Hydrogeology of the Sterkfontein Cave System, Cradle of Humankind, South Africa

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Abstract

A water level rise of almost 3 m in the space of two years in the Sterkfontein Cave system since late-2009 necessitated the re-routing of the tourist path through the cave to successively higher elevations on three occasions. It also raised concern for a possible association with copious acidic and sulphate-rich mine water drainage from the West Rand Goldfield (a.k.a. Western Basin) starting in early-2010, and the related threat to the UNESCO-inscribed fossil site. Although these circumstances have had little impact on the tourist value of the site, a prognosis of the impact on cave water level and quality is indicated by virtue of its karst setting and palaeontological significance. Historical and recent potentiometric data, together with ancillary hydrogeological and hydrochemical information acquired in the course of a water resources monitoring programme for the broader Cradle of Humankind World Heritage Site, provides new insight into the hydrogeology of the cave system. An improved understanding of the hydrophysical and hydrochemical response of the cave water system sheds light on the location of this system within the water resources environment. It is proposed that the present-day maximum cave water level is constrained to an elevation of ~1440 m above mean sea level. The recent electrical conductivity of 78 mS/m for cave water is 32% greater than the 59 mS/m recorded in mid-2010 and earlier. Similarly, the recent sulphate concentration of 161 mg/L is 178% greater than the 58 mg/L recorded before 2010. Compared to coeval values for ambient karst groundwater represented by the normative Zwartkrans Spring water, the magnitude of the increases in the springwater are similar, viz. 48% (from 84 to 124 mS/m) in salinity and 166% (from 154 to 409 mg/L) in sulphate. Although a distinct mine water impact is evident in both instances, the values indicate a muted impact on the cave water chemistry compared to the springwater. These and other documented observations better inform the threat from various poorer quality water sources to the fossil site in particular, and to the broader karst water resource in general. This contextualises concern for the hydro-environmental future of Sterkfontein Cave and other nearby fossil sites such as Swartkrans, Rising Star and Bolt's Farm. The dynamic response of the water resources environment to a variety of hydrological and hydrogeological drivers reinforces the need for monitoring vigilance across a range of disciplines.

Introduction

Sterkfontein Cave is arguably the most well-known fossil site on the Cradle of Humankind World Heritage Site (COH WHS) property. It is located in the Isaac Edwin Stegmann Nature Reserve on the farm Zwartkrans 172IQ, Gauteng Province, South Africa (Figure 1). The site shares the hydrosystem of the Zwartkrans Compartment karst basin with the Bolt's Farm, Rising Star, Swartkrans, Cooper's and probably as yet undiscovered fossil sites. The water level in the cave system has been the

