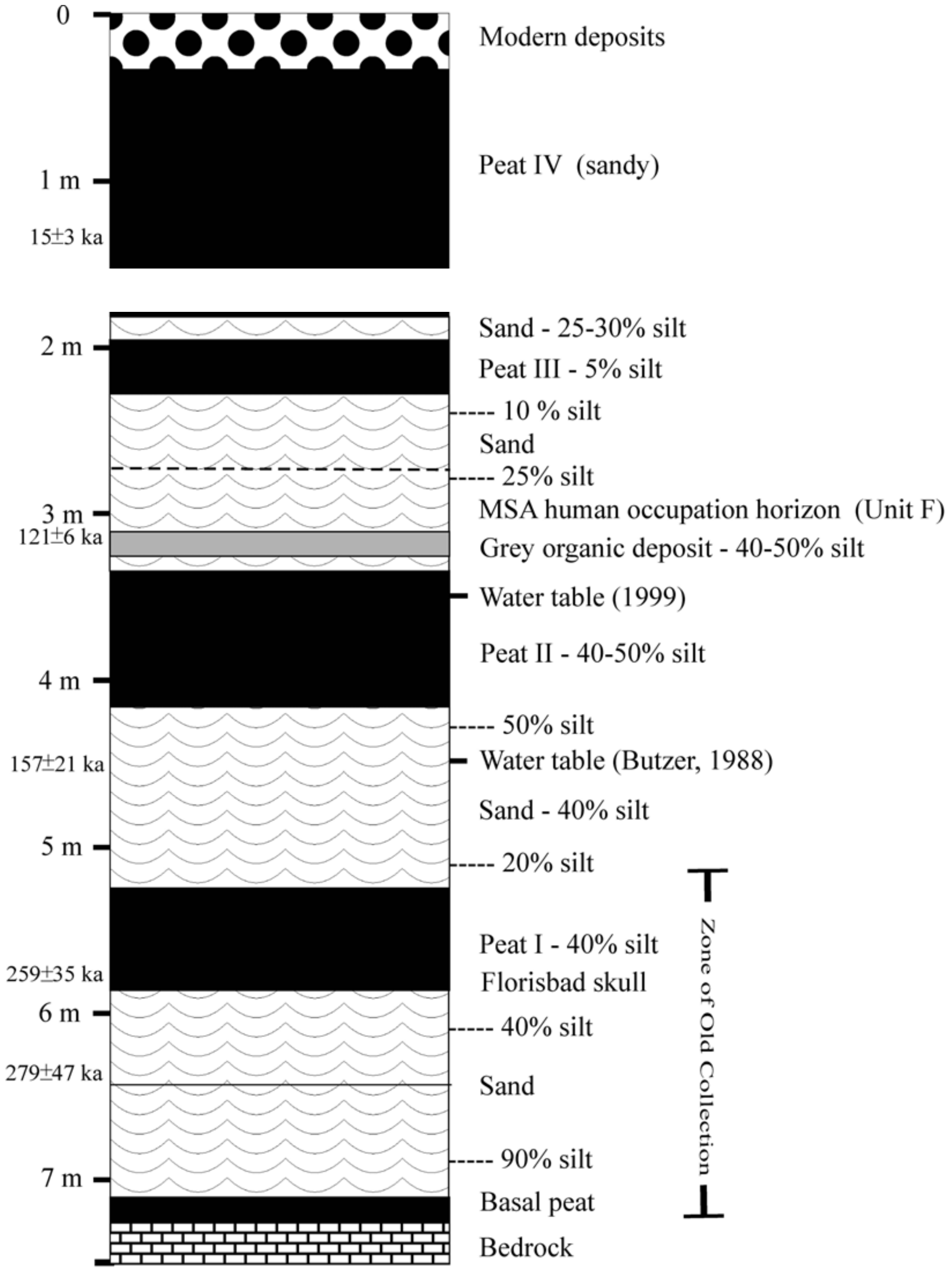


Figure 26. A simplified schematic cross-section of the Florisbad sedimentary deposits, showing the major organic-clay (peat) and sand horizons with their silt content (after Kuman *et al*., 1999 and Appendix IV). Dates in ka after Grün *et al*. (1996)

of compression due to the vertical weight of the sand is evident in Peat II in that the fine black clay fraction is being squeezed out from between the sand, and onto the walls of the pits. However, this oozing of the fine clay fraction may also partly be due to swelling pressure caused through the absorption of water by the clay. Load structures, mentioned by Brink (1987), are yet another aspect and indication of this compaction.

Meiring (1956) noted that in a part of the 1952 excavation (Figure 27), none of the layers were disturbed, but were continuous throughout the exposed face. This is further illustrated in Figure 28 which shows the upper section of excavation pit 2 (Figure 23). It was noted that because of the apparently relatively low spring-water temperature (29° C) over time, diagenesis of the underlying Ecca and Dwyka formation rocks had contributed little to the salinization of the site (Appendix I; Appendix II). Based on sand grain shape and surface features, Kuman *et al*. (1999) and Van Zinderen Bakker (1989) concurred that the spring sediments appear to be derived predominantly from an aeolian source, while sands from the lower levels, although showing signs of water transportation, were also originally of an aeolian nature. Grobler and Loock (1988a, 1988b) also stated that deposition was largely as a result of aeolian processes, and this was supported by Joubert and Visser (1991).

Dreyer (1938a) and Fourie (1970) both noted that the organic-clay layers became progressively more sandy towards the surface. Peat IV was described by Butzer (1984, 1988) as organic, grey to dark grey loam to loam clay, comprising abundant organic voids and abundant root casts, as well as black organic muck of semi-aquatic facies. Joubert and Visser (1991) described Peat IV as clayey silt, which contained the remains of reeds, disseminated carbonate and calcareous nodules. Peat IV, on the western wall of the excavation, is located just 650 mm below the soil surface and is 2.8 m thick. This can be compared to the relatively narrow thickness of the Peat IV layer reflected in the section in Figure 26.

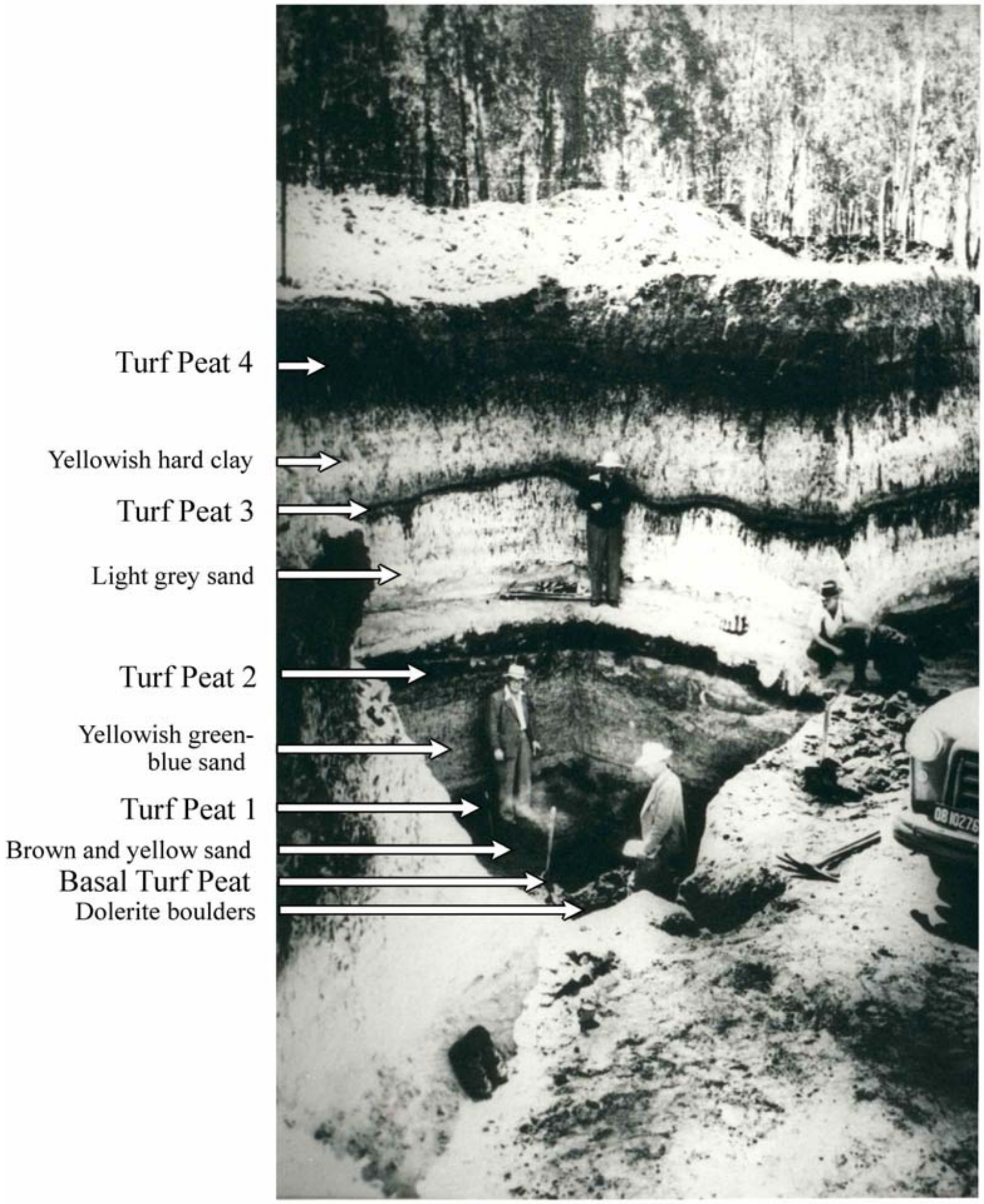


Figure 27. Meiring's 1952 excavations at Florisbad showing the continuous horizontal stratigraphy of the sediments at this particular location, with Meiring's 1952 original descriptions of the various layers (Photo: Meiring, 1956).

It should be born in mind that the degree of deposition and sedimentation was highly variable, being punctuated by high and low varying peaks. Variability would have been due to factors such as rainfall and temperature, which would have in turn related to wet and dry periods, wind, aeolian deposition, and the development of the lower organic layers, their decomposition and subsequent compression and compaction. Compression and compaction would also have been variable, dependant on the aforementioned factors, but they would also have been a varying constant in that they were always a present force, dependant on the accumulation of sediments and water. The lower deposits would always have been under the increasing effect of compression and compaction as the accumulation of the sediments and water increased. It would, at this time, be difficult to determine whether narrower layers of organic matter actually formed as narrow layers, or whether they were much wider layers that had been compacted over time. However, judging by the loose structure of Peat IV, I would appear that the lower organic layers were originally similar in their formation, and became compacted at a later stage.

As with the stratification of the sand dune, there are many variables influencing descriptions of the organic-clay layers. These include the composition and quantities of organic matter, clay, silt, and sand; the thickness of the overburden; the thickness of the section being examined; the location of the section being examined, the water content of the layer; as well as many other unknown historical factors. These factors can in turn be applied to variations in the salinization of the organic-clay layers as well as that of the groundwater. It is therefore apparent that from the above mentioned variables affecting the formation of the organic-clay layers, that the present state and composition of the organic-clay layers may vary considerably for samples from the same layer, taken at different localities. In relation to European peat, the Florisbad organic-clay layers in their current state, and in particularly the lower organic-clay layers, can hardly be classified as peat.

Figure 28.The upper section of excavation Pit 2 on the leeward face of the Florisbad sand dune showing the horizontal stratigraphy of the sedimentary deposits which appear to have been deposited in an aquatic environment (Photo: P.J. Holmes).

5.6.3 Salinization of the Organic-Clay Layers

Once water quality criteria for the spring- and groundwater were established (Appendix I), a number of questions arose. Firstly, why was the salinization of the groundwater generally so much higher than the spring-water? Secondly, how did the groundwater manage to maintain such a high salinity? Thirdly, what were the origins of the higher concentrations? Fourthly, why were there variations in TDS between high and average rainfall periods, and lastly, why was there a variation, contrary to norm, of lower ion concentrations with high rainfall, and higher ion concentrations with low rainfall?

For this part of the study, the Peat II organic/clay layer was sampled from the walls of the exploration pits (P2, X1; P3, X2; P4, X3) by first removing ±80 mm of the outer layer before taking the sample. In a similar manner, the north wall of the MSA excavation was sampled for Peat II (X1), while Peat IV was sampled from the west wall of the open excavation (X4; X5). In November 199 an additional Peat II sample was taken from Pit 2 (X6) for water extraction analysis by All Peat samples were analysed by the Department of Soil Science, University of the Free State, using Atomic Absorption and Spectrometric analysis. Sample X6 was analysed as a water extraction sample by the Institute of Groundwater Studies, University of the Free State, using ICP-OES analysis (Douglas, 2009). X-ray difractometric analysis was carried out by the Department of Geology, University of the Free State, on Peat II samples from Pit 2 as well as sand expelled from the spring eye. Sedimentation was used to separate and determine the clay fractions of Peat II. Rainfall was also examined over a 78 year period in order to determine whether rainfall had any effect on the long- and short-term quality of the spring-and groundwater.

Results of the organic-clay layers analyses are given in Appendix II, where Tables 1 to 4 show that the organic-clay layers were considerably more highly salinized than either the spring-, or the groundwater. Peat II results showed a higher ion concentration of 945% over the 1999 spring-water results, with Peat IV showing a maximum higher ion concentration of 902%. It was therefore determined that salinization was at its highest levels within the organic-clay layers (Table 14).

Table 14. The difference in ion concentrations (mg/*l*) between the groundwater and the Peat II organic-clay layer in exploration pits 2 and 4, during 1999, with a comparison to water from the spring eye (after Appendix I; Appendix II).

Ions (mg/*l*)	Spring eye	Ground-water Pit 2	Peat II Pit 2	Ground-water Pit4	Peat II Pit 4
Cl	1 442.00	1 104.00	7 727.40	5 648.00	12 513.00
Na	784.00	554.00	4 240.00	2 920.00	6 800.00
Ca	100.00	136.00	2 768.00	400.00	5 000.00
K	10.10	12.90	640.00	42.54	352.00
P	1.93	0.01	26.00	0.01	34.00
Mg	0.44	5.15	24.80	9.04	70.00
Zn	0.02	0.03	2.40	0.03	6.00
Total ions	2 338.49	1 812.09	15 428.60	9 019.62	24 772

Solutions to the above questions were initially investigated by referring to references relating to previous stratigraphic and sedimentary research at Florisbad (Dreyer, 1938a; Meiring, 1956; Fourie, 1970; Rubidge and Brink, 1985; Brink, 1987; Butzer, 1988; Van Zinderen Bakker. 1989; Joubert and Visser, 1991; Visser and Joubert, 1991; Kuman *et al*., 1999). This literature was reviewed for processes and scenarios which could possibly explain and support the results of a higher salinity for both the groundwater and organic-clay layers. The processes provided in the above references dealt fundamentally with the formation of the site, and were gradually disregarded, as none of these theories were considered capable of explaining the fact that the organic-clay layers and groundwater were so highly salinized. This was due to the fact that no previous chemical analysis was carried out by previous researchers on the organic-clay layers. Where previous spring-water analysis was available (Rindl, 1915, 1916, 1928; Fourie, 1970; Mazor and Verhagen, 1983, Appendix I), these were only been briefly mentioned by researchers such as Brink (1987) and Grobler and Loock (1988a), or ignored. Further to this, groundwater has rarely been mentioned in the literature (Brink, 1987), and it appears that the assumption has been made that all subterranean water at Florisbad originated from the springs.

A further literature review was focussed on specific chemical and physical processes which could possibly explain and support the results of a higher salinization in the groundwater and organic-clay layers (Hillel, 1971; Taylor and Ashcroft, 1972; Tóth, 1972; Blatt *et al*., 1972; Dykyjová, 1978; Tildon and Kadlec, 1979; Lakshman, 1979; Larcher, 1983;Wetzel, 1983; Brady, 1984; Bohn *et al*., 1985; Hatano *et al*., 1994; St-Cyr *et al*., 1997). From the above references, it was concluded that the process of salinization was the only process that could account for the high ion concentrations of the groundwater and organic-clay layers.

Processes and conditions which may result in the accumulation of salts in a semi-arid environment and therefore influence salinization include low rainfall and semi-arid conditions; strong winds with associated aeolian sand deposition; adsorption of ions by clay and organic material; evaporation, capillarity and matrix suction associated with high temperatures; and the accumulation of minerals through drainage impediment. The organic-clay layers at Florisbad also specifically accommodated two major components, namely, the organic matter and the clay. Owing to the many factors involved in the salinization process, and the complexity of the mechanisms involved, the process was examined in some considerable detail. This aspect of the study was also of prime importance in that the results would either support or disprove future hypotheses on the fossilization of faunal remains and formation of the site.

Organic matter, or humus, has a structure, or colloidal organization, of submicroscopic particles, referred to as colloids, which are somewhat larger than ordinary molecules and ions (Taylor and Ashcroft, 1972). The structure is similar to that of clay, with highly charged anions being surrounded by a large number of adsorbed cations (Brady, 1984). However, the negative charge on humus colloids are very much pH dependant with their absorptive capacity decreasing with a decrease in pH (Brady, 1984). On the other hand, under more alkaline conditions, the adsorptive properties of humus far exceed that of clays. With the Florisbad organic-clay layers having a pH range of 8.8 to 4.2 (Appendix II), it is apparent that the degree of salinization of the organic material will vary over the site, and will further vary with the quantities of organic material and clay present. Even before humus is formed through the decomposition of vegetable matter, hydrophytes (aquatic plants living in, or on, water), and in this instance particularly halophytes (plants capable of living in salt impregnated soils) and helophytes (plants growing in water saturated soil), have the ability to remove large quantities of salts from both the soil and water, as well as the ability to accumulate them at far greater concentrations than those in external solution (Larcher, 1983).

It is therefore apparent that the accumulated minerals absorbed and stored by halophytes and helophytes would contribute significantly to the salinization of the organic-clay layers when decomposition took place. These processes are not only applicable to halophytes and helophytes, as even micro-organisms such as phytoplankton and periphyton have the ability to absorb and accumulate salts (St-Cyr, 1997; Wetzel, 1983). This can be illustrated by the use of hydrophytes and halophytes in eutrophication control and bioaccumulation: examples are: the treatment of eutrophication caused by fertilizers and phosphorus-based insecticides in natural lake waters (Tóth, 1972); effluent wastewater treatment (Lakshman, 1979, Tildon and Kadlec, 1979); swine wastewater treatment (Hunt *et al*., 1993); dairy wastewater treatment (Davis *et al*., 1992); pulp wastewater treatment (Hatano *et al*., 1994); and mine wastewater treatment (Noller *et al*., 1994).

Clay particles are often negatively charged and capable of attracting positively charged ions to each colloid, therefore, having the ability to adsorb on their outer and inner surfaces large quantities of ions supplied by percolating water (Blatt et al., 1972; Brady, 1984). In many instances, salinization in clay beds is capable of concentrating minerals to such an extent, for example, through percolating water, that economic deposits of clay minerals such as gypsum, magnesite, halite, and mirabilite may form (Hillel, 1971; Blatt *et al*., 1972; Brady 1984). On the other hand, with the relative weakness of the structural unit bonds in clay, this may lead to the inter-layer ions being easily removed by percolating waters during diagenesis and weathering (Blatt *et al.*, 1972). This further supports the hypothesis that recharge water flushes elements from the organic-clay layers into the groundwater, rather than rainfall having a diluting effect (Appendix II). Montmorillonite clay found in the soil at Florisbad (Appendix II), characteristically absorbs water between successive layers, causing it to expand (Taylor and Ashcroft, 1972), and thus has a large specific surface area due to its lattice expansion and exposure of internal surfaces (Hillel, 1971). The potential and ability for both the organic matter and clay to attract, adsorb, concentrate and store ions in considerable excess to the groundwater, further supporting the choice of the salinization process.

Now that salinization processes had been resolved, the question arose as to the origin of the elements. Examination of the literature revealed that there were a number of possibilities for the origin of ions in the spring-, groundwater, clay and organic matter. Kent (1949) suggested that the high Na, Cl content of the spring water was due to the water having percolated through the underlying Dwyka Group rocks, as opposed to the Ecca Group rocks. Butzer (1988) on the other hand suggested that, due to the steady release of methane gas combined with the Na, Cl content of the water, the chemical had been derived from saline and carbonaceous facies of the Ecca Group rocks. Dolerite, which contains plagioclase, feldspar, and ferromagnesian silicate minerals may have contributed ions to the spring-water in the form of Fe, Si, Mg and Ca, but due to the low levels of these minerals in the spring water, this was not thought to be significant (Appendix I).

The influence of aeolian deposition from Soutpan on the salinization at Florisbad during average raingall periods can be gauged from water quality results obtained during the extremely high rainfall in February/March 1988. Soutpan was completely full for the first time in living memory, dissolving the salt deposits on the pan floor to produce very diluted standing pan water with an Na, Cl content of 9 135 mg/*l* (Douglas, 1992). This equated to the Na, Cl content of the pan water being 30.96% that of sea water, where the Na, Cl content of sea water was taken at 29 500 mg/*l* (Douglas, 1992). Seaman *et al*. (1991) recorded an Na, Cl level of 188 000 mg/*l*, with a TDS of 197 295 mg/*l* for Soutpan water, which may suggest that this sample was taken close to, or at, an evaporative salt dam. In this respect the result was converse, with the Na, Cl content of sea water being only 15.69% that of the pan water. This illustrates the large potentially large reserve of salts that are available from the pan floor sands for aeolian deposition by the north-west prevailing winds at Florisbad.

Salinization will also occur at the contact zone between the spring-water and groundwater, where there will be an exchange of ions through the process of miscible-displacement, through the intrinsic movement of molecules from areas of high to low concentration (Taylor and Ashcroft, 1972). Hydrodynamic dispersion resulting from the continually changing direction of the spring- and ground water through collision with pore walls will also either increase or decrease salinization (Taylor and Ashcroft, 1972). It is thought that the largest contribution of ions came from factors related to the environment of the area, particularly during semi-arid phases. In semi-arid environments, metallic cations do not to leach from the soil, and tend to dominate when pH values exceed 7.0. (Brady, 1984).

Figure 29 summarizes the processes and stages involved in the movement of ions and the salinization and fossilization of the Old Collection at Florisbad. The question of groundwater ion concentrations being higher during wet periods and lower during average rainfall periods, is explained by the organic-clay layers adsorbing ions during periods of low recharge, and the ions being flushed from these layers by the groundwater during periods of high recharge (Appendix II).

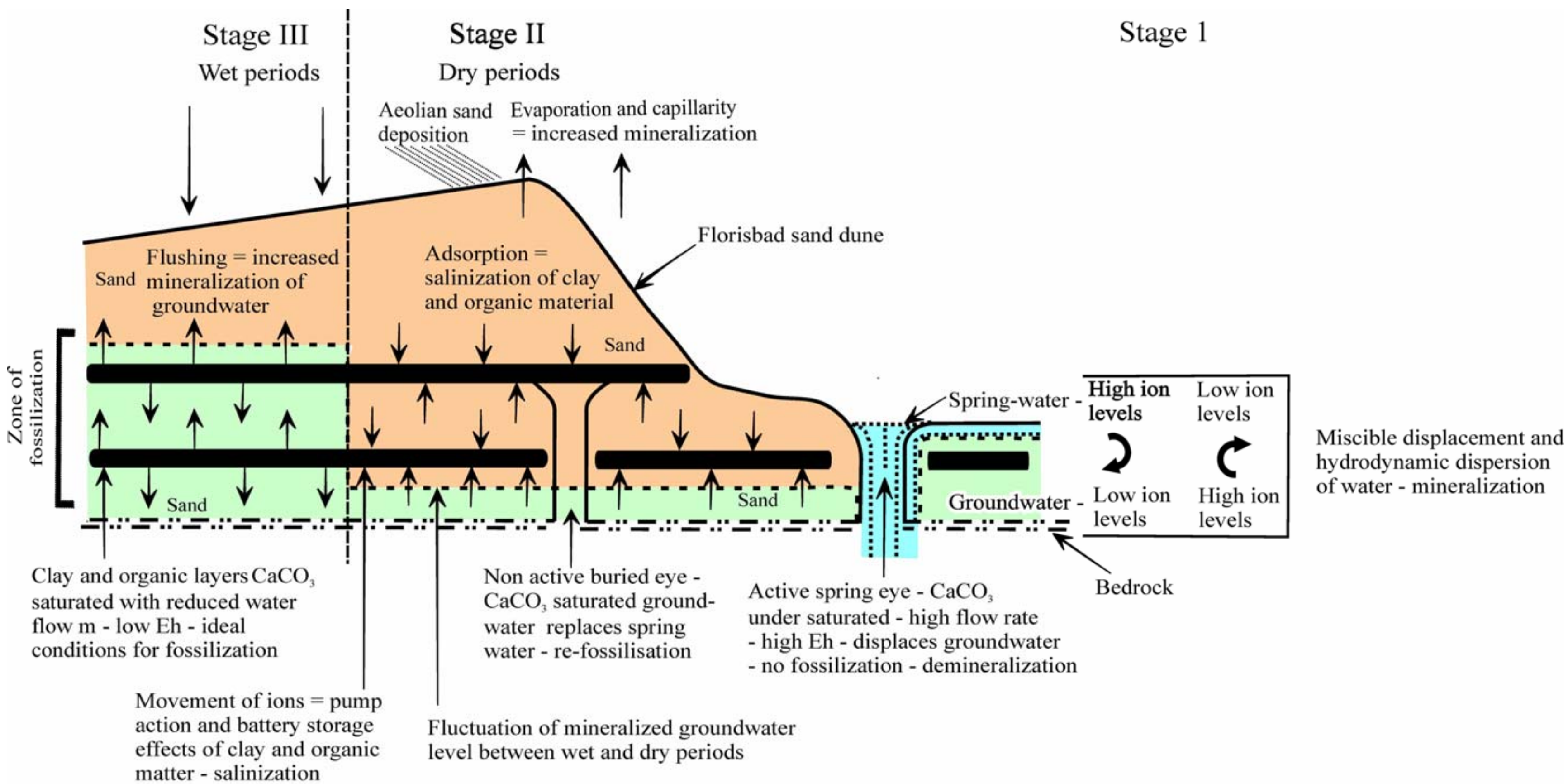


Figure 29. Schematic plan of the Florisbad spring site showing the movement of ions and the stages in the mineralization of the groundwater and the saliniztion of the organic-clay layers, as well as other factors contributing to the fossilization of the Old Collection (after Appendix 3).

In support of the above statements, Table 13 and Table 14 clearly shows the higher concentrations in TDS between the spring- groundwater, and organic–peat layers. Notable is the unexplained west to east increase in the TDS of both the groundwater and Peat II organic–clay layer. As there is a predominance of spring eyes in the area of the swimming pools, it is suspected that the spring water could be having a diluting effect on the TDS of groundwater in the area of the pits. This influence would appear to decrease from west to east, which in turn would affect the ion concentrations of organic-clay layer, Peat II.

5.6.4 A New Perspective on the Fossilization of Faunal Remains

It has now been established that the salinization of the organic-clay layers and groundwater was considerably higher than that of the spring-water, and that processes were in place to produce this higher salinization. The question then arose as to how these results could be applied to the physical environment at Florisbad and the fossilization of faunal remains. Because the large majority of fossils in the Old Collection have been recovered from what has been referred to as areas of spring activity (Brink, 1987), Brink (1987) stated, and it has generally been accepted, that the saline spring-water was responsible for fossilization. Brink (1987) also stated that, since the spring water was carbonate rich, the evidence of calcium carbonate deposition in the bones further suggested that they must have been in contact with the spring water for some time. It was further noted by Brink (1987) that, because fossilized remains of the Old Collection were found in areas of spring activity, post-depositional mechanical and chemical weathering clearly showed that the remains had become fossilized in a spring context. From a chemical perspective, Brink (1987) reported that the spring water has caused the remains of the Old Collection to be well preserved in a characteristic way.

Despite Brink's (1987) strong defence of fossilization in a spring context, he also admitted to some doubts in that, if the presence of calcite was an indication of contact with the spring-water in the original sedimentary environment, then the relatively low incidence of calcite concretions might contradict the assumption that the Old Collection was entirely derived from spring deposits. It was noted that only 48.2% of Bovidae and Hippopotamus bones showed signs of calcium carbonate deposition (Brink, 1987).

Further to this it was also suggested by Brink (1987) that acid groundwater mobilized carbonate cement from underlying geological strata, which may have become re-deposited, despite being of an alkaline nature. Due to already existing water analyses indicating the low Ca level of the spring-water, it is somewhat surprising that the spring-water had been credited with the fossilization of the faunal remains over the years.

The first spring-water analyses by Rindl (1915) clearly showed that the calcium level of the spring-water was very low at 93.42 mg/*l*. Subsequent water analysis by Fourie (1970), Mazor and Verhagen (1983) and Douglas (1992), all confirmed this level of calcium at <100 mg/*l*. Slight variations could be expected through technological advances in water analysis. Grobler and Loock (1988a) also mentioned the high sodium content associated with the low calcium content; if these authors are correct in their assumption that the low calcium content is due to cation exchange as the water moves through the underlying Ecca shales, then there is a strong probability that the spring-water has historically maintained a low calcium content. This would then tend to lend further credibility to the hypothesis put forward in this thesis that the spring-water could never have been responsible for fossilization. Dreyer and Lyle (1931) appear to have summed up the situation by stating that animal remains found at the base of such pipes (spring vents) seem to have been accumulated and concentrated by the sorting action of the spring water, with coarser materials finding the lower levels, and finer material filling the upper ends of the vents, but with no mention of any possible fossilization. Detailed chemistry of the spring-water, groundwater, and organic-clay layers is given in Appendices I and II, and summarized in Tables 12, 13 and 14.

Fourie (1970) found a lack of carbonate in the sands of the eyes, as well in the lower consolidated green sediment, although some evidence of calcite cementation was evident in the basal zone of Peat I. Butzer (1988) found no evidence of calcium carbonate in his 1982 explorations, noting that the spring water was under saturated in both calcium and bicarbonate. This led Butzer (1988) to conclude that any such carbonate enrichment must have originated from either direct, or indirect, external sources, other than the spring water. When the spring-water Ca concentration is compared to the Ca concentrations of the ground water, 82–400 mg/*l*, Peat II, 2 008–5 000 mg/*l*, and Peat IV, 6 720–7 360 mg/*l*, it is evident which environment is more suited to fossilization processes. Therefore, even early water analysis by Rindl (1915) with a value of 93.4 mg/*l* Ca, indicated that there was a primary under saturation of Ca in the spring-water impeding any fossilization processes.

There are a number of other important environmentally related factors, which are required individually, or as a combination of factors, in order to contribute to the effective precipitation of $CaCO_3$ and other authigenic minerals for fossilization. These include pH (alkalinity/acidity), Eh (oxidizing potential), P_{CO2} (partial pressure of carbon dioxide) P_{O2} (partial pressure of oxygen), DOC (dissolved organic carbon/hydrophobic acid), $CaCO_3$ saturation, the decomposition of aquatic plants including phytoplankton and periphyton, phosphates, biomineralization, capillarity (Appendix III). Krauskopf (1967) mentions that the solubility of $CaCO_3$ calcium carbonate is controlled by the pH of the environment, changes in temperature and pressure, as well as organic matter activity and decay. Freeze and Cherry (1979) stated that the solubility of carbonate minerals is largely dependent on the partial pressure of carbon dioxide (P_{CO2}), giving the range for the solubility of calcite and dolomite for natural groundwater's as P_{CO2} 10^{-3} bar and 10^{-1} bar. The solubility of calcite in water at 25° C, pH 7, 1 bar total pressure, and a P_{CO2} of 10^{-3} bar is 100 mg/*l*, while the solubility at a P_{CO2} of 10^{-1} bar is 500 mg/*l* (Freeze and Cherry, 1979). Most of the above factors can be related to the clay, organic matter and groundwater environment at Florisbad in providing a suitable fossilization process, while few of the above factors can be related, or applied, to the spring-water, or its immediate environment. The absence of factors such as the under-saturation of $CaCO_3$, Eh, and a restricted water flow in a spring context, may well be responsible for the de-fossilization of material (Appendix III).

In considering that a suitable environment had been established within the groundwater and organic-clay layers for fossilization, it was now important to ascertain why fossilization should be precluded from the effects of the spring water. It was concluded that there were considerably more reasons for precluding fossilization in a spring context than reasons for initiating fossilization. This conclusion was significant in that it would also indicate that the spring-water, which had developed over thousands of years, would never historically have been in a position to initiate fossilization. It was further concluded that the only way in which fossilization could have been initiated in a spring context would have been if the spring-water had have been supersaturated in $CaCO_3$, had an alkaline pH, with a considerably reduced flow rate. Although the spring-water has an alkaline pH, this is irrelevant on its own An important factor governing the presence of carbonates in natural waters is the presence of organic matter within a low Eh environment with a restricted water circulation (Blatt *et al*., 1972). A low Eh environment with a restricted water flow is necessary for the preservation of organic matter in order to prevent the complete oxidation of organic compounds to CO_2 and H_2O (Blatt *et al*., 1972).

An integral part of organic matter preservation is the formation of DOC, which is used as a food source by carbonate precipitating organisms, Conversely, DOC may act as a kinetic growth inhibitor, inhibiting calcite crystal growth (Hoch *et al*., 1998). This has been established in the Everglades, where the kinetic inhibition of DOC prevents any calcite precipitation in calcite supersaturated water (Hoch *et al*., 1998). A characteristic of the Florisbad spring vents are their white quartz sands (Brink 1987), which appear as such because all clay and organic material has been washed out of the vents by the spring flow. This was confirmed by spring eye sand analysis, which resulted in 0 % clay and organic matter content (Appendix II). Another aspect of carbonate precipitation which would have been negatively affected by the spring flow is Biomineralization. This is the process by which carbonate precipitating organisms sequester CO_2 into a solid carbonate mineral phase as a result of biomass degradation, also referred too as microbial carbonate precipitation (MCP) (Roh *et al*., 2001; Hammer and Verstraete, 2002). this processes could not have taken place in the spring vents in the absence of organic matter, and it is also questionable as to whether carbonate sequestrating organisms could survive in the flow from the eyes.

An important constituent in the formation of authigenic carbonate apatite is the presence of phosphates. Phosphates in the spring-water were very low, recording 1.93 mg/*l* in 1999 and 0 mg/l in 1988, while the groundwater registered between 0.010 and 0.100 mg/*l* (Appendix I). Water analysis by previous researches indicated no phosphates in the spring-water, further indicating that fossilization could not have taken place in the spring vents. The organic Peat II layers showed a large increase in phosphates to between 17 and 34 mg/*l* (Appendix II). Phosphates in the organic-clay layers peat were derived from three main sources.

- Dissolved oxygen in the water would have reacted with the organic material to release CO_2, which in turn would have lowered the pH and liberated phosphates (Karkanas *et al*., 2000).

- Carcass remains, which the Old Collection indicates were abundant, would have released phosphates before and after burial, resulting in a build up of phosphates within the sediments for fossilization. Brink (1987) and Kuman and Clarke (1986) both noted that fossilized bones showed signs of pre-burial weathering such as sun cracking, and horn cores riddled with grooves made by the horn moth Family Tineidae.

- Karkanas *et al*., 2000 noted that phosphates in caves were largely due to the oxidation of bird and bat guano, so the decomposition of waste material from the vast herds of game coming to drink at the site would have further released phosphates into the sediments during oxidation. See Chapter 6 for further discussion on this point.

The recorded pitting and hollowing, and the low incidence of calcium carbonate in, and on, *Hippopotamus amphius* and bovid bones (Brink, 1987), strongly suggests demineralisation. This would occur through the under saturation of calcium carbonate and phosphates in the spring-water, the dissolution of calcite, and abrasion weathering through spring-flow action. Although the incidence of pitting and hollowing is reportedly low, Brink, (1987) noted that many bones, which showed no evidence of calcite deposition on their outer surface, did have calcite development in the internal cavities. Brink (1987) considered pitting to have occurred through solution while the bones were still under-mineralised, and that pitting preceded calcite deposition. On the other hand, Brink (1987) believed that calcite deposition preceded hollowing, and that hollowing was the result of chemical solution, or weathering due to mechanical erosion through water action. Therefore, Brink (1987) seems to suggest that there was both calcite deposition and demineralisation within the spring vents. Dissolution is also expected to increase in sandy soils with flowing water (Hedges and Millard, 1995), and the relatively large surface area of bone apatite would account for its correspondingly high rate of dissolution in natural waters (Trueman and Tuross 2002).

The salinization and fossilization processes described here may have major implications at other archaeological sites in determining the fossilization of faunal remains, and history of the site. Butzer (1974) recorded eight distinct layers at the Cornelia fossil beds in the North-eastern Free State, of which six comprised clay and/or black organic soil. Brink and Rossouw (2000) recorded a massive yellow clay horizon beneath the Unit 1 described by Butzer (1974) both of which contained fossil remains and a salinization process similar to that at Florisbad cannot be ignored. It was noted from the literature that no other fossil sites in South Africa appear to have been examined in detail in order to determine the fossilization processes involved. It was suggested by Brouwer (1967) that many detailed palaeontological descriptions make no mention at all of the conditions under which fossilisation has taken place. Brink (1987) stated that, to fully understand a fossil fauna, it is important that the palaeoenvironment from which the fossil remains were extracted is better known.

5.6.5 A New Perspective on the Formation of the Florisbad Spring Site

There were two primary reasons for revisiting theories on the formation of the Florisbad spring site.

Firstly, on reviewing the literature, it became apparent that none of the previous theories on the formation of the spring site could explain the stratigraphy of the sediments in light of the salinization process. Three basic developmental theories have been put forward in the literature for the formation of the Florisbad spring site:

- The site developed from sand output from the springs, which was most probably derived from the underlying bedrock (Dreyer, 1938a; Brink, 1987; Butzer, 1988; Van Zinderen Bakker, 1989).
- Deposition at the site was directly related, or influenced by, water levels in the adjacent palaeolake (Soutpan), or the flooding of the palaeolake in four low water level phase and three high level water phases. (Visser and Joubert, 1991; Joubert and Visser, 1991; Henderson 2001a).
- Deposition was almost entirely derived from aeolian sources (Appendix I; Appendix II; Appendix III; Grobler and Loock, 1988a, 1988b; Van Zinderen Bakker, 1989; Kuman *et al*., 1999).

Some researchers have suggested that a larger body of water might have existed at Florisbad in the past (Brink, 1987; Kuman *et al*., 1999), with little detail, or evidence, being given regarding actual developmental processes of this larger water body.

Secondly, a number of unexplained, and debatable points were found in the literature to which it was hoped that answers could be provided. These included:

- The lack of upper red-brown sand units on the eastern side of the “mound” appeared to have either not been deposited, or had been eroded by spring discharge (Rubidge and Brink, 1985; Brink, 1987).

- A seven metres thick layer of clay deposits in the modified vlei area (Butzer, 1984), which could not have been deposited by the pre-existing weak ephemeral drainage line.
- Older aeolian deposits, evident over the rest of the site, are absent from the vlei area, while more recent aeolian deposits were present (Fourie, 1970). This represents an anomaly in relation to the area in general, where the erosion of the older aeolian sands along the vlei being seen as an uncharacteristic erosional character of the weak ephemeral drainage line.
- The extent of the dogleg in the ephemeral drainage line from the inter-dune valley, cutting across the natural drainage at almost right angles from NW to NNE, and back to NW. This could also be seen as being contrary to any erosional forces that could have been exerted by the weak ephemeral drainage line.
- The disproportionate width and depth of the vlei area in relation to the ephemeral drainage line.
- The erosion of the east bank of the vlei area.
- The pinching out of the Peat IV layer on the western wall of the excavation area.

Rubidge and Brink (1985) stated that the lithostratigraphy and depositional history of the Florisbad sediments were still in their initial stages of investigation, and that several models could be proposed. Rubidge and Brink (1985) also noted that, in their opinion, because the deposits were lithologically variable, they were the product of an unusual depositional environment. Kuman and Clarke (1986) noted that the sand, silts, and organic deposits were poorly understood and controversial. The above questions and statements therefore opened the way for further interpretations on the formation of the site. Unless otherwise referenced, all references are from Appendix IV, or are updated versions of the hypothesis put forward in Appendix IV.

It is hypothesized that the genesis of the Florisbad site occurred in a slightly undulating topographical environment, with the spring site evolving prior to the formation of the panveld. At the time a spring pan had probably already formed, but was covered by the migrating sand dunes, referred to as the south-east dune belt, resulting in a fossil pan being buried beneath the sand dunes. As the dune belt continued migrating in a south-easterly direction, so the fossil pan became uncovered and the springs again became active on the surface.

Figure 26, a simplified schematic cross-section of the sediments, has been replicated for convenience purposes from Figure 4, to be read in conjunction with Figures 30, 31 and 32. Figure 30 is the legend for the interpretation of the five developmental stages of the formation of the Florisbad spring site reflected as a plans in Figure 31, and as a cross-section in Figure 32.

Stage I: Growth of the modern spring pan would have initially developed along the form of the fossil pan, from when it would have developed its own character. This character would have been based on factors such as the large number of animals coming to drink at the pan, deflation during dry periods, aeolian deposition, growth and decay of aquatic vegetation, and the contribution of waste from the herds of animals. In the initial developmental stages it is suggested that the inter-dune ephemeral drainage line drained directly into Soutpan, and that it was later captured by the spring pan. This ephemeral drainage would also have contributed water to the pan during wet periods. Around about this time a sand dune started to develop on the south-eastern bank of Soutpan, derived from the deflation of the Soutpan floor and aeolian deposition from the surrounding area and further a field from the west and north-west.

Much has been made throughout this thesis regarding the influence of the vast herds of game roaming the Free State plains in the past, and the contribution they may have made to the erosional and chemical processes at the spring site. The contribution to erosion by the herds of game coming to drink at the spring site, in search of minerals such as salt, resulted in the breaking down of sand (weathering and deflation) and aquatic vegetation (decomposition) by trampling. These vast herds of game also contributed to the mineralization of the site through their waste contribution (decomposition, phosphates), and the contribution of carcases, either from carnivore killing, or natural causes (phosphates). But just how significant were these herds of game?

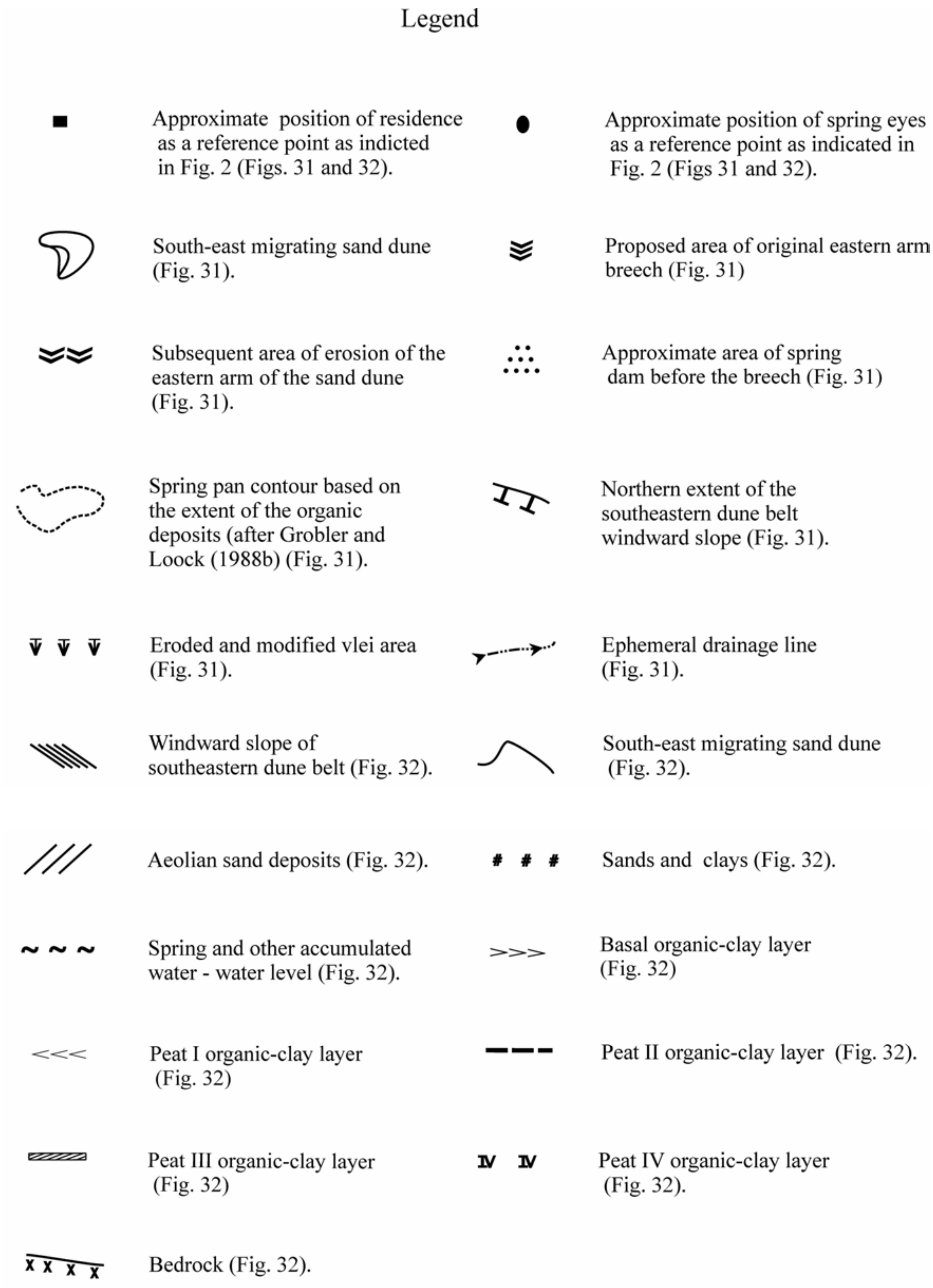


Figure 30. Legend for Figs 31 and 32.

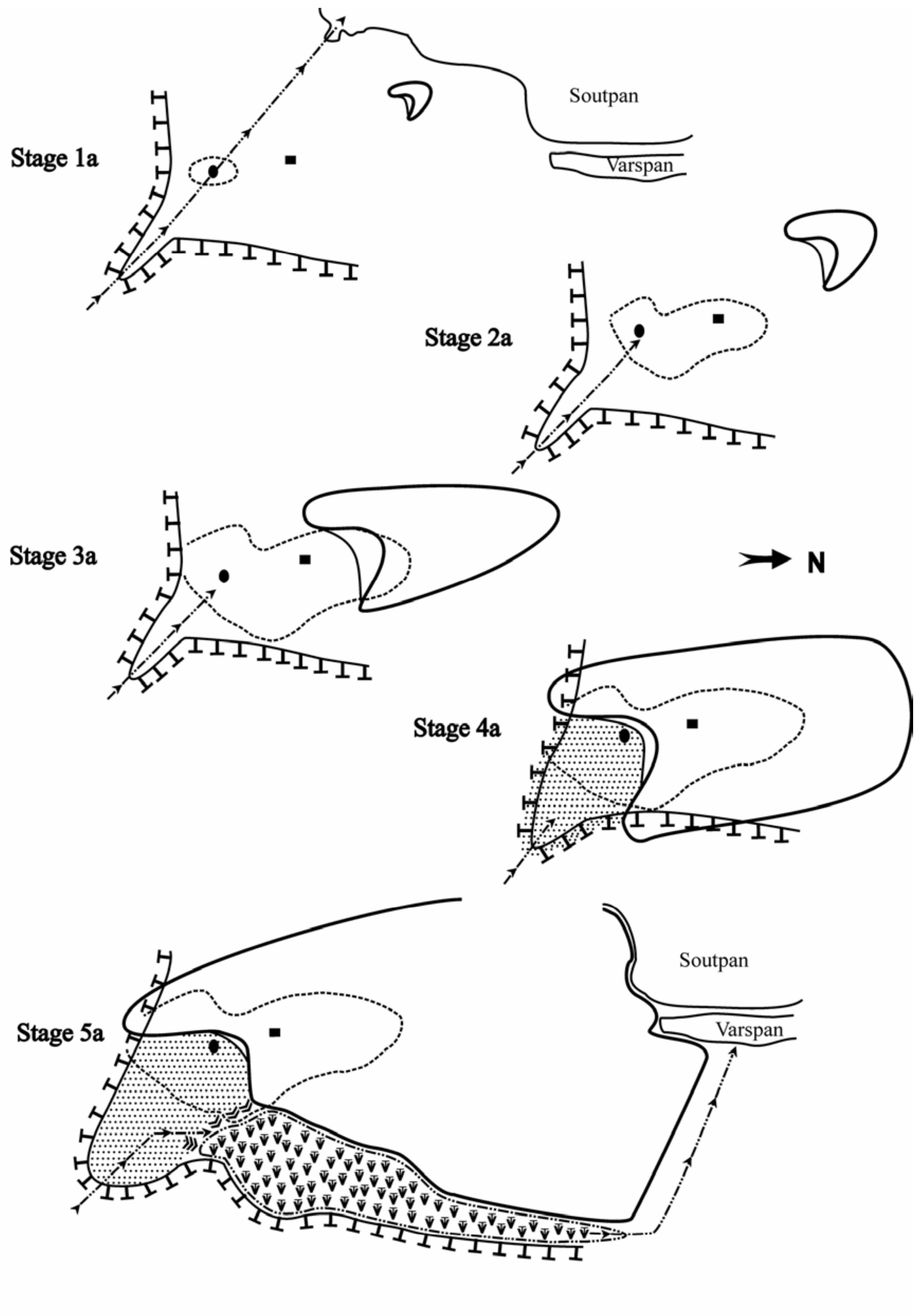


Figure 31. Schematic plan of the proposed developmental stages in the formation of the Florisbad spring site (after Appendix 4).

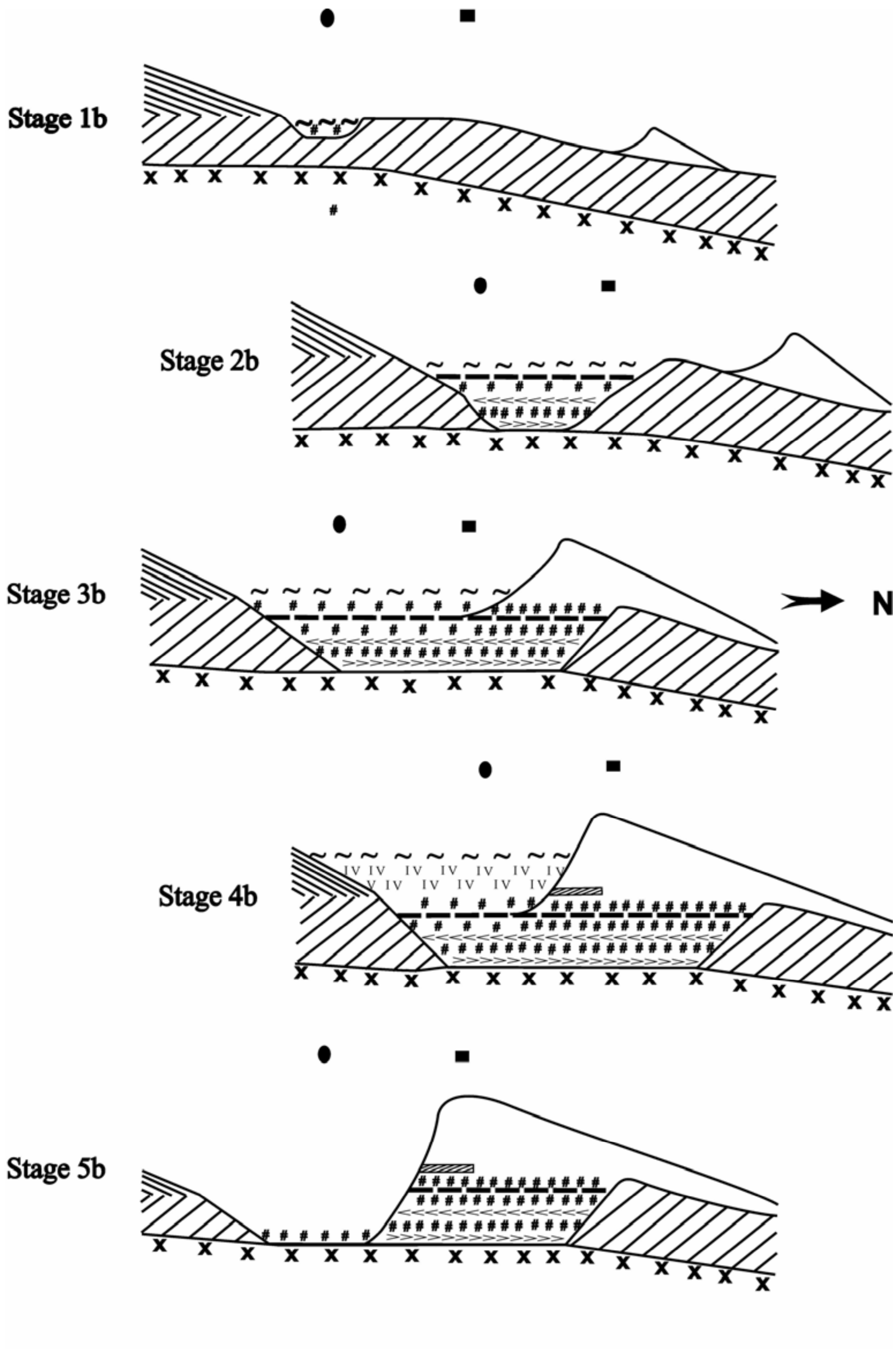


Figure 32. Schematic profile of the proposed developmental stages in the formation of the Florisbad spring site (after Appendix 4).

The size of the herds inhabiting these grass planes can be judged by the results of a hunt carried out by Prince Albert on the 24 August 1860. Some 1000 Baralong tribesman rounded up between 20 000 and 30 000 head of game at Bain's Vlei, 50 km south of Florisbad, and within an hour, the Prince and his hunting party managed to slaughter an estimated 5 000 animals (Anon [5], 1860; Anon [6], 1977). At the time, it was reported that the number of animals rounded up for the hunt was disappointing, and that the original number of animals driven onto the plain for the hunt should have been far greater (Anon [5]; 1860).

Stage II: The spring pan had grown considerably in size, and was now supporting aquatic dependant species such as hippopotamus (*Hippopotamus amphibious*), lechwe (*Kobus leche*), clawless otter (*Aonyx capensis*), and water mongoose (*Atilax paludinosus*) (Brink, 1987: Henderson, 1996; Henderson, 2001a). During wet periods, water was still being supplied by the ephemeral drainage. The now referred to Florisbad dune continued to grew in size from the deflation of Soutpan and other aeolian deposits, and started migrating in a south easterly direction towards the spring site.

Stage III: The Florisbad dune now started migrating across the spring pan and blanketing it with sand, thus halting any further northward expansion of the pan. The arms of the sand dune now also begun to encompass the spring pan from the sides, and thus a damming effect was began, with further expansion of the pan being restricted to areas east, west and south of the advancing dune. Fresh water was still being supplied by the ephemeral drainage during wet periods. At this time is it is thought that the distribution of organic matter, as indicated by Grobler and Loock (1988b) (Fig 9), was far more extensive to the east and west, but particularly so to the south. It is suggested that the Basal Peat, Peat I and Peat II were formed during the developmental stage of the spring pan

Stage IV: The arms of the sand dune now met and overrode the windward base of the south-eastern dune belt, cutting off any possible outlet for the spring pan water. The spring pan, which had always been seen as a drainage-impeded pan by the author, now became a dam, completely enclosed by sand dunes. As the dune migrated it continued to cover the spring pan and ride further up the windward slope of the south-eastern dune belt, allowing for a steady increase in the depth of the water through spring flow and the ephemeral drainage. Considering the thickness, friability, and relatively low degree of decomposition of the Peat IV layer (2.8 m wide, and to within 650 mm of the current surface) on the west wall of the excavations, the rise in the water level must have been relatively slow in order for these organic beds to have developed and attain this height (see Section 3.2 for aspects of palaeoclimate supporting this). The low degree of decomposition in Peat IV would also suggest that these reed beds had been, at various stages of their formation, returned to a dry state during their growth, which inhibited any further decomposition.

It is proposed that Peat III and Peat IV developed during this damming-up period, with Peat III possibly being isolated remnants of Peat IV which were left behind after Stage 5. The proposed area of the Florisbad dam is given in Figure 31, 4a and 5a.

Stage V: With the increase in sediments and height of the water, the pressure against the arms of the sand dune increased. The eastern arm of the sand dune was then breached, taking most of the eastern arm with it, with the contents (sediments and water) of the spring dam evacuating the site in a flash flood. This flash flood gouged out a new drainage line from the soft dune sands to form the vlei, with the force of the flash flood severely modifying the new drainage line into a deep, wide, flood plain. The contents of the spring dam were deposited along the length of the vlei, giving it its current form. As the force and volume of water decreased along the newly formed drainage line, the vlei began to narrow, and the flow took an almost right angle turn, resuming a natural north-westerly flow into Varspan. The large flat amphitheatre like area to the east and south-east of the springs is thought to be a part of the original base of the spring pan where it dammed up against the windward slope of the south-eastern dune belt.

Three possible scenarios are envisaged, all of which could have led to the final breach. One was that the water rose to a height where it began to flow over the top of the eastern arm of the sand dune and eroding the wall until it breached. A second alternative was that the weakest point in dune wall was the angle of repose of the eastern arm up against the south-eastern dune belt, which, as the height of the water increased, gave way under the pressure. The third alternative is that the herds of game coming to drink at the site, eroded the eastern arm of the dune to a stage where it began to overflow, and then as the arm further eroded, breached. Whichever of the fore mentioned scenarios was the cause of the breach, considerable erosion would have taken place on either side of the breach.

Perhaps one of the most compelling pieces of evidence supporting this formation hypothesis is the high salinization of the organic clay in the vlei. Table 15 gives a comparison of the degree of salinization between the Peat II samples from the north wall of the excavations (X1), excavation pit 2 (X2), excavation pit 4 (X3) (Figure 23), and the organic-clay from the vlei area, approximately 400 metres from the residence. This high salinization of the organic-clay from the vlei further supports the hypothesis that the deposits in the vlei originated from the dam site, and not Soutpan. Further support is provided by the total ion count for the vlei clay being close to the average (14 897 mg/*l*) for the three Peat II samples in Table 15, strongly suggesting that a homogeneous mixing of the sediments from the dam site had occurred on entering the vlei area. It is regrettable that samples could not be obtained from the Peat I layer as it is presumed that the TDS of this layer would be even higher than the Peat II layer.

In conclusion it can be stated that this research provides an alternative hypothesis on the formation of the Florisbad spring site, which accommodates all aspects of sedimentation and fossilization, as well as seriously questioning previous theories. Further to this, this hypothesis explains the eroding, or non-deposition, of the red-brown units on the eastern side of the site, the 7 m of clay deposits along the vlei area, the disproportionate size of the vlei relative to the ephemeral drainage line, and the changes in direction of the ephemeral drainage flow. Further evidence in support of this hypothesis is the deep cutting into the base of the windward side of the south-east dune belt on the eastern bank of the vlei, which would accommodate the directional flow hypothesis of a flash flood from the eastern wall of the dam site. The pinching out of the Peat IV layer on the western side of the site would be explained by the greatest erosion having taking place on the opposite, or eastern side of the dam site. This is because the western portion was furthest from these erosional forces, and the Peat IV layer remained largely intact on the western face. In Figure 32, 4b, it is not proposed that Peat IV completely filled the dam,

Table 15. A comparison between the salinization of the various Peat II samples and the organic-clay from the vlei area showing the high salinization of the vlei area organic-clay

	Sample X1 Peat II	Sample X2 Peat II	Sample X3 Peat II	Vlei Organic-clay
PH	8.30	8.00	6.65	6.75
Resistance (Ohm)	26	38	19	57
NaCl (mg/*l*)	2 427	11 967	19 193	8 499
IONS (mg/*l*)				
P	22	26	34	66
Cl	1 567	7 727	12 393	5 499
Ca	2 008	2 768	5 000	4 540
Mg	23	25	70	48
K	129	640	352	638
Na	860	4 240	6 800	3 000
Zn	1.2	2.4	6.0	5.5
Total ions	4 610.2	15 428.4	24 655.0	13 796.5

but reflects the Peat IV layer against the western wall of the dam. It was also proposed that much of the leeward side of the dune face slumped into the spring dam as it drained, having previously been supported by the roots of aquatic vegetation and the pressure of the water.

5.7 MEDICINAL PROPERTIES

Because Florisbad built up an extensive reputation for its healing properties over the years, and because Floris Venter became known as the "healer", a brief hypothesis on this aspect of the springs is provided. No scientific studies have been carried out on any possible medicinal properties of the Florisbad spring-waters, but as far back as the early 1900s, the Florisbad spring was included in papers and articles on the medicinal springs of South Africa (Rindl, 1915, 1916; Kent, 1948, 1964; 1971). In some instances the Florisbad spring water was reported to be particularly effective in the treatment of sciatica, and muscular and articular rheumatism (Rindl, 1916)

An examination of the water analysis tables in the appendices reveals that, with the exception of two elements, there appear to be no other elements indicating that the spring water could be medicinally beneficial. The two elements, about which much has been written and researched over the past 90 years, are strontium (Sr) and boron (B). Strontium had the fourth highest spring-water cation concentration of 2.8 mg/*l* after Na, Ca, and K, and B the fifth highest concentration of 2.6 mg/*l* (Appendix 1). In exploration pit 4, groundwater values for B increased by 288%, over that of the spring-water, to 10.1 mg/*l*, and Sr by 300% to 11.2 mg/*l* (Appendix I). Today Sr and B are credited in relieving the symptoms of inflammatory processes, immune function, calcium metabolism and absorption, osteoarthritis, rheumatoid arthritis, osteoporosis, and ruptured discs (Herbert, 1921; Hall, 1976; Skoryna, 1981; Marie *et al*., 1985: Marie and Holt, 1986; Gaby, 1994; Newham, 1994a, 1994b; Travers *et al*., 1990; Nielsen, 1992: Hunt and Idso, 1999; Henrotin *et al*., 2001; Meunier *et al*., 2002; Meunier *et al*., 2004).

It is postulated that, in line with the large ion concentration increases between the groundwater and organic-clay layers (Appendix II), Sr and B will exhibit a similar increases within the organic clay layers. There are a number of reasons for this supposition. Results from Boon and McIntyre (1968) gave evidence supporting the hypothesis of B incorporation in fine-grained sediments in amounts proportional to the salinity of the depositional environment, which was supported by Brooks and De Wall (1976). Shimp *et al*. (1969) noted that the B content was correlated with the content of clay finer than 2μ, and that this gave the best discrimination between marine and fresh water clays. Mortvedt *et al*. (1972) noted that the amount of B in the soil was directly proportional to the amount of organic matter, while Villumsen and Nielsen (1976) found variations in B being related to the content of montmorillonite in the clay fraction. Frederickson and Reynolds (1960) summarized the situation by stating that the B content in saline waters increased with the salinity of the water, finding that B in the clay mineral fraction of sedimentary rocks was associated with the clay mineral illite.

Large quantities of montmorillonite, illite, and organic matter are recorded from the Florisbad organic-clay layers, adding further support for salinization (Appendix II), and making them ideal depositories for Sr and B. The Florisbad organic-clay layers have an NaCl content of up to 65% that of sea water, compared to the 7.5% of the spring-water (Appendix II). It therefore seems feasible that the Sr and B content of the Florisbad clays would be considerably higher than that of the spring-water. Grobler and Loock (1988a) considered that, as the salts Ca^{2+}, Na^{+}, $SO^{2}{}_{4}^{-}$ and Cl^{-}, were derived from the underlying Ecca Group rocks, which were deposited in a marine environment, this may be an explanation for the presence and levels of Sr and B.

Newham (1994a) noted that in areas of the world where B intake was <1-2 mg per day, the incidence of arthritis varied between 20-70%, whereas, where B intake ranged from 3-10 mg per day, the incidence of arthritis varied between 0-10%. Although the effect of percutaneous absorption through the low levels of Sr and B in the spring-water is not known, the effects of percutaneous absorption through the higher organic-clay medium might be considerably higher. Two treatment methods were used at Florisbad. The first was to pack heated mud, taken from the vlei area below the swimming baths (Hendersom, Z. *pers com*. 20/09/2007) between two layers of newsprint, and pack these onto the patient's bodies (Henderson, 1995). The second was to submerge ones self in a mud pool located on the east side of the baths (Henderson, 1995). It is contended that it was contact with the organic-clay mud, which contained elevated levels of Sr and B, which brought relief from the arthritis. Although research has shown that Sr and B have definite anti-inflammatory benefits, the effects of these elements in the Florisbad context still need to be proven.

5.8 SYNOPSIS

In this chapter, as in previous chapters, a logical progression has been followed in dealing with the numerous components, or elements, contributing towards the formation and fossilization of faunal remains at the Florisbad spring site. These include providing a comprehensive background in order to bring together these elements in presenting a holistic picture of the spring site.

For example, in this chapter, the chemistry of the spring- and groundwater was examined in order to make a distinction between the two entities. This then led to an examination of the sedimentation, chemistry, properties, and processes, of the organic-clay layers in order to be able to relate the results to mineralization and salinization. The chemistry of the spring-, groundwater, organic-clay layers were then used to determine where the most likely areas for fossilization were likely to occur. From these results it was concluded that fossilization could not have occurred in a spring-water context, and that fossilization must have occurred in conjunction with the organic clay layers. As no available theories on the formation of the site could accommodate the various hypotheses that had been formulated from these results, an hypothesis was developed which would accommodate all the components. This hypothesis further resulted in a number of previously unanswered questions in the literature, regarding the formation and morphology of the site, being resolved, and new evidence being established in support of these answers. It is therefore concluded that a comprehensive and holistic picture on the formation and fossilization of faunal remains at Florisbad has been achieved.

Problems with the dating of the Florisbad deposits has been emphasised in these chapters. However, published data on the most recent dating of fossils from one of the excavation pits, indicates equal compaction and compression of the sediments over time and depth. This was questioned, in that logically, the deeper sediments should be far more compacted than the shallower sediments, with time and decomposition increasing the compaction. This issue is examined in some detail and discussed in Chapter 6.

Chapter

6

The use of Mathematical Trend Lines in Evaluating ESR and OSL Dating at Florisbad

CHAPTER 6

THE USE OF MATHEMATICAL TREND LINES IN EVALUATING ESR AND OSL DATING AT FLORISBAD

6.1 INTRODUCTION

Florisbad is a significant archaeozoological and archaeological site situated in the Free State Province of South Africa (28° 46'S 26° 04'E). The importance of Florisbad lies in three main areas. Firstly, the discovery of the Florisbad hominid cranium by Prof. T.F. Dreyer in 1932 and the description thereof in 1935 (Dreyer, 1935), brought recognition to Florisbad. Currently the cranium is electron spin resonance (ESR) dated at 259 ±35 ka, (Grün *et al*., 1996). Secondly, a vast collection of faunal fossil remains, and artefacts, representing some 26 species, referred to as the Old Collection, (Brink, 1987), which represents the Florisian-Cornelian faunal boundary with a possible age of *c*. 400 ka (Klein, 1984). Thirdly, a Middle Stone Age Human Occupation horizon, which reflects a temporary butchering site, and which has delivered butchering tools as well as fossils (Brink, 1987; Henderson, 2001a, 2001b; Brink and Henderson, 2001).

While researching various aspects of the Florisbad spring site, such as water quality (Appendix I), salinization (Appendix II), and the formation of the site (Appendix IV), a paper on the most recent dating of the spring sediments by electron spin resonance (ESR) and optically simulated luminescence (OSL) was examined. It became apparent that the ages given for the sediments at descending depths in profile (*a*) Grün *et al*. (1996), did not conform to a logical progressive compaction of the sediments with increasing depth, and increased overburden, over time. That variation in factors such as, climate, spring flow, vegetation growth, aeolian deposition, and decomposition, would all equal–out to produce two depositional zones of almost equal thickness and compaction, over almost equal time frames, at different depths, was considered highly improbable. Due to the high water table, which fluctuates between three and four metres above basement, it was not possible to obtain more detailed compaction data. A major influence on the compaction of the basal sediments would have been the decay and compaction of the organic material, with decomposition also contributing to greater compaction. Before the build up of aeolian sand deposits, the basal organic layers were in all probability wide porous bodies similar to the present day extremely porous Peat IV organic layer, which extends from just below the surface down to 2.8 m on the western wall of the excavations (Appendix IV). Despite its depth, the Peat IV layer is only covered by a few centimetres of overburden, which was not enough to cause any compaction. The basal sediments would have gone through processes of vegetation build-up during wet periods, being covered and compressed by the mass of aeolian deposits during dry periods, and decomposition during further wet periods. These cyclic events would have compacted and compressed the basal organic layers into the narrow bands they are today. It is therefore evident that the basal sediments have not only been in existence longer due to their position in the sequence, but have also been exposed to a longer and greater period of compaction and compression than the upper sediments.

The effect of compaction and compression is further illustrated in the Unit F horizon where the mass of sediments was so great that the compacted skull and pelvis of *Hippopotamus amphibius* have been ascribed to this process (Henderson, 1996; 2001a). These remains were located approximately 3.0 m from the surface, and are less mineralised than remains from the Old Collection (Brink, 1987). The compaction of bones has not yet been recorded from the Old Collection, which in some instances is nearly twice the depth of the Middle Stone Age (MSA) Occupation Horizon from surface. This would indicate that the Old Collection remains possibly became fossilised before any significant build up of sediments occurred. It is obvious from the excavations on the west wall that there has been an insufficient build-up of aeolian deposits above the porous Peat IV layer, or a prolonged aquatic environment which lasted long enough to decompose this organic layer (Appendix IV). If the compaction of *H. amphibius* bones occurred at a depth of just over 3.0 m, then the compaction of vegetation and sands at more than double this depth must have been far greater.

Grün *et al*. (1996) noted several inconsistencies in the dating and gave possible reasons for these inconsistencies in their paper. After examination of the available data, it was decided to find a method with which to test these results. It was decided that the application of mathematical trend lines would reflect increased, or linear, compaction of sediments with depth, over time. This paper is not an attempt to establish new ages for the Florisbad sediments and faunal remains, but rather a novel hypothetical exercise to examine the factors that may have influenced the most recent ESR and OSL ages, and to test the validity of these ages against the use of mathematical trend lines. The use of mathematical trend lines in evaluating sedimentary compaction and compression with depth does not appear to have been previously examined.

6.1.1 Stratigraphy

The stratigraphy of the Florisbad spring site is important because it holds the key to the history of the site, including aspects such as its formation and the preservation and fossilization of both the Old Collection and the MSA human occupation horizon. For example, it is here contended that the spring water was never responsible for the fossilization of the faunal remains, but that a multifaceted salinization process within the sediments, particularly in the clay and peat layers, ionically enriched the groundwater to the extent that the latter became responsible for fossilization (Appendix III). Further more, it has been contended that the spring water is responsible for the demineralisation of previously fossilized material (Appendix III).

The stratigraphy of the site is complex, poorly understood and controversial, while being the product of an unusual depositional environment, which can be described by several depositional models (Rubidge and Brink, 1985). Hypotheses on the formation of the site include the accumulation of spring sand, vegetation and aeolian deposits (Brink, 1987), the transgression and regression of waters from the nearby palaeolake (Soutpan) in a number of high and low cycles (Visser and Joubert, 1991), and the release of detrial sand from the underlying Ecca shale and dolerite (Dreyer, 1938a; Butzer, 1988). Other researchers have proposed that sand at the site is almost entirely of an aeolian nature (Grobler and Loock, 1988b; Loock and Grobler, 1988; Joubert and Visser, 1991; Van Zinderen Bakker 1989; Kuman *et al*. 1999). It is proposed here that the spring mound is a sand dune which moved in a south-easterly direction from Soutpan over the site, resulting in a damming of the site by the dune, and that the spring pan and sand dune were inextricably interlinked in the later stages of the site's formation (Appendix IV). It is also contended that the spring itself contributed little to the depositional process (Appendix IV). See Chapter 5, Figure 26, for detail on the stratigraphy discussed in this chapter.

Adding to the complexity of the site is the considerable variation in both horizontal and lateral stratigraphy. This is confirmed by the results of an auger-drilling programme where a low degree of correlation was noted even between adjacent boreholes (Rubidge and Brink, 1985). From previous experience, large diameter auger drilling is considered an imprecise method for determining stratigraphic sequences in soft sediments, as opposed to core sampling. This is because the details of narrow layers may become distorted through the action of the auger bit, as well as through the feed rate of the auger bit, which in turn may result in compaction and distortion. Due to this, the accuracy of depth between boreholes may also come into question. Variations in the stratigraphy of the site can be attributed to several factors. These include the non-uniform deposition of aeolian sands and the uniform deposition of sediments in the aquatic environment of the spring pan, with further variations being due to alternating wet and dry periods. The reworking and remixing of sediments through the eruption and migration of spring eyes (Dreyer, 1938a; Brink, 1987) would have also played a significant role in the redistribution of material within the spring deposits and sand dune.

Other potential influencing factors are earthquake-induced liquefication (Visser and Joubert, 1990) and the stratification of the sand dune through its growth and migration, with the further effect of stabilization by plants during wet periods and aeolian deposition during dry periods (Appendix IV). It is suggested that erosion was a major force during the early history of the site when deflation was dominant for much of the time, but as from the time of the formation of the dam, erosion decreased, and was minimal. Investigations by the author suggest that the effects of erosion have generally been relatively minor since the formation of the dam, with the site being in a largely aggrading state (Appendix I). This was largely due to aeolian sand deposition and the sand dune moving over the site (Appendix IV). With the exception of severe erosion when the dammed site drained itself (Appendix IV), other erosion factors may have included the early periods of deflation of the spring pan by wind during dry periods, which in turn may have been offset by a relatively constant spring flow, and some erosion by the large herds of animals coming to drink at the pan (Appendix I; Appendix IV).

It should be born in mind that the degree of deposition and sedimentation was highly variable, being punctuated by high and low varying peaks. Because sediments are deposited on top of one another over time, this presents a continual stratigraphic unit in situ, so there is no way of determining whether non depositional periods (events) actually occurred. It must therefore be assumed that deposition was more or less continuous to varying degrees. Variability would have been due to factors such as rainfall and temperature, which would have in turn related to wet and dry periods, wind, aeolian deposition, and the development of the lower organic layers, their decomposition and subsequent compression and compaction. Compression and compaction would also have been variable, dependant on the aforementioned factors, but they would also have been a varying constant in that they were always a present force, dependant on the accumulation of sediments and water. The lower deposits would always have been under the increasing effect of compression and compaction as the accumulation of the sediments and water increased. It would, at this time, be difficult to determine whether narrower layers of organic matter actually formed as narrow layers, or whether they were much wider layers that had been compacted over time. The latter is thought to have been the case.

6.1.2 Archaeozoological and Archaeological Deposits

Two distinct archaeozoological and archaeological deposits, which are separated from each other both horizontally and vertically, have been excavated at Florisbad. The first is referred to as the Old Collection, and comprises a basal accumulation of fossilized faunal remains in the areas of spring activity, representing a death assemblage resulting from largely carnivore killings. Bones in this assemblage are characterized by evidence of hyena damage, the unbroken state of the bones, and to a much lesser extent, porcupine gnawing, with no indication of cut marks (Brink, 1987; Brink, 1988). The Old Collection, or spring assemblage, is of particular importance in that it represents the type collection of the Florisian Land Mammal Age (Klein 1984). The second collection is the Middle

Stone Age (MSA) occupation horizon which occurs approximately 3.5 metres above basement, where species diversity and numbers are far less than in the basal Old Collection, and represents a butchering site (Brink, 1987; Henderson, 2001a; Brink and Henderson, 2001). Artefact and faunal remains were found in horizontal deposits on a sandy substrate, which appear to have been deposited in an aqueous environment with little disturbance (Meiring, 1956; Henderson, 2001a; Brink and Henderson, 2001; Kuman and Clarke, 1986; Kuman *et al*., 1999). In this assemblage, signature marks on bones indicate slicing, scraping and cutting, as well as bone-breaking, with no signs of carnivore damage (Brink, 1987; Henderson 2001). MSA faunal remains are also very friable in relation to remains from the Old Collection (Brink, 1987).

6.1.3 Previous Dating

Since the early 1950s, numerous attempts have been made to determine the age of the Florisbad sediments by ^{14}C radiocarbon dating methods (Libby, 1954; Broecker, 1956; Barendsen, 1957; Beaumont and Vogel, 1972; Beaumont *et al*, 1978; Kuman and Clarke, 1986), as well as by Amino Acid Racemization (Protsch, 1974), uranium series (Clarke, 1985), and thermoluminescence (Joubert and Visser, 1991) methods. Earlier ^{14}C dating of the more recent deposits, from Peat II to the surface, has provided ages for these upper sediments, although the method has also provided considerable variation in the ages for these horizons. ^{14}C dating of the basal deposits and Peat II layer provided ages greater than the limit of the ^{14}C method. Previous ^{14}C dating of the Unit F horizon was >43.7 ka whereas the MSA marker bed, below Unit F, was dated at >47.2 ka (Kuman and Clarke, 1986), with these ages being well below recent ESR and OSL ages This paper is based on ESR and OSL dating results only.

ESR dating on tooth enamel and OSL dating on quartz separates from sediment samples have recently been employed (Grün *et al*., 1996) in an attempt to obtain more accurate dating. ESR results placed the Florisbad hominid tooth at 259 ±35 ka, Peat I from near the base of the formation at 281 ±73 ka, and the sandy MSA Unit F horizon near the middle of the formation at 121 ± 6 ka (Grün *et al*., 1996).

6.2 METHODS

6.2.1 The Application of Trend Lines to the ESR and OSL Ages

The ESR and OSL age estimates used in Figures 33, 34, and 35 were taken from Grün *et al* (1996), and from the following profiles: (*a*) *Age estimates for the third test pit*, and (*c*) *Schematic diagram of the basal part of the sediment sequence at Florisbad with a reconstruction of the spring vent yielding the hominid fossil (not to scale)*. These results have been tabulated in Table 16 where the disjunction of ages and degree of age spreads in (a) are evident. The ages and the depth of the samples were plotted on graphs with the *y* axis representing age, and the *x* axis representing the depth. Linear (*Lin*), logarithmic (*Log*) and exponential (*Exp*) trend lines were then applied to the resultant graphs. It was hypothesised that the trend lines would reflect the varying degrees of compaction and compression over depth and time.

Hypothetical

It was further hypothesised that, over depth and time *Lin* would reflect equal compaction at all levels; that *Log* would reflect greater compaction in the basal levels, with a gradual decrease towards the surface; and that *Exp* would reflect an increasingly greater compaction with depth, with a gradual decrease towards the surface. A best fit to data between the original data (graph) and the trend line was tested against R^2 values, where R^2= 1.0 indicates a perfect fit. Although 3 decimal places may not be justifiable, they are given as an indication of variance where results are very close. Should this hypothesis be valid then the application of trend lines to ESR and OSL dating at Florisbad could be used as a tool for validating and evaluating these ages.

Two approaches were adopted in this study in order to validate ESR and OSL dating against depth and time. In the first approach, the ESR MSA dating was used as a variable with the basal age remaining as a constant, and in the second approach, the basal ESR dating was used as a variable, with the ESR MSA dating remaining as a constant. In both instances, the near-surface ESR dating remained as a constant. The ESR dating used at 1.75 m from surface was obtained from the mean of the two samples taken from (*a*) as 25 ka. However, this age, was considered far too old for this shallow depth, and should

Table 16. A tabular interpretation of age estimates and depths for the third test pit of Grün *et al*. (1996) (illustration (a)) (refer to Pit 2 for this study) and the lower spring sediments, Grün *et al*. (1996) (illustration (c)).

Illustration (a)			Illustration (c)		
Depth Metres	Method	Ages and age spreads in ka	Stratigraphic layer	Method	Ages and age spreads in ka
0					
0.2	^{14}C	1			
0.5	^{14}C	1			
1.3	ESR	15 ±3			
2.2	ESR	93			
			White sand layers	OSL	146 ±15
			White sand layers	OSL	128 ±22
3.4	OSL	143 ±7			
4.0	ESR & OSL	145 ±40			
			Above Peat II -MSA	OSL	133 ±31
			Above Peat II –MSA	ESR	121 ±6
4.2	ESR & OSL	127 ±33 MSA			
4.4	ESR	181 ±44	Olive-green sand	OSL	157 ±21
4.7	ESR	159 ±21			
5.5	ESR	250 ±25	Peat I –OC	OSL	281 ±73
			Florisbad Skull -OC	ESR	259 ±35
6.2	ESR	140 ±10	OC		
6.2	ESR	202 ±28			
6.4	ESR	129 ±9	Brown sand –OC	OSR	279 ±47
7.0	ESR	152 ±8	OC		
7.0	ESR	177 ±5			
7.4	ESR	198 ±46	OC		

MSA = Middle Stone Age Occupation Horizon
OC =Old Collection

possibly be in the range of *c* 8.8 to 11.7 ka, as indicated by the trend lines. This was confirmed by ^{14}C radiocarbon dating taken from Kuman & Clark (1986), who recorded ages for the sand above Peat III (Table 4) as ranging from 4.37 ±.07 ka to 11.7 ±.1 ka, and Peat IV (Table 4) as 3.55 ±.6 ka to 5.53 ±.8 ka. With the MSA horizon in (*a*) having an ESR dating spread of 125 ka, the mean of the ESR dating of 121 ka in (*a*) and (*c*), and the OSL dating of 133 ka (*c*) at the same level, was used. This resulted in an MSA age of 127 ka. Although the oldest age of *c*. 255 ka in (*a*) is situated 2 m above the base, it has been presumed that this sample was repositioned by some physical force and that it represents the basal sediments, based on the oldest age coming from the greatest depth. The mean age of the two samples in (*a*) was taken as representing the basal sediments at 250 ka. Although this is not the oldest age recorded in the paper, the age fits within the ranges for older ages from the spring sequence (*c*) as well as the 259 ± 35 ka of the hominid tooth (Grün *et al*., 1996).