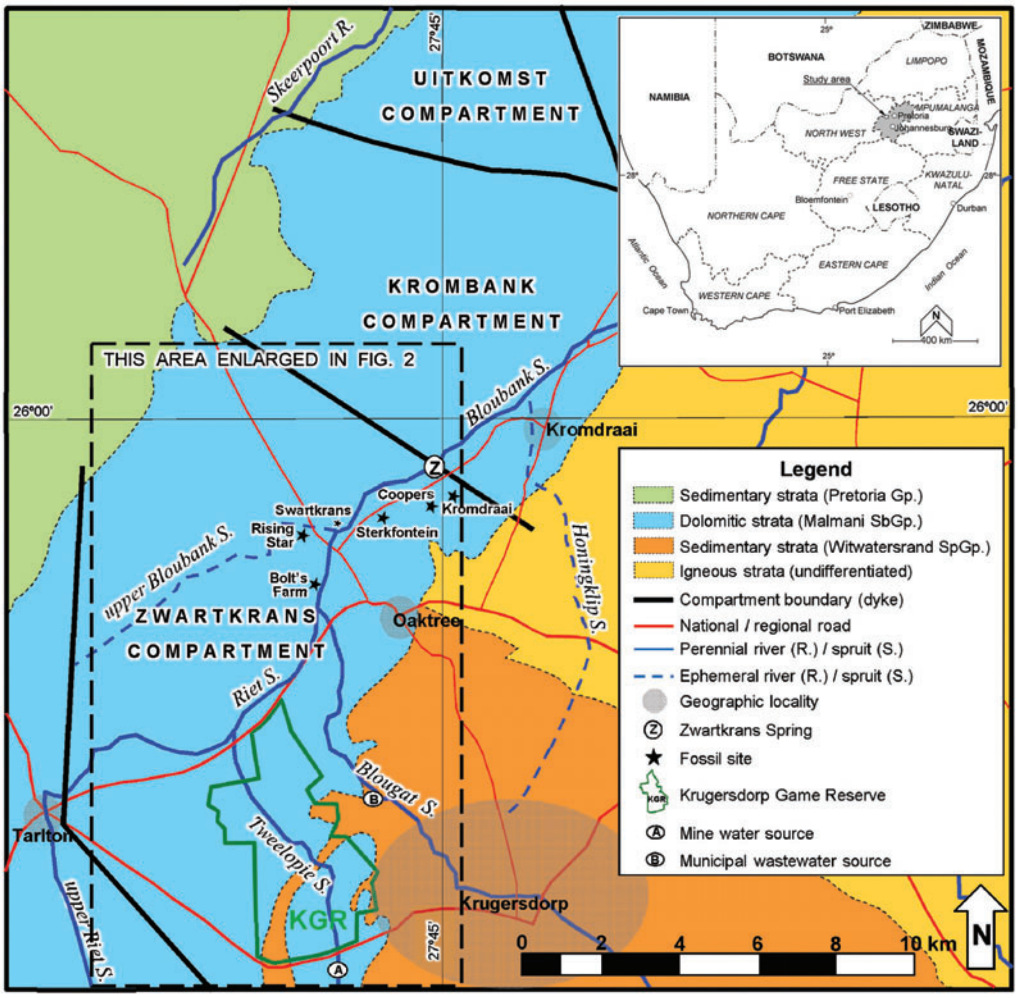

Figure 1. *Map of the study area showing features of relevance, e.g. geography (water sources, localities), geology (dolomite footprint), hydrology (surface water drainages) and palaeontology (fossil sites); inset map shows position on the northwest margin of Gauteng Province (shaded area), South Africa.*

subject of debate and at least some confusion in the comparatively recent past. The water level in question is that associated with what Martini et al. (2003) refer to as the 'Main Lake'. This is the most readily accessible water body in the cave system, and marks the furthest and deepest point of the tourist route through the cave. Wilkinson (1973) refers to this water body as the 'Lake', a convention followed in this paper.

A rise in Lake water level of almost 3 m in the space of some two years, commencing in late-2009, caused Maropeng āAfrika (the authority responsible for managing the tourist component of the site) to re-direct the tourist route through the cave to a successively higher elevation on three occasions. This has had little impact on the tourist value of the site, as visitors still descend to and view the Lake after passing the location of the 'Little Foot' (StW 573) find. The circumstances have, however, raised concern for their possible association with copious acidic and sulphate-rich (amongst other quality concerns such as metals) mine water drainage from the West Rand Goldfield (a.k.a. the Western Basin), and in particular for the impact on cave water chemistry in the context of karst dissolution, speleogenesis, cave ecosystems and the preservation of fossil deposits. The concern has been compounded by a poor understanding of the ambient hydrogeology which, until recently, relied on sparse and weakly constrained historical data. This paper presents a revised and more robust understanding of the hydrogeology and hydrochemistry of the cave system and host water resources environment. Based on both historical and 'new' hydrogeological and hydrochemical data, prevailing concerns for the fossil site are contextualised within the framework of observed impacts on the broader water resources environment.

Setting and background

The Zwartkrans Compartment karst basin lies to the north of the actively draining Western Basin that straddles the sub-continental divide formed by the Witwatersrand at an elevation of ~1720 metres above mean sea level (mamsl). It receives allogenic runoff from the non-karst watershed area via various headwater tributaries of the Bloubank Spruit, most notably the Tweelopie, Riet and Blougat Spruit drainages (Figure 2). Itself an upper tributary of the Crocodile River, the Bloubank Spruit drains in a northeast direction past the cave site. The Zwartkrans Spring located ~1400 m northeast of the cave (Figure 2), serves as natural outlet for the ~9800 ha karst basin. The basin comprises Neoarchaean (~2.65 to 2.50 Ga) carbonate strata (mainly dolomite) of the Malmani Subgroup in the Chuniespoort Group at the base of the Transvaal Supergroup succession of Vaalian (~2.65 to 2.05 Ga) clastic and chemical sedimentary strata and volcanic rocks (Visser, 1989; Eriksson et al., 2006). The carbonate strata cover 26% of the hydrological catchment (identified as Quaternary basin A21D by RSA, 2000), and 35% of the total karst footprint (~27 850 ha) of the COH property, which extends into Quaternary basin A21G to the north.