While ages from Grün *et al*., (1996) are related to tooth samples, it is assumed that these tooth samples are directly related to the sediments in which they were found within the profiles. Therefore, in this context, tooth sample and sediment are synonymous, as in basal and lower sediments.

There is no available stratigraphic cross-section for test pit 3, and in any event, because the test pit was located relatively high up on the dune face, a cross-section would not reflect many of the stratigraphic sequences referred to in the paper. This is due to the considerable horizontal variation in the stratigraphy, and may largely reflect the internal structure of the dune. Any cross-section reflecting a specific section of the site would also only reflect that particular sequence and would be of relatively limited application in the context of the site as a whole. Figure 28 provides a simplified cross-section illustrating the major stratigraphic sequences discussed in this paper. The deposits are composed entirely of sands, clay, and peat-like layers, with no rock formations, thus making compaction and compression major contributing factors to the formation of the sequence.

6.3 RESULTS

The ESR dating of the deposits from ref. 2 (*a*) were plotted against depth (Figure 33). It was clear from the resulting graph that the MSA and basal deposits were almost equidistant apart, possibly indicating that compaction and compression in the upper layers was almost equal to that in the basal layers over two separate time periods. In order to test this assumption, *Lin*, which would corresponded to a uniform rate of compaction throughout the sequence, was applied to the data (Figure 33. An R^2 value of 0.991 was obtained (Figure 33; Table 17, (a)), clearly showing that the data taken from 2 (*a*) represents equal compaction of both the upper and basal sediments, as well as suggesting an equal formation time.

In order to test the statements on the age of the deposits being related to greater compaction with increased depth, *Log* applied to the above data (Figure 33; Table 17, (a)). As predicted, the trend line indicated greater compaction occurring in the basal levels, and gradually decreasing towards the surface. Because of the close fit of data to *Lin*, *Log* did not fit the data at R^2= 0.95. *Log* provided an age for the basal deposits similar to that of the original data of *c*. 260 ka, and a basal MSA occupation horizon age of *c*. 90 ka. When the *Log* MSA age of *c*. 90 ka was substituted for the ESR MSA age of *c*. 127 ka, the fit to data improved to R^2= 0.997, with an even lower *Log* MSA age of *c*. 82 ka, and a *Log* basal deposit age of *c*. 253 ka (Table 17, (b)). The best *Log* fit to data was obtained by regressing the ESR MSA age to 78 ka (R^2= 0.999). with the basal deposits remaining at 250 ka (Figure 33; Table 17, (c)). As the ESR MSA age was regressed, so the R^2 value of *Lin* decreased away from the original R^2= 0.991, to R^2= 0.919, while at the same time the *Log* MSA age increased towards R^2= 1.0. These results tend to contradict the supposition of equal compaction times for the upper and basal sediments, by confirming a higher degree of compaction and compression in the basal sediments. To further test the statements on the age of the deposits being related to greater compaction with increased depth, *Exp*, which would indicate an increasingly greater degree of compaction at depth, was applied to the above data (Figure 33). As was expected, because of the aforementioned close fit of data of *Lin, Exp* did not fit the data either, with a best fit to data of R^2= 0.790 (Figure 33; Table 17, (a)). *Exp* also suggested that the Unit F, horizon was considerably younger at *c*. 60 ka, with the basal deposits being older at *c*. 465 ka. This *Exp* MSA age is somewhat older than the previously mentioned >43 to 47 ka ^{14}C ages. When the *Log* MSA age of *c*. 90 ka was substituted for the ESR MSA age of *c*. 127 ka, *Exp* provided a much improved R^2 value of 0.803 with an MSA age of *c*. 55 ka and a basal deposit age of *c*. 409 ka (Table 17 (b)). The basal deposit age is in partial agreement with the suggestion that the Florisian-Cornelian faunal boundary could be as old as *c*. 400 ka (Klein, 1984). There were only small changes in *Exp* ages when the MSA age of 78 ka was substituted for the *Log* 127 ka (Figure 34; Table 17, (c)). The R^2 value improved slightly to 0.807 with an *Exp* MSA age of *c*. 52 ka and a *Exp* basal deposit age of *c*. 386 ka.

In addition to the above, if it were assumed that the ESR MSA age of *c* 127 ka was correct, and held as a constant, the application of *Log* extends the basal deposits to *c* 420 ka (R^2= 0.999) (Figure 35; Table 17, (d)). The application of an *Exp* to the ESR MSA age of *c* 127 resulted in the age of the basal deposits being extended to *c* 690 ka (R^2= 0.843) (Table 17, (d)). It was found to be pointless applying further basal deposit age variables to the exponential trend line, as MSA and basal deposit ages advanced rapidly to the point of being meaningless (Table 17, (e)), while the logarithmic trend line had already provided a best fit to data. By substituting the deepest average age of 200 ka from (*a*) for the basal deposit age, all the MSA ages decreased from 127 ka (*Lin* 120 ka, *Log* 82 ka, *Exp* 55 ka), with an increase in the R^2 values (*Lin* R^2= 0.994, *Log* R^2= 0.885, *Exp* R^2= 0.7617). When using the oldest age of 281 ka from (*c*) for the basal deposit age, all MSA ages increased relative to (*a*) (*Lin* 150 ka, *Log* 100 ka, *Exp* 63 ka), with a decrease in the *Lin* R^2 value (R^2= 0.979), and an increase in the *Log* and *Exp* R^2 values (*Log* R^2= 0.976, Exp R^2= 0.803). These increased R^2 are an indication that both the MSA and basal deposit ages should be higher than either the *Log* or *Exp* ages in Table 17, (g), with a further indication that Figure 35; Table 17, (d), is most probably the more correct. A trend line takes into account all data, and smoothes this out over the entire data range. Trend lines do not take into account variables such as time frames, aeolian deposition, aquatic growth, decomposition and compaction, as well as wet and dry periods. This would imply that in Figures 33 and 34, compaction might have tracked neither *Log* nor *Exp* exactly. It is therefore proposed that compaction occurred somewhere between *Log*

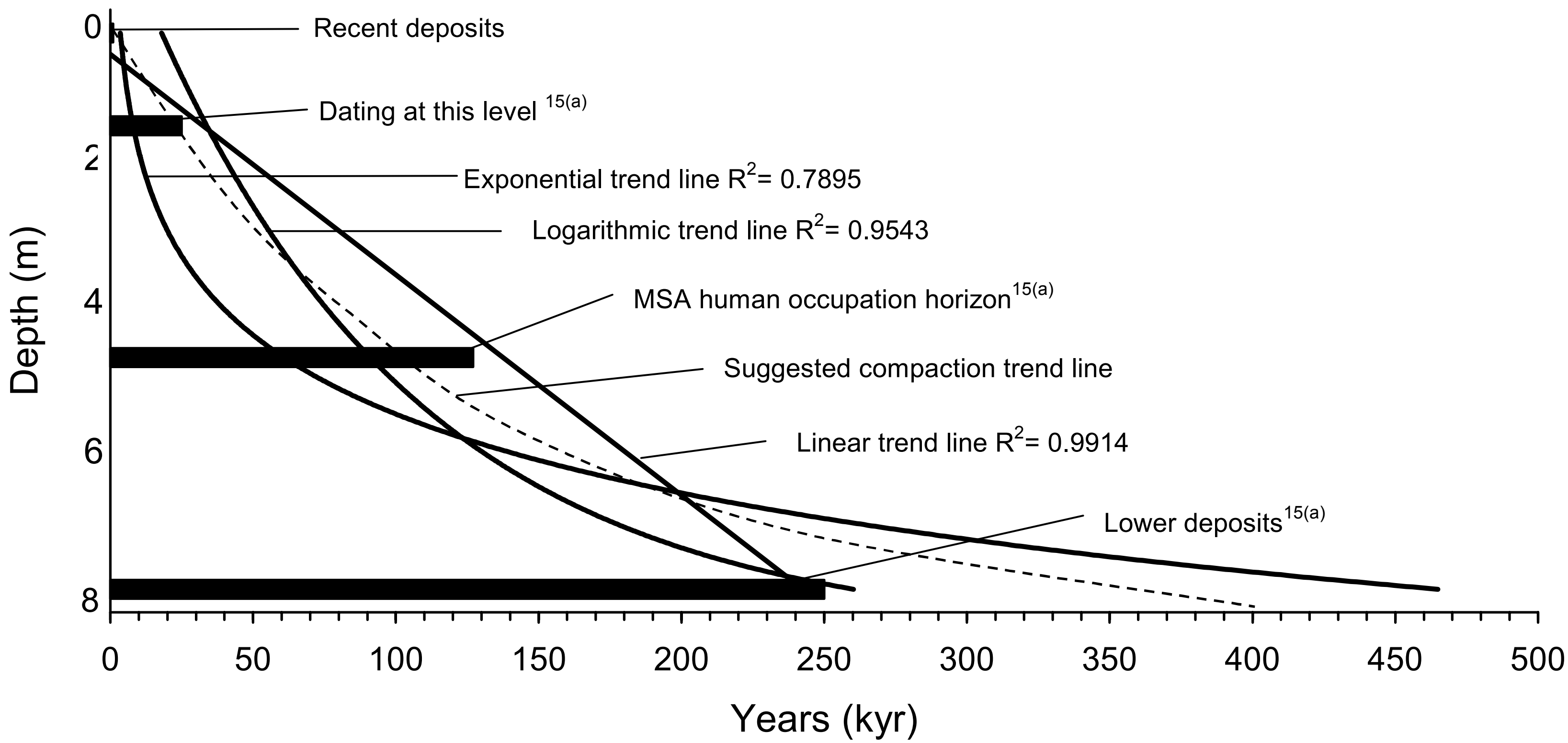


Figure 33. A plot of linear, logarithmic, and exponential trend lines to recent Florisbad ESR and OSL ages from Test Pit 3 over a depth of 8 metres. The best fit to data of a trend line is represented by R^2= 1 0. 15(a) = Grün *et al*. (1996) (a) represents the illustration identifier where data has been tabulated in Table 16.

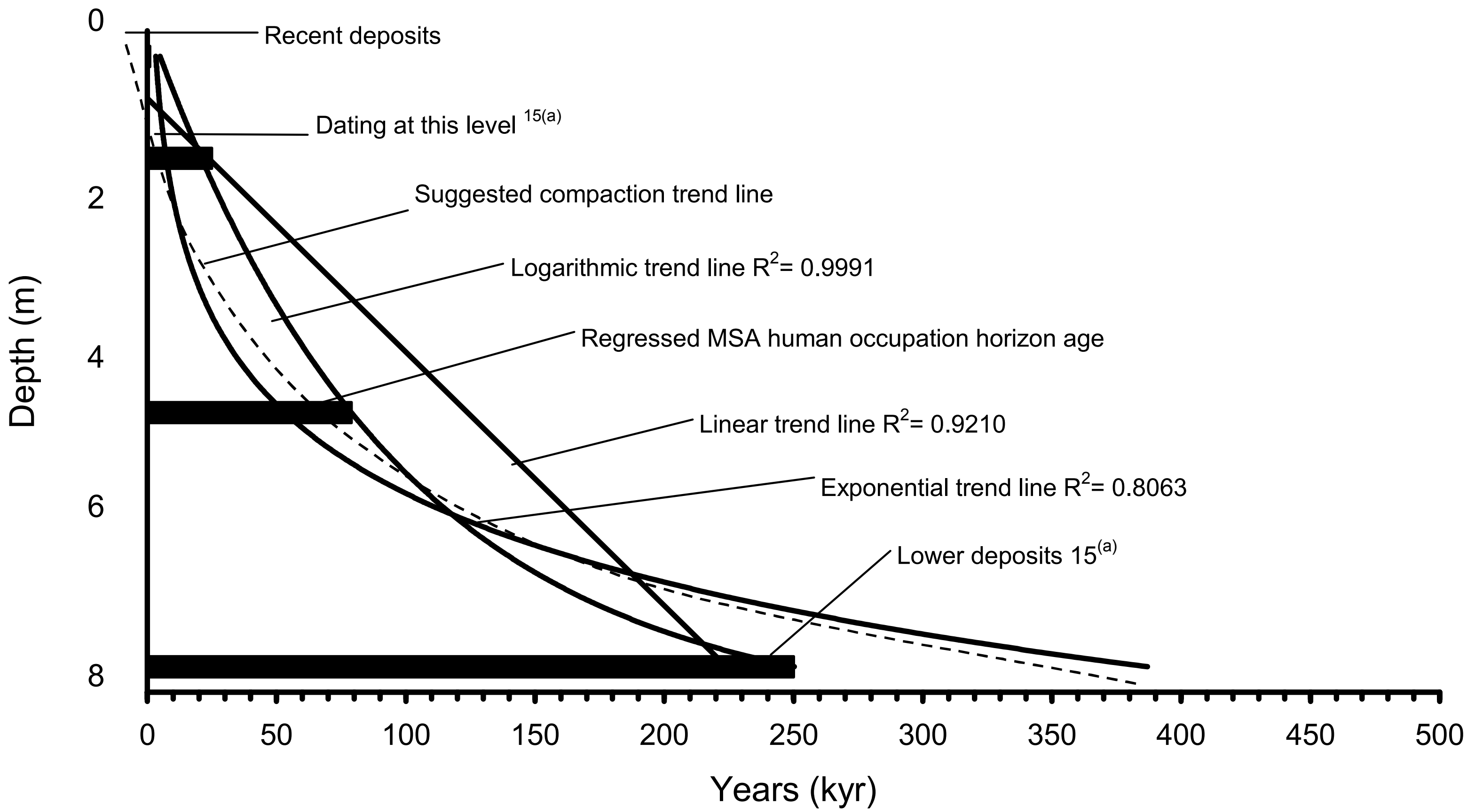


Figure 34. A plot showing the effect on linear, logarithmic, and exponential trend lines by regressing the MSA age to 78 ka and holding the basal sediment age at 250 ka. The best fit to data of a trend line is represented by R^2= 1 0. 15(a) = Grün *et al*. (1996) (a) represents the illustration identifier where data has been tabulated in Table 16.

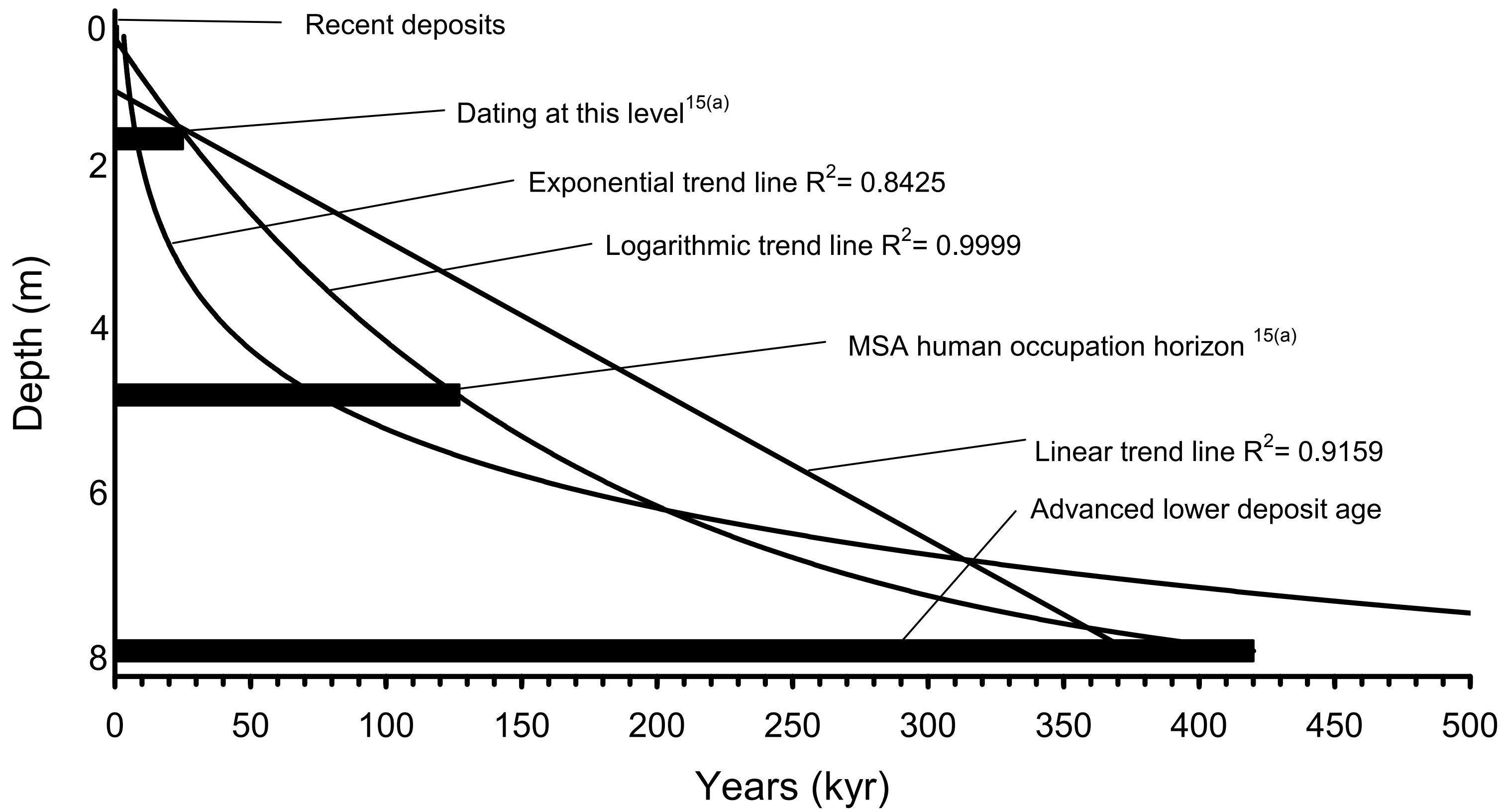


Figure 35. A plot showing the effect on linear, logarithmic, and exponential trend lines by extending the age of the basal sediments to 420 ka and holding the MSA age at 127 ka. The best fit to data of a trend line is represented by R^2= 1 0. 15(a) = Grün *et al*. (1996) (a) represents the illustration identifier where data has been tabulated in Table 16.

Table 17. A summary of the various ages and depths discussed in text and the affect of applying linear, logarithmic and exponential trend lines to this data

Text Ref,	MSA age x plot (ka)	B. Dep. x plot (ka)	Linear trend line MSA age (ka)	Linear trend line B. Dep age (ka)	Linear trend line R^2	Logarithmic trend line MSA age (ka)	Logarithmic trend line B. Dep age (ka)	Logarithmic trend line R^2	Exponential trend line MSA age (ka)	Exponential trend line B. Dep age (ka)	Exponential trend line R^2
a (Fig. 33)	127	250	137	242	0.9914	93	260	0.9543	60	465	0.7895
B	90	250	128	229	0.9446	82	253	0.9967	55	409	0.8029
c (Fig. 34)	78	250	125	222	0.9187	80	250	0.9991	52	386	0.8065
D (Fig. 35)	127	420	205	375	0.9159	127	420	0.9999	75	690	0.8425
E	127	1825	770	1460	0.7547	430	1740	0.9500	127	2170	0.9240
F	127	200	120	200	0.9939	82	200	0.8850	55	200	0.7617
G	127	281	150	281	0.9793	100	281	0.9756	63	282	0.8028

B. Dep. Age = Basal deposit age

and *Exp*, more or less following the suggested compaction trend lines. It is proposed that compaction increases from the surface downwards, initially following *Exp*, but tending towards *Log* near the middle of the sequence, although at the same time mirroring the *Exp* gradient. With increasing depth, the suggested compaction trend line then tends back towards *Exp*, but then mirrors the *Log* trend line gradient, taking a path between the two. This would suggest that, by maintaining the age of the basal deposits at 250 ka, the age of the Unit F, horizon could then fall between the *Log* MSA age (*c*. 84 ka) and the *Exp* age (*c*. 56 ka), at 70 ± 12 ka (Table 17). The crossover point between *Log* and *Exp*, which remains fairly constant at about 2 m above basement in Figures 33, 34, and 35, is significant. From current investigations, it is contended that from the genesis of the spring to *c*. 200 ka (Figure 35), the site existed as a large drainage impeded spring pan (Appendix IV; Kuman *et al*., 1999). From *c*. 200 ka to the present, it is further contended that there was a considerable increase in the rate of aeolian deposition and a corresponding increase in the rate of compaction in the basal sediments due to an increased mass of aeolian sand moving across the site in the form of a sand dune (Appendix IV).

6.4 DISCUSSION

The ESR age of a sample normally lies somewhere between the estimates of two hypothetical models based on the rate and manner of the early, and linear (continuous) uranium uptake by an object (Grün and Thorne, 1997). These models are however only relevant on U absorbing material such as teeth and bone. There are inherent problems with particularly the ESR dating method used, and the results may reflect a lack of information regarding aspects such as the formation, history and chemistry of a site. Chemical reactions, for example, may influence the distribution of radioactive elements and subsequently the ages obtained by ESR and thermoluminescence dating (Mercier *et al*., 1995), with the degree of chemical reaction varying over the site, and even within samples. ESR and uranium-series dating also depend upon knowledge of the manner of uranium uptake into the bone (Millard and Hedges, 1999) and, more specifically, that the sample must be from a closed system with uranium being present at the time of deposition (Trueman and Tuross, 2002). A number of factors may have influenced the Florisbad uranium-dependent dating methods as well as indicating historically low radioactivity levels. These include very low scintillation counts of between 9 and 18 counts per minute for the area in general, including calcrete exposures, pit-waters and spring-water, as well as extremely low levels of stable and radioactive isotopes (Appendix I; Mazor and Verhagen, 1983). Uranium levels for tooth enamel (3–101 ppb) and dentine (10–268 ppb) in the ESR analysis were also low (Grün *et al*., 1996).

Libby (1954) provided an age of >41 ka for the Peat I layer at Florisbad but apparently this greater-than-age was not based on the limitations of the methods used. According to Oakley (1955), the Florisbad deposits were accumulated by spring waters possibly arising from an underlying Palaeozoic formation, which included Ecca coal measures. The peat unquestionably contains Pleistocene plant material with the possibility that it also contains Palaeozoic carbon (Oakley, 1955). Should this `dead` carbon have a high percentage, the point of no measurable radio-activity would be reached within 41 ka (Oakley, 1955). This was interpreted by Oakley (1955) that Libby (1954) found no radio-activity in the Peat I layer.

The accumulation of uranium in enamel and dentine, over time, can result in uncertainties in estimating age (Grün *et al*., 1990a), and it is evident that age uncertainty in ESR results is strongly dependent upon the circumstances of burial (Rink, 2001). Relatively high uranium concentrations in both enamel and dentine may result in considerable differences between linear and early uptake results, making it difficult to estimate ages (Grün *et al*., 1990b) due to the lack of differentiation between the two uptake modes. For example, uranium can show steep concentration gradients in bone and teeth with uranium gradients decreasing towards the cortex, as can be illustrated by the difference in uranium concentrations between enamel and dentine (Williams, 1988; Grün *et al*., 1990a; Grün *et al*., 1990b; Janssens *et al*., 1999; Rink, 2001). The low U and Th content at Florisbad, and the use of the saturated content for the duration of the burial period by Grün *et al*. (1996), may have contributed to a greater source of scatter in their data. It has been proposed that the U content at Florisbad has always been historically low, and therefore there would have been little variability over time, and this would have been reflected in all the sediments through out the sites depositional history.

Results may be further influenced by the degree of bone preservation, with bones in sandy soils usually remaining more porous and attaining a more rapid equilibrium with the environment in which they were buried, but with a shallower concentration gradient (Millard and Hedges, 1995; Hedges and Millard, 1995). For example, no firm ages could be established for ESR dating on tooth enamel from Kromdraai B, due to high uranium concentrations in the enamel and dentine, neither could the reliability of determining isochron age estimates (Cunroe *et al*., 2002). Good bone preservation also usually results in low uranium levels while poor preservation usually leads to high uranium levels (Janssens *et al*., 1999). Bone preservation will of course largely be dependent on the hydrological environment and history of the site, as well as the location of bones within the sediments. While much attention has been given to the Florisbad fossils, no detailed studies or information on the historical chemical and physical processes of fossilization have been published.

Post-depositional sample enrichment by uranium may also affect the magnitude of ESR signals. Uranium concentrations for the Florisbad tooth enamel and dentine can be considered low at 3–101 ppb and 10-268 ppb, respectively (Grün *et al*., 1996). Post-depositional sample enrichment by uranium was not regarded as a problem at Border Cave because of what were considered negligible amounts of U in the teeth (Grün *et al*., 1990b) (enamel <10–460 ppb and dentine <10–440 ppb), but which were still higher than the Florisbad values. These low uranium values for the Border Cave teeth indicated that there had been no mobilization of uranium to influence the ESR result, this probably being due to the cave floor deposits having been in a dry state throughout their history (Grün *et al*., 1990b).

Environmental conditions may have implications at Florisbad because, despite long dry periods, the site was also inundated with water for long periods due to the possible uninterrupted spring flow from an apparently large aquifer, about which little is known (Appendix I). This could have had a major effect and influence on the mobilization of any uranium present. Furthermore, uranium may well have been flushed at a considerably higher rate from bones that were in close proximity to the spring vents (Appendix III). Variations in uranium concentrations over the site were possibly further influenced by the

Eh of the peat deposits, which tend to bind uranium, producing low solution concentrations (Janssens *et al*., 1999), and good bone preservation (Millard and Hedges, 1995). Despite the many ideal factors involved at Border Cave, such as low uranium concentrations and dry conditions, it was still concluded that ESR age assessments of certain significant occurrences, such as the basal part of the SAS member, and the Howiesons Poort artefact occurrence, could not yet be resolved (Grün *et al*., 1990b). In contrast to this, at Klasies River Mouth it was noted that the high uranium values might have influenced ESR results, making it difficult to estimate age (Grün *et al*., 1990b). This is of interest because the Florisbad enamel and dentine uranium values are the lowest of the three sites, suggesting that the Florisbad ages should be the more accurate. Notwithstanding apparent inconsistencies and other possible problems associated with the dating and methods employed at Florisbad, the authors still had confidence in the reliability of the methods (Grün *et al*., 1996).

In ref. 2 (*c*) OSR results give the Unit F horizon as 133 ±31 ka and the sands above Peat II as ascending in age by a probable 55 ka, from 106 ka to 161 ka. This is questioned when considering that spring activity probably never occurred in these upper levels, as is indicated by the unbroken and uniform stratigraphy of the sediments in the MSA horizon (Kuman and Clarke, 1986; Kuman et al., 1999; Henderson, 2001a). Another issue is the 100–350 ka age spread for hippopotamus (*Hippopotamus amphibious)* and wildebeest (*Connochaetes gnou*) teeth from the spring material (Grün *et al*., 1996; Brink, 1997), which makes the Unit F horizon hippopotamus remains older than the youngest of the spring collection specimens. However, this spread could well be attributable to spring action. Contrary to the confidence expressed in the reliability of the methods, it was noted that ESR and OSL ages from the third test pit only confirmed the general age span of the site, and were considered too imprecise to provide any further insight (Brink, 1997). Reasons given for some of these anomalies were large errors in the OSL results due to saturation problems, and that the large age spread of ESR ages, in both the third test pit and spring material were due to re-working of the sediments by spring action (Grün *et al*., 1996). Columns of sand resulting from the spring vents are distinguished by their whiteness (Dreyer, 1938; Brink, 1987), because they have been flushed clean of clay and organic material (Appendix III). There appears to be no evidence of such columns having been recorded in the third test pit, and therefore no real evidence of spring action. This would imply that if any disturbance has occurred within the sediments, this must have had other origins. It is suggested that these large ESR and OSL age spreads, of up to 125 ka at the MSA level levels, are more likely to be as a result of the internal stratification of the sand dune and debris it accumulated as it migrated across the site (Appendix IV), and not spring action.

The location and sampling of the third test pit may also have been partially responsible for the perception of skewed data and erroneous dating. For example, it is possible that test pit 3 was not located over an area where early remains had been deposited. Had there been remains of earlier specimens in the area of the test pit, then the previously suggested spring activity (Brink 1987; Grün *et al*., 1996) should have made it even more likely that earlier remains would have been sampled. Furthermore, the size of the spatial sampling area of the test pit, in both horizontal and vertical dimensions, may have been too limited, and presented too small a window for the sampling of early remains.

The application and comparison of other age-depth models was well beyond the scope of this study, and this study should be seen as an alternative method. There are a number of age-depth models based on linear regression, splines, and linear interpolation (Bennett, 1994), fuzzy regression (Boreux *et al*. 1997), Bézier curves (Bennett and Fuller, 2002), mixed-effect models (Heegaard, 2003), polynomial regression or cubic splines, and monotonic smoothing splines (Enters *et al*., 2007), which can be used. These can be integrated into programs such as INTCAL98 and INTCAL04 (Reimer *et al*. 2004), and the Bayesian model, OXCAL 4.0 (Ramsey, 1995; Ramsey, 2001). Age–depth models are only meaningful when using calibrated radiocarbon dates where different approaches may produce very different results for an age at a particular depth, a difference in their respective sedimentation rates, as well as their uncertainties attached to the estimates (Bennett, 1994; Bennett and Fuller, 2002; Telford *et al*., 2007; Enters *et al*., 2007). It should be noted that different calibration curves are separately applicable to the northern and southern hemispheres due to differences in the reservoirs of ^{14}C in the two hemispheres (McCormac *et al*., 2004). Cal ka BP dating of from 0-12.4 cal ka BP are based primarily on dendrochronological dating by ^{14}C tree ring data measurements, as in the SHCAL04 calibration set of McCormac *et al*. (2004) (0-11 cal ka BP) for the southern hemisphere. Marine dating, which covers the calendar time span of 10.5- 26 cal ka BP, is derived is primarily on data derived from U/Th dated corals and foraminifera as in the MARINE04 calibration curves (Hughen *et al*., 2004).

BP is the abbreviation for "Before Present" and represents uncalibrated data, while cal BP represents data calibrated from calibration curves onto a calendar time scale expressed as BC/AD. "Present" is taken as 1950, being the year in which calibration curves for radiocarbon dating were established, and also precedes large scale testing of nuclear weapons which altered the global ratio of ^{14}C to ^{12}C. The BP scale may typically have uncertainties high enough that the difference between 1950 and the actual present year may be insignificant (Anon [7] 2008). From the literature it would appear that there are two distinct schools of though on the use of calibrated (cal BP, AD/BC) and uncalibrated (BP) ^{14}C data. One school of thought can be typified by the archaeological and geological community who would like to know the age of an item, or sediment, is by using uncalibrated ^{14}C radiocarbon years (BP) data. The other school of thought is typified by the theoretisists who find it more logical and convenient to work with calibrated dates, as on a calendar time scale, or calendar age range (cal BP, AD/BC). An argument for the use of calibrated ages is that BP gets all sorts of uses in different contexts, and that age is also a variable quantity (Ramsey, C.B. pers com. 24 February 2009).

Even although theoretisists emphasise the use of calibrated ^{14}C radiocarbon dates in age-depth models, they also concede that calibrated dates are probabilistic and add an extra level of complexity (Bennett, 1994; Sewell, 1998), as the resulting probability distributions are not Gaussian (Telford et al, 2004). By including the stratigraphic relationship of the dated samples, information from short-lived isotopes, and other factors the uncertainty of using calibrated dates can be considerably reduced (Enters *et al*., 2007).

The question remains as to whether the application of calibrated correction curves to the Florisbad BP ages is warranted, and what might be achieved from such an exercise.

Because cal ka BP dates only provide good accuracy up to 26 cal ka BP (Reimer, *et al*. 2004), this would only accommodate approximately the top two metres of the Florisbad deposits down to roughly the Peat III horizon. This would imply that calibrated age-depth models would possibly also only be accurate to this level, but this was not investigated in detail. To date, much of the interesting material that has been excavated at Florisbad lies below these dates, and at this time, there does not appear to be sufficient justification for testing these complex exercises, which would appear to be incapable of producing any meaningful results.

Sedimentation is another important factor in age-depth determinations. Telford *et al*. (2007) noted that the number of dates needed to construct an age-depth model for a sedimentary sequence will depend on the required precision and complexity of the sedimentation rate, and that in order to achieve accurate, high-precision, many more dates may have to be used than is currently the norm.

6.5 SYNOPSIS

It is apparent from the results in Figures 33 and 34 that the ESR and OSL ages of the Florisbad sediments, as reported in ref. 2 (*a*), do not conform to a process of increased compaction and compression with depth, as has been shown by the application of *Lin*, *Log* and *Exp*. As indicated by *Lin*, and considering the many variables involved, it is highly improbable that the upper and basal sediments accumulated and became equally compacted and compressed within almost identical time frames, at different depths. This leads to the conclusion that either one, or all three of the ages drawn from ref. 2 (*a*), and used in Figure 33, may be in error. Errors in the dating may have been further aggravated by factors such as inherent problems with the dating methods, and a lack of information regarding the formation, history and chemistry of the site. As regards chemistry, there could be grounds for the possible low, or even non-existent, occurrence of uranium for early uptake, with the effects of spring flow and the peat deposits influencing results negatively. The degree and type of bone preservation along with environmental factors, and about which little detail appears to be known, may also have had a considerable influence on results. These aspects emphasise the importance and necessity of having a more holistic knowledge and understanding of the many aspects associated with a particular site, and that by simply applying a dating method to a sample, may not be sufficient for accurate dating.

If the original ages drawn from ref. 2 (*a*), and used in Figure 33, are taken to be incorrect in relation to their depth, it must then also be presumed that although *Log* and *Exp* reflect a more accurate version of compaction, they may also be slightly in error due to the many previously mentioned variables. According to the above arguments, if the Unit F, horizon ESR age of *c*. 127 ka is found acceptable, then according to *Log*, the Unit F age would reduce to *c* 90 ka, with the basal deposits increasing slightly to *c*. 260 ka (Figure 33; Table 17, (a)). *Exp* indicates a greater increase in the age of the basal deposits to *c*. 465 ka with a conversely greater decrease in the Unit F age to *c* 60 ka (Figure 33; Table 17, (a)). Alternatively, if the basal deposits age of 250 ka is found acceptable, then the Unit F horizon must be much younger at *c*. 79 ka, as indicated by *Log*, or even *c*. 52 ka as indicated by *Exp* (Figure 34, Table 17, (c). However, both suggested compaction trend lines indicate a Unit F age of *c*. 95 ka and a basal deposit age of *c*. 380 ka. What Figures 33 and 34 also show, is that compaction could not have occurred at equal rates in both the upper and basal sediments.

Log in Figure 35 reflects the best R^2 fit to data, with the Unit F age at 127 ka, and the basal deposit age advanced to 420 ka, indicating that the ESR 250 ka basal deposit age used in Figures 33 and 34 may not be a true reflection of the age of the basal deposits at this depth. The 420 ka age also has better agreement with the oldest possible age recorded for the site, namely *c*. 354 ka, and the minimum suggested compaction trend line ages (*c*. 380 ka) in Figures 33 and 34. Additionally, the 420 ka age also corresponds closely to the proposed Florisian–Cornelian boundary at *c*. 400 ka (Klein, 1984). The ages suggested by the suggested compaction trend line in Figures 33 and 34 of 92 ±12 ka for the Unit F horizon and 400 ±20 ka for the basal deposits, seem to best reflects the ages of the sediments at Florisbad, but more work is needed to confirm this.

In light of the above, and regardless of the Unit F horizon having the proposed smallest error value (Grün *et al*., 1996), this small error value should not necessarily be seen as reflecting the accuracy of the dating. Given the recent dating of the Florisbad hominid cranium at *c*. 259 ka (Grün *et al*., 1996), the arguments advanced in this paper regarding the chemistry of the site and possible flaws in the dating methods may help to explain how such a very early age was claimed for a specimen with such an advanced morphology (Brink, 1997).

It is concluded that mathematical trend lines could be used to evaluate and validate ESR and OSL ages at Florisbad, with the trend lines confirming that the ages in ref 2 (*a*) must be suspect, and in error, when related to depth. On the other hand, where it was indicated that ages and dating might be more correct, for example, where the oldest recorded age was placed at the greatest depth, the trend lines closely confirmed these ages and depths. It is also evident that the Unit F and basal deposit trend line ages reflect the general trend of the ESR and OSL ages, but with the trend lines strongly suggesting that the age of Unit F could possibly be younger, while the basal deposit age could possibly be older. From this it can be stated that the trend lines confirmed two opposing scenarios at Florisbad and can therefore be seen as a tool in evaluating ESR and OSL ages under these circumstances.

Chapter 7

Discussion and Conclusions

CHAPTER 7

DISCUSSION AND CONCLUSIONS

The initiation of this research was as a result of spring and excavation pit water samples taken during a high rainfall period in 1988. These results recorded a significant difference in water quality between the spring- and excavation pit water. In 1999, during an average rainfall period, it was decided to confirm the previous results by re-sampling the spring and excavation pit water, and to expand the water sampling to other excavation pits in the area of the springs. The spring water had previous been analysed by Fourie (1970), so the results would not only expand the spring-water database, but also establish a database for the exploration pit waters. In addition to this, comparative results would be available for both an exceptionally high rainfall period as well as an average rainfall period.

The TDS of the original exploration pit water showed a 58% decrease between the 1988 high rainfall period and the 1999 average rainfall period. This was contrary to the norm where high rainfall would be expected to create a dilution effect on the excavation pit water. This incongruity provided motivation to examine other aspects of the sedimentation in order to resolve the issue. The increase and variability of the composition of the pit waters, and the stability of the spring-water, is evident in the Stiff diagrams (Figure 25). This evidence then also provided a strong argument for two separate hydrological entities, namely, the spring-water and groundwater (exploration pit water). This was the first time that the spring-water had been analysed in such detail, and the first time that the groundwater had been examined.

Further to the chemical evidence, the quality of the two hydrological environments was also examined in relation to rainfall. It was established that neither long- nor short-term, rainfall had much affect on the quality of the spring-water due to the presumed large size, and distant location, of the spring aquifer. It is proposed that the spring aquifer effectively stabilizes any fluctuations in the spring-water quality. However, short-term rainfall had a decided affect on the quality of the groundwater by increasing TDS during high rainfall periods and reducing TDS during average to dry periods.

Of prime importance was the establishment of causes and processes for the higher salinization of the groundwater. Peat II samples were taken for analysis from the wall of the excavations and the excavation pits, while Peat IV samples were taken from the west wall of the excavations. Analysis of the organic-clay layers showed even more elevated TDS levels than the groundwater, providing evidence of an even higher degree of salinization in these layers. Causes and processes that would account for this higher salinization of the organic-clay layers were then sought. Primarily, the adsorption of ions occurs due to the presence of both clay and humus, the latter being derived from the decomposition of halophytes, phytoplankton, periphyton, the waste from the large herds of animals, which previously visited the site, and the decomposition of carcases.

Clay and humus have a similar structure where highly charged anions attract, adsorb, large numbers of cations. It was concluded that the organic-clay layers at Florisbad acted as storage zones, with ions being adsorbed and concentrated during periods of average and low rainfall, and then being flushed from the organic-clay layers by percolating rain and groundwater during periods of high rainfall. Other factors contributing to the salinization of the organic layers included salts from aeolian deposition, particularly from Soutpan; the accumulation of salts through impeded drainage and a closed pan environment; capillarity; illuviation; the lack of leaching in a semi-arid environment; and a degree of adsorption through the miscible displacement with the spring-water. The organic-clay layers therefore had the ability to adsorb, accumulate, and retain ions at appropriate pH and Eh levels which were within the required dissolved organic carbon content and P_{CO2} conditions. These processes further occurred in the presence of Ca and P with a restricted water circulation. Salinization is a complex process with the above mentioned factors being found to be interrelated and often dependent on one another in creating an ideal equilibrium for fossilization to take place.

This was the first time that the organic-clay layers at Florisbad or any other southern African archaeology site had been analysed from a salinization aspect, and as a potential source of mineralization for the fossilization of faunal remains. The quality of the hydrological and sedimental environments in relation to one another, and particularly the salinization of clay and organic matter, could be important in understanding fossilization process when applied to investigations at other fossil sites where clay and/or organic matter occur. It is strongly suggested that, based on these results, the basal peat layer and Peat I where, at one time, appreciably more highly mineralised than even the Peat II layer, allowing for the fossilization of the Old Collection.