A Western Basin Technical Working Group (WBTWG) meeting held on 28 February 2007 at Rand Uranium in Randfontein was informed that the Zwartkrans Spring had stopped flowing, and that any streamflow in this vicinity represented surface runoff. This was in response to a query regarding the water level in Sterkfontein Cave (reportedly 1436 mamsl) compared to the 1439 mamsl reported elevation of the spring (JFA, 2006; Krige and Van Biljon, 2006; Van Biljon, 2006). These circumstances were put forward as evidence that the water level in the cave system could not rise more than 3 m,

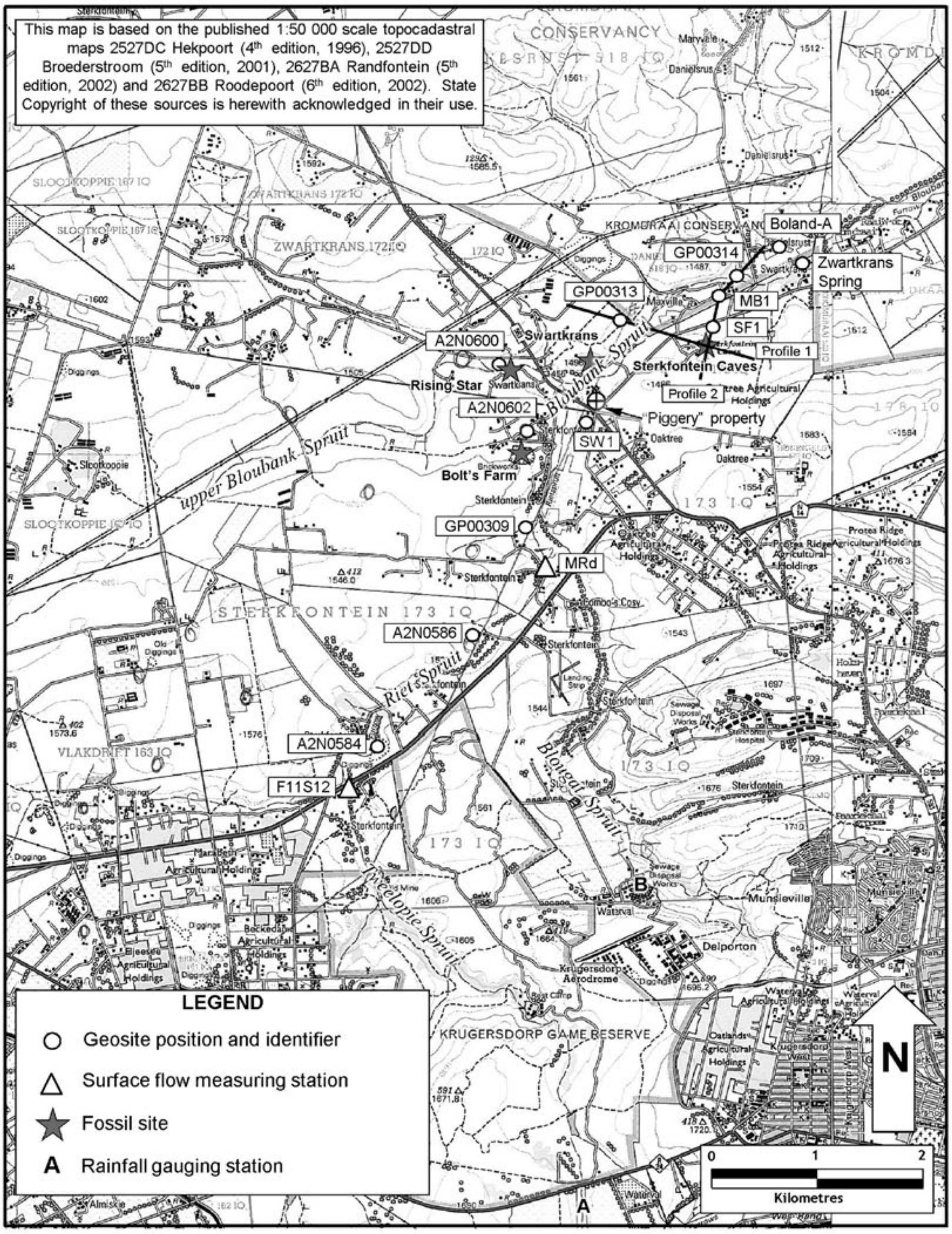

Figure 2. *Map of geosites and other features referenced in this paper, of particular relevance being the rainfall gauging station localities **(A)** and **(B)** also associated with the sources of poor mine and municipal wastewater respectively, the surface water drainages and the profile positions that intersect the Sterkfontein Cave fossil site.*

i.e. to the elevation at which groundwater would drain from the karst basin via the spring. The owner of the property on which the spring is located, however, subsequently confirmed that the spring had never stopped flowing since taking occupation in 1980 (Roos, pers. comm. 2011). This has been the case to the present. The indication that the spring must occupy a lower elevation than 1436 mamsl initiated a re-assessment of the circumstances that inform the Lake water level and associated hydrogeology (Hobbs, 2011).