It appears to have been generally accepted that the salinized spring-water was responsible for the fossilization of faunal remains at Florisbad (Brink, 1987, 1988), with much unsubstantiated evidence having been provided in support of this. It was put forward that the spring-water was carbonate rich and, with evidence of calcium carbonate deposition, the bones must have been in contact with the spring water for some time (Brink, 1987). Brink (1988) also noted that such prolonged contact with the spring-water had caused the remains of the Old Collection to be well preserved in a characteristic way. Post-depositional mechanical and chemical weathering was cited as further evidence that faunal remains had become fossilized in areas of spring activity. Butzer (1988) on the other hand stated that the spring-water was under-saturated in both calcium and bicarbonate and that any carbonate enrichment must be derived from either direct, or indirect external sources, other than the spring-water. This further supported the salinization of the organic-clay layer hypothesis.

An investigation of previous water analysis (Rindl, 1915; Fourie, 1970; Mazor and Verhagen, 1983; Douglas, 1992) indicated that there was insufficient Ca, $CaCO_3$ and P in the spring-water for the initiation of fossilization. This substantiated Butzer's (1987) statement of under-saturation, and was further verified by the analyses and results of this study. It was concluded that under saturation would have been applicable even in an historic sense because, historically, the spring vents, whether active or inactive, contained no clay or organic material to adsorb, accumulate, or retain the ions necessary for fossilization. To all intents and purposes, in the context of salinization and fossilization, the spring vents were found to be sterile. Furthermore, the high Eh environment of the spring flow would have inhibited precipitation and rapidly oxidized any organic material.

Contrary to the statement that post-depositional mechanical and chemical weathering clearly showed that the remains had become fossilized in a spring context, analysis of data indicated that post-depositional mechanical and chemical weathering, as well as the low incidence of calcite on, and in, bones (chemical weathering), reflected that bones from the spring vents have been in a state of demineralisation. At a high pH and low Eh, calcite and hydroxyapatite would be preserved, while at a low pH and high Eh, they would be dissolved and mobilized. The former conditions are also necessary for the production of DOC, which in turn acts as a food source for carbonate precipitating micro-organisms. However the level of DOC may be critical as over saturation may act as a kinetic growth inhibitor on calcite crystals, with a fine balance having to be reached between the production, and consumption of DOC by $CaCO_3$ precipitating micro-organisms and plant decomposition in order to prevent DOC acting as a $CaCO_3$ inhibitor. It is suggested that these conditions could not have existed in either the active or non-active spring vents due to a lack of organic material. Although the spring water may have been conducive to the preservation of calcite and hydroxyapatite at pH 8.0, it would not have been conducive for the dissolution and mobilization of calcite and hydroxyapatite in order for initialise fossilization.

In summary, the spring vents and their immediate environment are devoid of all requirements and processes for the initiation of fossilization. Brink (1987) did however note that if the presence of calcite was an indication of contact with the spring-water in the original sedimentary environment, then the relatively low incidence of calcite concretions might contradict the assumption that the Old Collection was entirely derived from spring deposits. It was noted that only 48.2% of Bovidae and hippopotamus bones showed signs of calcium carbonate deposition (Brink, 1987).

Rubidge and Brink (1985) considered the lithostratigraphy and depositional history of the Florisbad sediments as still being in their initial stages of investigation, and felt that several models could be proposed. In their opinion, because the deposits were lithologically variable, they were seen as being the product of an unusual depositional environment. Kuman and Clarke (1986) also noted that the sand, silts, and organic deposits were poorly understood and controversial. The above questions and statements therefore opened the way for further interpretation on the formation of the site.

Based on contemporary theories on the formation of the Florisbad archaeological site, as well as the results of this study, it was concluded that aeolian deposition and sedimentation alone could not account for the formation and the current morphology of the immediate spring site and that erosion played a major role. It was further concluded that previous theories on the formation of the spring site could neither explain adequately the stratigraphy of the site, nor accommodate the fossilization process within the current stratigraphy. The three main theories put forward for the formation of the spring site were that:

- Deposition was almost entirely derived from aeolian sources (Appendix I; Appendix II; Appendix III; Grobler and Loock, 1988a, 1988b; Van Zinderen Bakker, 1989; Kuman *et al*., 1999).
- Development and formation of the spring site was from sand output from the springs, which most probably originated from the underlying bedrock (Dreyer, 1938a; Brink, 1987; Butzer, 1988; Van Zinderen Bakker, 1989).
- That deposition was fluvial, being directly related to four low water level phases and three high level water phases (flooding) in the adjacent palaeolake (Soutpan) (Visser and Joubert, 1991; Joubert and Visser, 1991; Henderson 200).

Based on data from the water analysis and salinization processes, as well as an intensive study of the morphology of the region and the site, along with literature reviews, it was concluded that a new hypothesis was required in order to incorporate new evidence into the formation of the spring site. This hypothesis would have to account for, and accommodate, not only the unexplained morphological anomalies mentioned below, but at the same time be capable of accounting for the fossilization of faunal remains at least up to the level of the MSA Human Occupation Horizon. As part of this formulation considerable attention was paid to the western Free State panveld and dune development, with a new hypothesis on the development of the Florisbad sand-dune being formulated. These unexplained morphological features included:

- The lack of upper red-brown sand units on the eastern side of the "mound" appeared to have either not been deposited, or had been eroded by spring discharge (Rubidge and Brink, 1985; Brink, 1987).
- The seven metres thick layer of clay deposits in the modified vlei area (Butzer, 1984), which could not have been deposited by the pre-existing weak ephemeral drainage line.
- The older aeolian deposits, evident over the rest of the site, were absent from the vlei area, while more recent aeolian deposits were present Fourie (1970).
- The extent of the dogleg in the ephemeral drainage line from the inter-dune valley, cutting across the natural drainage at almost right angles from NW to NNE, and back to NW.
- The disproportionate width and depth of the vlei area in relation to the ephemeral drainage line.
- The erosion of the east bank of the vlei area.
- The pinching out of the Peat IV layer on the western wall of the excavation area.

A hypothesis was formulated whereby a dam was formed at the spring site by a migrating sand dune coming to rest against a previously establish dune belt. This dam slowly filled with sediments until the water level reached the top of the eastern arm of the dune. The water then broke through the eastern arm of the dune creating a flash flood, with the contents of the dam evacuating to create a wide vlei area. It should be pointed out that since the publication of Appendix IV, this hypothesis has been considerably revised in the introductory part of the thesis due to further investigation, and should be taken as contemporary

This hypothesis proposes that an area of fossilized faunal remains will extend, and correspond, to the final size of the spring pan, before the migrating sand dune finally covered it. Therefore, the probability of extensive areas under the sand dune containing concentrations of fossil faunal remains is relatively high. It is also proposed that fossils located on the pan floor, within the area of the dam, would have been washed into the vlei, suggesting that fossil faunal remains may also occur along the bottom of the vlei.

From the results of the vlei organic-clay analysis, it can also be assumed that any faunal fossil remains found in this environment will be in a relatively well preserved state. Because these fossils would have been disseminated over a large area, they would have been concentrated to a far lesser degree than those on the original pan floor. It should be stated that the Florisbad sand dune probably had little influence on the fossilization of faunal remains, with the exception of any minerals it may have contributed to the salinization process. However, the damming of the site, the erosion of the eastern arm of the dune, and subsequent flash flood, severely affected, and changed, the local morphology of the site and surrounding area. It is believed that this theory offers a viable explanation for the formation of the spring site in a manner which accommodates the current morphology, the morphological anomalies, as well as providing an environment conducive for the fossilization of faunal remains.

Induce

There are numerous archaeozoological sites in the western and south-western Free State which are related to springs, pans and alluvial river terraces. It has been shown here that the source water of some of these springs is similar to that of Florisbad, which would imply that they too have a very low Ca content. This would then also preclude faunal remains at these sites from being fossilized in a spring or river context, and possibly clay and/or organic matter is the source of mineralization. It is therefore contended that most, if not all, faunal fossil remains of the Florisian – Conelian time period in the Free State were fossilized in either spring pan, or marsh contexts. It is also contended that the vast quantities of faunal remains at Florisbad, as well as at other sites in the area, were largely as a result of cyclical epidemic periods of bovine and equine diseases such as foot and mouth disease, anthrax, and African horse sickness.

Dating of the Florisbad sediments and deposits has been a contentious issue over the years, and sometimes for valid reasons. From the many reasons given in Chapter 6 for discrepancies that might occur in radioactive dating methods, the extremely low presence of radioactive isotopes, and the lack of information on the chemistry and history of the Florisbad site, are most probably the most conspicuous. Grün *et al.*, 1996 also noted that all teeth in the spring collection had low uranium concentrations. It was noted that the ages of the faunal remains and sediments given by Grün et al. (1996), were incongruous, and did not follow a logical ascending chronological progression from the top of the sediments down to bedrock. For example, the oldest age recorded was located some 2 m above the bedrock, while some of the much younger MSA ages were recorded well below the 2 m level. Further more, age scatter, or age spread, of up to 125 ka occurred in a narrow horizon of less that a metre wide in the area of the MSA Human Occupation Horizon, at approximately the mid-way point in the profile.

A plot of age against depth from the excavation pit of Grün et al. (1996) suggested that the narrow human occupation horizon was positioned at a mid point in the profile. On a linear scale the implication was uniform compaction of sediments throughout the profile, with only *c*. 62.5 ka being available for the basal sediments to accumulate, and another *c*. 62.5 ka for the upper sediments to accumulate. This plot left no room for the possibility of greater compression and compaction with depth, particularly regarding the decomposing deeper organic-clay layers. Considering these factors, an experimental exercise was initiated whereby mathematical trend lines were applied to the data presented by Grün et al. (1996). Another application of this exercise was that data could be manipulated and varied, as in keeping one age fixed and varying another age to obtain different ages.

The incongruous ages of Grün et al. (1996) were more acceptable in the reconstruction of the lower sediments around the spring vents where age distribution could be attributed to the redistribution of material from spring activity at different levels. However, although it was stated that these age spreads in the exploration pit were also associated with a reworking of the sediments by spring action, it is questioned whether spring action within the dune itself ever ascended to the height of the MSA horizon. The unbroken horizontal stratigraphy of the exploration pit and the MSA excavations, do not suggest spring activity at these levels. That the excavation pit was opportunely located directly over a spring eye is highly improbable as there is no mention of any larger fossils other than tooth fragments. It is proposed that the incongruity in the ages of the small tooth fragments, and their position within the dune, may have been as a result of changes in the internal stratigraphy of the dune as it moved across the site. Rubidge and Brink (1985) found that over most of the site that rapid facies changes and the lateral wedging out of layers, over very small distances within the dune, made lateral correlation extremely difficult. Rubidge and Brink (1985) stated that this unconformity was as a result of micro-environments created by a number of spring eyes giving rise to ponds, channels, and over bank deposits which moved laterally in time. This is in agreement with the hypothesis proposed in Chapter 5, section 5.6.5, but this situation would only have occurred at the lower levels, and not have extended to the higher levels, that is, with the exception of the dam itself.

As a linear trend line disputed the ages of the excavation pit ages derived from Grün *et al*. (1996), deliberation was given to applying logarithmic and exponential trend lines to the data. This, it was anticipated, would reflect increasing compaction with depth. The best fit to data of R^2=0.9999 was obtained by the application of a logarithmic trend line with the MSA age fixed at 127 ka, and the basal deposit age extended to 420 ka.. By using a suggested trend line, a basal deposit age of 400 ±20 ka was obtained while the MSA age was reduced to 92 ±12 ka, which may not be an unreasonable estimate. Application of an exponential trend line produced no best fit to data values higher than R^2=9240, where, with the MSA age fixed at 127 ka, the basal deposit age was far outside any practical limit at 2170 ka.

The application of trend lines confirmed the age incongruities in the excavation pit. By taking the youngest ages in the MSA spread of Grün *et al*. (1996) for the exploration pit at *c*. 115 ka, and the maximum age for the basal deposits at 354 ka, these ages closely approximated the ages of the suggested trend line where the maximum MSA age was 104 ka and the minimum basal deposit age was 380 ka. The suggested trend line was developed in an attempt to smooth out some of the variances which may have occurred in the stratigraphy

In summary, the chemical database for the spring-, groundwater and organic clay deposits were considerably expanded on, leading to a better understanding of the hydrological and sedimentary environments at Florisbad. This in turn resulted in a scientifically supported hypothesis being formulated that states that the fossilization of faunal remains occurred in conjunction with the organic-clay layers and groundwater, and not in a spring context, as previously proposed. It was further concluded that the spring-water is responsible for the demineralization of faunal remains, as opposed to their fossilization. The hypothesis on the formation of the Florisbad site accommodated and supported the depositional history and stratigraphy of the site, as well as accommodating the fossilization of faunal remains. Further to this, the hypothesis on the formation of the spring site also addressed, and provided, explanations for a number of previously unexplained morphological aspects of the site. While a number of hypotheses have been put forward over the years in an attempt to explain the history of the development and formation of the Florisbad spring site, and its fossilization processes, it is hoped that this research will go some way in contributing to a more holistic and better understanding of the site.

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Appendices

Appendix

I

The quality of the Florisbad spring-water in relation to the quality of the groundwater and the effects of rainfall.

Douglas, R. M. 2006.

Water SA 27: 39-48.

The quality of the Florisbad spring-water in relation to the quality of the groundwater and the effects of rainfall

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Abstract

The spring-water, groundwater and rainfall were examined as part of a study to determine aspects of the environment in which faunal remains at Florisbad were fossilised. A 1988 analysis of the Florisbad spring and exploration pit-water showed a 27% higher TDS in the pit-water after a high rainfall period, despite the two being located only 22 m apart. An extended qualitative water-sampling programme in 1999 confirmed a difference, but in this instance the TDS of the same exploration pit was 49% lower than that of the spring-water after a low rainfall period. This was contrary to the norm where high recharge usually results in low TDS and low recharge results in high TDS. Results also showed extreme TDS variations of up to 6 times higher between individual pit-waters 54 m apart. The fluctuation in the quality of the pit-water, in relation to the stable spring-water, led to the conclusion that the two should be separate entities. It was further concluded that the mineralisation of the pit-water originated either directly, or indirectly, from a source other than the spring-water. Long-term rainfall appears to have only a slight effect on the quality of the spring-water and possibly no effect on the quality of the pit-water, while short-term rainfall appears to have little effect on the spring-water quality, but has a decided influence on the pit-waters. Aspects relating to water quality and water monitoring at Florisbad are discussed and a comprehensive historic record of the spring-water quality and composition is also given. Investigations on the origin/s of the groundwater mineralisation are continuing. The results clearly indicate that the spring-water does not currently carry sufficient mineralisation for fossilisation.

Introduction

Florisbad spring is located 49 km north-west of Bloemfontein and is located on the eastern boundary of the Western Free State panveld. The topography of the area is slightly undulating with occasionally washes from infrequent runoff. The residence at Florisbad is located on the op of a lunette, which has been formed from aeolian sand deposited by the prevailing north-west wind. Drainage is from south to north with a vlei draining to the north from the third swimming pool. The northern tip of the farm incorporates a part of an extensive salt pan, Soutpan.

The 500 mm isohyet passes slightly to the east of Florisbad with an average 78 year rainfall of 496 mm. Annual rainfall is extremely variable with a maximum of 957 mm in 1988 and a minimum of 271 mm in 1965. The flow rates of the spring, as given in the literature, are possibly not very reliable and vary from 18.8 m^3/h (Grobler and Loock, 1988) to 159.3 m^3/h (Kent, 1948). It has been suggested that seismic activity has played a role in the flow rate of the spring over time, as well as in the migration of spring eyes. During the September 1912 earthquake at Fauresmith, a new spring eye appeared at Florisbad. Water flow was said to have increased and gas, sand, artefacts and fossils were expelled from the new eye (Anon, 1980).

Loock and Grobler (1988) stated that the basement rocks of the area were of the Ventersdorp Supergroup overlying older granite and gneiss. This basement is in turn overlain by a Permian Age Karoo sequence of the Ecca Group into which dolerite dykes and sills have intruded (Brink, 1987; Loock and Grobler, 1988). It is at such a dolerite intrusion that the Florisbad spring has formed. The surface geology is composed of an unconsolidated mantle of red-yellow and pale bleached aeolian sand of varying depth (Brink, 1987; Loock and Grobler, 1988). There is no outcropping of bedrock formations on the farm, while calcrete horizons that have been exposed through erosion, occur in the vlei draining from the spring site.

No specific research has been carried out on the Florisbad spring aquifer and therefore factors such as size, recharge, storage capacity and abstraction are unknown. Grobler and Loock (1988) postulated that the intake of the Florisbad aquifer was located 30 km north of Florisbad at Basberg where permeable Beaufort sandstone occurred at an elevation of 150 m above Florisbad. The possibility that the intake area of the spring may lie equidistant to the south-east of Florisbad, in the hills north of Bloemfontein, was also suggested by Grobler and Loock (1988). It was calculated that if the water intake area was located at Basberg, the water would have to descend to the contact zone between the Karoo and the Upper Ventersdorp basement rock at approximately 500 m, in order to reach a temperature of 32°C to 33°C, and issue at 29°C (Grobler and Loock, 1988). There, however, appears to be some uncertainty as to the travel time that recharge water could take to travel from the intake area, through the aquifer, to the spring eyes. Grobler and Loock (1988) calculated anything from 160 to 16 000 years, with a probable 1 600, years for the water to travel this distance.

With the exception of intermittent spring-water sampling (Rindl, 1915; Fourie, 1970; Mazor and Verhagen, 1983; Douglas, 1992), a study by Grobler and Loock (1988) on the characteristics and genesis of the spring, and some hdydrological data supplied by Fourie (1970), the hydrological environment at Florisbad has been largely ignored. This is perhaps somewhat surprising considering the amount, and diversity, of palaeontological, archaeological, geological, and other research, that has been centred on the spring and its associated fossil remains (Brink, 1987; Douglas, 1992). The association between the spring-water and fossil remains probably originated from reports of fossil finds when the swimming pools

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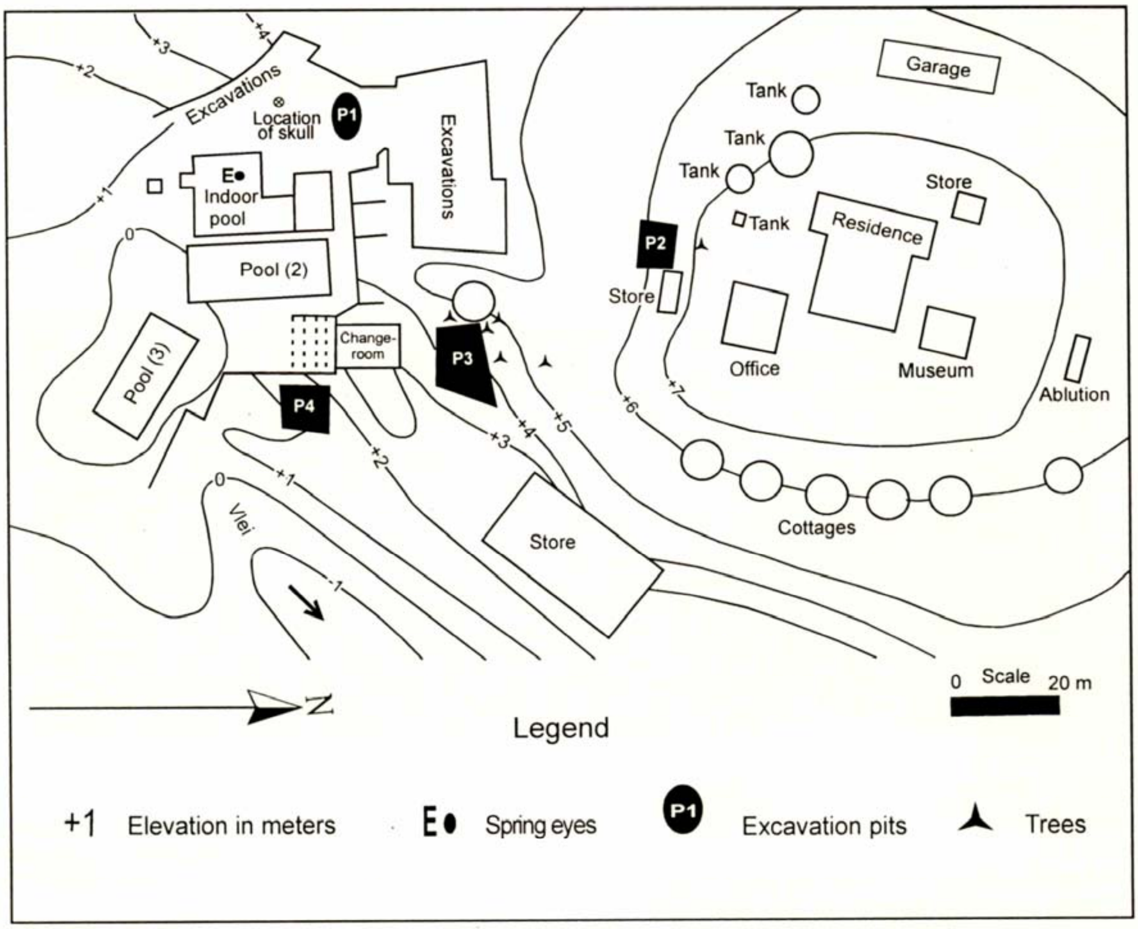

Figure 1
Map of the Florisbad mound showing the water sampling sites

were enlarged in 1912 as well as from fossils expelled during the earthquake of the same year.

In June 1988, various water bodies at Florisbad Research Station were sampled as part of an ecological investigation into the herpetofauna of Florisbad farm (Douglas, 1992). In this study, water quality was examined in terms of saline tolerance, habitat utilisation, and breeding by amphibians (Douglas, 1992). A re-examination of these results showed that the water from one of the exploration pits, located only 22 m from the spring eyes in the indoor pool, had a 27% higher total dissolved solids (TDS). It was thought that the TDS of the pit-water could well have been in a diluted state due to a high recharge from one of the highest rainfall periods recorded. Similarly, during drier periods, the TDS could be increased further by evaporation.

Brouwer (1967) noted that many detailed palaeontological descriptions make no mention at all of the conditions under which the fossilisation has taken place. The purpose of this study was to establish a basis for an ongoing study to determine aspects of the environment in which faunal remains at Florisbad were fossilised and the relationship and effects of that environment on fossilisation. It was felt that further investigations were warranted in order to investigate statements by Brink (1987; 1988) that the slightly alkaline nature of the spring-water was the reason for the good faunal preservation; that the chemical nature of the spring-water had caused remains to be preserved in a characteristic way; and that faunal remains of the Old Collection had became fossilised in contact with the spring water. Some of the other objectives of this preliminary qualitative investigation were to confirm, or refute, the difference between the Douglas (1992) spring and pit-water TDS; to obtain more information on the quality of the pit-waters and to examine a number of factors which may have had an influence on the pit-water values. Yet another important aspect was to provide a background for further investigations into the mineralisation of the pit-waters. As sampling took place over both wet and dry periods some aspects of the possible effect of rainfall could also be examined. Because only sporadic interest has been shown in the spring-water quality over the years it was considered important to add the results of Douglas (1992) (unpublished), and this study, to the published historic record of spring-water quality, as well as to collate some of the existing hydrological information.

Sampling and water quality results

In August 1999 samples of the spring-water (E99) as well as the water in the four exploration pits (P199, P299, P399, P499) were analysed. The locations of the sample sites are given in Fig. 1, where "E" represents the spring eyes in the indoor bathhouse and "P" represents the exploration pits. The year in which the two sets of samples were taken is represented by either "88" or "99". All water samples, including E88 and P188, were analysed by the Institute for Groundwater Studies, University of the Orange Free State, Bloemfontein. In Tables 2 and 3, only factors and ions common to both the 1988 and 1999 analyses have been used in the compilations.

Rainfall figures for the period January 1922 to September 1999 were recorded at either the Florisbad Weather Station or Florisbad Research Station. Missing data was taken from Glen Agricultural College Weather Station records, 32 km east of Florisbad. To illustrate the variation in the occurrence of ions between the spring- and pit-water, the relative ion increases and decreases between the eye and P1, and between the pits, were plotted (Fig. 2). The possible influence of long-term rainfall on the ion concentrations of the spring- and pit-waters were examined from annual rainfall figures plotted for the ten-year period prior to the 1999 and 1988 sampling periods (Fig. 3). Average three-monthly rainfall was also plotted for the 24 months prior to each of the aforementioned sampling periods in order to determine the possible effects of short-term rainfall (Fig. 4). Annual rainfall for the year prior to the 1999 and 1988 sampling periods was then plotted against the total anion and cation concentrations for both the spring- and pit-water (Fig. 5).The 1999 spring-water temperature was taken at a depth of 2.2 m inside one of the eye vent pipes in the indoor pool. Temperatures were also taken at various depths and locations in the indoor pool itself and it was found that the temperature in the indoor pool was 0.9°C to 1.2°C lower than in the vent pipes. However, this range may not be as great in mid-summer when the ambient air temperature is higher.

TABLE 1
Results of the 1999 Florisbad spring and groundwater analyses

	E99	P199	P299	P399	P499
pH	8.91	7.88	7.65	7.66	8.14
p-Alk	7.00	0.00	0.00	0.00	0.00
m-Alk	12.00	171.00	138.00	116.00	92.00
EC (mS/m)	388.00	220.00	325.00	750.00	1 419.00
Salinity*(%)	6.36	3.04	4.71	12.13	24.48
NaCl**(%)	7.55	3.61	5.59	14.39	29.04
Scintillation (CPS)	10-14	9-12	14-17	12-14	15-18
ANIONS (mg/ℓ)					
Chloride	1 442.000	685.000	1 104.000	2 824.000	5 648.000
Nitrite	0.010	0.010	0.010	0.010	0.010
Bromide	19.306	8.462	15.855	36.014	71.964
Nitrate	0.733	1.186	0.901	2.335	4.807
Phosphate	1.929	0.100	0.010	0.010	0.010
Sulphate	1.500	2.600	6.600	2.700	81.600
Fluoride	5.500	9.950	6.030	5.260	10.410
Silica	6.570	7.790	8.240	10.700	0.000
CATIONS (mg/ℓ)					
Ca.	100.000	82.000	136.000	257.000	400.000
Mg	0.440	3.910	5.150	7.510	9.040
Na	784.000	380.000	554.000	1 420.000	2 920.000
K	10.100	8.620	12.900	28.200	42.540
Al	0.054	0.139	0.076	0.038	0.041
Fe	0.014	0.106	0.152	0.058	0.033
Mn	0.003	0.021	0.788	0.464	0.016
Cr	<0.010	<0.010	<0.010	<0.010	<0.010
Zn	0.016	0.026	0.030	0.040	0.034
Cu	0.004	0.008	0.014	0.017	0.017
Cd	<0.006	<0.006	<0.006	<0.006	<0.006
B	2.588	1.735	2.003	6.414	10.158
Sr	2.790	1.890	2.950	6.380	11.200
Co	<0.003	<0.003	<0.003	<0.003	<0.003
As	<0.015	<0.015	<0.015	<0.015	<0.015
La	<0.050	<0.050	<0.050	<0.050	<0.050
Be	<0.010	<0.010	<0.010	<0.010	<0.010
Se	<0.020	<0.020	<0.020	<0.020	<0.020
Li	0.347	0.451	0.272	0.787	2.310
Ba	0.126	0.090	0.167	0.548	0.468
Ni	<0.006	<0.006	0.016	0.018	0.020
Mo	0.002	0.009	0.004	0.003	0.012
V	<0.025	0.026	0.030	0.043	0.058
Ag	<0.010	<0.010	<0.010	<0.010	<0.010
Total ions	2 383.867	1 204.209	1 862.352	4 613.933	9 223.282

* and ** see end of Table 4

Results

The 1999 spring (E99) and pit-water results (P199, P299, P399, P499) are presented in Table 1 and shows the differences in quality between the spring and the pit-water, as well as the difference in water quality between the various pits. The TDS of P199 was 49% lower than E99 while the TDS of P299 was 22% lower than E99. However, TDS increased by 94% for P399, and 287% for P499, as compared to E99. There was also a TDS increase between P199 and P299 of 55%, between P199 and P399 of 283%, and between P199 and P499 of 666%. Of interest was that the increase in the groundwater TDS between P299, P399 and P499 was almost lineal in a west to east direction. Factors possibly influencing these values are discussed further on. These variations in groundwater, in relation to the relatively stable spring-water, would strongly suggest that the pit-waters are not directly related to the spring-water and that the spring- and pit-water could be seen as separate entities.

The most notable ion increase between P199 and P499 was that of SO_4, which showed a 30-fold increase. Other noteworthy increases were Br 750%, Cl 725%, salinity 705%, Na 668%, EC 545%, Sr 494%, B 485%, Ba 420%, Li 412%, and Ca 387%.

Table 2 gives the comparative results between the spring-water E99 and E88, and the pit-water P199 and P188. In order to accentuate any quality changes, the percentage increase, or decrease, of ions and other factors is also given. From the TDS it was clear

TABLE 2
Variations in water quality between the Florisbad spring-water (E99 and E88) and pit 1 groundwater (P199 and P188) over a ten-year period

	E99	E88	% Inc/ Dec	P199	P188	% Inc/ Dec
pH	8.91	9.34	-5	7.88	7.01	+12
p-Alk	7.00	11.25	-38	0.00	-	-
m-Alk	12.00	23.72	-49	171.00	119.46	+43
EC	388.00	398.00	-3	220.00	493.00	-55
Salinity*(%)	6.36	5.96	+7	3.04	7.34	-59
NaCl**(%)	7.55	7.07	+7	3.61	8.73	-59
ANIONS (mg/ℓ)						
Chloride	1 442.000	1 361.32	+6	685.000	1 670.81	-59
Nitrite	0.010	0.00	+1000	0.010	0.00	+1000
Bromide	19.306	4.34	+345	8.462	5.05	+68
Nitrate	0.733	0.24	+205	1.186	0.44	+170
Phosphate	1.929	0.00	+192993	0.100	0.51	-80
Sulphate	1.500	1.55	-3	2.600	24.41	-89
CATIONS (mg/ℓ)						
Ca.	100.000	98.360	+2	82.000	168.570	-51
Mg	0.440	0.502	-12	3.910	5.736	-32
Na	784.000	725.060	+8	380.000	903.330	-58
K	10.100	9.110	+11	8.620	12.460	-31
Al	0.054	0.150	-64	0.139	0.150	-7
Fe	0.014	0.215	-93	0.106	3.999	-97
Mn	0.003	0.012	-75	0.021	1.141	-98
Cr	0.010	0.124	-92	0.010	0.114	-91
Zn	0.016	0.181	-91	0.026	0.178	-85
Cu	0.004	0.059	-93	0.008	0.065	-88
Cd	0.006	0.061	-90	0.006	0.050	-88
B	2.588	1.620	+60	1.735	1.807	-4
Total ions	2 362.713	2 202.904	+7	1 173.939	2 798.820	-58

* and ** see end of Table 4

that the spring-water remained relatively stable between 1988 and 1999, with only a 7% increase in ions. Table 2 also confirmed the variability of the pit-water quality, not only over a period of time, but also at the same site, with P1 showing a 58% decrease in ions between 1988 and 1999. Notable from both Tables 1 and 2 was that while the calcium level in the spring-water remained constant at about 100 mg/ℓ, there was a considerable fluctuation of calcium in the pit-water, with P188 being 106% higher than P199. Although some ions showed considerable increases and decreases, these were often from a very low base.

The comparative results between the spring-water E99 and the pit-water P199, and between the spring-water E88 and the pit-water P188, are presented in Table 3. Of particular interest is that not only does Table 3 show the differences between the spring and pit-water quality, but also the fluctuation of pit-water ions either side of the relatively stable spring-water. Contrary to the 27% higher TDS of P188 over E88, the TDS of P199 was 49% lower than E99. Sulphate, magnesium, iron and manganese values were also considerably higher in P188 relative to both E88 and P199. This supported the theory that if the pit-water was directly related to the spring-water, then there should have been a greater parity between the two.

Figure 2 gives the relative number of ions with increased and decreased concentrations between E99 and P199, P199 and P299, P299 and P399 and P399 and P499. In conjunction with Table 1, it will be noted that between 41% and 53% of ions increased their concentrations relative to the previous sample. With the exception of the slight jump in ions with increased concentrations between E99-P199 and P199-P299, other increases remained constant in a west to east direction. Although there was a 50% decrease in TDS between E99 and P199, 41% of P199 ions had higher concentrations than E99. In most instances these increases were not confined to specific ions and the increases varied between ions and samples. This indicates that TDS does not always give a true picture of water quality, particularly when minor salts are involved.

Average rainfall over the 78-year period was 496 mm, with a maximum annual rainfall of 957 mm in 1988 and a minimum of 271 mm in 1965. Even over this short time span, extremes in wet and dry periods are apparent at Florisbad. Annual rainfall for the ten-year period prior to the 1999 and 1988 sampling is presented in Fig. 3. Despite a high of 957 mm being recorded within the 1988 ten-year period, the pre ten-year average was 459 mm, while the pre-1999 ten-year average was 515 mm, or 12% higher. This higher rainfall corresponded to a 7% increase in ions between E88 and E99. Although this might be indicative of a possible longer-term rainfall effect on the spring-water, it may not necessarily reflect periods of high rainfall within the ten-year periods examined. It may on the other hand reflect influences over even longer or shorter periods, depending on factors such as the size and recharge area of the aquifer, which are not known. Contrary to this increase, the ion levels between P188 and P199 decreased by 58% over the same period, possibly indicating that longer term rainfall had no effect on ion concentrations in the pit-waters.

Rainfall for the 24 months preceding the 1999 and 1988 sampling was plotted as three-month averages (Fig. 4) and the results found to be contrary to those obtained over the ten-year periods. The 1988 24-month average rainfall was 35% higher than the 1999 period, while the 1988 12-, 9-, 6-, and 3-month averages were 73%, 93%, 433% and 2 150% higher, respectively, than for the corresponding 1999 periods. In order to illustrate the effect, and relationship, between rainfall and ion concentrations, the 1988 and 1999 annual rainfall figures were plotted against total ions for E88 and P188, and E99 and E199 (Fig. 5). Figure 5 shows that the 42% decrease in annual rainfall between 1988 and 1999 had very little influence on the TDS of the spring-water, with only a slight converse 7% increase in ions between E88 and E99 (see also Table 2). Results also showed that the 73% increase in rainfall between 1999 and 1988 corresponded to a 138% ion increase in P1.

TABLE 3
Variations in water quality between the 1999 Florisbad spring-water (E99) and pit 1 groundwater (P199), and between the 1998 spring-water (E88) and pit 1 groundwater (P188)

	E99	P199	% Inc/ Dec	E88	P188	% Inc/ Dec
pH	8.91	7.88	-12	9.34	7.01	-33
p-Alk	7.00	0.00	-100	11.25	-	-
m-Alk	12.00	171.00	+1325	23.72	119.46	+404
EC	388.00	220.00	-43	398.00	493.00	+24
Salinity*(%)	6.36	3.04	-52	-5.96	7.34	+23
NaCl**(%)	7.55	3.61	-52	7.07	8.73	+23
ANIONS (mg/ℓ)						
Chloride	1 442.000	685.000	-52	1 361.32	1 670.81	+23
Nitrite (NO_2)	0.010	0.010	0	0.00	0.00	0
Bromide	19.306	8.462	-56	4.34	5.05	+16
Nitrate	0.733	1.186	+62	0.24	0.44	+83
Phosphate	1.929	0.100	-95	0.00	0.51	+5100
Sulphate	1.500	2.600	+73	1.55	24.41	+1475
CATIONS (mg/ℓ)						
Ca.	100.000	82.000	-18	98.360	168.570	+71
Mg	0.440	3.910	+789	0.502	5.736	+1043
Na	784.000	380.000	-52	725.060	903.330	+25
K	10.100	8.620	-15	9.110	12.460	+37
Al	0.054	0.139	+157	0.150	0.150	0
Fe	0.014	0.106	+675	0.215	3.999	+1760
Mn	0.003	0.021	+600	0.012	1.141	+9408
Cr	0.010	0.010	0	0.124	0.114	-8
Zn	0.016	0.026	+63	0.181	0.178	-2
Cu	0.004	0.008	+100	0.059	0.065	+10
Cd	0.006	0.006	0	0.061	0.050	-18
B	2.588	1.735	-33	1.620	1.807	+12
Total ions	2 362.713	1 173.939	-50	2 202.904	2 798.820	+27

* and ** see end of Table 4

Conversely, the 42% decrease in rainfall between 1988 and 1999 corresponded to a 58% decrease in ions in P1. This would tend to indicate that the short-term rainfall has a far greater influence on the pit-water ions than long-term rainfall.

The intermittent Florisbad spring-water analyses (Rindl, 1915; Fourie, 1970; Mazor and Verhagen, 1983; Douglas, 1992) over the past 84 years are presented in Table 4 and provide a modern historic account of the spring-water composition and quality. Under Rindl (1915), methane, hydrogen and nitrogen results have been incorporated from Rindl (1916). As previous researchers have examined different aspects of the spring-water, the compilation in Table 4 gives a far broader and more detailed picture of the spring-water composition and quality than would be obtained from the specific studies. The mineral analysis of the sand by Fourie (1970) has been included because some minerals not included in the various water analyses in the sands, and conversely, some minerals in the water analyses were found in the sands. This then also provides a more comprehensive record of the mineralisation of the spring site.

In relation to the fossilisation of faunal remains it is apparent at this stage of the investigation that there is insufficient mineralisation in the spring-water, particularly Ca, $CaCO_3$ and Si (Table 4), for fossilisation. However, on the other hand it would appear that the groundwater may well carry sufficient $CaCO_3$ for this process. This aspect, and the possibility of the spring-water having a historically low mineral content, is being investigated further.

Discussion

Flow rates of the spring given in the literature appear to be inconsistent, irregular and the reliability of the data uncertain. This is due to one or more of the following factors: not being able to ascertain when or where the measurement was taken; the sampling method not being mentioned; and the use of different sampling methods. Although the flow rate of 112.5 m^3/h given by Rindl (1915) was not actually taken by him, and no mention was made from where he obtained this figure, he thought this figure to be exaggerated. The 49.4 m^3/h to 55.9 m^3/h flow rate given by Fourie (1970) was determined from what is presumed to be two of the three pools, although it is indicated on his map that the sampling site was at the outlet of the third pool. Grobler and Loock (1988) state that, based on unpublished data with no mention of who took the measurements or how the measurements were taken, the flow rate increased from 4.5 m^3/h to 18.8 m^3/h after the 1912 eruption of the new spring eye. Kent (1948) gives the highest flow rate of 159.3 m^3/h, but again with no details.

As the monitoring of spring flow and groundwater levels are important for estimating recharge, if the recharge area were known, the most practical site for monitoring both these factors at Florisbad would be the overflow, or discharge, from the third swimming pool. The three pools at Florisbad are interconnected with the accumulated water discharging from the top of the third pool into the vlei to the north. Douglas (1992) noted that there were minor differences in ion concentrations between the indoor pool and pool 3, with a 1.5% TDS increase between the two. The water in pool 3 therefore represents the average accumulation of spring-water from approximately 21 eyes of various sizes, and it may be more correct to use the water from this pool for analysis and flow rates, rather than water from the indoor pool. While Bredenkamp (2000) stated that spring flow corresponded surprisingly well to average rainfall over several years prior to a specific month, the term several years would be relative and dependant on factors relating to intake rainfall, size of the aquifer, abstraction and flow rate, none of which are known for Florisbad. Because there have been no studies, or monitoring of the hydrological environment at Florisbad, the majority of parameters given by Bredenkamp (2000) for effective groundwater monitoring such as, hydrological modelling, ground-

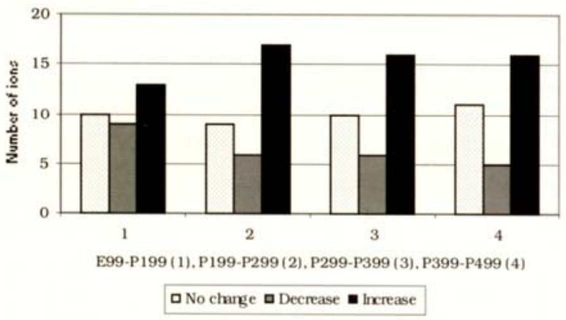

Figure 2
The relative number of ions with increased and decreased concentrations between E99 and P199, P199 and P299, P299 and P399, and P399 and P499

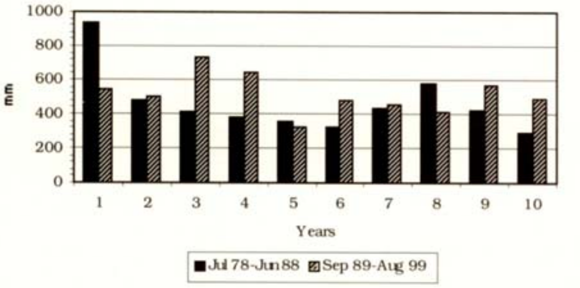

Figure 3
Ten-year average rainfall for Florisbad prior to the June 1988 and August 1999 water sampling

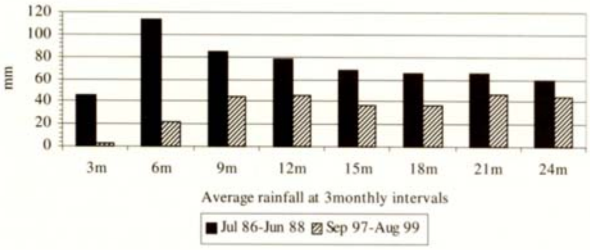

Figure 4
Average three-monthly rainfall for Florisbad for the 24 months prior to the June 1988 and August 1999 water sampling

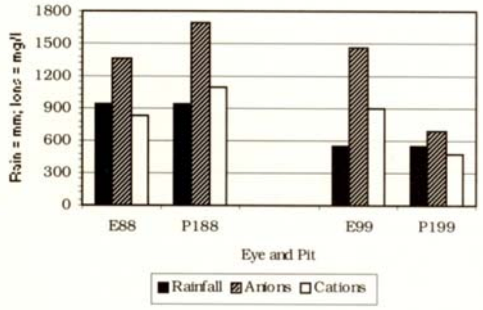

Figure 5
Annual June 1988 and August 1999 rainfall for Florisbad in relation to the anion and cation concentrations of the spring- (E) and pit-water (P1)

water recharge and groundwater management, do not exist. This leaves considerable scope for further research.

Numerous interpretations have been put forward relating to the stratigraphy, sedimentology and formation of the Florisbad mound (Fourie, 1970; Rubidge and Brink, 1985). These interpretations have ranged from the sands having largely originated from the spring (Dreyer, 1938; Brink, 1987; Butzer, 1988), to deposition as a result of sedimentation in a flood plain environment (Butzer, 1984; Visser and Joubert, 1991; Joubert and Visser, 1991), to deposition as a result of largely aeolian processes (Grobler and Loock, 1988; Kuman et al., 1999). In the final analysis it will be found that all these factors will have contributed in some way or another to the formation of the mound. From the above interpretations it is clear that there are very few extensive homogeneous strata, while there are many smaller lens-type layers (Butzer, 1984; 1988; Rubidge and Brink, 1985; Brink, 1987). This interrupted stratification would greatly assist in the percolation of recharge water.

The chemical data of groundwater samples reflect the hydrological interactions that occur from the time of recharge to the time of sampling (Bredenkamp, 2000). Groundwater chemistry is also determined firstly by rainfall and secondly by geochemical interactions that occur firstly in the unsaturated zone as the water filtrates down to groundwater level, and secondly, in the saturated aquifer matrix, and thirdly, when pollutants enter the aquifer (Bredenkamp, 2000). In relation to the Florisbad groundwater, recharge rainfall would gather some salts from the largely aeolian sand layers, which are partly derived from the deflation of Soutpan to the north-west, as well as salts deposited by capillarity at or near the surface. It is not, however, thought that the contribution of these salts could account for the high TDS of the groundwater.

The contribution of minerals from the Ecca shale and dolerite, on which the Florisbad mound and groundwater lies, should also be considered. Dolerite, being largely silicate based, is composed largely of plagioclase, feldspar and a ferromagnesian silicate mineral such as biotite, hornblende, pyroxine or chrysolite (Ford, 1966). Some small mineral contribution may have been made by the dolerite in the form of K, Fe, Mn, Mg and Ca, but this would probably be at low levels. Kent (1949) felt that the high NaCl of the spring-water was more related to the water having percolating down through the underlying Dwyka series, rather than the overlying Ecca series. Any such contribution would have taken place in an environment where the spring-water had been heated to between 32°C and 35°C, as well as possibly having been in contact with the Ecca/Dwyka series for hundreds of years. Over this time span, and under these conditions, the spring-water has still not managed to attain concentration levels near to those of many of the groundwater samples. As the groundwater environment at Florisbad is far cooler, and is also regularly recharged by rainfall, it is felt that diagenesis of the underlying Ecca shale possibly contributes very little to the groundwater concentrations. Should Kent (1949) be correct about the NaCl contribution arising in the Dwyka series, then the mineral contribution to the groundwater by the underlying Ecca shale could be minimal, suggesting some other source.

The determination of recharge using the Cl ratio method has been described by Bredenkamp (2000). Bredenkamp (2000) specified that this method was dependant on the Cl concentrations of rainfall that were sampled in the area being accurately determined, and that no Cl had been added to the water by dissolution of aquifer material, and that no pollution has occurred. The application of this method may be problematic in the Florisbad context, whether it is being applied to the spring-water or the groundwater. At Florisbad there are no Cl values available for rainfall, and because of the high spring Cl values derived from the aquifer material, and the even

TABLE 4
The composition and variation of the Florisbad spring-water over an 84 year period, including minerals identified in the sands

	RD-1999	RD-1988	M&V-1983	GP-1970 Water	MR-1915	GP-1970 Sand
pH	8.91	9.43	-	8.30	-	-
P-Alk	7.00	11.25	-	-	-	-
M-Alk	12.00	23.72	-	-	18.69	-
Alkalinity	-	-	41.00	-	-	-
EC (mS/m)	388.00	398.00	-	-	-	-
Salinity (%)*	6.36	5.96	6.14	6.00	5.80	-
NaCl (%)**	7.55	7.07	7.29	7.12	6.90	-
Hardness – temp. ($CaCO_3$)	-	-	-	177	-	-
Hardness – perm. ($CaCO_3$)	-	-	-	40	-	-
Absorption O2 (4 h at 27°C)	-	-	-	0.08	-	-
Temperature (°C)	29.00	-	29.00	28.88	29.05	-
Density	-	-	-	-	1.0015	-
Scintillation (CPS)	10-14	-	-	-	-	-
ANIONS (mg/ℓ)						
Chloride	1 442.000	1 361.32	1 350.00	1 321.00	1 304.578	-
Nitrite	0.010	0.00	-	0.00	-	-
Bromide	19.306	4.34	-	-	-	-
Nitrate	0.733	0.24	-	0.00	0.380	-
Phosphate	1.929	0.00	-	0.00	Trace	-
Sulphate	1.500	1.55	65.00	16.00	2.282	-
Fluoride	5.500	-	-	6.00	-	-
Iodine	-	-	41.00	0.25	Trace	-
Carbon	-	-	-	0.00	-	⊗
Bicarbonate	-	-	-	49.00	18.690	⦸
Silica	6.570	-	-	-	21.858	⊗
CATIONS (mg/ℓ)						
Ca	100.000	98.360	93.00	87.00	93.424	⊗
Mg	0.440	0.502	7.00	0.00	0.794	⊗
Na	784.000	725.060	800.00	778.00	744.047	⊗
K	10.100	9.110	8.00	8.00	-	⊗
Al	0.054	<0.150	-	7.00	0.854	⊗
Fe	0.014	0.215	-	2.00	0.334	⊗
Mn	0.003	0.012	-	0.00	-	⊗
Cr	<0.010	0.124	-	-	-	-
Zn	0.016	0.181	-	-	-	⊗
Cu	0.004	0.057	-	0.00	-	-
Cd	0.006	0.061	-	-	-	⦸
B	2.588	1.620	-	-	-	-
Ba	0.126	-	-	-	0.737	⊗
Li	0.347	-	-	-	0.693	⦸
Co	<0.003	-	-	-	-	⦸
As	<0.015	-	-	-	-	⦸
La	<0.050	-	-	-	-	-
Be	<0.010	-	-	-	-	⦸
Se	<0.020	-	-	-	-	-
Sr	2.790	-	-	-	-	⊗
Ni	<0.006	-	-	-	-	⦸
Mo	0.002	-	-	-	-	⦸
V	<0.025	-	-	-	-	⊗
Ag	<0.010	-	-	-	-	-
Sb	-	-	-	-	-	⦸
Bi	-	-	-	-	-	⦸
Ge	-	-	-	-	-	⦸
Pb	-	-	-	-	-	⊗
Sn	-	-	-	-	-	⦸
Ti	-	-	-	-	-	⊗

higher Cl values of the groundwater, these would effectively mask any changes in Cl brought about by rainfall. As was pointed out by Bredenkamp (2000), this method is particularly applicable to dolomitic aquifers where the Cl values are low and any increases or decreases in Cl would be easily detectable.

In considering problems experienced with the copious amount

TABLE 4 (continued)						
	RD-1999	RD-1988	M&V-1983	GP-1970 Water	MR-1915	GP-1970 Sand
GASES % (F = mg/ℓ)						
Oxygen	-	-	-	3.6	F0.615	-
Carbon dioxide	-	-	-	-	F3.241	-
Sufurated hydrogen	-	-	-	-	F0.275	-
Hydrogen	-	-	-	0.0	10.000	-
Nitrogen	-	-	-	27.4	18.500	-
Methane	-	-	-	6.9	71.500	-
STABLE AND RADIOACTIVE ISOTOPES						
Hydrogen dD (‰)	-	-	-36.50	-	-	-
Oxygen dD^{18}O (‰)	-	-	-6.82	-	-	-
Tritium (TU)	-	-	0.30	-	-	-
^{14}C (pmC)	-	-	0.00	-	-	-
^{13}C (‰)	-	-	0.00	-	-	-
OTHER						
Sodium chloride (NaCl)	-	-	-	-	1 890.815	-
Sodium nitrate ($NaNO_3$)	-	-	-	-	0.521	-
Ammonium bicarbonate (NH_4HCO_3)	-	-	-	-	2.730	-
Lithium chloride (LiCl)	-	-	-	-	4.202	-
Calcium chloride ($CaCl_2$)	-	-	-	-	242.233	-
Calcium sulfate ($CaSO_4$)	-	-	-	-	2.499	-
Calcium bicarbonate $Ca(HCO_3)_2$	-	-	-	-	21.060	-
Magnesium chloride ($MgCl_2$)	-	-	-	-	3.109	-
Barium sulfate ($BaSO_4$)	-	-	-	-	1.256	-
Ferrous bicarbonate $Fe(HCO_3)_2$	-	-	-	-	1.064	-
Aluminum oxide (Al_2O_3)					1.610	-

* Based on the salinity of sea water = 35000 mg/ℓ ** Based on the NaCl of sea water = 29500 mg/ℓ

⊗ Minerals occurring in the sands ⃠ Minerals tested for but not found in the sands

RD-1999 = (This study); RD-1998 = Douglas (1992); M&V-1983 = Mazor and Verhagen (1983); GP-1970 Water and GP-1970 Sand = Fourie (1970); MR-1915 = Rindl (1915; 1916;1928).

of groundwater during the various excavations, and the amount of water issuing from the spring eyes, drilling results have indicated that the occurrence of groundwater in the immediate area may not be as plentiful as might be expected. From Fourie's (1970) tables it was deduced that four out of eleven boreholes were dry up to a depth of 34 m, that six of the holes produced slightly saline water, and that only one bore hole produced very saline water. This would somewhat contradict Grobler and Loock (1988) who stated that the whole area had a high water table from which the water emanates. This is presumably a reference to problems with the groundwater flow in the earlier excavations. From the literature it would also appear that two auger drilling programmes have been carried out at Florisbad. The first must have been prior to Fourie's (1970) study, and the second in 1981 (Rubidge and Brink, 1985). Unfortunately there has been no attempt to correlate and examine data from these two programmes in the light of the bed-rock profile and water table.

The water table in the immediate area of the spring is effectively represented by the level of the groundwater in the exploration pits, and is supported by Brink's (1987) statement that earlier excavations were at times carried out below the water table. Due to human influences over the years, such as the construction of the swimming pools, which included discharge pipes, it is difficult to denote an accurate level to the spring eye. Fourie (1970) recorded two large and seven small eyes in the indoor pool, two large and seven small eyes in the first outdoor pool, and one large and three small eyes in the second outdoor pool. For the purpose of this study the zero level of the spring eye has been taken as the level of the water in the indoor pool. The depth of the overburden and depth to the water table, as given in the literature is often of little value because the exact locality of where the measurements were taken are not known. In 1982, Butzer (1984) put the water table at 1.7 m above the basement and about 390 mm below Peat II.

Taking the indoor pool spring eyes as zero, the water table in P1 was 200 mm higher than the spring-water; in P2, 300 mm higher; and in P3, 500 mm higher. In P4 the water table was 450 mm lower than the spring-water. This was an indication that the water table was at a fairly even level over most of the area, rising slightly to the north and east, and then dropping to the south. It will be noted from the contours in Fig. 1 that it is only between P3 and P4 that there is any real correlation between the water table and the topography. The water table would therefore appear to correspond more to the bedrock profile, rather than the topography. As the water table is currently almost level with the top of Peat II in pits P2 and P3, in relation to Butzer's (1984) observations, the water table is now approximately 650 mm higher than in 1982. The pre-1982 average ten-year rainfall, which was 524 mm, was higher than either the pre-1988 or pre-1999 ten-year average rainfall, may be yet another indication that long-term rainfall has little influence of the level of the water table. Bredenkamp (2000) stated that groundwater level was linearly related to average rainfall over a number of preceding years.

It could well be expected that, in an area of less than 1000 m^2, that the TDS of the groundwater would be relatively uniform particularly if there was any influence from the stable spring-water. With variations of up to 666% between P199 and P499 it appeared that some other factor(s) might be involved, and the possible effect of evaporation from the pits, coupled with an associated concentration of ions, was considered. Evaporation from P299 and P399 was not thought to be a factor as P299 was a narrow excavation, 6.70 m deep, and largely protected from influences such as wind and extended periods of direct sunlight, although some evaporation would have still taken place. While P399 was shallower at 4.20 m, it was shaded by very large eucalyptus trees, and covered with iron sheeting just above the water table.

A siphoning effect from the eucalyptus trees was considered a possibility. The groundwater TDS increase between P299, P399 and P499 was, however, almost lineal with no anomaly evident at P399. The number of individual ion concentration increases and decreases between these pits also remained constant (Fig. 2). This would tend to indicate that the trees did not exert any undue influence on the groundwater TDS. Pit P499, with its high TDS, was shallower at 2.95 m, and the sides of the pit had eroded to form a very exposed, open basin-shaped excavation. This pit was directly exposed to the sun for long periods of time, but there was no influence from large trees. Evaporation, as a cause of the high values in P499 appeared to be a distinct possibility. However, the water table in P199, in the main excavation area, was even shallower at 450 mm, with erosion again having resulted in a very exposed open basin-shaped excavation. Although this was the most exposed of all the pits, contrary to P499, the TDS in P199 was the lowest of the four pits. Based on these considerations it was felt that neither evaporation nor a siphoning effect from the trees were factors in influencing the TDS of the groundwater in the pits.

The higher SO_4 and Cl values in the groundwater of a rural environment such as Florisbad could possibly be interpreted as an indication of pollution. However, Fig. 2, in conjunction with Table 1, shows that ion increases were not confined to SO_4 and Cl alone. Between 41% and 53% of ions showed an increase in their individual concentrations between the pits from west to east. The springs at Florisbad were closed to the public in April 1980 when they were handed over to the National Museum for research purposes. Therefore, for the past 20 years the influence and effect of rural activities on pollution has been minimal, with a permanent manager and visiting researchers being the only residents. The toilets in the change-room near P3 and P4 were seldom, if ever, used during the first 10 years of occupation by the National Museum, and have been out of order for the remaining period. Sewage from the residence drains to the north away from the exploration pits. The large number of groundwater ion concentration increases, which cannot all be related to possible pollution, coupled with the SO_4 values being so much lower than the Cl values, would tend to suggest that pollution has had little influence and that some other factor or mechanism might be responsible for the higher pit-water values. The apparent spring-water SO_4 increase between 1915 and 1983 (Table 4) is also disproportionately higher than the Cl increases and not thought of as being pollution-related. Along with the high 1983 F value, the importance of regular water monitoring is emphasised as a tool in confirming or disproving such anomalies.

Based on the water and rainfall analyses it was concluded that there was a definite association between the groundwater ion values and short-term rainfall. The pronounced influence of short-term rainfall may partially be due to the shallow nature of the site, bed-rock-wise, as well as the porosity of the aeolian sand. However in the Florisbad context, the results of this study were contrary to the originally proposed theory that high recharge would dilute groundwater concentrations, as well as being contrary to that of established theories. Established theories state that if TDS of groundwater were low, this was an indication of a high rate of groundwater renewal, while conversely, if the groundwater TDS were high, then the rate of groundwater renewal was low (Bredenkamp, 2000). This theory is also applicable to surface waters (Kruger and Lubczenko, 1994) with the latter two statements both supporting the original theory. However, the results at Florisbad clearly showed that during periods of high recharge the TDS of the groundwater in P1 was higher, and conversely, during periods of low recharge the TDS of the groundwater in P1 was lower. These results were not only in relation to P1, but also in relation to P1 and the spring-water. Why the Florisbad results should be contrary to what appears to be the norm is not known at this stage, but it could indicate that there are some other factors or mechanisms involved. This is also the subject of ongoing investigations.

Results have indicated that long-term rainfall probably has little effect on the ion values in the groundwater. However, long-term effects would be more difficult to evaluate as long-term rainfall would dissipate with the groundwater over a relatively short period, and unless an accumulation effect could be determined, any effects of long-term rainfall would be smoothed out by shorter-term rainfall. The effects of long-term rainfall on the spring-water would be difficult to determine at this stage. Besides the intake area of the aquifer and the flow of the spring being unknown, the travel time for recharge to travel through the aquifer to the spring eyes, is also unknown. In any event, the travel time of water from the intake to the spring eyes would be considerably longer than the ten-year period examined here.

The spatial variability of rainfall through characteristic isolated storms in semi-arid areas could detract from using rainfall figures from distant monitoring sites as data in analytical methods. A comparison of rainfall figures for a 62-year period between Florisbad and Glen Agricultural College, 32 km south-east of Florisbad, showed that the rainfall average was 15% higher at Glen (Douglas, 1992). Bredenkamp (2000) noted that monitoring at the Grootfontein aquifer implied that variability in monthly rainfall was largely homogenised in spite of a large variability in daily rainfall. Despite considerable variations in previously mentioned annual rainfall at Florisbad, variability in annual rainfall will also be homogenised over the long term as illustrated by the 78 year average of 496 mm in relation to the 500 mm isohyet. Although this homogenisation of rainfall results may well be applicable to monthly and annual rainfall, Bredenkamp (2000) still felt that there was an uncertainty and unreliability associated with the use of records from remote monitoring points.

In relation to the large herds of animals that previously roamed the area, the potability of the Florisbad spring-water in relation to quality is of some importance. Smit (1977), for example, noted that Kalahari animals could tolerate water with a TDS of 6 000 mg/ℓ. Kruger and Lubczenko (1994) give the salinity tolerance (approximate to TDS) of drinking water for pigs and milking cows as <3 200 mg/ℓ; dry dairy cows and horses <4 500 mg/ℓ; beef cattle <5 760 mg/ℓ, and sheep 6 400 mg/ℓ. Although the Florisbad spring-water is the most saline of South African spring-waters (2 378 mg/ℓ TDS), the above tolerances would clearly indicate that the spring-water is actually relatively sweet and potable for most animals. In contrast to the salinity of spring-water, after the heavy 1988 rains, water in the salt pan just north-west of Florisbad, registered a TDS of 10 346 mg/ℓ, representing a salinity of 29.56% of sea water (Douglas, 1992). This was the first time in living memory that this much water had accumulated in the pan, and at these levels the Florisbad spring-water would have been far more attractive to the herds of game. Seaman et al. (1991) recorded 197

295 mg/ℓ TDS for the salt pan, or 564% the salinity of sea water, and an NaCl content of 188 000 mg/ℓ. These concentrations are approaching that of the Dead Sea with a TDS of 250 000 mg/ℓ. As no precise locality was given for these samples, the values could possibly only have been obtained if they were taken directly from one of the very saline boreholes supplying the evaporative salt dams, or from one of the evaporation dams itself.

It was postulated by both Fourie (1970) and Butzer (1988) that the saline nature of the spring-water, combined with the steady release of methane gas, was an indication of the saline and carbonaceous facies of the underlying Ecca shale. Coal and carbon bearing layers in the Ecca shale were also given as a possible source of methane gas by Fourie (1970). Grobler and Loock (1998) felt that the Ecca shale was deposited under marine conditions with abundant concretions, many of which had a core of organic material and some pyrite, and that these were possibly the origins of the methane gas. Both Rindl (1916) and Fourie (1970) suggested that the carbonaceous turf and peat layers at Florisbad may also have contributed to the production of methane gas. This theory was however discounted by Brink (1987), who stated that gas had been observed emanating from bedrock during test excavations. Other probable reasons for discounting the possibility of the methane gas being associated with the peat beds are, the clarity of the spring-water and the shallow depth of the bed rock, which would not allow for any significantly thick layers of organic matter in the immediate vicinity of the spring.

Mazor and Verhagen (1983) did not record any of the carbon isotopes ^{14}C and ^{13}C, and stated that the tritium value of 0.3 TU could effectively be regarded as zero. Florisbad was not included in the study by Mazor and Verhagen (1983) on the stable hydrogen and oxygen isotopic composition of adjacent rivers. Results however indicated that the springs examined by Mazor and Verhagen (1983) were recharged either by direct rain infiltration, or were recharged during a possibly cooler climatic period with isotopically lighter rains. Mazor and Verhagen (1983) concluded that all the springs they examined were of meteoric origin and that temperatures, salinity, and stable oxygen and hydrogen isotopic composition showed no correlation with radiocarbon age. It can therefore be presumed that this may be applicable to Florisbad as well and that radiocarbon dating of the water would not be an effective method of dating.

It is apparent from the results presented here, and the review of previous investigations, that there are many areas where further research is required in order to obtain a better understanding of the hydrological environment at Florisbad. For example, there is still a lack of understanding regarding the relationship between rainfall and water quality. The application of the cumulative rainfall departure technique (CRD), as given by Bredenkamp (2000), may provide more information on many aspects of the spring. Explanations to questions such as, why there is such a considerable difference between the water quality of the groundwater over such short distances, why high recharge is resulting in high TDS values when this appears to be contrary to the norm, and what are the possible origins of the high mineral content of the groundwater, are currently under investigation.

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Appendix

II

Salinization of the Florisbad organic layers, clay, and groundwater.

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SALINISATION OF THE FLORISBAD ORGANIC LAYERS, CLAY AND GROUNDWATER

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ABSTRACT

Douglas, R.M. 2001. Salinisation of the Florisbad organic layers, clay and groundwater. Navors. nas. Mus., Bloemfontein 17(1): 1-24 . The salinisation of the organic layers and clay at Florisbad were examined in relation to the spring- and groundwater as part of an ongoing study to determine aspects of the environment in which the fossilisation of faunal remains took place. Previous water analyses at Florisbad showed that the TDS of the groundwater in the exploration pits was up to 287% higher than that of the spring-water and that there were variations of up to 666% TDS between the individual pit-waters. In order to establish the origin of this higher mineralization, Peat II and Peat IV samples were analysed and compared with previous ground- and spring-water results. Results showed the ability of the organic layers and clay, at or above the water table, to attract and carry ions at levels 751% above those of the groundwater, and 954% above that of the spring-water. The groundwater appears to be enriched through miscible displacement with the spring-water and the accumulation of organic layer and clay ions, which are flushed from the aforementioned by periods of high rainfall. There was also as much as a 435% variation in TDS within the same organic layer. The uptake and accumulation of ions by the halophytes, and their subsequent decomposition is considered a primary contributor to the salinisation process. Organic rich layers located above the water table were also found to be mineral enriched and it would appear that this is largely as a result of the presence of halophytes and capillarity. Wind-blown salts from Soutpan and the surrounding area have also contributed to the salinisation of the clays and organic material. These processes may have applications at other fossil sites where clay and organic deposits are found in combination with a relatively mineral rich hydrological environment. In order to understand the processes involved theories on the formation and current state of the Florisbad organic layers, the salinisation of clay and organic material, and a number of alternative perspectives on the organic layers are discussed. **(Florisbad, salinisation, organic material, clay, groundwater, spring-water, fossilisation)**

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CONTENTS

INTRODUCTION

The spring- and groundwater quality

A number of intermittent spring-water analyses over the years (Rindl 1915, 1916; Fourie 1970; Mazor & Verhagen 1983; Douglas 1992; 2001) have shown that the Florisbad spring can be classified a warm water mineral spring with the highest NaCl content (2 226 mg/l NaCl) of all South African mineral springs. Despite Van Zinderen Bakker (1989) stating that considerable attention had been paid to the hydrology of the spring site, Douglas (2001) felt that, with the exception of these various spring-water analyses, very little attention has been paid to the hydrological environment at Florisbad. The above analyses have also shown that the mineral content of the spring-water has remained fairly constant and stable over an 84 year period, with slight increases in Cl and Ca, and Na remaining fairly constant (Douglas 2001). Contrary to this, Douglas (2001) showed that the TDS (total dissolved solids) of the groundwater in the exploration pits was up to 287% higher than that of the spring-water with variations of up to 666% TDS between the individual pit-waters.

Douglas (2001) pointed out a number of factors which indicated that the salinisation of the pit-waters was not directly related to the mineralization of the present day spring-water and that some other factor/s or mechanism/s might be involved. Due to a very heavy 1988 rainfall period, Douglas (1992) felt that the high TDS recorded in the pit-water (P188=2 798 mg/l) may well have been in a diluted state due to heavy recharge, while in drier periods the TDS could increase even further due to evaporative concentration and other factors. This was supported by Bredenkamp (2000) who stated that high recharge usually resulted in low TDS, while low recharge usually resulted in high TDS. However, changes that were occurring in the Florisbad groundwater were found to be contrary to the above, as was shown by the considerably drier 1999 rainfall period where a 58% lower TDS was recorded in the same pit-water (P199=1 174 mg/l) (Douglas 2001).

During the drier 1999 period the TDS of the same pit-water (P1) was also 50% lower than that of the spring-water (E99=2 363 mg/l) (Douglas 2001). However, some of the other groundwater TDS (P399=4 614 mg/l and P499=9 223 mg/l) were still much higher than that of the spring-water, with up to a 666% variation in TDS between pits (Douglas 2001).

From the rainfall figures for these periods it was concluded that short-term rainfall had a decided effect on ion concentrations in the pit-water, with high rainfall resulting in a higher pit-water TDS, and low rainfall resulting in a lower pit-water TDS (Douglas 2001).

Other factors that may have had an influence on TDS variations in the groundwater were also examined and effectively ruled out (Douglas 2001). These included the high SO_4 and Cl values which might have suggested some type of pollution, the possible contribution of minerals from the underlying Ecca shale and dolerite, the possible effect of evaporation, and a possible siphoning effect from nearby trees (Douglas 2001).

The higher and lower TDS of the pit-waters in relation to the stable spring-water, the variation in TDS within a specific pit-water, and the variation in TDS between pit-waters, were the primary factors for considering that the salinisation of the groundwater must be related to a source other than the spring-water (Douglas 2001). Another indication that the salinisation of the groundwater was possibly not related to the spring-water, was that, despite Fourie (1970) finding Peat I to be calcareous, Butzer (1988) noted that there was no calcium carbonate in his 1982 excavation profile, and that the spring-water was under saturated in calcium and bicarbonate. This was supported by the results of Douglas (2001). Although the presence of fossil remains support the presence of calcium carbonate, Butzer (1988) conclude that any such carbonate enrichment must derive from either a direct, or indirect, external sources.

Theories on the formation of the Florisbad organic layers

Theories on the formation, stratigraphy and sedimentology of the Florisbad spring site, including the organic layers, are well documented in the literature. Dreyer (1938) suggested that Florisbad deposits represented the sand output of the spring, while Butzer (1988) felt that the quartz grains represented detrial sands released from the underlying Ecca shales and subsurface dolerite. Van Zinderen Bakker (1989) partly supported this theory by noting that SEM (Scanning Electron Microscope) studies showed that sand from the 2.15 m and 4.00 m levels was not of an aeolian nature, but that the surface sands were. Grobler & Loock (1988a, 1988b) felt that deposition was largely as a result of an aeolian process, which was supported by Joubert & Visser (1991). SEM analysis of sand samples by Kuman *et al.* (1999) seemed to confirm that the spring sediments were derived from an aeolian source.

Dreyer (1938) felt that alternating increases and decreases in the spring flow were responsible for the stratification of the organic layers and sand layers. He considered a decreased flow from the eye as allowing for luxuriant plant growth, while an increased flow hindered growth by covering the plants with sand. Brink (1987) saw the spring site developing from sand being brought to the surface from a spring vent which became enlarged due to chemical and mechanical actions, with vegetation developing around the margins of the spring pool. He then saw the vent becoming blocked and the size of the mound increasing due to a combination of factors such as, choking vegetation, the deposition of windblown sediments and a diminishing supply of, what he referred to as, groundwater. Brink (1987) felt that after this closure the spring may have moved laterally along the bedrock fissure to form a new passage which may have cut through existing strata to form another sand unit which would have partly overlapped the mound created by the original vent. These processes would, he contended, have repeated themselves to form alternate layers of organic and non-organic deposits such as those found at Florisbad.

However, the flooding of Soutpan, a large salt pan lying aproximately 1.5 km north of Florisbad, and the associated flooding of the spring site, appears to be one of the more popular theories for the formation of the organic layers. Butzer (1984) postulated that sedimentation had occurred in a flood plain environment and suggested that Peat I had a similar semi-aquatic accumulation as Peat II. Van Zinderen Bakker (1988) agreed with this, stating that the intervals in the organic layers were due to rises in the level of water in the pan (Soutpan) and the water table, and that during stable spring production, vegetation around the spring produced organic deposits, the so-called peat layers.

In similar vain, Visser & Joubert (1990) suggested that sedimentation and soil horizons formed during low lake (Soutpan) levels with the deposition of shoreline sands and lake bottom silts occurring during wet periods and the resultant flooding of the spring area. Joubert & Visser (1991) felt that the organic-rich waxy clay formed in a marsh along the lake margin during transgressive phases, presumably due to rises and falls of the Soutpan waters. They further stated that sedimentation at Florisbad was cyclical and directly related to changes in climate over the period, representing fluvial, shoreline, or aeolian deposits, and suggested that sedimentation occurred during three high and four low phases of the nearby lake complex (Soutpan).

State of the present day organic layers at Florisbad

There appears to be little contention in the literature that "peat" has been used as a convenient term for describing the four organic rich horizons at Florisbad. These horizons were described by Van Zinderen Bakker (1988) as so-called 'peat' layers of carbonaceous clay and silt, and by Butzer (1988) as organic horizons where the 'peat' lacked definition in terms of interwoven vegetation structure. The quantities of sand deposited within these layers would also tend to make the term "organic rich" more appropriate. The lack of interwoven vegetation structure, as noted by Butzer (1988), was particularly noticeable in Peat II, but was very high in Peat IV. The author supports the point of view that the "peat" layers should rather be seen as layers of organic rich material, clay and sand. A simplified stratigraphic profile of a section of the western wall of the Florisbad excavation is presented in Figure 1 and shows the main organic horizons (after Kuman & Clarke 1986).

Various authors have described the state of the present day organic layers. Fourie (1970) mentioned a thin basal peat layer resting directly on the dolerite. Peat I, which is located below the water table was described by Dreyer (1938) as being bituminous and by Meiring (1956) as waxy looking and free of modern tree roots. Fourie (1970) described Peat I as being waxy with scattered pieces of decomposed wood where the fibrous structure could still be identified.

Peat II is currently at the same level as the water table in pits 2, 3 and 4. Meiring (1956) described Peat II and III as being heavily contaminated with modern tree roots. The Peat II sampled in this study did not, however, reflect any heavy contamination of modern tree roots, vegetation or fibrous structure. Butzer (1984, 1988) saw Peat II as an organic, very dark grey sandy loam layer, interbeded with dark grey loam and pockets of light grey loam. Organic material comprised what Butzer (1984, 1988) referred to as vegetation structures

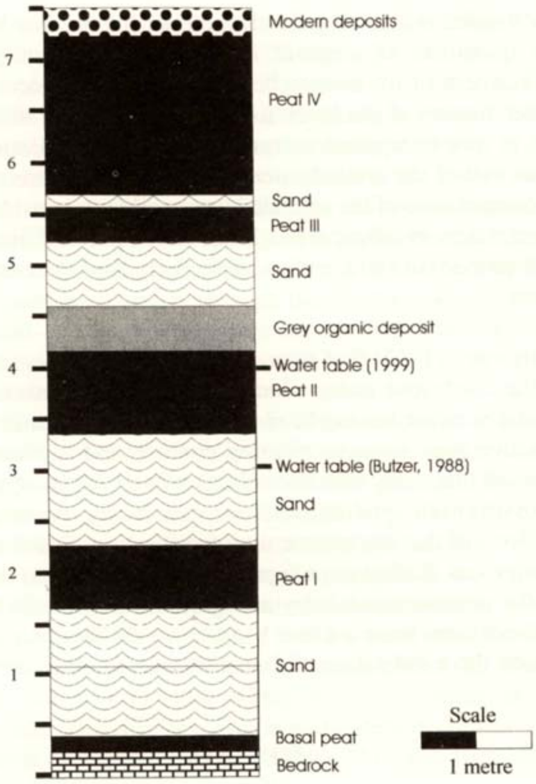

Figure 1: A simplified stratigraphic profile of a section of the western wall of the Florisbad excavation showing the main organic horizons (after Kuman & Clarke 1986).

with carbonised and fine vertical roots, and a lack of interwoven vegetation structure. The Peat II examined in this study was of a grey sandy/clay nature with a very fine black clay fraction which had the consistency of light grease. The effect of compression due to the vertical weight of the sand is evident in Peat II in that the fine black clay fraction is being squeezed out from between the sand and onto the wall of the pits. However, this oozing of the fine clay fraction may also be partly due to swelling pressure caused through the absorption of water by the clay. Load structures, mentioned by Brink (1987), are yet another aspect and indication of this compaction.

Dreyer (1938) and Fourie (1970) both noted that the organic layers became progressively more sandy towards the surface. Peat IV was described by Butzer (1984, 1988) as organic, grey to dark grey loam to loam clay, comprising abundant organic voids and abundant root casts, as well as black organic muck of semi-aquatic faces. Joubert & Visser (1991) described Peat IV as clayey silt, which contained the remains of reeds, disseminated carbonate and calcareous nodules. Peat IV, on the western wall of the excavation, is located just 650 mm below the soil surface and is 2.8 m thick.

There are many variables influencing descriptions of the organic layers. These include the composition and quantities of organic matter, clay and sand; the thickness of the overburden; the thickness of the section being examined; the location of the section being examined, the water content of the layer; as well as many other unknown historical factors. These factors can in turn be applied to variations in the salinisation of the organic layers and clay as well as that of the groundwater. It is therefore apparent from the above that the present state and composition of the organic layers will vary considerably, even for samples from the same layer, taken at different localities. In relation to European peat, the Florisbad organic layers in their current state, and in particularly the lower organic layers, can hardly be classified as peat.

It was noted by Brouwer (1967) that many detailed palaeontological descriptions make no mention at all of the conditions under which fossilisation has taken place. This is probably due to these conditions never having been examined in any detail. Although Brink (1987) stated that fossilisation took place in a spring context, and examined a number of related aspects, the purpose of this study was to establish a basis for an ongoing study to determine aspects of the fossilisation environment not previously investigated, as well as the relationship and effects of that environment on fossilisation. Based on the above findings, it was decided to carry out further investigations on the possible origin of the higher ion concentrations in the organic layers, clay and groundwater. It was also felt that the results of this study could not only have a direct bearing on the fossilisation of faunal remains at Florisbad, but also on the fossilisation of faunal remains at other sites.

METHODS

Although the so-called "peat" layers are considered here as organic rich layers, due to the fact that these layers have been previously designated as Peat I, II, III and IV in nearly all publications, this terminology has been continued here. The location of Peat samples (X), the exploration pits (P), and the spring eye (E), are given in Figure 2. Peat II samples X1, X2 and X3, as well as Peat IV samples X4 and X5, were all taken in September 1999, while Peat sample X6 was taken in November 1999. Peat II, sample X1, was located at the base of the north wall of the main excavation above the water table; sample X2 at the level of the water table in Pit 2 (P2); and sample X3 at the level of the water table in Pit 4 (P4). Peat IV, sample X4, was taken from the top third of the layer on the west wall of the excavation, and sample X5 from the bottom third of the same layer above the water table. Sample X6 was taken at the same locality as X2.

As no defined Peat III layer was observed during this study, Peat III has not been included. The exclusion of Peat III is justified by the fact that according to Kuman & Clarke (1986) it comprises up to 95% sand and is often difficult to distinguish due to variations in the amounts of organic matter. With the sand layer between Peat IV and Peat III often being less sandy (75%) than Peat IV, as well as being as narrow as 50 mm in places (Kuman & Clarke 1986), it is possible that the two layers might merge in certain localities.

Peat samples X1 to X5 were analysed by the Department of Soil Science, University of the Free State, using Atomic Absorption and Spectrometric analysis. Peat sample X6 was analysed as a water extraction sample by the Institute of Ground Water Studies, University of the Free State, using Inductively Coupled Plasma Optical Emission Spectrometry (ICP).

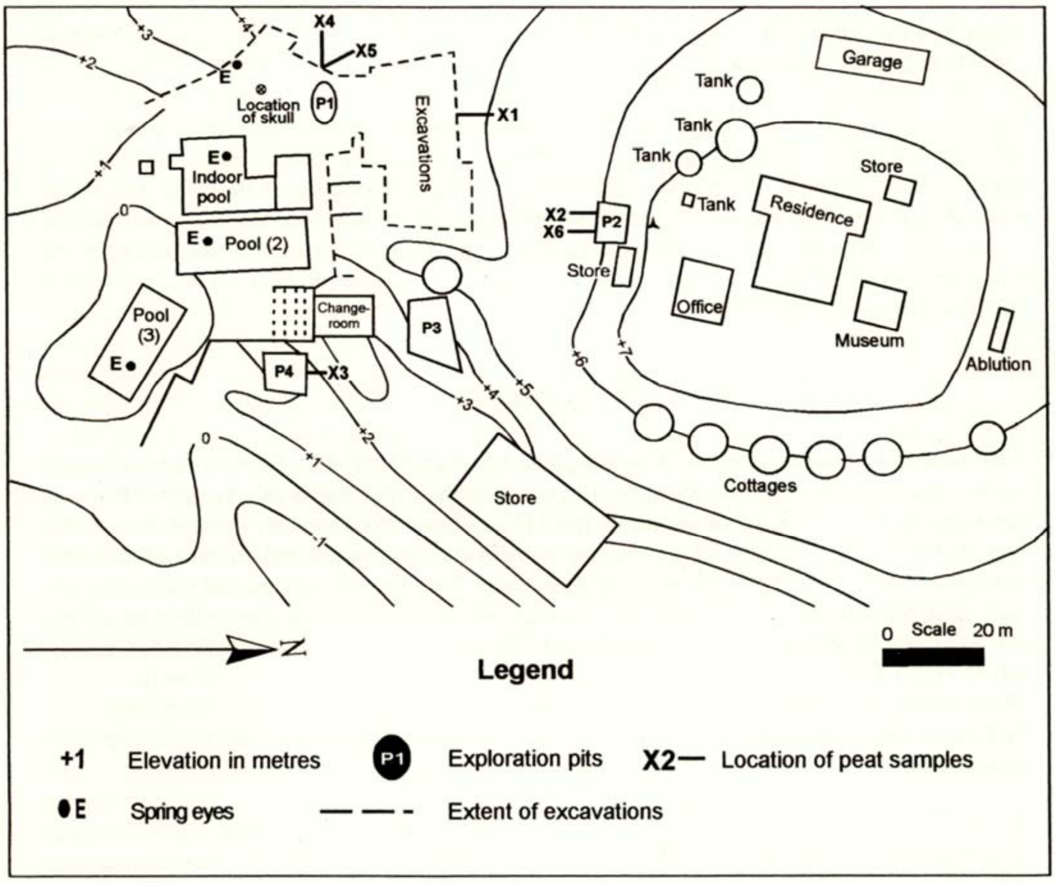

Figure 2: Map of the Florisbad spring site showing the excavations, peat sample sites and exploration pits.

All the 1988 and 1999 water samples referred to were analysed by the Institute of Ground Water Studies, using ICP analysis. For comparative purposes only, ions from the 1999 water samples (P199, P299, P499 and E99) (Douglas 2001) common to the peat analyses have been used in the tables. For more comprehensive spring- and groundwater water analysis results the reader is referred to Douglas (2001), and for water analysis of other water bodies at Florisbad, Douglas (1992).

X-ray difractometric analysis was carried out by the Department of Geology, University of the Free State, Bloemfontein, on a Peat II sample from Pit 2 as well as on a sample of sand expelled from the spring eyes in the indoor swimming pool.

Estimates of the porosity of Peat IV and the clay content of Peat II were also determined. The porosity of Peat IV was determined by taking three 150 mm x 25 mm cores, dissolving them individually in water, and drying them at 37 °C. The average dry volumes, subtracted from the original core volumes, resulted in an estimated porosity percentage. The quantity of light clay material for Peat II was determined by dissolving three equal peat samples in water and vigorously agitating them in measuring cylinders, allowing the fractions to settle out and then measuring the proportions.