The re-assessment followed a holistic approach in recognition of the typically close interaction between surface water and groundwater in a karst environment. It has contributed to resolving earlier disparities, and provides a better understanding of the recent Lake water level and quality response, and their prognosis. More importantly, it contextualises concern for the future of this and neighbouring fossil sites in the broader water resources environment.

Lake water level

As the most obvious groundwater element of the cave system, the re-assessment commenced with an evaluation of available Lake water level data. Time constrained (once-off) data and information from the late-1960s and early-1970s provided a historical 'snapshot' perspective. Groundwater level monitoring carried out in the last decade better defined the temporal response in Lake water level.

Historical context

An early reference to the Lake water level is found in Cooke (1969), who observed that “The water table is regional and the underground lake is 186 feet below the crest of the hill with the ape-man quarry.” Wilkinson (1973) reported an elevation of 1485 mamsl for the top of the 'hillock'. Martini et al. (2003) placed the top of the 'small hill' at 1491 mamsl. The elevation of the hillcrest above the cave system has a spot height of 1487 mamsl on the 1:10 000 scale orthophoto map 2627BA5 Sterkfontein (RSA, 2003). Google Earth® shows a highest elevation of 1489 mamsl in this area.

It is apparent that using the top of the hill as datum for deriving a Lake water elevation, is unsatisfactory under circumstances where this datum differs by 6 m between the referenced sources. Nevertheless, and in order to derive comparative Lake water elevations from the historical data, the authors accept the orthophoto-based elevation of ~1487 mamsl for the hilltop datum. Other elevational data used in this study (e.g. the spring and borehole collar elevations) also stem from this source. Using this datum, the depth of 186 feet (56.7 m) below the hillcrest reported for the Lake by Cooke (1969) places the contemporary Lake water elevation at ~1430 mamsl. Similarly, the depth of 48 m reported by Wilkinson (1973) gives a coeval Lake water elevation of ~1439 mamsl. Martini et al. (2003) reported the depth of the water table in the cave as ~50 m below the surface, placing the contemporary Lake water elevation at ~1436 mamsl. It will be shown that a Lake water elevation of ~1435 mamsl can be postulated for ca. 2003, indicating that the Martini et al. (2003) observation is consistent with more recent values.

The common hilltop datum of ~1487 mamsl constrains the historical Lake water elevation to a range of 9 m. Although not impossible, it is highly unlikely that the Lake water level rose by some 9 m between the Cooke (1969) and Wilkinson (1973) observations. Potentiometric changes of this and greater magnitude and rapidity in highly transmissive karst aquifers (Hobbs, 1988) are usually observed only in the presence of anthropogenic impacts such as large-scale and intense groundwater use (abstraction) or engineered dewatering. The historical water level data are therefore best seen in relative and qualitative rather than absolute and quantitative terms. As will be shown, however, they are in the appropriate range.

Recent context

In a significant contribution on the genesis of Sterkfontein Cave, Martini et al. (2003) observed that the Lake water level fluctuation was within a range of ~2 m, and varied gradually in response to an antecedent rainfall pattern. Further, that speleothems were corroded in the interval up to 6 m above the Lake water level, and that no speleothems occurred in the 2 to 3 m interval above the Lake surface. The latter observation indicates that this interval represents the most aggressive in terms of carbonate dissolution, at least in the more recent speleogenesis of the cave system.

A December 2007 survey of the datum inside Sterkfontein Cave from which the Lake water level is determined, placed it at an elevation of 1437.94 mamsl (Krige, 2009). This is 14.43 m below a previously reported elevation of 1452.37 mamsl (JFA, 2006). Applying the difference to the Lake water elevation of 1450.88 mamsl reported by JFA (2006), yields an elevation of 1436.45 mamsl. This is similar to the 1436 mamsl reported at the WBTWG meeting held some ten months earlier. As a consequence, an earlier unchallenged hypothesis that the Lake water level was associated with a perched aquifer (Krige and Van Biljon, 2006) to explain the Lake water elevation of 1450.88 mamsl, was also revoked (Krige, 2009; Krige and Van Biljon, 2010). The revised Lake water elevation accords with the conclusion by Martini et al. (2003) that water level differences between various water bodies (up to 30 reported 'static' pools) in the cave system amount to decimeters rather than the up to 10 m reported by Wilkinson (1973).