As Cl was not determined as part of the peat analysis, the average Na:Cl ratio was calculated from 15 water analyses at Florisbad (Douglas 1992, 2001). This resulted in an average Na:Cl ratio of 1.828:1 (STD = 0.1095, Average Deviation = 0.0828), which was then applied to the Na values of the peat samples, in order to approximate the Cl values. For comparative purposes, the salinity of samples in the tables are given as a percentage of sea water salinity (SW) at 35 000 mg/l TDS, and the NaCl of a sample is given as a percentage of SW NaCl at 29 500 mg/l.

RESULTS

The results of the peat analyses, X1 through to X5, are presented in Table 1 and are largely self-explanatory. Results clearly show the considerable differences in ion values between the various peat samples. For example, the TDS of Peat II, sample X3 in pit 4, was 435% higher than Peat II, sample X1, on the north-north wall of the excavation, and 60% higher than that of Peat II sample X2, in Pit 2. This lower TDS for X1 was possibly due to a low clay and humus content, as well its location above and beyond the influence of the groundwater. Also noticeable was the higher TDS of 23 438 mg/l for sample X4, from the top of Peat IV, when compared to the TDS of 14 438 mg/l for sample X5 from the bottom of the same layer. These high TDS were of interest because the distance from the top of Peat IV to the groundwater was 3.7 m, while the bottom of the layer was 90 cm above the water table.

A comparison between Peat II samples X2 from Pit 2, X3 from Pit 4, and the 1999 water analyses for these two pits (Douglas 2001) are presented in Table 2. In order to make any differences in quality more comparative and visible, percentage increases, or decreases, have been given. Peat II, sample X1, was excluded from Table 2 as it was located above the water table and therefore no comparative water analysis was available. As was expected, the test water extraction analysis for sample X6 from Pit 2, produced results similar to the P299 water analysis results. This method would appear to be of little value in determining the ion values of the organic layers as it only records ion values of the interstitial soil water. Table 2 also shows the capacity of Peat II to attract and carry ions in excess of the groundwater. The TDS of sample X2 was 751% higher than the corresponding P299 groundwater, while the TDS of sample X3 was 173% higher than the corresponding P499 groundwater. There was also a 435 % TDS variation between Peat II samples X1 and X3. Also noticeable was the lower pH of the peat in relation to the groundwater. From these results it was apparent that the organic layers were in some way responsible for the high salinisation of the groundwater.

In order to illustrate the ion carrying capacity, and difference in quality, of the peat in relation to the spring-water, Peat II samples X1, X2 and X3 are presented together with the 1999 spring-water analysis in Table 3. The percentage increase, or decrease, has again been given in order to make any changes more visible and comparative. Sample X2 and X3, which were taken at the same level as the water table, had a TDS of 560% and 954% respectively, higher than the spring-water. Sample X1, which was above the water table,

Table 1: Results of the Florisbad Peat II and Peat IV analyses.

	X1-Peat II Exc. N. Wall	X2-Peat II Pit 2	X3-Peat II Pit 4	X4-Peat IV Exc. Top W. Wall	X5-Peat IV Exc. Bot. W. Wall
pH	8.30	8.00	6.65	5.20	4.18
Resistance (Ohm)	26.00	38.00	19.00	44.00	88.00
NaCl (mg/l)	2 427.35	11 967.40	19 193.00	16 370.50	6 943.35
Salinity % (of SW)	13.17	44.08	70.44	66.96	41.25
NaCl % (of SW)	8.23	40.57	65.06	55.49	23.54
IONS (mg/l)					
P	22.00	26.00	34.00	21.00	17.00
Cl	1 567.35	7 727.40	12 393.00	10 570.50	4 483.35
Ca	2 008.00	2 768.00	5 000.00	6 720.00	7 360.00
Mg	23.40	24.80	70.00	30.60	22.80
K	129.00	640.00	352.00	294.00	93.00
Na	860.00	4 240.00	6 800.00	5 800.00	2 460.00
Zn	1.20	2.40	6.00	1.40	1.70
Total ions	4 610.95	15 428.60	24 655.00	23 437.50	14 437.85

Table 2: The difference in quality between groundwater sample P299 and Peat II sample X2, between groundwater sample P499 and Peat II sample X3, and between groundwater sample P299 and the water extraction sample X6 from P2.

	P299 Water	X2 Peat II Pit 2	% Inc./ Dec.	P499 Water	X3 Peat II Pit 4	% Inc./ Dec.	X6-Peat II Pit 2 Water Ext.	% Inc./ Dec. Between P299 (Col 2)
pH	7.65	8.00	+4.58	8.14	6.65	-18.30	7.87	+2.88
Resistance (Ohm)	-	38.00	-	-	19.00	-	-	-
EC (mS/M)	325.00	-	-	1 419.00	-	-	276.00	-15.08
Salinity % (of SW)	5.18	44.08	+750.97	25.77	70.44	+173.34	4.58	-2.76
NaCl % (of SW)	5.62	40.57	+621.89	29.04	65.06	+124.04	5.29	-5.37
IONS (mg/l)								
P	0.01	26.00	+259 900.00	0.01	34.00	+339 900.00	0.00	-100.00
Cl	1 104.00	7 727.40	+599.95	5 648.00	12 393.00	+119.42	956.00	-13.41
Ca	136.00	2 768.00	+1 935.29	400.00	5 000.00	+1 150.00	30.00	-77.94
Mg	5.15	24.80	+381.55	9.04	70.00	+673.34	3.00	-41.75
K	12.90	640.00	+4 861.24	42.54	352.00	+727.46	11.90	-7.75
Na	554.00	4 240.00	+665.34	2 920.00	6 800.00	+132.88	603.00	+8.84
Zn	0.03	2.40	+7 900.00	0.03	6.00	+19 900.00	0.07	+133.33
Total ions	1 812.09	1 5428.60	+751.43	9 019.62	24 655.00	+173.35	1 603.97	-11.49

Table 3: The difference in quality between the spring-water and Peat II samples X1, X2 and X3.

	Eye 1999	X1-Peat II Exc. N. Wall	% Inc./ Dec.	X2-Peat II Pit 2	% Inc./ Dec.	X3-Peat II Pit 4	% Inc./ Dec.
pH	8.91	8.30	-6.85	8.00	-10.21	6.65	-25.36
Resistance (Ohm)	-	26.00	-	38.00	-	19.00	-
Salinity % (of SW)	6.68	13.17	+97.16	44.08	+559.88	70.44	+954.49
NaCl % (of SW)	7.55	8.23	+9.01	40.57	+437.35	65.06	+761.72
IONS (mg/l)							
P	1.93	22.00	+1 039.90	26.00	+1 247.15	34.00	+1 661.66
Cl	1 442.00	1 567.35	+8.69	7 727.40	+435.88	12 393.00	+759.43
Ca	100.00	2 008.00	+1 908.00	2 768.00	+2 668.00	5 000.00	+4 900.00
Mg	0.44	23.40	+5 218.18	24.80	+5 536.36	70.00	+15 809.09
K	10.10	129.00	+1 177.23	640.00	+6 236.63	352.00	+3385.15
Na	784.00	860.00	+9.69	4 240.00	+440.82	6 800.00	+767.35
Zn	0.02	1.20	+5 900.00	2.40	+11 900.00	6.00	+29 900.00
Total ions	2 338.49	4 610.95	+97.18	15 428.60	+559.77	24 655.00	+954.31

Table 4: The difference in quality between the spring-water and the Peat IV layer samples X4 and X5.

	Eye 1999	X4-Peat IV Exc. Top	% Inc./ Dec.	X5-Peat IV Exc. Bot.	% Inc./ Dec.
pH	8.91	5.20	-41.64	4.18	-53.08
Resistance (Ohm)	-	44.00	-	88.00	-
EC (mS/M)	388.00	-	-	-	-
Salinity % (of SW)	6.68	66.96	+902.40	41.25	+517.51
NaCl % (of SW)	7.55	55.49	+634.97	23.54	+211.79
IONS (mg/l)					
P	1.93	21.00	+988.08	17.00	+780.83
Cl	1 442.00	10 570.50	+633.04	4 483.35	+210.91
Ca	100.00	6 720.00	+6 620.00	7 360.00	+7 260.00
Mg	0.44	30.60	+6 854.55	22.80	+5 081.82
K	10.10	294.00	+2 810.89	93.00	+820.79
Na	784.00	5 800.00	+639.80	2 460.00	+213.78
Zn	0.02	1.40	+6 900.00	1.70	+8 400.00
Total ions	2 338.49	23 437.50	+902.25	14 437.85	+517.40

and contained far less organic and clay material than either X2 or X3, reflected this in having a TDS only 97% higher than the spring-water.

The quality between the two Peat IV samples, X4 and X5, and the spring-water are presented in Table 4. From this table it can be seen that samples X4 and X5, both of which were well above the water table, had ion concentrations 902% and 517%, respectively, higher than the spring-water. Also of note was the lower pH of the Peat IV samples, relative to both Peat II and the spring-water.

The x-ray difractometric analysis on Peat II in Pit 2 showed that it was composed of quartz, feldspar and clay, with the clay fraction comprising illite and montmorillonite. The light clay fraction comprised 46% of Peat II, while the heavier clay fraction, quartz and feldspar, accounted for the remaining 54%. X-ray difractometric analysis on the accumulated sand from the spring eye showed the clay content to be 0%, quartz 85% and feldspar 8-10%. Other minerals included rutile, ilmenite, magnetite, zirconium, garnet, tourmaline, epidote, pyribole as either amphibole or pyroxene, fossilised wood fragments and fossilised seed. A quantity if fossilised teeth and bone was also recovered from the sand of the spring eye.

DISCUSSION

The salinisation of the clays and organic material

Douglas (2001) noted that there must be some other factor/s or mechanism/s responsible for the salinisation of the groundwater. The results of this study support this view and indicate that it is the salinisation of the clay and organic material which is largely responsible for the salinisation of the groundwater. The attraction of ions to clay particles and organic matter, and their resultant concentration, is a complex mechanism that has been explained in various ways by different authors. In light of that which this study proposes, it is important to briefly examine a few of the mechanisms and processes involved in the salinisation of the clay and organic matter in order to substantiate the results.

It is apparent from the variation in the TDS of the pit-water, in relation to the stability of the spring-water TDS, that there is a constant process of miscible displacement between the two hydrological environments. Two influential processes which occur simultaneously during miscible displacement are, diffusion, caused by the intrinsic movement of molecules from areas of high to low concentration, and hydrodynamic dispersion, caused by the continually changing direction of a liquid through its collision with pore walls (Taylor & Ashcroft 1972). However, these processes will in turn also be influenced by factors such as pore size and the resultant degree of restriction to the flow of the water, with an osmotic effect developing when this flow becomes restricted (Taylor & Ashcroft 1972). Miscible displacement will also be influenced and affected by variations in rainfall.

This and other previously mentioned studies, clearly show that the Florisbad organic layers comprise clay, humus, sand and minerals. Brady (1984) saw the colloidal clay fraction falling into two basic groups, inorganic clays comprising silicate clay minerals, and organic clays comprising humus. Silicate clay and organic matter are extremely important soil components in that they constitute the chemically active portion of the soil (Taylor & Ashcroft 1972). Not only does the clay fraction designate a range of particle sizes, usually smaller than 2μ, but also designates a large group of minerals which occur as highly structured microcrystals of submicroscopic, or colloidal size (Hillel 1971, Taylor & Ashcroft 1972). The nature of silicate clay, unlike sand, is highly dynamic, and may change considerably in response to changes in environmental conditions, with most of the physical and chemical changes taking place between the colloids and the solution (Hillel 1971; Taylor & Ashcroft 1972). Silicate clay particles may have internal surfaces formed between the plate like crystals, or multiple-stacked composite layers, and these internal surfaces may exceed external surfaces (Hillel 1971; Blatt *et al.* 1972; Brady 1984). Montmorillonite clay, found in the Florisbad soil, has one of the largest specific surface areas due to its lattice expansion and resultant exposure of internal surfaces (Hillel 1971). Montmorillonite also characteristically absorbs water between successive layers causing the minerals to expand in the *c*-axis, while illite does not absorb interlayer water and is therefore nonexpanding (Taylor & Ashcroft 1972).

All solid substances have unsatisfied bonds at their surfaces with the number of bonds being particularly large for amorphous and crystalline materials with a high surface area (high surface/volume ratio) (Blatt *et al.* 1972). Micelles, or minute silicate clay colloids, are negatively charged, attracting hundreds and thousands of positively charged ions to each colloidal crystal (Brady 1984). The clay particle is thus able to attract and hold an enormous numbers of cations on its negatively charged surface (Brady 1984). Clay minerals and organic matter therefore have the capacity to adsorb on their outer and inner surfaces large amounts of ions that may be supplied by percolating waters (Blatt *et al.* 1972), and therefore interact in miscible displacement.

Silicate clays serve as a resin with the exchange of ions being serviced by the flow of water. The attraction of ions to clay was described by Blatt *et al.* (1972) as salt sieving, and the following are some of the salient points made by him. Illite structures, for example, have internal charge deficiencies that are balanced by the absorption of exchangeable cations. The rate of water flow through the clay layer, and thus the rate of ion deposition, will be influenced by the depth of the layer and its degree of compaction. Increased compaction results in fixed negative charges on adjacent clay particles becoming so close together that anions in solution on the underside of the membrane are repelled and must be retained there. Adsorbed cations are still free to move between adjacent exchange sites and through the membrane, but this latter movement must be compensated by an electrically equivalent inflow from the reverse direction. This movement also results in hydrogen, which is the most easily diffusible cation, to diffuse to the underside of the membrane, creating a pH difference between the underside and topside of the membrane. The significance of this is that there is most probably a change in subsurface pH, which would continually effect the solubility of minerals, particularly carbonates.

Butzer (1988) noted that the pH values in his cross sections showed no systematic correlation to the presence of organic horizons through the Florisbad deposits. He suggested that the pH curve was a long-wave response to fundamental environmental parameters such as variations in vegetation cover, precipitation and temperature. However, the resultant variable pH changes caused by the previously mentioned movement of hydrogen ions between adjacent exchange sites through the clay membrane, and particularly in the humus, may better explain why there was no correlation between the pH values and the organic layers. The variable pH values in the tables, for both the peat and groundwater, would tend to support this.

Another significant aspect of the ionization of clay is that, relative to other structural forces, the bonding forces between interlayer cations and the site of the charge deficiency is largely ionic, nondirectional, and therefore weak (Blatt *et al.* 1972). Blatt *et al.* (1972) concluded that in many instances clay beds are essentially pure deposits of clay minerals, and that due to the relative weakness of the structural unit bonds, percolating waters might easily remove interlayer ions during diagenisis and weathering. This would also support the theory that recharge rainfall flushes ions from the mineral rich clay and organic layers into the groundwater resulting in an increased TDS, rather than rainfall having a diluting effect which would result in a lower TDS (Douglas 2001).

Organic soil colloids, or humus, have a similar colloidal organisation to that of clay, with a highly charged anion being surrounded by a large number of adsorbed cations (Brady 1984). However, the negative charge on humus colloids is very much pH dependant (Brady 1984). Hydrogen, which is tightly bound under strong acid conditions, is not easily replaceable by other cations, resulting in its adsorptive capacity being very low at low pH (Brady 1984). Rises in pH allow the hydrogen to be replaced by other cations, such as calcium and magnesium, and under these more alkaline conditions the absorptive capacity of humus far exceeds that of layer silicate clays (Brady 1984).

The result of these processes is therefore a continual diffusion of ions between the spring- and groundwater, when the groundwater ion levels are low, and a converse diffusion of ions from the groundwater to the spring-water away from the eyes, when ion levels in the groundwater are high. These processes are in turn influenced by the flushing of ions from the clay and organic material during periods of increased rainfall and recharge, and conversely by the increased adsorption of ions by the clay and organic material during periods of decreased rainfall. This would explain why the TDS of the groundwater, in relation to recharge, is contrary to the authors original proposal as well as that of Bredenkamp (2000). It is clear from the spring eye water analysis that these processes have little effect on the water from the exposed eyes per se, with the strong spring flow possibly masking any changes. Therefore, any changes in the quality of the spring-water will largely be limited to subterranean spring-water in a zone some distance from the eyes.

The ability of halophytes to remove salts from both soil and water is so considerable that the application has, and is, being researched in many areas. Many researchers have examined the nutrient uptake of littoral plant communities (Dykyjová, 1978), the treatment of wastewater such as effluent (Lakshman 1979: Tildon & Kadlec 1979) and the control of eutrophication due to fertilizers and phosphorus-based insecticides (Tóth 1972). Constructed wetlands in bioremediation research has been used in diverse areas such the purification of dairy wastewater (Davis *et al.* 1992), mine wastewater (Noller *et al.* 1994), domestic wastewater (Juwarkar *et al.* 1995), swine wastewater (Hunt *et al.* 1993) and pulp mill wastewater (Hatano *et al.* 1994). Other areas of research include salt bioaccumulation (the removal of salts by halophyte species) in reclaiming saline soils, studying select halophytes that accumulate salt in their tissue for harvesting, and biomonitoring of metals and pollution.

It is therefore apparent that the presence of quantities of halophytes in the composition of the Florisbad mound is one of the most important factors contributing to the salinisation of the organic layers, clay and groundwater. This is because plant cells, particularly in the vacuoles, are capable of taking up ions against a concentration gradient and accumulating them at concentrations far greater than those in the external solution (Larcher 1983). In other words, plants can only draw water from the substrate if they can produce an osmotic potential lower than that of the soil solution (Larcher 1983). The cells of a plant are also capable of taking up preferential nutrient ions which the plant requires, and therefore cations are preferred to anions (Larcher 1983). The process becomes even more involved, with some cations being accumulated in higher concentrations than others, and other ions being accumulated in ratios specific to a particular family, or even a species, of plant (Larcher 1983). In grasses salt concentrations may be lower than in dicotyledonous, plants because a low osmotic potential is often maintained in the sap by the storage of soluble carbohydrates along with the salts (Larcher 1983).

Metals taken up by the roots of submerged macrophyte species represents the bioavailable, free-metal ion concentrations in the sediment interstitial water, as well as any metal contamination in the water columnt (St-Cyr, 1997). Metal contamination in the water column may also be taken up by submerged leaves and stems (St-Cyr, 1997). Although metal concentrations in the sediment may over-estimate the actual bioavailable metal concentrations for benthic organisms, including plants (St-Cyr, 1997), ion concentrations in the sediment are a strong indication of the ion potential available to the groundwater through flushing and miscible displacement.

Beside the accumulative potential of plants such as reed and rushes, phytoplankton and periphyton should not be ignored when considering ion uptake by aquatic plants and their subsequent mineral contribution to the sediments when they die. Phytoplankton, for example, are able to take up metals from water and bind them on their surface to produce exudates with metal-complexing properties (St-Cyr, 1997). Periphyton, on the other hand, is the interface between the substrata and the surrounding waters and consequently, influences the biogeochemistry and the dynamics of ecosystems (Wetzel 1983). Because of its sedentary nature, periphyton is a good indicator of local conditions with the micro-algal fraction playing a potentially important role in the trophic transfer of contaminants (St-Cyr, 1997) and can exhibit high concentration factors for many metals (Newman & McIntosh 1989).

Although halophytes are capable of taking up and accumulating salts in order to withdraw osmotically bound water from the soil, the accumulation of salts in the plant cells could result in a reduction in yield, or even toxic effects such as death (Larcher 1983). Halophytes have therefore developed various mechanisms whereby salt content can be regulated by the plant in order to avoid the adverse effects of high concentrations. These mechanisms include salt filtration by the plasmaemma of the roots; salt-transport prevention, where salts are limited to certain parts of the plant; and salt elimination, where the use of salt secreting glands and the shedding of various heavily salt loaded plant parts are brought into play. Yet another protective mechanism is that of succulence, where plant cells continue to absorb water in order to limit the salt concentrations and expand at the same time (Larcher 1983).

Many of the above mentioned processes eliminate minerals back into the environment, and although these may only be in small quantities, most of the accumulated minerals are released when the plants die, or parts of the plant are shed (Larcher 1983). This takes the form of leaching from the accumulated detritus. The release of minerals through decomposition, combined with the contribution of animal waste, would have lead to the further eutrophication of the spring pan water and contributed to the salinisation of the decomposed organic matter and clay.

The organic layers and clay may have been further mineral enhanced because calcium and other metallic cations do not leach from the soil in arid and semiarid environments and tend to dominate with pH values exceeding 7.0 (Brady 1984). Although Florisbad appears to have gone through long wet periods, more recent dry periods as well as intermittent dry periods throughout its history, would have contributed to a build up of metallic ions in the spring site sands. Drainage impediment in arid and semi-arid environments also tends to result in a predominance of alkaline salts, such as sodium ions, in the soil, and may equal, or even exceed, those of the adsorbed calcium ions (Brady 1984). The latter situation could well indicate that the Florisbad spring pan was a self-contained, drainage-impeded, pan, which was never influenced by any rises or falls in the water level of Soutpan. As Soutpan is also a drainage-impeded pan, these aforementioned situations may well explain the predominance of ions such as Na in the pan. Much work is still needed to fully understand the processes involved.

Additional perspectives on the Florisbad organic layers

Based on the results of this study, the study by Douglas (2001), and observations at Florisbad, some additional perspectives on the Florisbad organic layers are presented.

There is a strong possibility that due to the movement and remixing of the sediments by tectonic, hydrological and physical factors interpretations of the spring stratigraphy, based on auger samples and excavations (Butzer 1984; Rubidge & Brink 1985; Brink 1987; Grobler & Loock 1998b; Visser & Joubert 1991), may in many instances be questionable. Rubidge & Brink (1985), who based their interpretations on sediment colour and grain size parameters from auger drilling results, noted that there was a surprisingly low degree of correlation between even adjacent boreholes and that the deposits showed considerable lateral facies changes. This high degree of disconformity of the spring sediments is consistent with that noted by Deacon (1970) at the Amanzi spring site in the Eastern Cape. Rubidge & Brink (1985) concluded that drilling had been of limited value although it did confirm that the organic rich layers were confined to the central part of the site. This would tend to support the theory that re-mixing had taken place, as in an aquatic environment it could be expected that sedimentation would have occurred in fairly even layers over a large area. Despite the strong evidence of re-mixing by the aforementioned authors, Kuman *et al.* (1999) felt that, based on grain size distribution, the sediments had not been reworked after deposition.

Judging by present day plant production in the saline vlei waters, which drains from the springs, northwards, past prolific plant production at the Florisbad spring site is hardly questionable. In more recent times it was reported that when Floris Venter arrived at Florisbad in 1835, the spring eye was so heavily overgrown with vegetation, that a path had to be cut open in order to gain access to the spring (Anon 1980). The present day helophytes in the vlei, which may also be termed halophytes, include a variety of rushes and sedges such as *Schoenoplectus* (*Scripus*) *triqueter*, *Typha capensis*, *Juncus kruassii* and *Juncus rigidus*, while grasses such as *Cynodon dactylon* grow in dense profusion along the immediate bank of the lower vlei (Douglas 1992). Other plants such as *Conyza bonariens*, *Mestocema* spp. and *Mesembryanthemum* spp. also grow in the drier reaches of the vlei (Douglas 1992).

Douglas (1992) noticed the possible effects of eutrophication at the lower end of the vlei where the flow of water was being restricted by dense stands of reeds and rushes, as well as a build-up of thick detritus rafts. Of five water samples taken for analysis along the length of the vlei, the sample at the lower end of the vlei gave the highest TDS of 2 600 mg/l (Douglas 1992). Although this only represented an increase of 18% TDS over the TDS of the spring-water (2 202 mg/l), 75% of individual ion levels increased by an average of 6 207%. The most notable increases were Mn (45 358%), Al (4 969%), Fe (4 842%), Mg (542%) and Ca (50%) (Douglas 1992). This would strongly suggest that the accumulated detritus and its decomposition was contributing to the eutrophication of the water.

The Irish Peatland Conservation Council (IPCC) states that European peat formation occurs when the complete breakdown of plant material by bacteria and fungi is inhibited by a poorly oxygenated and waterlogged environment as a result of high rainfall, low temperatures, and a subsequent low evaporation rate (Anon 1996). The formation of European peat is therefore dependent on the rate of plant production exceeding that of plant decomposition through the incomplete decay of the plant material and comprises 90% water and 10% solid material (Anon 1996). Under the above conditions the IPCC gives the formation of two types of peat. Firstly, there is bog peat, which is usually formed when the water source is in the form of mineral-poor rainwater (ombrotrophic) with a pH 3.2 to 4.2, and beds varying from 2 to 12 m in thickness. Secondly, there is fen peat, which is formed when the water source is mineral-rich groundwater (minerotrophic) with a pH 7 to 8, and occurs in layers up to 2 m thick. The sphagnum moss in live European peat beds has a considerable effect on pH, as these plants absorb cations and release hydrogen ions into the water, making the water more acidic and less suitable for micro-organisms which break down the dead plant material (Anon 1996).

If the Florisbad organic layers had been formed under conditions similar to European peat, then the Florisbad organic layers would represent periods of high rainfall, low temperatures and a low evaporation rate. There seems to be consensus in the literature that this was most probably not the case. It would therefore appear that the Florisbad organic layers were formed under much higher temperatures with the slowly rising water level in the spring pan, provided by the spring and supplemented by rainfall, compensating for the low European temperatures and evaporation factor. This is also similar to the situation that is developing at the lower end of the vlei where there is a build up of dead vegetation, an increase in the water level, and the beginnings of Florisbad type "peat" formation. Grobler & Loock (1988a) noted a similar situation arising at a freshwater spring west of Soutpan.

Plant production at the spring site may have at times also been adversely affected in drier periods by a toxic accumulation of ions in the ever-increasing quantities of clay and organic matter at the bottom of the spring pan. This was illustrated during the 1988 exceptionally high rainfall period when water filled Soutpan pan for the first time in living memory (Douglas 1992). Soutpan, which normally has only a few scattered bushes and grasses growing on it, produced an extraordinary profusion of luxuriant vegetation in the standing water. This proliferation of plant growth could be seen as a direct result of the dilution of toxic ion levels by the mass of fresh water which entered the pan. The water in the pan registered 10 365 mg/l TDS (Douglas 1992), indicating that halophyte plant production is possible at this level, but may cease at levels somewhat higher than this. In Table 2 it can be seen that the water in Pit 4 (P499) registered 9 019 mg/l TDS, with Peat II from the same pit (X3) registering 24 655 mg/l TDS. Kruger & Lubczenko (1994) gave 2 000 mg/l TDS as being the level for high tolerance crops, while >3 200 mg/l TDS was considered generally too saline for agricultural purposes. Judging from the Soutpan scenario, as well as the TDS levels in some of the pit-waters, it is possible that any increase in the TDS of the groundwater to over 10 500 mg/l TDS in dry periods, could have resulted in the environment becoming too toxic for normal halophyte production.

The deflation of Soutpan would also have been assisted by the large herds of animals roaming the area. These herds, who most probably used Soutpan as a source of salt and minerals, would have trampled the sand and salt to fine dust, making them more easily transportable by wind. Enhanced chemical weathering due to the high concentrations of salt in the pan may also have assisted the weathering process (Grobler & Loock 1988a). Considering the amount of sand that was transported by wind to form the lunette and surrounding dunes, salts from Soutpan and the surrounding area must have also been blown into the spring site. Therefore, the minerals deposited with the sand into the spring site, and onto the lunette by the prevailing north-west wind, will also have contributed to the salinisation of the spring mound. Over time, rainfall would have leached salts from the lunette into the groundwater, thereby contributing to the minerals available for uptake by the clay and humus.

The transportation of pollen, other vegetation material and sand, to the spring site by wind, appears to have been considered on a very localised scale by some authors. Van Zinderen Bakker (1989) suggested that due to the prevailing north-west wind, pollen found in the spring sediments was largely a reflection of vegetation changes which had taken place in Soutpan. In considering the size of Soutpan, as well as the obvious long-tern deflating nature of the pan, Soutpan must have basically been devoid of vegetation for considerable periods of time. Again, when considering the wind energy involved in transporting the vast quantities of sand within the area, it seems apparent that much of this sand must have come from distances far greater than that of Soutpan. Modern day dust storms in the Free State are a clear indication of this large-scale transportation of sand over great distances. Although Van Zinderen Bakker (1989) saw pollen in the spring sediments as reflecting vegetation changes in Soutpan, it is felt that pollen in particular would have reflected, and represented, vegetation changes over a far greater area than that of Soutpan.

During prolonged dry periods, the fresh peat beds at the spring site would have slowly been covered with layers of largely aeolian sand, and minerals, both from the pan and the larger surrounding area. Standing vegetation would also have provided a trap for the wind blown sand, with the vegetation gradually being covered during these drier periods. At the same time the vegetation would have begun to be compressed by the weight of an ever-increasing sand layer. While the production of plant material may have been halted, the decomposition of plant material would have slowly continued beneath the sand, helping to decrease the thickness of the peat beds even further. As previously mentioned, this compacting process is sill in progress and can be seen where the very fine clay fraction is being squeezed out from between the coarser sand fraction on some of the exposed Peat II surfaces. It is postulated that when the organic layers were originally formed, Peat I and II were most probably similar in composition and appearance to the present day Peat IV, before decomposition and compaction compressed them to their current state.

Viewing aspects of the formation of the Florisbad spring site in isolation may also be applicable to the previously mentioned flooding of the palaeopan (Soutpan) theory (Butzer 1984; Van Zinderen Bakker 1988; Visser & Joubert 1990; Joubert & Visser 1991). In this regard Peat IV is of interest because, relative to Peat II, it is a fresh layer, as indicated by its porosity and lack of decomposition and compaction. This layer was approximately 2.8 m thick at the sampling site, indicating that it must have been in a saturated developmental stage over a very long period in order to attain this height. This would also mean that the water level at the site must have, over an extended time, gradually and continually risen to at least the top of Peat IV in order to allow for a build up of sand and plant production to this level. It is postulated at this time that this rise in the water level was in no way related to the flooding of the palaeopan.

In considering the present day flat topography of the area, an even flatter topography would possibly have existed in the past before erosion, tectonic and aeolian factors reshaped the landscape. Had the water level from the flooding of Soutpan, who's floor is about 30 m below the top of Peat IV, reached the top of Peat IV in this flat environment, it would have effectively created a massive lake covering a large part of the central-western Free State. However, it should be remembered that the differentiation between the top of the mound and the Soutpan floor would also have been less in the past. Joubert & Visser (1991) suggested that the waxy organic rich clay would have formed in marshes along the palaeolake margin. Even if the possibility of a massive lake is discounted, a smaller lake should then have resulted in the organic layers extending over a much greater area than just the spring site, as well as in more or less the same sequences and thickness as those found at Florisbad. Again this does not appear to be the case. The punctuated flooding theory would also imply that an extended palaeolake must have existed in stages for extremely long periods in order to allow for the development of Peat I and II, as well as for Peat IV to attain its present height.

The texture of Peat IV is fairly uniform from top to bottom with possible shorter intermittent dry periods being reflected by a more sandy composition. The current state of Peat IV would tend to indicate that, after a slow but steady rise in the water level, in which the Peat IV formed, the water level must have dropped rather rapidly in order to arrest plant production and decomposition, leaving Peat IV in its current state. Another factor, which possibly influenced the current state of Peat IV, is the regional vegetation cover, which would have proliferated during the relatively wet period of the spring sites development. Based on the feeding niche of the extinct springbok, *Antidorcas bondi*, Brink & Lee-Thorpe (1992) suggested that during the Late Pleistocene the grasslands of the area were far more productive than at present, with year-round production occurring. Grass cover would therefore have decreased the aeolian transport of sand and therefore the quantities deposited on the lunette and spring site. The result of this is that a sand overburden of only 650 mm has been deposited on the west wall of the excavation above Peat IV. Unlike with the lower organic layers, the compacting effect of the sand build-up on Peat IV has been minimal, allowing it to retain its present porous state. As previously mentioned, at this stage of the investigation it is not thought that there was any significant cyclical flooding of Soutpan which resulted in the subsequent flooding of the Florisbad spring site. It is also felt that if any such flooding did take place, that this was not responsible for the formation of the organic layers.

Results presented in Table 1 show that, despite Peat IV being situated above the present water table, it is also highly mineralised. The TDS of sample X4, from the top of Peat IV was 62% higher than sample X5, taken from the bottom of Peat IV. The TDS of X4 and X5 were also 902% and 517% respectively higher than the spring-water TDS (Table 4). The overall high TDS values of Peat IV are seen largely as an indication of the capacity of the halophytes to accumulate, concentrate and store, ions by the aforementioned processes. When the water table was high, salinisation of the clay and humus would have taken place through the absorption of ions from the spring- and groundwater, decomposition of the halophytes, as well as from minerals in the aeolian sand. After the drop in the water level, salinisation would have continued largely through the decomposition of the halophytes, the process of capillarity, and to a somewhat lesser extent, from the minerals in the aeolian sands.

These TDS values, particularly those in sample X4, can partially be explained by the aforementioned process of capillarity. Capillarity is a process which occurs when the top layer of soil is relatively wet and the evaporation of water into the atmosphere reduces soil wetness and increases matric suction at the surface (Hillel 1971). The process also occurs where the groundwater table is relatively close to the soil surface and water moves by capillarity to the surface where it evaporates, leaving behind its salts (salinisation) (Hillel 1971; Bohn *et al.* 1985). Capillarity is also reliant on the conductivity of the soil in relation to the surface evaporation rate (Hillel 1971) and will increase in soils with a high clay and humus content. In this instance, salinisation may be appreciable even where the water table is several meters deep and will increase where the groundwater is brackish and the evaporation potential is high (Hillel 1971). This is particularly so in semi-arid environments where evaporation exceeds precipitation and calcium carbonates form as a result of a high evaporation rate (Hillel 1971) making Peat IV an ideal medium for such processes. Capillarity can therefore be seen as a reverse leaching of the soil, where the minerals in the lower soil levels are brought to the surface by capillarity and evaporation, as opposed to the minerals in the top soil being leached downward by continual heavy precipitation.

The effect of evaporation and matric suction can be gauged from rainfall and evaporation figures taken from Glen Agricultural College, 32 km east of Florisbad, as part of Douglas' (1992) study. Rainfall for January, February and March 1992, was 66.5 mm while the mean evaporation rate for the same period was 315.0 mm. This then gives a rainfall deficit of 248.5 mm for the period. Van Zinderen Bakkker (1989) gave an average annual rainfall deficit for eight stations, with records varying from 18 to 47 years, of 1260 mm. Bohn *et al.* (1985), for example, illustrated that where the water table was 900 mm below the surface, the effects of salt deposition began at about 400 mm where the electrical conductivity (EC) was 0.2 ms/m, and this increased to EC 70 ms/m at 50 mm. The aforementioned scenarios are very much the case at Florisbad and capillarity and salinisation have obviously both played a major role in the mineralization of the upper Peat IV horizon (sample X4).

Possible application at other fossil sites

The results of this study may well have applications at other fossil sites such as Vlakkraal south of Florisbad and the Cornelia fossil beds in the north-eastern Free State. In the case of the Cornelia beds, for example, clays are associated with all the fossil occurrences and also occur in most of the other layers (Butzer 1974). A reconstruction by Butzer (1974) of the five layers identified by Van Hopen (1930) at the Cornelia-Uitzoek site showed that four of these layers comprised clay and/or black organic soil. Fossil layer (b), for example, was identified as a clay layer overlain by a massive bed of grey clay (Butzer 1974). Butzer (1974) later divided these beds into eight layers, of which six comprised clay. More recently, Brink & Rossouw (2000) recorded another massive yellow clay horizon beneath the Unit 1 described by Butzer (1974), both of which contain fossil remains.

Although no mention has been made of spring activity at the Uitzoek site, the question remains as to whether the groundwater at Uitzoek is mineral rich, or not. Should it be found that the groundwater is mineral enriched, then it must be asked, from where has this enrichment originated? At present it does not appear as if probable salinisation of the clay and organic material is in any way related to spring activity and, from the results of this study, it would not appear that the Ecca shales are responsible either. This leaves the alternative of a pan system which later developed into a marsh, or wetland, during wet periods. What is apparent at Uitzoek is, that the clay and organic matter appear to have stored and supplied the minerals for the fossilisation of faunal remains.

CONCLUSIONS

It can be concluded that the initial salinisation of the groundwater at Florisbad has occurred through the miscible displacement of the spring-water with the groundwater. It is also apparent that where the salinisation of the groundwater is in excess to that of the spring-water, this excess has been derived largely from sources other than the spring-water, or the underlying bedrock. From the evidence presented in this paper, it is clear that one of the primary sources is that of the clay and organic layers which have attracted and concentrated ions from the groundwater, and that these concentrations will vary in direct relationship to the quantities of clay and organic matter present. The accumulation and storage of minerals by the halophytes, and their subsequent release of minerals though decomposition has played a major role in salinisation at Florisbad. Other sources, such as the accumulation of salts in a semi-arid environment, aeolian deposition, capillarity, and the drainage impediment of the spring pan, have all contributed to the salinisation of the site. As these processes would have been in operation since the genesis of the spring, it is more than likely that fossilisation would have taken place even if the spring water never carried sufficient minerals for fossilization.

Contrary to the norm, the TDS of the groundwater at Florisbad increases due to rainfall, flushing ions from the organic layers and clay, while the TDS of the groundwater decreases in dry periods due to the attraction and concentration of ions by the organic layers and clay. This then produces an environment where, when there is relatively mineral rich groundwater in combination with decomposing halophytes and clay, the opportunity for fossilisation increases considerably.

The author does not feel that the above theories on the flooding of the palaeopan provide adequate, or appropriate, explanations for the formation of the Florisbad organic layers, and that the organic layers must have formed under an alternative set of conditions. This is primarily due to the fact that it does not seem probable that the water level in the palaeopan could have risen to the level of the top of Peat IV. It is also postulated that when the organic layers were originally formed, Peat I and II were most probably similar in composition and appearance to the present day Peat IV, before decomposition and compaction compressed them to their current state.

The results of this study may well have implications for the fossilisation of faunal remains at Florisbad, where it has been proposed that it is the spring-water which has been responsible for fossilization. This study may also have applications at other fossil sites where there is a relatively mineral rich hydrological environment combined with clay and/or organic deposits. Salinisation processes discussed in this paper may also be applicable to coastal sites where a relatively mineral rich marine environment could exert a far greater influence over both salinisation and fossilisation. Other aspects on the fossilisation of faunal remains and the formation of the Florisbad site and organic layers form part of an ongoing investigation.

OPSOMMING

As deel van 'n studie om aspekte van die omgewing waarin die fossilering van diere-oorblyfsels plaasgevind het, te bepaal, is die versouting van Florisbad se organiese lae en klei in verhouding tot die bron- en grondwater ondersoek. Die analise van water by Florisbad het bewys dat die TDS van die grondwater in die eksplorasieputte tot 287% hoër was as dié van die bronwater, met variasies van tot 666% tussen die water uit die verskillende putte. Om die oorsprong van die hoër mineralisasie te bepaal, is monsters van Peat II en Peat IV geanaliseer en met vorige resultate van bron- en grondwater vergelyk. Dit het bewys dat die organiese lae en klei, bo of onder die watertafel, die vermoë het om ione te lok en op vlakke 751% bo die van grondwater en 945% bo die van bronwater te dra. Dit wil voorkom of die grondwater versout is deur die verplasing van ione met die bronwater, asook deur die opgaar van ione in die organise lae en klei wat deur hoër reënval uitgewas word. Daar is bevind dat soveel as 435% variasie in die TDS binne dieselfde organiese lae was. Die ophoping en dra van ione deur halofiete en die verrotting van die plante word as een van die belangrikste faktore in die versoutingsproses beskou. Organiese lae bo die watertafel is ook hoogs mineraal-verryk en dit blyk asof dit 'n gevolg van die teenwoordigheid van die halofiete en ook kapillêre werking is. Sand en minerale ingewaai vanaf Soutpan en die wyer omgewing het ook tot die versouting van die klei en organiese lae bygedra. Die prosesse kan aanwendings hê in ander gebiede waar fossiele saam met 'n mineraalryke hidrologiese omgewing gevind word. Om bogenoemde prosesse beter te verstaan, is aspekte soos teorieë oor die formasie en huidige toestand van die organiese lae en die versouting van die organiese lae en klei, asook alternatiewe perspektiewe oor die organiese lae, bespreek.

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Appendix

III

Is the spring-water responsible for the fossilization of faunal remains at Florisbad, South Africa.

Douglas, R. M. 2006.

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Is the spring water responsible for the fossilization of faunal remains at Florisbad, South Africa?

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Abstract

It is has been suggested that faunal remains at Florisbad were fossilized in a spring context due to the mineralized spring water. However, the environment conducive to the precipitation of $CaCO_3$ and other authigenic minerals was formed largely through the salinization of the organic layers and clay, and the mineralization of the groundwater. Factors contributing to this favorable environment include: $CaCO_3$ saturation, pH, the decomposition of halophytes, Eh, rainfall, biomineralization, and aeolian deposition. With the exception of pH, none of the above factors feature in a spring context, with evidence suggesting that the spring water may historically never have carried sufficient minerals for fossilization, and that contact with the spring water may actually have resulted in the demineralization of previously fossilized material. In light of this evidence, it is concluded that the fossilization of faunal remains at Florisbad took place in a sedimentary organic matter and clay environment and could not have taken place in the spring vents where there is an undersaturation of Ca.