A groundwater level measurement in borehole SF1 (Figure 2) on the cave property in July 2006 (Holland, 2007) returned a value of 17.00 m below surface (mbs). For an interpolated surface elevation of ~1454 mamsl obtained from orthophoto map 2627BA5 (RSA, 2003), this depth gives a water table elevation of ~1437 mamsl. The <1 m difference with the revised Lake water elevation suggests that the groundwater rest level in borehole SF1 and the cave system represents a low-gradient contiguous potentiometric surface. A water level depth of 17.4 mbs in SF1 in October 2007 (Hobbs and Cobbing, 2007) gives a water table elevation of 1436.6 mamsl, i.e. within 0.2 m of the revised earlier Lake water elevation of 1436.45 mamsl. These results establish the utility of a groundwater level measurement in borehole SF1 as a suitable proxy for the contemporary Lake water level.

There is little doubt that borehole SF1 is the 'pumphouse borehole' referred to by Wilkinson (1973; Figures 7.2 and 7.4), and for which a water level depth of 60 m below the 'hillock' datum is reported. This is 12 m deeper than the contemporary Lake water depth (48 m) below the 'hillock' datum. The difference represents an exceedingly steep hydraulic gradient of 0.04 (4:100 or 2.3°) over the ~300 m distance between the Lake and the borehole. Hydraulic gradients of this magnitude in highly transmissive karst aquifers of the South African interior are uncommon. The more recently recognised low-gradient contiguous potentiometric surface between the cave system and borehole SF1 is a truer reflection of this hydraulic parameter. These circumstances again illustrate the relative rather than absolute value of the Lake water elevations derived from the historical (Cooke, 1969; Wilkinson, 1973) data.

Groundwater drainage

The potentiometric response pattern of the Lake water level is a function of the groundwater drainage pattern in the vicinity of the cave system. Groundwater drains naturally from the Zwartkrans karst basin via two routes. The most conspicuous of these is the ~136 litres per second (L/s) Zwartkrans Spring (Figures 1 and 2) located at an elevation of ~1433 mamsl (Hobbs, 2011) on the right bank of the Bloubank Spruit. The reported yield is determined from the difference in flow measurements made in the stream reach within ~10 m upstream and downstream of the outlet of the spring into the stream, as the outlet itself does not facilitate reliable flow measurements. The discharge of the spring is considered to be comparatively steady, reflecting minor to moderate variability (~80 to 286 L/s) in the medium- and long-term. This accords with the 'regular discharges at resurgences' reported by Martini (2006) as a characteristic of the Transvaal Supergroup karst formations.

The second drainage route is inconspicuous groundwater resurgence in the channel of the Bloubank Spruit upstream of the spring. The magnitude of this contribution varies with the distance upstream of the spring at which the water table intersects the stream channel. In mid-2007, this position was located northeast of the cave a distance of ~800 m upstream of the spring. By mid-2011 it extended to a position north of the cave ~1500 m above the spring. The upstream extension of the gaining reach accompanied the ~3 m rise in water table observed in the Lake water level. It has since maintained this position with comparatively minor up- and downstream migrations as defined by a <1 m fluctuation in water table. These circumstances are illustrated in Figure 3, which also indicates the existence of a nick-point in the stream profile opposite the cave. It is hypothesised that this feature marks the upstream limit of the gaining stream reach.

Groundwater resurgence in the stream channel upstream of the spring contributes >200 L/s to the discharge of the Bloubank Spruit (Hobbs, 2013). In January 2013 the contribution from a reach of only ~400 m upstream of the spring was measured at ~124 L/s. A contribution of this magnitude to the ~136 L/s of the spring, and assuming a similar spring discharge in the mid-1980s, approximates the discharge of ~258 L/s reported by Bredenkamp et al. (1986) for the spring. Bredenkamp et al. (1986) do not define the location of their springflow measurement(s) and make no mention of an instream contribution to streamflow at this position. It is presumed, therefore, that the value of ~258 L/s was obtained in the stream channel downstream of the spring. The recent groundwater discharge from the Zwartkrans basin of ~336 L/s equates to ~29 million litres per day (ML/d) or 10.6 million cubic metres per annum (Mm^3/a).