Keywords: Florisbad; Faunal remains; Fossilization; Groundwater; Sediments; Spring water; Processes; Mechanisms; Factors

Introduction

It is well known that the sedimentary environment in which material was deposited will determine the type and intensity of the physical and chemical processes. With only Brink (1987) having put forward any theory on the Florisbad fossilization process, without any detailed examination of the factors and processes involved, further examination of these processes was considered warranted in light of new evidence.

The Florisbad spring site (28°46′S 26°04′E), with its archaeozoological and archaeological excavations, is located 49 km north-west of Bloemfontein, Free State Province, South Africa. The 500-mm isohyet passes slightly to the east of Florisbad with a 78-year average annual rainfall of 496 mm, with annual rainfall has varying from a maximum of 957 mm in 1988 to a minimum of 271 mm in 1965 (Douglas, 2001a). Royer (1999) found a significant correlation between the presence of carbonate horizon bearing soils and a mean annual precipitation of <760 mm. The Florisbad spring originates from an aquifer of unknown location and size and does not appear to have had any thermal history. Grobler and Loock (1988) estimated that in order for the spring water to issue at 29°C, recharge water must have percolated to a depth of approximately 500 m and reached a temperature of 32° to 35°C. The flow rates of the spring, as given in the literature, are possibly not very reliable and vary from 18.8 m^3/h (Grobler and Loock, 1988) to 159.3 m^3/h (Kent, 1948).

Florisbad is the type site for an important accumulation of fossil remains representing the Florisian Land Mammal Age, which falls between the Middle and Late Pleistocene. Electron Spin Resonance (ESR) dating has placed faunal remains into two basic categories, an earlier Middle Stone Age (MSA) Old Collection dated at 281,000 ± 73,000 ESR yr, and a later MSA human occupation horizon dated at 121,000 ± 6000 ESR yr (Grün et al., 1996). Six mammalian Orders, represented by 27 species, have been recorded from the Old Collection. Of these, five species are extinct, with the rest being distinguishable from extant species by anatomical features (Brink, 1988). Occurring within the lower sediments is the Florisbad hominid skull dated at 259,000 ± 35,000 ESR yr (Grün et al., 1996). Besides a wealth of artefacts, Brink (1987) also recorded four mammalian Orders, represented by eight species, from the MSA human

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occupation horizon. This paper deals only with factors affecting the preservation of the Old Collection.

Brink (1987) noted that the fossilized bones of the Old Collection were found in the area of spring activity and that post-depositional mechanical and chemical weathering clearly showed that they became fossilized in a spring context. It was further stated by Brink (1987) that because the spring water was slightly alkaline, "the chemical nature of the spring water caused the remains of the Old Collection to be preserved in a characteristic way," resulting in their good preservation. Brink (1988) expanded on this by stating that faunal remains became mineralized in contact with the spring water in vertical bodies of sediment, referred to by Dreyer (1938) as "extinct spring eyes" and by Deacon (1970) as "vent structures." These spring vents did not penetrate to the surface and consisted of a basal accumulation of artefacts, bones and coarse sand, capped by pure quartz sand (Brink, 1987).

The fossilization of faunal remains in the spring vents was further supported by what was referred to as the post-depositional formation and modification of calcite on, and within, the bones (Brink, 1987, 1988). Of the Old Collection, 48.2% of Bovidae and *Hippopotamus amphibius* bones showed signs of calcium carbonate deposition with Brink (1987) further noting that, since the spring water was carbonate-rich, the bones must have been in contact with the spring water for some time. However, and perhaps more importantly, Brink (1987) conceded that if the presence of calcite was considered to be an indication of contact with the spring water, then the relatively low incidence of calcite concretions might contradict the assumption that the Old Collection was entirely derived from spring deposits. Pitting and hollowing were also given as further indications of chemical action by the spring water, with pitting occurring as a result of chemical solution in an aqueous environment prior to mineralization, and hollowing occurring through chemical solution after mineralization (Brink, 1987).

Permineralization in the form of calcite and authigenic apatite precipitation takes the form of small calcite crystals forming even coatings in the bone cavities and sponge tissue, as well as concretions found on the external surfaces (Brink, 1987). A few larger calcite crystals were also recorded by Brink (1987) in the bone cavities and sponge tissue. As the fossilization of faunal remains is dependant on the precipitation of authigenic minerals, which require certain favorable conditions for precipitation, the aim of this paper is to examine factors, conditions and processes conducive to the precipitation of authigenic minerals in a groundwater and sedimentary context. Trueman and Tuross (2002) noted that the preservation of bone requires the growth of authigenic apatite in contact with groundwater. Conversely, factors that would preclude and inhibit the precipitation of authigenic minerals in a spring water context, and therefore fossilization, are also examined.

Background to the quality of the spring and groundwater

A review of water quality and salinization processes at Florisbad by Douglas (2001a,b) indicate that, when combined and examined in the context of fossilization, they provided an entirely new perspective on the fossilization process at Florisbad. Douglas (2001a) established that the groundwater in Pit 1 (Fig. 1), 22 m from the spring eye (SEW, Fig. 1) had a TDS (total dissolved solids) 50% below than that of the spring water during a normal rainfall period (Table 1). Analysis of the same groundwater during a high rainfall period indicated that the TDS of the groundwater was 27% higher than that of the spring water (Douglas, 1992). This was contrary to what can be considered the norm, where due to high recharge, there is usually a decrease in TDS due to a diluting effect and an increase in TDS with low recharge (Bredenkamp, 2000; Douglas, 2001b). If the groundwater in Pit 1 was under the influence of the spring water, then the TDS of the groundwater in this pit should have remained in equilibrium with the spring water during both high and normal rainfall periods. This led to the conclusion that if the TDS of the pit waters could be either higher, or lower, than that of the spring water, over such a short distance, then the two hydrological environments could not be interrelated and should be seen as separate entities. Other anomalies which appeared were; the TDS of groundwater in the pit furthest from the main spring eyes (Pit 4) (Fig. 1) was 289% higher than that of the spring water, variations in TDS of up to 683% occurred between individual pit waters (Pit 1 and Pit 4) only 54 m apart, and TDS increases of up to 943% occurred between Peat 2 (Pit 4) and the spring water (Douglas, 2001a,b). Fluctuations of this magnitude over such a small area would suggest that mineral deposition and fossilization must have been just as variable. Between the two study periods the ion content of the spring water remained stable and little changed (Table 1).

The stratigraphy of the sediments comprises alternating horizontal layers of mainly aeolian sand and compacted organic layers, suggesting that these were laid down during alternating wet and dry periods (Fig. 2). In the Florisbad context, the aeolian contribution must have been significant when considering that the Florisbad mound is actually a sand dune, and that the area to the south and south-east of the spring site also comprises sand dunes. Perhaps, the primary consideration at any potential fossilization site is the existence of micro and macro environmental conditions, as well as the properties conducive for the precipitation of authigenic minerals.

Factors conducive to fossilization in a groundwater/sedimentary context

The sedimentary environment at Florisbad presents ideal conditions for salinization and the formation of authigenic minerals due to large quantities of clay and organic material, together with the sites' aquatic history and relatively high water table. Without the unique presence and stratigraphy of the sediments, factors such as the saturation and precipitation of $CaCO_3$ would not have been able to take place. At Florisbad, the clay and organic layers act as a battery, accumulating, adsorbing and storing of ions from the groundwater during low rainfall periods and releasing them during periods of high rainfall (Douglas, 2001b). Since the earliest investigations at

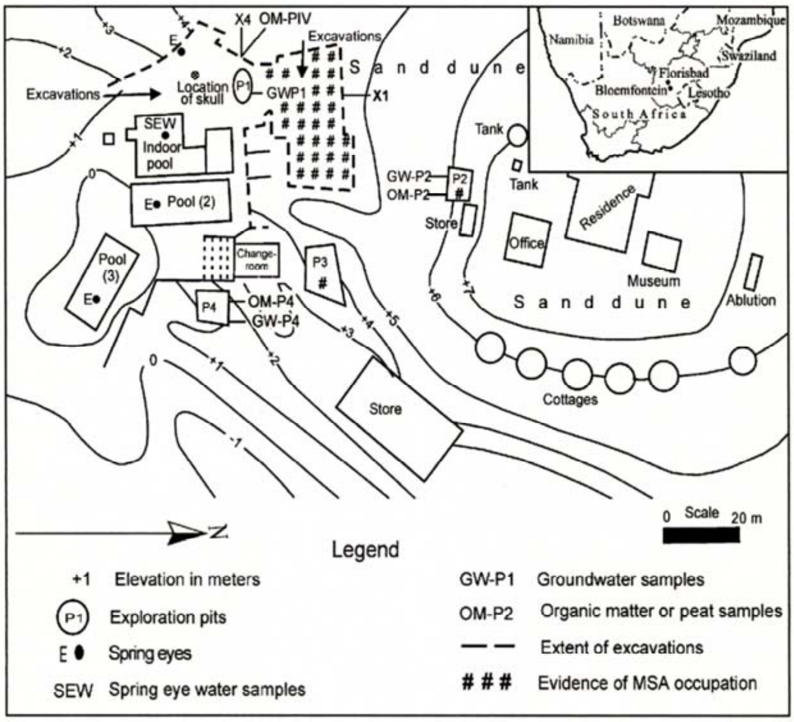

Figure 1. The location of the Florisbad spring site and a plan of the site showing the location of the spring water, groundwater and peat samples, exploration pits and extent of the present excavations (after Douglas, 2001a,b).

Florisbad, "peat" has been used as a convenient term for describing the various organic-rich layers (Douglas, 2001b), and this has been followed here. The position of the peat bodies within the sequence is given in Figure 2, with the salinization process at Florisbad being illustrated in Figure 3 and discussed further on in the paper.

The solubility of calcium carbonate is controlled by the pH of the environment, changes in temperature and pressure, as well as organic matter activity and decay (Krauskopf, 1967). Hornberger and Brady (1998) noted that the solubility of bicarbonate concentrations for calcite in pure water at PCO_2 10^{-3} was 83 mg/l, and at PCO_2 10^{-1}, 370 mg/l. Freeze and Cherry (1979) gave the solubility of calcite in natural groundwater with a temperature of 25°C, pH 7, 1 bar total pressure, and PCO_2 10^{-3} as 100 mg/l, while solubility at PCO_2 10^{-1} increased to 500 mg/l. A reasonable free ion calcite saturation level was considered by Ferguson (1991) as 150 mg/l, while Coetzee et al. (1966) used supersaturated solutions of 150 mg/l Ca for precipitation purposes.

Calcium showed one of the most relevant ion increases between the spring water, groundwater and organic layers (Table 1). Between the spring (SEW Norm) and the groundwater (GW-P4 Norm) in Pit 4, Ca increased from 100 mg/l to 400 mg/l, while between the groundwater (GW-P4 Norm) and the Peat II layer (OM-P4 Norm) in the same pit, Ca increased from 400 mg/l to 5000 mg/l (Douglas, 2001a,b). There was an even greater Ca increase at the bottom of the Peat IV layer (OM-PIV, Fig. 1), with Ca increasing to 7360 mg/l (Douglas, 2001b) through possible evaporation and capillarity. Another indication of the high levels of calcium in the groundwater was that of m-ALK, as $CaCO_3$. The m-ALK of the groundwater in Pit 1, during the normal rainfall period, showed a 1325% increase over that of the spring water (m-ALK 12) to m-ALK 171 (Douglas, 2001a). Conversely, Ca fell by 51% in Pit I (GW-P1 Wet) between the 1988 high rainfall period (169 mg/l) and 1999 normal rainfall period (GW-P1 Norm) (82 mg/l), bringing the Ca level in the groundwater to below that of the 1999 spring water level of 100 mg/l (SEW Norm) (Douglas, 2001a). This could be seen as the 1988 groundwater (GW-P1 Wet) having a 106% higher Ca concentration than the 1999 groundwater (GW-P1 Norm) when ions were being flushed from the clay and humus during the high rainfall period. The low Ca level of the groundwater during the normal rainfall period would support the theory that the clay and humus were adsorbing and accumulating ions through salinization. All indications are that the Ca saturation levels of the Florisbad organic layers, and to a lesser extent the groundwater, are well within limits for Ca precipitation.

Calcium carbonates and phosphates become more soluble with reducing pH which can be brought about by water percolating through organic rich soils and rainfall. At Eh (oxidizing potential) 0 and pH 8.0, calcite and hydroxyapatite will be preserved, while organic matter would decompose, while at pH 7.1, calcite and hydroxyapatite would be dissolved. Alternatively, the oxidation of organic matter by the dissolved oxygen in the water would release CO_2 and phosphates, making the water capable of reacting with calcite and altering it to hydroxyapatite (Karkanas et al., 2000). Because this reaction uses acid, the solution becomes alkaline, favoring the stabilization of carbonates. As the sediment and groundwater Ca

Table 1
Variations in spring and ground water quality at Florisbad during normal (rainfall for 12 months proceeding sampling 545 mm) and wet (rainfall for 12 months proceeding sampling 957 mm) rainfall periods

	SEW	SEW	GW-P1	GW-P1	GW-P2	OM-P2	GW-P4	OM-P4
	Norm	Wet	Norm	Wet	Norm	Norm	Norm	Norm
pH	8.91	9.34	7.88	7.01	7.65	8.00	8.14	6.55
p-Alk	7.00	11.25	0.00	–	0.00	–	0.00	–
m-Alk	12.00	23.72	171.00	119.46	138.00	–	932.00	–
EC (m/SM)	388.00	398.00	220.00	493.00	325.00	–	1419.00	–
IONS (mg/l)								
Chloride	1442.00	1361.32	685.00	1670.81	1104.00	7727.40	5648.00	12,393.00
Nitrite	0.01	0.00	0.01	0.00	0.01	–	0.01	–
Bromide	19.31	4.34	8.46	5.05	15.86	–	71.76	–
Nitrate	0.73	0.24	1.19	0.44	0.90	–	4.81	–
Phosphate	1.93	0.00	0.10	0.51	0.01	26.00	0.01	34.00
Sulphate	1.50	1.55	2.60	24.41	6.60	–	81.60	–
Ca	100.00	98.36	82.00	168.57	136.00	2768.00	400.00	5000.00
Mg	0.44	0.50	3.91	5.74	5.15	24.80	9.04	70.00
Na	784.00	725.06	380.00	903.33	554.00	4240.00	2920.00	6800.00
K	10.10	9.11	8.62	12.46	12.90	640.00	42.54	352.00
Al	0.05	0.15	0.14	0.15	0.77	–	0.04	–
Fe	0.01	0.22	0.11	4.00	0.15	–	0.03	–
Mn	0.01	0.01	0.02	1.14	0.79	–	0.02	–
Cr	0.01	0.12	0.01	0.11	0.01	–	<0.01	–
Zn	0.02	0.18	0.03	0.18	0.03	2.40	0.03	6.00
Cu	0.01	0.06	0.01	0.07	0.01	–	0.02	–
Cd	0.01	0.01	0.01	0.05	0.01	–	<0.01	–
B	2.59	1.62	1.74	1.81	2.00	–	10.16	–
Total ions	2362.73	2202.85	1173.96	2798.83	1839.20	15,428.60	9188.09	24,665.00

As a comparison, the results of organic material from some of the exploration pits are also given. Samples as located in Figure 1: SEW = Spring eye water sample from indoor pool; GW-P1, 2, and 4 = Groundwater samples from excavation pits 1, 2 and 4; OM-P2 and 4 = Organic material samples from excavation pits 2 and 4.

saturation levels are well within limits for Ca precipitation, with the exception of sample OM-P4 Norm (Table 1), the pH of all other samples were above pH 7.0, with almost the entire site providing suitable conditions for calcium carbonate precipitation on the basis of pH. The gradual build-up of phosphates over time from various sources, such as carcases and excretement from large herds visiting the watering hole, could have provided for the later formation of authigenic phosphate apatite and the fossilization of bone.

The importance of salinization in the clay and organic layers is illustrated by colloidal particles of clay and humus having the capacity to adsorb large quantities of ions. Ions may be supplied by percolating water, and in many instances, clay and humus act as depositories for minerals (Hillel, 1971; Blatt et al., 1972; Brady, 1984). Salinization is capable of concentrations to such an extent that economic deposits of minerals, such as mirabilite, gypsum, magnesite and halite, may form. The term clay was omitted by Kuman and Clark (1986) who discussed the composition of the various Florisbad layers in terms of sand and silt. As Kuman and Clark (1986) did not define silt in terms of particle size, weathering or plasticity, it is presumed that this silt fraction contained a considerable quantity of clay. With the exception of only one of the sixteen layers described by Kuman and Clark (1986) as comprising silt/clay, the average silt/clay content for the profile was 36%. Douglas (2001b) recorded a 46% clay fraction in the Peat II layer, comprising both illite and montmorillonite. Therefore, the presence of clays and organic material at Florisbad may have a significant effect in inducing mineral deposition.

Another important factor governing the presence of carbonates in natural waters is the presence of organic matter within a low Eh environment, and a restricted water circulation (Blatt et al., 1972). Nearly all carbonate deposits are organic in nature, with the preservation of the organic matter being dependent on a low Eh environment in order to prevent complete oxidation of the organic compounds to CO_2 and H_2O (Blatt et al., 1972). The ability of halophytes such as reeds and rushes, as well as other aquatic vegetation such as phytoplankton and periphyton, to produce an osmotic potential lower than that of the soil solution results in them taking up, accumulating and concentrating ions at levels far greater than those in the external solution, even in ratios specific to a particular family or even species (Larcher, 1983). Conversely, halophytes are capable of employing mechanisms to counteract the toxic effects of salt accumulation through the elimination and deposition of minerals back into the aquatic environment (Larcher, 1983), while having the ability to adsorb and release minerals during various stages of decomposition. Although metal concentrations in the clay and organic sediments may overestimate the actual bioavailable metal concentrations for benthic organisms, including plants (St-Cyr et al., 1997), ion concentrations in the sediment are a strong indication of the ion potential available to the groundwater (Douglas, 2001b).

Due to the amounts of organic matter in the Florisbad sediments, DOC (dissolved organic carbon) most probably also plays an important role in the precipitation and formation of authigenic carbonated minerals, in that it serves as a food source for carbonate precipitating organisms. DOC, also

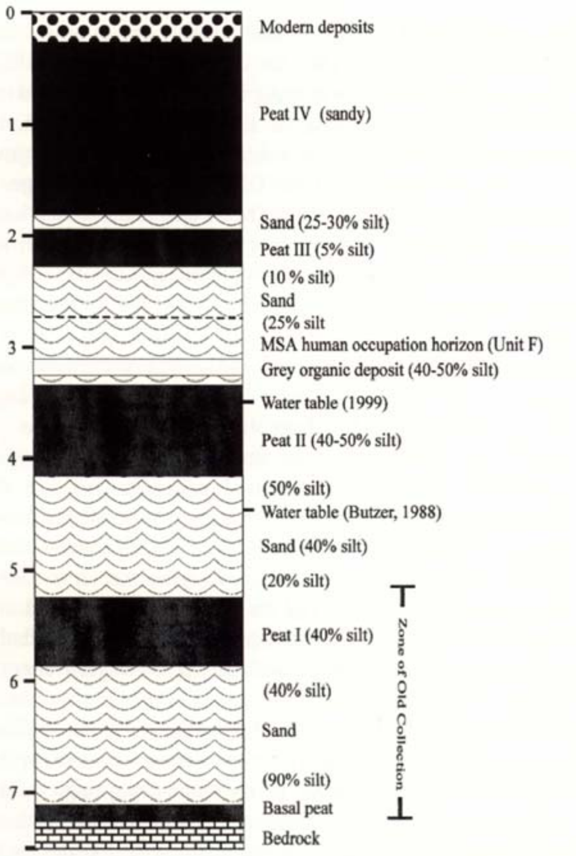

Figure 2. A simplified schematic cross-section of the Florisbad deposits (after Kuman et al., 1999) showing the main organic and sand horizons and their silt (clay) content.

referred to as hydrophobic organic acid, or humic acid, is the organic carbon present in water, comprising fatty acids, alkalines, sugars and complex polymeric molecules. These hydrophobic acids are formed through the decay of plankton and plant material, leaching and precipitation, resulting in the excretion of extra-cellular material in the form of dissolved humic and fluvic acid. This material breaks down to CO_2, partly by bacteria, which use the DOC as a food, energy and nutrient source. Alternatively, it has been shown that DOC may act as a kinetic growth inhibitor, inhibiting calcite crystal growth. For example, although calcite supersaturation has been observed in the Everglades, kinetic inhibition by DOC prevents any precipitation from occurring (Hoch et al., 1998). The coatings by DOC at low negative surface charge, when (theta) is close to 1, explains the reduction of the precipitation rate of close to zero even when the system is supersaturated (Lebrón and Sáurez, 1998). Of particular interest is the fine balance which must be obtained between the production, consumption and build-up of DOC by $CaCO_3$-precipitating microorganisms, and plant decay, in order to allow for fossilization, without a build-up of DOC acting as a $CaCO_3$-precipitation inhibitor.

In examining the fossilization process at Florisbad, it is apparent that the processes involved here are extremely complex, with diagenesis being influenced by numerous factors and mechanisms. Factors and mechanisms, such as pH, Eh, the presence of calcium and phosphate, DOC, biomineralization, salinization and PCO_2, are often interrelated, and in some way interdependent on each other for the formation of authigenic minerals and the success of fossilization. Other factors that play a role are rainfall, wind and aeolian deposition, evaporation, capillarity, illuviation and the presence of halophytes, macrophytes, phytoplankton and periphyton. Mineralization and salinization processes may have been even further influenced and enhanced because calcium and other metallic cations do not leach from the soil in arid and semiarid environments, tending to dominate where pH values exceed 7.0 (Brady, 1984). Drainage impediment in arid and semiarid environments also tends to result in a predominance of alkaline

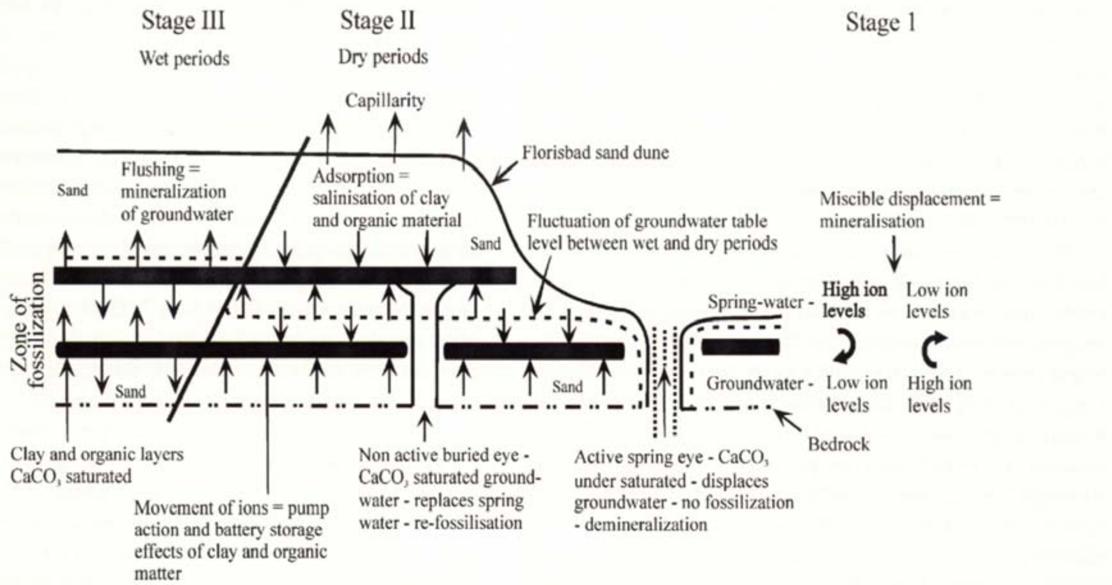

Figure 3. A schematic plan reflecting the movement of ions and stages in the mineralization, salinization processes in the fossilization of the Old Collection at Florisbad.

salts (Brady, 1984), as does the proposal by Douglas (2001b) that in the past the Florisbad spring pan was a self-contained, drainage-impeded, pan and was never influenced by any rises or falls in the water level of the near by Soutpan (a large salt pan 1.5 km north-west of the spring site).

Factors precluding fossilization in a spring context

The only manner in which carbonate precipitation could have been initiated in a spring context would have been if the spring water was supersaturated in $CaCO_3$ with an alkaline pH, and a considerably reduced flow rate. Despite it having been stated that the Florisbad spring water was carbonate rich (Brink, 1987), there are a number of references indicating that the spring water and sand layers are low in $CaCO_3$. These include water analyses by (Rindl, 1915; Fourie, 1970; Mazor and Verhagen, 1983; Douglas, 1992, 2001a), all of which indicated undersaturated Ca levels. A compilation of spring water results over the past 84 years showed that Ca levels remained constant and at relatively low levels, never exceeding 100 mg/l (Douglas, 2001a). Of particular importance is that Butzer (1988) found no evidence of calcium carbonate in his 1982 excavation profile, noting that the spring water was undersaturated in both calcium and bicarbonate, concluding that any such carbonate enrichment must therefore derive from either direct, or indirect, external sources other than the spring water. Fourie (1970) found a lack of carbonate present in the sands of the eyes as well as in the lower consolidated green sediment, although he did find some evidence of calcite cementation in the basal zone of Peat 1.

The characteristic white quartz sands of the spring vents could not have played a role in fossilization because all clay or organic material has been washed out by the spring flow, leaving them, to all intents and purposes, sterile. Spring eye sand analysis has given a 0% clay and organic matter content (Douglas, 2001b), indicating that the spring eye sands do not have an ion adsorption, carrying or retention capacity for the storage of minerals for salinization, or for the precipitation of authigenic calcium carbonate minerals. In contrast to this, Kuman and Clark's (1986) layer "p," which also lies just above bedrock at approximately the same level as the spring eyes, but away from the main area of spring activity, contains between 60 and 90% silt.

This raises the important question as to whether the $CaCO_3$ levels of the Florisbad spring water, based on current-day ion levels, have always been historically undersaturated, stable and incapable of mineral deposition, or was the spring water more highly mineralized in the past? One of the primary reasons for suggesting that the spring water has remained stable and weakly mineralized is based on the historic relationship between the quality and the potability of the spring water. Although the Florisbad spring water is the most saline of South African spring waters at 2362 mg/l, it can be regarded as relatively sweet and potable for most animals (Douglas, 2001a). Judging from the quantities of fossilized faunal remains found at the Florisbad spring site, and the time span over which they have accumulated, this was obviously a preferred drinking site. It was also possibly one of only a few potable drinking sites within the immediate area, particularly during dry times. In order to make it such an attractive drinking site, not only for the animals, but also for human habitation, the TDS levels of the spring water must have been relatively low, even back to the genesis of the Old Collection assemblage. It would seem incongruous that the MSA inhabitants would have repeatedly established themselves at the spring site, even for short periods (Kuman et al., 1999; Henderson, 1995) if the water were not potable and had to be carried any distance from another source.

No evidence could be found in the literature where cool water springs, originating from shallow aquifers, have been responsible for the fossilization of mammal remains. A historically deeper hot water spring could have provided alternative processes for increased mineralization and the fossilization of faunal remains. Because of the possible shallow depth of the existing aquifer (500 m), the temperature of the water would never have risen much higher than it is today, which would also have had an influence on pH and the solubility of minerals in the rock formations through which the water passed. There is also no evidence, at this stage, to suggest that the Florisbad spring ever had deeper origins or was ever of a thermal nature, because many of the characteristics associated with thermal springs, even cooler ones, are not evident. In this regard, there is no evidence of travertine-depositing, sinters, silicified or otherwise mineralized microorganisms such as cyno-, filamentous-, and other bacteria, microstromatolites or microbial mats. High rates of mineralization are often exhibited in hydrothermal environments, favoring microbial fossilization and providing a rich storehouse of palaeobiologic information (Farmer, 2000). Allen et al. (2000) noted that organic matter is rare in carbonate sinters deposited at temperatures above 30°C because decomposition rates are very high in thermal environments. The presence of the organic layers at Florisbad indicates that, historically, temperatures were in all probability never much higher than at present (29°C).

Further historical indications of the spring water being relatively mineral poor are: the presence of the well-defined organic layers throughout the sequence, organic material within the sand layers, the presence of freshwater gastropods (Joubert and Visser, 1991; Visser and Joubert, 1991), and the long-term habitation of the site by largely aquatic dependant species such as hippopotami (*Hippopotamus amphibious*), lechwe (*Kobus leche*), water mongoose (*Atilax paludinosis*) and the clawless otter (*Aonyx capensis*) (Brink, 1987, 1988). Not to be ignored is the possible size of the Florisbad spring aquifer. Although no firm evidence exists at present, it is felt that the spring aquifer has always been of some considerable size, effectively stabilizing the mineral content of the spring water over long periods of time by evening out both long- and short-term mineral anomalies within the aquifer (Douglas, 2001a).

With regards to the original hypothesis that the spring water was solely responsible for fossilization, it can be concluded from available data that only one factor can be seen as being conducive for $CaCO_3$ precipitation from the spring water, and that is pH. As there appears to be no other processes, or other factors conducive for spring water fossilization, pH then becomes irrelevant on its own. Here, it can be noted that pH is often less important than the saturation level of $CaCO_3$. The undersaturated levels of $CaCO_3$ in the spring water therefore negates the statement by Brink (1987) that because the spring water was slightly alkaline, "the chemical nature of the spring water caused the remains of the Old Collection to be preserved in a characteristic way," resulting in their good preservation. Possibly, the only beneficial effect of the alkaline spring water was to mix with more acidic groundwater in the vicinity of spring eyes, elevating the pH of the groundwater to make it more conducive for $CaCO_3$ precipitation in a sedimentary context.

Discussion and conclusions

In light of the evidence presented here, it is apparent that, historically, the Ca, $CaCO_3$ and P levels of the spring water have never been high enough for the precipitation of carbonate and phosphate minerals. Even if $CaCO_3$ levels had been higher in the past, the high Eh environment created by the spring flow would have inhibited the precipitation of minerals. In addition to this, any influence the groundwater may have had on active spring vents would have been rapidly diluted by the spring flow. It would also have been impossible for fossilization to have taken place in the spring sands because the sands of the spring vents contain no organic matter or clay, and are effectively sterile, with no ability to adsorb, accumulate, carry or retain minerals for the precipitation of carbonate and phosphate minerals.

Although fossil remains were found in extinct spring vents as well as in areas of spring activity, this was largely due to all previous excavations having being carried out in areas of spring activity, with previous excavators not having recorded finds in situ (Brink, 1987). Had fresh remains become directly incorporated into the spring vents, it could be expected that abrasion through the agitation of quartz particles by the spring flow would have been considerably higher than the 29.6% recorded by Brink (1987), weathering fresh bones away before fossilization could have taken place. It is therefore contended that it was the migration and eruption of new spring eyes through layers of previously fossilized material on the original pan floor that dislodged fossilized material, effectively trapping it within the confines of the vents where no further fossilization would have taken place. Contrary to Brink's (1987) statement, evidence suggests that post-depositional mechanical and chemical weathering, such as hollowing, pitting, as well as the low incidence of calcite crystals found on and in bones, is actually the result of previously fossilized material becoming demineralized in the active spring vents. When spring vents ceased to flow, the mineralized groundwater must have replaced the spring water in the areas of the vents and demineralization arrested, with the possibility of calcite and authigenic apatite being re-deposited.

Of particular interest is the relatively low incidence of calcium carbonate deposition in and on the *Hippopotamus amphibious* and bovid bones (48.3%) from the Old Collection. This, combined with further evidence of hollowing (51%), and pitting (Brink, 1987), clearly indicates the dissolution of calcite and authigenic apatite caused by the spring flow, and an undersaturation of calcium carbonate and phosphates in the spring water. Phosphates in the spring water were non-existent during the high rainfall period, rising to only 1.93 mg/l in the normal rainfall period (Table 1 and Douglas, 2001b), making the precipitation of carbonate phosphates unlikely. The groundwater is essentially also devoid of phosphates, while there was a significant phosphate increase in the organic layers. Conditions for a low Eh environment appear to have been met in all the lower Florisbad organic layers through the action of slowly percolating groundwater, but these conditions would not have been met in the vicinity of the spring eyes. This provides further evidence that fossilization could not have taken place under the influence of the relatively fast flowing, undersaturated $CaCO_3$, spring water, where organic matter would have been rapidly oxidized and/or flushed from the sand. Hedges and Millard (1995) also expected dissolution to increase in sandy soils with flowing water.

It is evident that the extent of the reworked sediments will be determined by the historic migration and extent of active eyes, and not be restricted by the location of the current spring eyes and fossil vents. The implication is that large areas of the site, far greater than have been excavated at present, have been reworked through spring activity. Evidence of this was provided by Grün et al. (1996), who stated that in test Pit 3 (P2 in Fig. 1), some 80 m from the eyes of the indoor pool, the age scatter of ESR (Electron Spin Resonance) results indicated reworking of the sediments by spring action within the layers. Dreyer (1938) noted that spring gravel had been retrieved from an 18.5-m borehole sunk next to the residence at the highest point of the dune, again implying that spring activity previously extended well to the north of the current eyes.

Figure 3 illustrates and summarizes the mineralization and salinization processes within the Florisbad mound. In the contact zone between the spring water and groundwater, miscible displacement will be a continuous process (Fig. 3, Stage I) with ions moving from the spring water to replace ions in the groundwater, which are being adsorbed by the clay and organic material during drier periods. This process is reversed when ions in the sediments are flushed into the groundwater by periods of rainfall. The clay and organic material can therefore be likened a battery, accumulating, adsorbing and storing ions from the groundwater during low rainfall periods (Douglas, 2001b) (Fig. 3 Stage II). During high rainfall periods, when the groundwater flow increases and begins to rise and percolate, the accumulated ions are flushed from of the organic and clay layers back into the groundwater (Fig. 3 Stage III). This could also be likened to an ion pump. These latter two stages reflect variations in hydraulic potential over time, but only gradual in space (Hedges and Millard, 1995). The hung water table within the dune is related to the various layers of compacted clay and decomposed organic material that restrict the drainage of groundwater to the lower levels, as well as restricting the mixing of the spring and groundwater away from the spring vents. The diagonal line has been used to separate the normal and high rainfall period mineralization and salinization processes.

Brink (1987) suggested that calcium carbonate cements in the underlying geological strata were dissolved by acidic groundwater, mobilized and re-deposited, although the evidence presented here does not tend to support this theory, particularly the elevated pH of the spring water. To argue that the spring water may have had a high carbonate and/or phosphate content in the past is at this stage unfounded and lacking any evidence. A number of factors, including the possible contribution of minerals from the underlying Ecca shale and dolerite rocks, evaporation and pollution, all were ruled out as having had any influence on the high ion concentrations in the groundwater and sediments (Douglas, 2001a). The primary sources of pedogenic calcium are wind-carried particles, usually in the form of carbonates, and dissolved Ca^{2+} in rain water (Royer, 1999). Therefore, with the long history of aeolian deposition at Florisbad, the aeolian sand, clay and salts, which were blown by the north–west prevailing wind into the spring site from Soutpan and the surrounding areas, must have contributed significantly to the mineralization of the site (Douglas, 2001b).

Because of the number of variables involved in the fossilization process at Florisbad, it is clear that the degree of fossilization will have varied both laterally and vertically over the entire site. The original area of fossilization is most probably delineated by the old spring pan and by the extent of the organic deposits as defined by Grobler and Loock (1988). Due to the water table having restricted a more extensive examination of the basal layers, it is suspected that the existing Old Collection represents but a fraction of the true accumulation of faunal remains. The Florisbad sedimentary and groundwater environments would appear to meet all the requirements and conditions necessary for the precipitation of authigenic carbonate and phosphate minerals, and the permineralization of faunal remains. An analysis of REE and other trace elements could also provide further evidence of fossilization occurring in association with the groundwater. It is concluded that fossilization at Florisbad took place within the organic sediments and clay and under the influence of the groundwater, with the potential for, and degree of, fossilization increasing proportionately to the quantities, proximity and degree of salinization of the clay and organic matter.

The salinization and other processes discussed here may have applications at other fossil sites where organic material and clay occur, as well as at coastal sites where there is a relatively mineral rich marine environment that may exert an even greater influence over both salinization and fossilization. This is particularly true in light of the observation by Deacon (1998, 2001) that it was a South African pattern that Acheulian sites were associated with wetland habitats such as springs, pans, and along river valleys.

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Appendix

IV

Formation of the Florisbad spring and fossil site – an alternative hypothesis.

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Formation of the Florisbad spring and fossil site — an alternative hypothesis

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Abstract

In considering that fossilization could not have taken place in the areas of the spring vents at Florisbad, and that the spring-water was not responsible for fossilization, this hypothesis provides an alternative environment for the fossilization of faunal remains. Based on examination of the site, studies on various aspects of the site, as well as previous hypotheses on the formation of the site, an alternative theory on the formation of the Florisbad spring site is presented. It is contended that a large spring pan developed around the springs, which was later largely covered, and transformed into a dam, by the migration of the Florisbad sand dune. When the water rose to near the top of the Florisbad dune dam, the eastern arm of the dune was breached, evacuating the contents of the dam in a flood of water and sediments. This caused the ephemeral drainage line to the east, to be modified into a flat-bottomed marshy vlei, where the majority of the dam sediments were deposited. Previously largely unanswered and debatable questions such as the erosion of the eastern side of the spring site, the 7 m of clay deposits recorded in the modified vlei area, and aspects of the Peat IV layer are addressed. The importance and influence of the panveld and sand dunes in the formation of the spring site are also examined.

Keywords: Florisbad; South Africa; Spring site; Formation; Alternative hypothesis; Mechanisms

1. Introduction

1.1. Background

The Florisbad spring site (28°46′S 26°04′E) is located 49 km north-west of Bloemfontein, Free State Province, South Africa, with the 500 mm isohyet passing slightly to the east. The 78-year annual rainfall average was 496 mm, with annual rainfall being extremely variable [8]. A maximum annual rainfall of 957 mm was recorded for 1988 and a minimum of 271 mm in 1965 [8]. Florisbad is the type-site for an important accumulation of fossil remains representing the Florisian Land Mammal Age, which falls between the Middle and Late Pleistocene. Faunal remains fall into two basic categories, an earlier Middle Stone Age (MSA) Old Collection, Electron Spin Resonance dated at 281 000 ± 73 000 yr, and a later MSA human occupation horizon dated at 121 000 ± 6000 yr [18]. Six mammalian Orders, represented by 27 species, have been recorded from the Old Collection, with five species being extinct, and the rest being distinguishable from extant species by anatomical features [1,2]. Occurring within the lower sediments was the Florisbad hominid skull dated at 259 000 ± 35 000 yr [18]. Besides a wealth of artefacts, four mammalian Orders, represented by eight species, have been recorded from the MSA human occupation horizon [1].

The basement rocks of the area are of the Ventersdorp Supergroup overlying older granite and gneiss, which in turn is overlain by a Permian Age Karoo sequence of the Ecca Group shale into which dolerite dykes and sills have intruded [1,26]. It is at such a dolerite intrusion that the Florisbad springs have formed. The springs originate from an aquifer of unknown origin and depth, with it being estimated that in order for the spring-water to issue at 29 °C, recharge water would have had to have percolated to a depth of approximately 500 m

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and reaching a temperature of 32–35 °C [8,15]. There is no outcropping of bedrock on the farm while the surface geology is composed of an unconsolidated mantle of red-yellow and pale bleached aeolian sand of varying depth [1,26].

Florisbad is located on the eastern boundary of the western Free State panveld, with many researchers having sought to explain the origins and formation of this panveld [6,14,17,25,27,28]. The panveld covers an area of 1227 km^2, with some 18 803 pans having been recorded [13], with a concentration high of 82.1 pans per km^2 [7]. The derangement of drainage as a possible cause for the formation of the pans was suggested by Geyser [14], but Le Roux [25] disagreed with this, stating that it could not account for the wide distribution of pans, and that wind was the only agent that could have been responsible for the formation of most pans. Grobler et al. [17] regarded pans formation as being largely a result of climatic causes and stated that whatever the causes, the hydrodynamic equilibrium of the streams was altered from an actively degrading drainage to an aggrading situation. When dry periods occurred, prevailing winds then deflated the sediments, forming hollows, which were the initial sites of the pans [17].

The panveld was examined within a morphotectonic framework by Marshall [27], where it was proposed that the panveld was developed in the Miocene, but that the landscape cycle had been disturbed by structural upheavals in the Pliocene. These upheavals resulted in what was described as a tectonically disturbed major palaeodrainage system, resulting in the down warping of the Palaeo-Kimberley River [27]. The formation of the pans was seen as the result of this disrupted palaeodrainage system, with further modification taking place largely through deflation [27]. Van Zinderen Bakker [32] was of the opinion that Jurassic dolerite intruded the Karoo beds and formed sills and dykes, and that during subsequent erosion periods, pans such as Soutpan, a large salt pan north-west of the spring site (Fig. 1), were formed.

It is recognized that the southern African dune system and the Sahel of West Africa were both established during multiple arid phases since the last interglacial [31]. Therefore, climate plays a critical role in dunefield dynamics and the interplay between dune surface erodibility and atmospheric erosivity [5, 31], and thus the mobility of sand dunes. Newly formed crescent dunes, or lunettes, in the Florisbad area were formed on

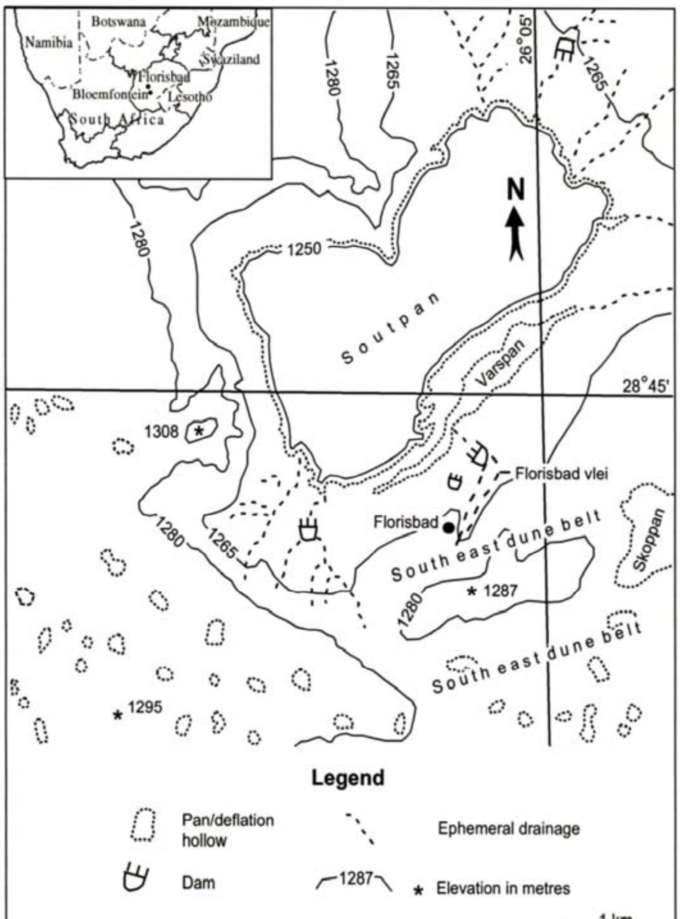

Fig. 1. The location of Florisbad and a map of the surrounding area showing the topography and other features as discussed in the paper.

the lee side of deflation hollows by fine unconsolidated material being blown by a prevailing north-west wind [26,32]. Such dunes are easily recognizable on aerial photographs, and in the field, from their peculiar vegetation cover, colour, shape and topographic expressions [26]. The huge Soutpan lunette, which drapes over the Florisbad site, varies from other dunes in that it has a higher brackish soil content that has been derived from Soutpan [26].

Rubidge and Brink [30] described the lithostratigraphy and depositional history of the Florisbad sediments as still being in their initial stages of investigation, and that several models could be proposed, with a need for testing at the site. Kuman and Clark [23] noted that the spring cycles, which are recorded in a repeated alternation of sands, silts and organic deposits, are poorly understood and controversial. The stratigraphy and sedimentology for a relatively small part of the Florisbad spring site have been documented, and it is hoped that this hypothesis will go some way in resolving some of the issues.

1.2. Previous hypotheses

Brink [1] believed that the spring site developed through sand being brought to the surface by spring vents with vegetation developing around the margins of the spring pools. The vents then became blocked, with the size of the mound increasing due to a combination of factors such as choking vegetation, the deposition of windblown sediments and a diminishing supply of, by what was referred to as, groundwater (spring-water) [1]. After this closure, the spring eye moved laterally along the bedrock fissure to form a new passage, which cut through the existing strata to form another sand unit, partly overlapping the mound created by the original vent [1]. It was contended that these processes would have repeated themselves to form alternative layers of organic and non-organic deposits [1].

The site was described by Butzer [4] as a 7 m mound of spring beds interbedded with organic intrusions, and only partly covered by aeolian material. The site was seen as developing through spring flow, which was determined by a deep-seated regional aquifer and fluctuations in recharge, with the bulk of quartz grains being represented by detrial sands originating from the underlying Ecca shale and subsurface dolerite, through which the spring waters had passed [4]. Sandy pools developed at the spring site, with vegetation and peaty organic horizons developing as vegetation encroached on less active springs during periods of low discharge [4]. Vegetation was later submerged by periods of more active spring discharge, and subsequently buried by spring sediments [4]. Dreyer [11] suggested that the Florisbad deposits represented sand output from the spring and considered the Florisbad "mound" as being formed by the sand from a huge (unknown) eye beneath the highest point of the "mound". This would suggest that the spring pan extended well to the north of the current eyes.

It has been contended that a close relationship and correlation existed between the spring sedimentation and the shoreline position of the adjoining palaeolake complex (Soutpan) and that this relationship played a significant role in the modification of the spring mound sediments [21,22,24,34]. It was proposed that deposition at Florisbad was directly related to the palaeolake levels, which reflected climatic conditions at the time [22,34]. This involved cyclic sedimentation with soil horizons forming during arid stages when palaeolake levels were low, while deposition of palaeolake bottom silts occurred during wet periods when the spring area was flooded by the palaeolake [22,34]. These cyclic transgression and regression sequences of the palaeolake shoreline were translated into four low water level phases and three high phases [22]. It was noted that water-logged conditions had existed at levels higher than the existing water table, and that load structures had been identified as high as the top of Peat II [1]. These load structures as well as faunal evidence led to the suggestion of the existence of a large water body in the past, but whether or not this was related to the water levels in the palaeopan still needed to be established [1].

Sedimentary deposits to the east of the spring site have been referred to as the "lacustrine sequence" [22,34] because of their presumed association with the Soutpan complex. Based largely on the presence of freshwater gastropods in the clay facies, these sediments were interpreted as being palaeolake bottom deposits directly related to Soutpan [22]. It was also suggested that during humid phases the Soutpan palaeolake complex enlarged to more than twice its present size [34]. Despite the flooding of the palaeopan theory, Visser and Joubert [34] noted that the distribution and lateral variation of the organic-rich deposits at the spring site reflected water-logged, or bog conditions, on a poorly drained flood plain, reflecting the local influence of constant freshwater discharge at the spring. Fourie [12], who originally described these deposits and the freshwater gastropods, did not see them as being related to the palaeolake.

Kuman et al. [24] suggested that a number of micro-environments existed in response to the waxing and waning of the spring, as well as the expansion and contraction of the nearby palaeolake. Based on grain sample size and distribution, it was concluded that most of the sediments not only accumulated under uniform, low-energy, subaqueous environmental conditions, but also under several composite geomorphic regimes, and had not been reworked after deposition [24]. It was further noted that the sand layers were usually more than 10 cm thick, with an absence of thin layers, as in annual or shorter, cyclical deposition systems [24]. The sharp bedding contacts of the Pleistocene levels were seen as indicating sudden changes in environmental conditions due to the possible increase and decrease in the spring-water discharge, with the stratigraphic layers reflecting changes in the environmental water regime, from open dam conditions to vegetated marshland [24].

Meiring [29] noted that in part of the 1952 excavation, none of the layers were disturbed, but continuous throughout the exposed face. Both Kuman et al. [24] and Douglas [8] believed that the historic palaeotemperature of the spring-water was similar to the present, with Kuman et al. [24] finding no evidence of stromolites, chemical, or biochemical precipitation, as in hot spring environments. Douglas [8,9] noted that because of the low temperature of the spring-water over time, as well as its low mineral content, diagenesis of the underlying Ecca and Dwyka formations had contributed little to the mineralization of the site. Based on sand grain shape and surface features, Kuman et al. [24] and Van Zinderen Bakker [32] concurred that the spring sediments appear to be derived predominantly from an aeolian source, with even sands from the lower levels, which showed signs of water transportation, being originally of an aeolian nature [32].

2. Hypothesis

A plan of the Florisbad spring site is presented in Fig. 2 with contours indicating the sand dune. Owing to the complex stratigraphy of the site any cross-section would only represent that particular section, therefore, Fig. 3 represents a simplified cross-section indicating the major horizons discussed in the paper. Fig. 4 represents a schematic plan of the developmental stages of the Florisbad spring site, 1a–5a, while Fig. 5 provides a profile of the same developmental stages from a south to north perspective. The legend for Figs. 4 and 5 is presented in Fig. 6, while the residence at Florisbad, which is located on the top of the Barchan dune, and spring eye, are indicated in Figs. 4 and 5 as reference points.

Where the spring eyes originally surfaced is considered irrelevant in the context of this hypothesis as evidence indicates [11,18] that the spring eyes have, over time, surfaced over a fairly extensive area within the confines of a greater spring pan. A greater spring pan can be roughly delineated by the extent of the organic deposits in Fig. 4 (Stage 4a), after Grobler and Loock [15]. It is proposed that, for the purpose of this hypothesis, the spring eyes were originally located, more or less, in the area of current spring activity, or possibly slightly to the north. This is close to the base of the windward slope of a line of sand dunes lying south and south-east of the spring site, and referred to here as the south-east dune belt. It is also assumed that the spring aquifer and springs originated around the time of the tectonic disturbances [27], prior to the formation of the panveld. These tectonic disturbances then produced a landscape conducive for further later modification by wind into a panveld.

The effect of tectonic disturbances can be illustrated by an earthquake, which occurred at Fauresmith, 130 km west of Florisbad in 1912, when it was reported that the Florisbad spring discharge increased fourfold after the quake [16]. Therefore, if the aquifer and springs evolved before the formation of the panveld, this could imply that there was already an existing spring pan when the south-east dune belt formed to the north-west. This dune belt would have migrated over the spring site, effectively blanketing the area of spring activity to create a fossil spring pan below the sand dunes. Additional wind activity would have continued to shift the dune belt further south-east, re-exposing the old spring site and fossil pan.

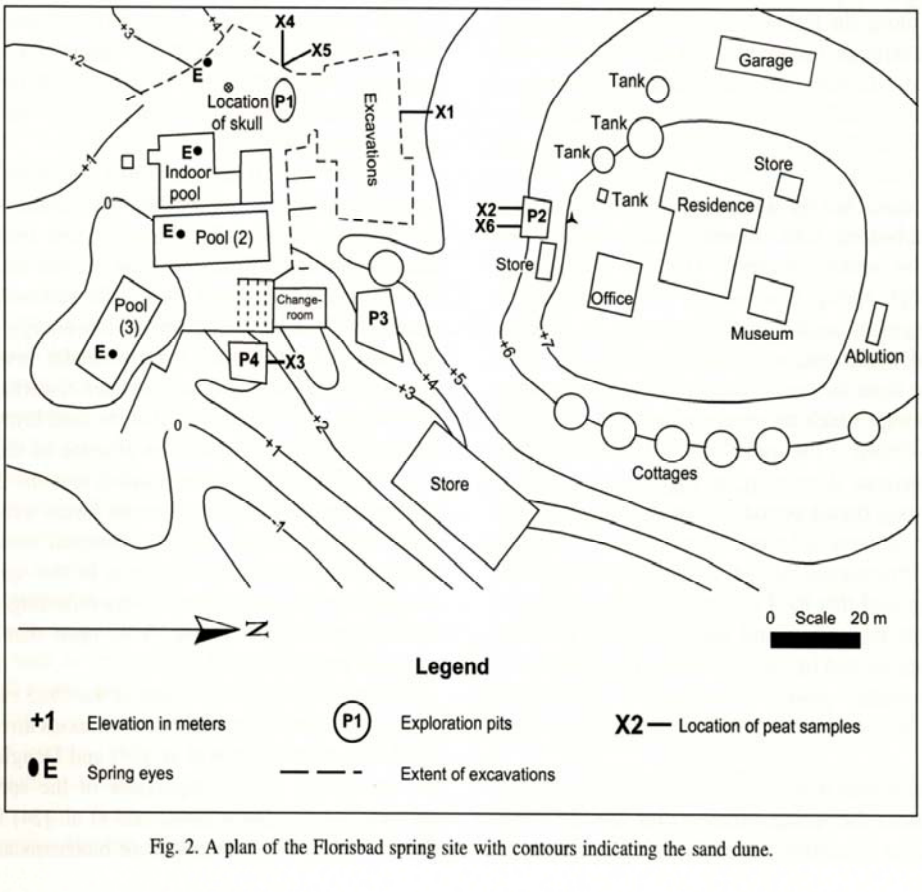

Fig. 2. A plan of the Florisbad spring site with contours indicating the sand dune.

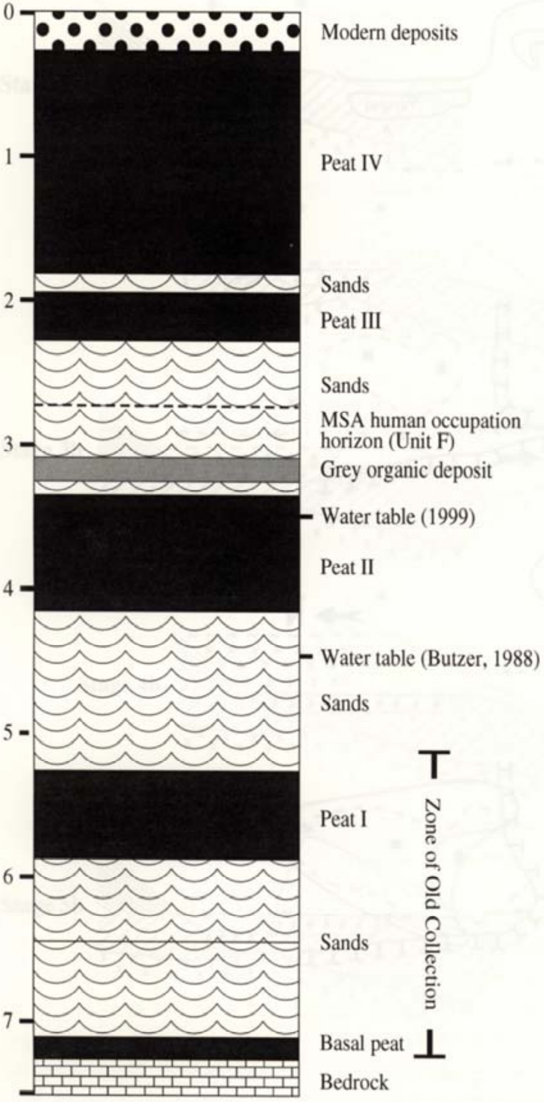

Fig. 3. A simplified cross-section of the Florisbad deposits towards the west wall of the excavations (see text) indicating the major stratigraphic layers as discussed in the paper.

At this time the springs would have resurfaced, and through deflation and spring action, the fossil pan would have started to take on its pre-existing form.

A number of variables can be associated with the progression of the sand dunes over the site, but these are not thought to have influenced the final result. For example, the distance between the south-east dune belt and the Florisbad dune, as they moved south-eastward, would have determined the growth and ultimate size of the spring pan. Had the two sets of dunes been close together, the spring pan would not have had time to grow significantly between the passing of the south-east dune belt, and the Florisbad dune. The opposite would have been true had this distance between the two sets of dunes been greater. Another unknown factor is whether the two sets of dunes were formed during the same dry period, or did the south-east dune belt become stabilised during one wet period, with the Florisbad dune being formed during a subsequent dry period. This would also have had an effect on the ultimate size of the spring pan. Figs. 4 and 5 illustrate the latter proposal, while if the former is considered more correct, then the Florisbad dune in the illustration should simply be seen as being much larger than illustrated.

2.1. Stages 1a (Fig. 4) and stage 1b (Fig. 5)

Stage 1a (Fig. 4) and Stage 1b (Fig. 5) make the presumption that the modern spring pan originated where the pre-existing springs resurfaced within, or close to, a fossil pan, or deflation hollow, after the windward side of the south-east dune belt had moved over the site. Further growth of the modern pan would then have roughly followed that of the fossil pan until it began to develop its own character in the new environment. The modern spring site would have gradually become enlarged through large numbers of animals drinking at the site, as well as through deflation. Douglas [9] saw the bottom sediments as being built up largely through the deposition of aeolian sands and clays, some spring sand, the trampling of sand by herds of animals, the contribution of waste material from such herds, and the growth and decay of aquatic vegetation. It is thought that, at this stage, the spring pan may have been somewhat larger than illustrated.

2.2. Stages 2a (Fig. 4) and 2b (Fig. 5)

As the spring pan gradually increased in size due to the above-mentioned factors, so did the Florisbad dune to the north-west of the site, migrating steadily south-eastwards towards the spring pan. Although there must have been extended dry periods in order to allow for the continued formation of the Florisbad dune, and its migration, there must also have been wet periods which allowed for further vegetative stabilization of the south-east dune belt. By this time the spring pan had grown to some considerable size, supporting aquatic dependant species such as hippopotamus (*Hippopotamus amphibius*), lechwe (*Kobus leche*), clawless otter (*Aonyx capensis*) and water mongoose (*Atilax paludinosus*) [1,20,21].

2.3. Stages 3a (Fig. 4) and 3b (Fig. 5)

At this stage the Florisbad dune started migrating across the spring pan, blanketing it, and halting any further northward expansion. With the arms of the dune now beginning to encompass the spring pan from the sides, the initial stage of a damming effect was begun, resulting in the spring pan having to extend itself eastwards, southwards and westwards. This expansion possibly caused some seepage and overflow into the ephemeral drainage line, but this is not thought to have been significant. As the Florisbad dune continued south-east, further expansion of the spring pan resulted in it capturing the ephemeral drainage to the east, thus increasing the flow of fresh water into the spring pan during wet periods. It is suspected that

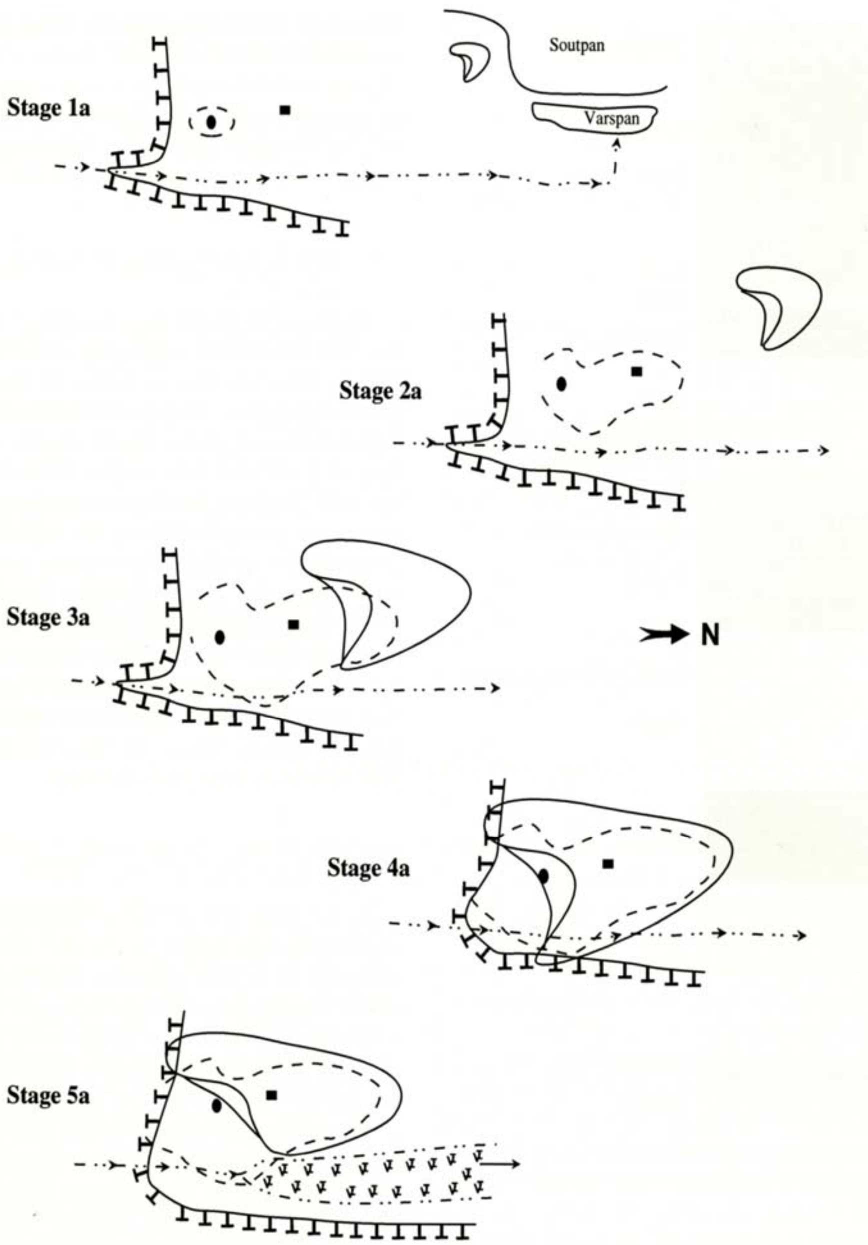

Fig. 4. A schematic plan of the proposed developmental stages of the Florisbad spring site (not to scale).

at this stage the size of the spring pan was considerably greater than the distribution of organic matter as indicated by Grobler and Loock [15] in Fig. 4, Stage 4a, and that the final spring pan extended further south, east and west. This would have been because drilling programmes would have picking up fairly significant accumulations of older organic material, but not more recent expansions to the pan where organic matter may not yet have had time to accumulate in significant quantities.

2.4. Stages 4a (Fig. 4) and 4b (Fig. 5)

Eventually both arms of the migrating Florisbad dune met, and over-rode, the windward base of the south-east dune belt, completely cutting off any possible outlet for the spring pan water. Although the spring pan was always a self-contained, drainage-impeded pan, at this stage it became a dam, completely enclosed by the dunes. As the Florisbad dune continued its south-eastwards migration, covering more and more of the original spring pan, the surface area of the spring pan continued to contract, with the dam becoming relatively small in relation to the spring pan. This resulted in a corresponding slow increase in the depth of the water through contributions by the spring flow and the ephemeral drainage line. Judging by the thickness and relatively fresh condition of the Peat IV layer, there is not thought to have been any rapid rise in the water level, but rather a gradual process allowing for the

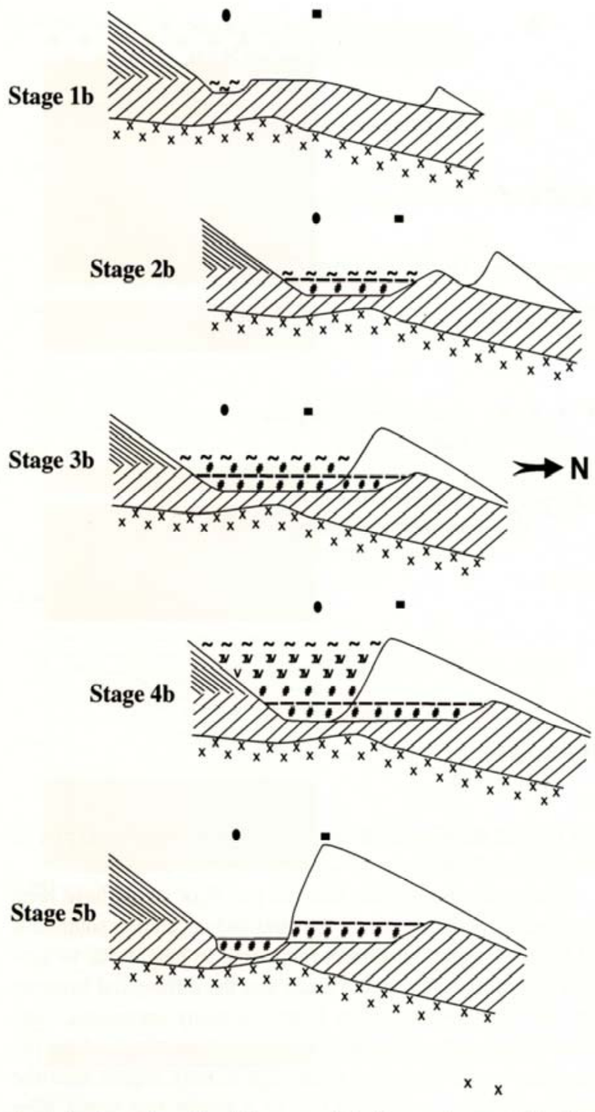

Fig. 5. A schematic profile of the proposed developmental stages of the Florisbad spring site (not to scale).

slow build up of aquatic vegetation in a marshy environment. When the progress of a sand dune is halted, it can still be deepened and expand at right angles to the prevailing wind [17]. The arms of a Barchan dune might also extend to such an extent that their total obstructive power becomes equal to that of the middle of the dune [19]. This is what appears to have occurred when the arms of the Florisbad dune came to rest up against the windward face of the static south-east dune belt to form a dam.

2.5. Stages 5a (Fig. 4) and 5b (Fig. 5)

As the water and sediments increased in depth, so did the pressure against the arms of the now almost stationary Florisbad dune. At this stage three possible situations arose, with all having essentially the same result. The first possibility was that the water rose to such a height that it began to flow over the top of the eastern arm of the dune, eroding the wall until it was breached. Secondly, the weakest point in the Florisbad dune wall was provided by the angle of the eastern arm against the south-east sand dune belt, eventually breaching as the water neared the top of the wall. Thirdly, the herds of animals drinking from the site may have eroded part of the eastern dune arm to the extent where it eventually overflowed and then burst. In any event, the water and sediments then evacuated the eastern side of the dam in a flash flood, eroding away most of the eastern arm, and draining down the already existing ephemeral drainage line of the dune valley. This resulted in the ephemeral drainage line being gouged out from the point of breach, severely modified, and the sediments from the spring dam being re-deposited to form a typical wide, flat-bottomed flood plain, now referred to as the vlei.

When the water evacuated the dam site to the east, it also took with it most of the sediment contained in the dam as well as a portion of the leeward face of the Florisbad dune. However, as the western wall of the dam was furthest from the breach, erosion in this area was at its lowest, leaving this area largely intact with an impression and record of sedimentation similar to that in Fig. 5 (Stage 4b), up against the face of the dune. A thick layer of basal sediments would also have been left behind at the dam site, but quantities of fossil and artefact material would also have been washed out and redistributed in the sediments along the length of the modified vlei.

3. Discussion and conclusions

The hypothesis put forward here would question previous depositional theories put forward by Brink [1], Butzer [4] and Dreyer [11]. Overlapping, convex mounds created by sand from the spring eyes and vegetation would have created a distinct disjunction between the organic layers as each convex mound formed over the other. On the other hand, the organic layers have formed in extended horizontal sheet like layers, signifying a larger aquatic environment. There also appears to be a general consensus in the literature that the majority of spring site sands are of aeolian origin, and not from the spring eyes. That some authors feel that the Florisbad dune is composed almost entirely of quartz sands derived from the underlying rock formations, and deposited by the springs, must be questioned. It is argued that when the springs erupted through the overlying aeolian sediments, the spring flow flushed all clay and organic material from these sands, leaving behind the heavier quartz grains. As vents, which were later capped, did not penetrate to the surface [1], they were to all intents and purposes contained in a completely enclosed environment. This would mean that the existing aeolian sand above the eyes would have had to be replaced by spring sands in the closed environment of the vent. If this is in fact what occurred, what then happened to the aeolian sand? It is therefore contended that the quartz sand in the vents is mostly washed aeolian sand.

This then brings into question the status of the spring mound. It is maintained that, from evidence presented here,

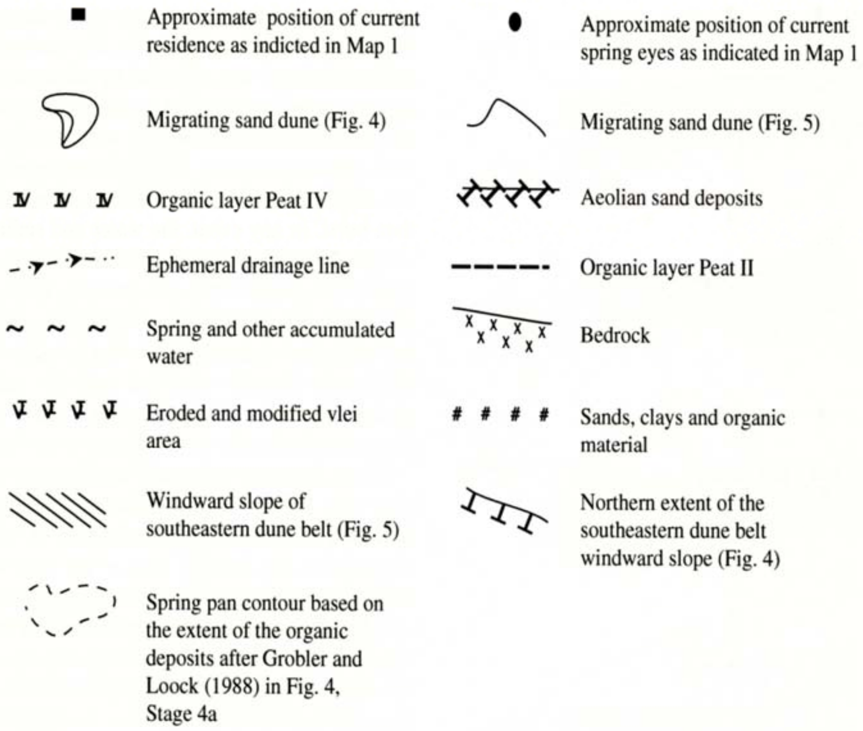

Fig. 6. Legend for Figs. 4 and 5.

there is no such feature as a "spring mound" at Florisbad. The term "spring mound" does not appear to have been defined in the literature, with the possible exception of those few researchers who feel that the Florisbad sand dune is composed entirely of material emanating from the springs. Therefore, literature references to the "spring mound" may refer to either material emanating from the springs, the sand dune, accumulated deposits from the time of the spring pan and dam, or deposits from the palaeopan. It is maintained that there was never sufficient material emanating from the springs to produce a "spring mound" of any type, and that the spring site deposits are almost entirely a result of aeolian deposition and the growth and decomposition of aquatic vegetation resulting from the spring pan, and later the spring dam.

Should the flooding of the palaeopan theory be a consideration [22,34], including the suggestion that the palaeolake may have enlarged to twice its size [34], this would have meant that the water in Soutpan would have had to rise to at least the level of the top of Peat IV (Fig. 5, Stage 4b) to ensure its formation. An even flatter past topography was pointed out by Grobler et al. [17], who stated that, even prior to the formation of the pans, the area was already in a state of low topographic relief. It has been noted that there was only a 19-m decrease in elevation across the panveld from Florisbad westward, over a distance of some 120 km [7]. In fact, there was most probably not much difference between the level of the spring site and the Soutpan floor when the weathering process began. When considering an even flatter topography in the past, it is evident that any significant increase in the water level of Soutpan would have resulted in a sheet of water spreading far beyond the Florisbad spring site, covering a large part of the western Free State, of which there is no evidence.

During this period, the Soutpan pan floor would have been continually deflating, while the sand and vegetation south-east of Soutpan would have been increasing in height due to aeolian deposition. This would mean that the differential between the two levels would have been constantly increasing, suggesting that each subsequent high water level phase of the palaeolake would have had to be significantly higher than the previous phase in order to rise to a height that could have flooded the spring area. In any event, after the Florisbad dam had formed, water from the palaeolake would not have been able to enter the spring site unless it rose to a height higher than that of the Florisbad dune. Currently the top of the Florisbad dune is approximately 25 m above the Soutpan floor. Had water from the palaeolake in fact flooded the spring site at this stage, it would not have been able to escape, and Peat IV would not have been able to develop under consistent high water levels. Kuman and Clark [23] stated that most material was in the fine sand to silt range, indicating hydraulic transportation of material by suspension under low-energy conditions, and not as fluvial bedload. With so many pans in the area, any such flooding would have filled these pans and other low lying areas with relatively fresh water, which would not support the concentration of game at Florisbad. Brink [1] noted that evidence suggests the existence of a large body of water, but whether this was related to higher water levels of the palaeopan still needed to be established. It is proposed that this large body of water represented the spring pan.

It is clear that major erosion has occurred on the eastern side of the site, and yet, little mention of this is made in the literature. However, this erosion provides the greatest support for the breaching of the dam theory. Brink [1] and Rubidge and Brink [30] noted that the upper red-brown sand units were absent and suggested that these had either not been deposited, or been eroded away by spring action. This hypothesis confirms that these units were in fact eroded away when the arm of the dune was breached, and not by spring action. To the east of the site Butzer [3] recorded a 7-m-thick layer of clayey sediments along the stream (modified vlei) draining from the spring to the north, and interpreted this as being part of an extensive marsh terrain. However, the current theory would indicate that this thick layer of clay was a result of the spring dam sediments being flushed from the spring site and re-deposited in the current modified vlei area. The modified vlei extends onto the neighbouring farm, where it turns westwards, continuing towards Varspan, a considerably smaller, less saline pan, lying east of Soutpan. In all probability Varspan was partially formed as a result of the deposition of sediments from the spring site. The floor of Varspan is covered by green grey gley [15], which could correspond to the 110-cm-thick argillaceous green sand layer [34], and other similar layers, at the spring site [34]. These could have either flowed into Varspan with the initial flash flood, or later been washed down the drainage line of the vlei by rainfall.

The effect of flash flood erosion on the ephemeral drainage line is also consistent with the above theory. When the profiles of the ephemeral drainage line, above and below where the proposed breach occurred, are compared, it is clear that below the proposed breach, the current wide flat profile of the vlei is totally uncharacteristic of erosion in the upper reaches of, what is in effect, a small drainage line, particularly over such a short distance. It is also improbable that such a marshy terrain could have developed in the sandy ephemeral drainage line, but this would have been possible on the flood plain of the modified vlei. In any event, the ephemeral drainage line in its original state would not have had the capacity, or the velocity, to transport such large quantities of material in suspension. Besides this, the area above the proposed breach is composed almost entirely of aeolian sands, with no reserves of clay material.

The initial main thrust of the flash flood through the breach would have been along the eastern-bank of the ephemeral drainage line, where it has cut into the bank leaving a steep eastern embankment. This embankment was clearly much deeper cut when the breach occurred, but has over time, gradually built up with sand from higher up the ephemeral drainage line, aeolian deposits, and the growth and decay of aquatic vegetation. On the other hand, the far more gradual sloping western bank, on the inside bend of the vlei, is consistent with load deposition of the spring pan sediments during flooding. As the initial force of the flash flood abated downstream, the flow would have followed the natural drainage line towards Varspan.

In light of the breaching of the eastern wall of the dam, it is proposed that the lacustrine deposits on the eastern side of the spring site are in no way related to rises and falls of the nearby palaeolake. If rises and falls in the palaeolake were responsible for the depositional sequences at the spring site, because of the aquatic environment, these should be replicated in the lacustrine deposits, or visa versa. According to Fourie [12], well defined turf and clay layers were absent from the lacustrine deposits, with these sediments being described as either sandy clay or clayey sand. From borehole results, Kuman and Clark [23] concluded that the fewer organic layers and greater quantities of clay indicated a more deeply flooded terrain. This would again support the breaching of the dam hypothesis where the sand, clay and organic material would have been mixed on evacuating the site. Another factor supporting the breached wall theory is the abundance of freshwater gastropods and terrestrial snails found in the vlei deposits [12,22,34]. These were found in the basal portion of the sand layer of the vlei, with numerous bones in the lower calcrete band, but not in either the upper or middle bands [12]. This would suggest that the heavier shell and bone remains, flushed from the spring site, settled out in the muddy environment when the dam wall breached.

Contrary to this, Visser and Joubert [34] interpreted two lacustrine layers as being organic rich, and despite the prominence and importance of the organic-rich layers as marker horizons at the spring site, these were described as being unreliable. However, these two organic-rich lacustrine layers [34] do not correspond horizontally to any of the spring site organic layers, which they should have, if all were laid down under the same aquatic environmental conditions, namely, the flooding of the palaeopan. There is also no explanation as to the considerable difference in height (5–7 m) between the top of the lacustrine deposits and top of the spring deposits, namely, Peat IV, with neither of the lacustrine limestone layers being reflected in the spring deposits. However, references to the lack of sedimentary structures, poorly sorted material, water-reworked silt, and flooding, which were put down to rises and falls in the palaeolake [34] may well reflect the results of the breaching of the dam wall and deposition of sediments in the vlei. The five species of freshwater gastropods (four genera) recorded by Visser and Joubert [34], were also only found in the lower limestone marker at the bottom of the sequence.

The gradual rise of water to the top of Peat IV after the formation of the dam is also supported by the current theory. In order for Peat IV to have come within 650 mm of the surface and attain a thickness of approximately 2.8 m on the western wall of the excavation, it must have been in a saturated developmental stage over a long period of time [9]. This would mean that there must have been a corresponding gradual rise in the water level in order to allow for the growth and formation of the Peat IV layer. It is clear that the north to south dip of the contact between Peat IV and the underlying sand against the western wall of the excavation represents the formation of peat against the leeward face, or western arm of the Florisbad dune. The pinching, or lensing out, of Peat IV to the south probably represents greater erosion on the deeper fractions of this layer when the eastern wall was breached.

The stratigraphy of the Florisbad spring site is complex, with Rubidge and Brink [30] noting a low degree of correlation between adjacent auger boreholes, stating that the deposits showed considerable facies changes that were inclined to wedge out laterally over very short distances. This lack of correlation was confirmed by Kuman et al. [24]. Rubidge and Brink [30] further concluded that, because the deposits are lithologically variable they were a product of an unusual depositional environment at the site. Besides the accurateness of interpreting large diameter auger samples, particularly where narrow, rapidly changing, and interrupted horizons are being examined, it is hardly surprising that correlation was found to be difficult.

Firstly, there is the horizontal stratification of the sediments resulting from aeolian deposition and the aquatic environment of the spring pan, and later the spring dam. Secondly, there is the angle of internal cross-stratification of the sand dune where the angle of repose on the leeward slope is usually between 30 and 35°, but as the dune migrates, the angle of internal stratification decreased down the windward slope. Thirdly, as the water in the dam began to rise in Stage 4a, the rising water may have caused the steep leeward face of the Florisbad dune to slump into the dam, decreasing the angle of the leeward slope. This would also have resulted in the dam floor being raised, and dam sediments being allowed to built up horizontally where the dune face once was. It is therefore proposed that the horizontal sedimentary layers currently visible on the northern face of the excavations will pinch out further north where the leeward face of the dune will be encountered. Finally, there is the reworking and re-mixing of the sediments through the migration and eruption of spring vents through already established layers, as well as possible earthquake induced liquefaction mentioned by Visser and Joubert [33]. In parts of the site the influence of more than one of these factors may have come into play.

The influence of the south-east dune belt, the Florisbad dune, and the formation of the panveld and spring pan, are all considered to have played major roles in the formation of the site, but appear to have been largely ignored in most other theories. Development of the western Free State panveld is also considered crucial in the formation of the original Florisbad site, with faults and fissures developed during periods of tectonic activity resulting in the establishment of the Florisbad aquifer and spring. Although the extent of the Florisbad aquifer is unknown at this time, it is thought to be of some considerable size, possibly having a number of recharge points [8]. Because of its suspected large storage capacity, the spring flow would not necessarily have corresponded directly to alternating wet and dry periods [8]. Therefore, the storage capacity of the aquifer, when fully charged from wet periods, would allow for spring activity well into dry periods, and possibly even through to the following wet period if the two periods were close enough to each other [8]. This continuous supply of water may be yet another reason why the site was a preferred drinking site. It is thought that the large size of the aquifer evened out factors such as spring flow and mineralization over time, with rainfall having little significant effect on these factors [8]. Butzer [4] noted that spring cycles varied considerably in terms of energy, but did not necessarily wax and wane in a predictable fashion.

While there are many factors contributing to the formation of the modern spring site, the development and migration of the sand dunes are undoubtedly one of the most important. Besides the Florisbad dune being responsible for the damming of the spring pan, had the south-east dune belt not been stabilised at that specific time in its current location, it would not have been in a position to retard the migration of the Florisbad dune. This would have meant that, had the Florisbad dune migrated a further 50–100 m south-east, it would have completely covered the spring eyes, effectively re-capping them, and possibly preventing any spring flow. Under these circumstances there may never have been a spring site with associated fossils at Florisbad.

Perhaps the most significant aspect of this hypothesis is that it allows for the development of a sedimentary sequence, with a resulting environment, conducive to the fossilization of faunal remains. This is particularly pertinent in light of Douglas [10] stating that fossilization could not have taken place in spring vents or areas of spring activity, as previously thought [1,2], because amongst other factors, the spring-water was currently and historically under saturated in $CaCO_3$, with a high Eh due to the spring flow. Douglas [10] showed that fossilization actually took place in the organic and clay rich sediments, in conjunction with highly mineralized groundwater.

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