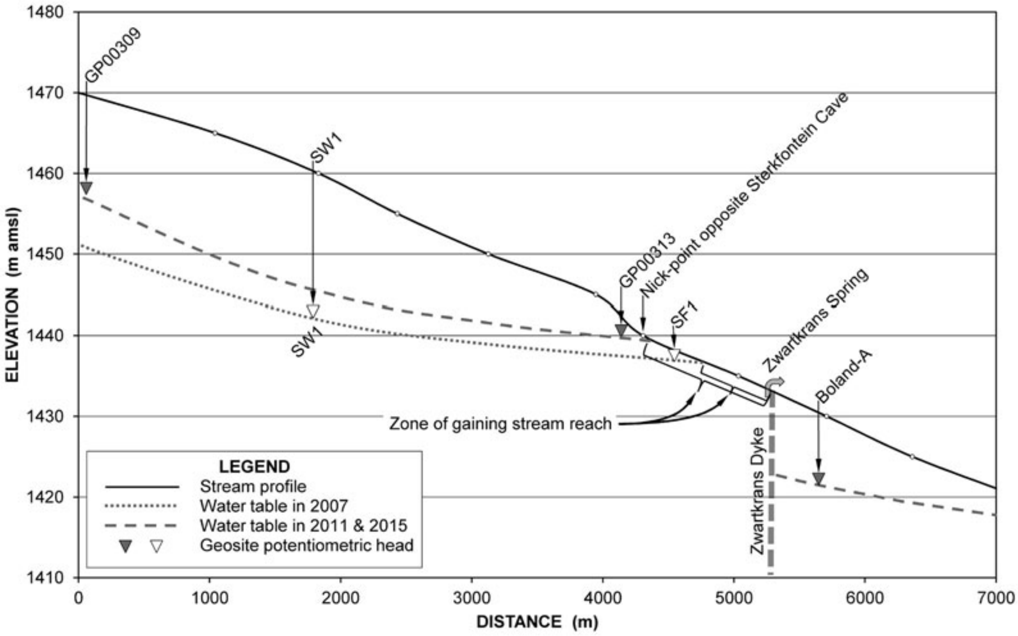

Figure 3. *Bloubank Spruit stream profile bracketing the Sterkfontein Cave system, showing the water table position on two occasions in the recent past, and the zone of gaining stream reach in relation to the cave site and the Zwartkrans Spring (see Figure 2 for geosite localities).*

Potentiometric response

Groundwater levels typically reflect the response of an aquifer to rainfall recharge as a gain evident in rising water levels, whereas losses associated with natural drainage and abstraction (amongst others) manifest as declining water levels. In the severely disrupted surface and subsurface mining environment of the Western Basin, rainfall recharge produces an unequivocal increase in mine water discharge (Hobbs, 2013; 2014), sometimes exceeding the capacity of the treatment plant constructed to manage this outflow. [Note: Commissioned in mid-2006 with a capacity of ~12 ML/d, this increased to ~24 ML/d in mid-2012, and to its current ~34 ML/d in mid-2013]. At such times, excess raw (untreated) mine water enters the Tweelopie drainage where it joins the treated/neutralised mine water discharged to the downstream environment. The precipitation pattern therefore influences the potentiometric response (and indirectly also the hydrochemical response) in the receiving Zwartkrans Compartment karst aquifer. The rainfall record as gauged on the SW margin of the study area is shown in Figure 4. The combination of data for two stations (A and B in Figures 1 and 2) allows the construction of a composite record dating back to the 2001 hydrological year (see https://water.usgs.gov/nwc/explain_data.html for definition), pre-dating by some two years the start of mine water discharge in late-August 2002. The daily precipitation record has been aggregated to both an annual and seasonal value, as it is not intended here to analyse the minutiae of the karst potentiometric response to rainfall-driven recharge.

A water level (data) logger installed in the Lake in May 2005 by the then Department of Water Affairs and Forestry (DWAF), yielded information on the Lake water level response pattern to July 2007 (Figure 5). The hydrograph reveals three periods of steady decline at a rate of between 0.08 and 0.06 m per month, the last following a rise of 0.5 m in the 2007 summer. The latter is considered to reflect recharge generated in the wetter than average 2006 hydrological year (Figure 4). This also suggests a delayed potentiometric response, by some six to nine months, of the karst aquifer to natural (primarily rainfall-driven) autogenic recharge, as the mine water discharge at the time amounted to only ~5 ML/d (Hobbs et al., 2011).

The record of water level measurements obtained with a standard handheld 'dipmeter' in borehole SF1 is shown in Figure 6. The early trend continues the decline previously recorded in the cave water level (Figure 5) with a linear recession rate of 0.03 m per month. The slower rate of decline reflects the exponential 'decay' in potentiometric head. The recession terminates somewhere in the period between the October 2007 and February 2010 manual measurements. Based on the observations of cave guide K Mangole (pers. comm., 2010), this most probably occurred in late-2009. When asked in mid-May 2010 about the rise in Lake water level, Ms Mangole estimated this to be about 1 to 2 feet (0.3 to 0.6 m) since late-2009. This is in fair agreement with the 0.6 m rise observed in borehole SF1 between February and June 2010, and the 0.4 m rise in borehole MB1 (Figure 2) between February and May 2010 (Hobbs et al., 2011). The overall rise in Lake water level amounted to 2.8 m by June 2011, stabilising at an elevation of some 1439.3 mamsl before starting to fall in mid-2012. The second decline again reflects an average linear recession rate of 0.03 m per month (Figure 6).

Compared to the modest 0.5 m rise in the 2007 summer (Figure 5), the much greater rise in water level in the 2010 and 2011 hydrological years was driven mainly by the uncontrolled discharge of mine water to the environment in the wetter than average 2010 summer (Figure 4). Starting on 30 January 2010 (Hobbs, 2013), this discharge far exceeded the ~12 ML/d capacity of the mine water treatment plant. Mine records show

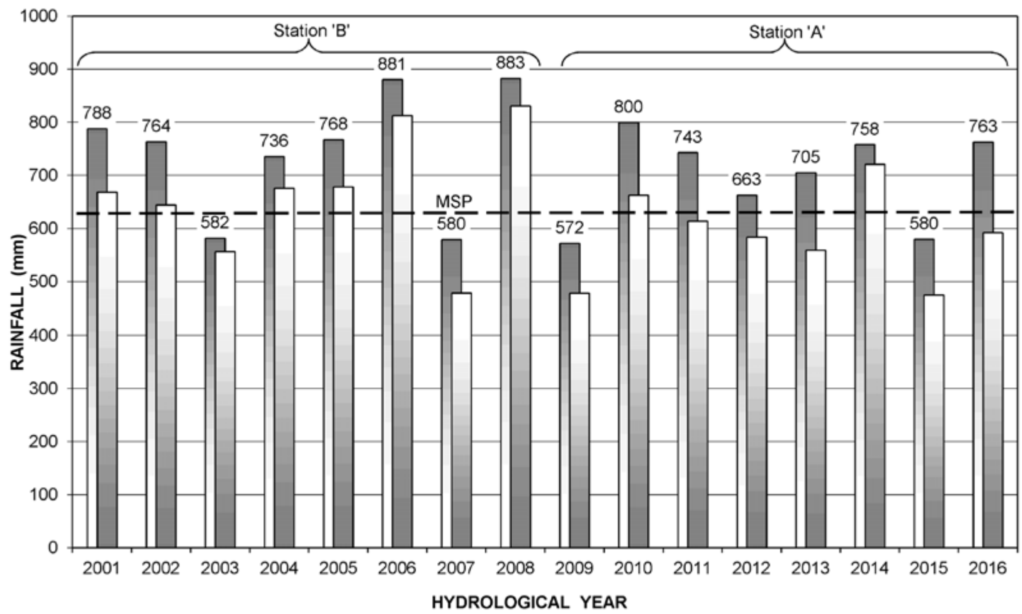

Figure 4. *Composite recent annual and overlapping summer (wet season) rainfall record for stations **(A)** and **(B)** (Figures 1 and 2) along the southwest margin of the study area; MSP denotes mean summer precipitation (627 mm).*

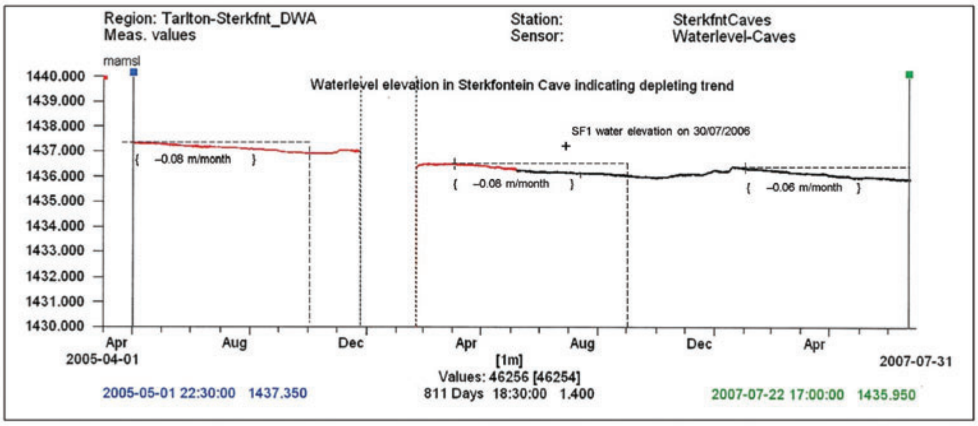

Figure 5. *Semi-continuous hydrograph of the Lake water level captured over a period of 29 months by a data logger installed in Sterkfontein Cave in May 2005 (use of image courtesy of DWS).*

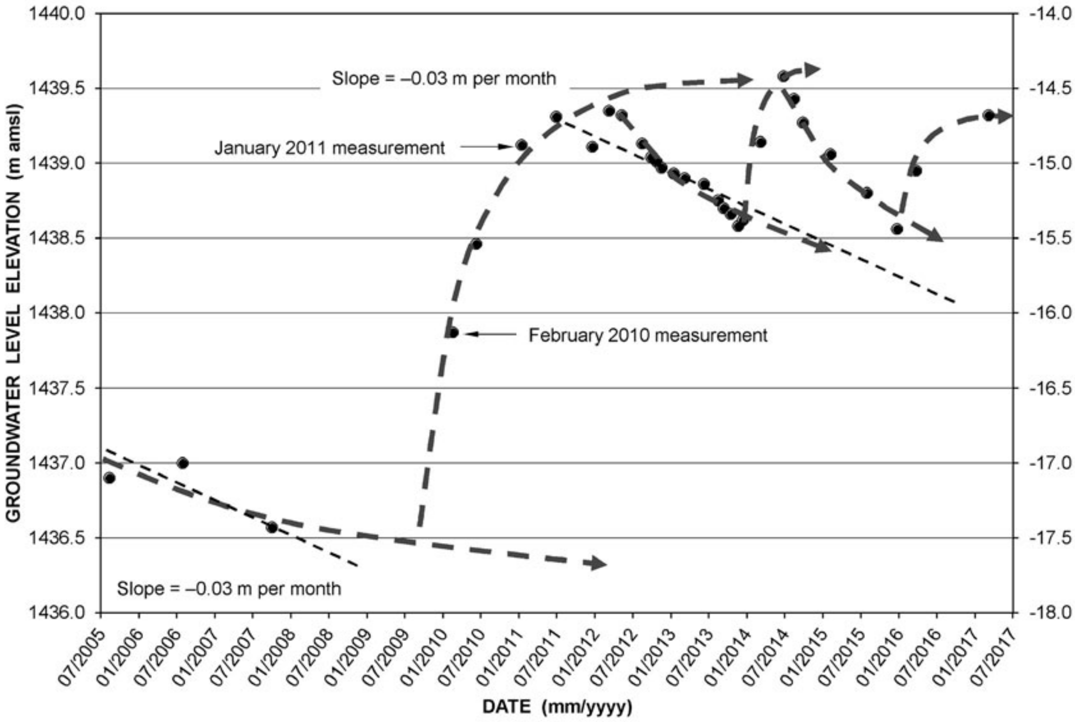

Figure 6. *Hydrograph of manual water level measurements in borehole SF1 (Figure 2) serving as a proxy for the Lake water level.*

that discharges in excess of 60 ML/d were recorded through to mid-2010.

The more recent water level rise reflects the influence of the wet 2014 summer rainfall season (Figure 4), and indicates stabilisation of the Lake water level at an elevation of around 1439.5 mamsl similar to the previous maximum. Representing a quite rapid rise of ~1 m in less than six months, the mean rate of 0.16 m per month is similar to the 0.15 m per month reflected in the earlier rise between the February 2010 and January 2011 measurements (Figure 6). The shorter delay of some three to six months indicates a quicker positive potentiometric response to allogenic recharge in the karst environment. It will be shown that allogenic recharge is focussed discretely but intensely on losing stream reaches, and therefore has a greater pro rata impact on proximate groundwater levels than that associated with spatially distributed (wide-area) autogenic recharge.

The long-term hydrographs of monitoring boreholes A2N0600 and A2N0602 (see Figure 2 for positions) and of borehole SF1 presented in Figure 7, reveal the following salient circumstances:

- the most recent groundwater elevation in boreholes A2N0600 and A2N0602 are the highest in the 27-year period of record (Figure 7) and, in the case of A2N0602, exceeds by ~3 m the highest pre-decant elevation recorded ca. 2000;
- the ca. 2003 groundwater elevations (~1439 mamsl) in borehole A2N0602 are the lowest in the 27-year record going back to 1989; and
- a similar pattern and trend is replicated in each hydrograph,

although the SF1 hydrograph values lie on average 4.3 m below those of A2N0602.

The veracity of the regular separation between the A2N0602 and SF1 potentiometric values finds support in two photographs presented by Van Biljon (2006). Taken of the Lake water level in September 1984 and February 2004 respectively, a comparison shows that the February 2004 level is some 1.8 m lower than in the earlier photograph. If the 2004 level is placed at ~1435 mamsl (Figure 7), then the 1984 elevation must be ~1437 mamsl. Applying the 4.3 m separation with the A2N0602 elevation gives a value of ~1441 mamsl for the latter. The archived data record for A2N0602 shows a similar elevation in mid-1987, some three years later than the 1984 photograph. The agreement supports the utility of the historical A2N0602 data as a proxy for the SF1 groundwater elevation and, by association, the historical Lake water elevation. This is illustrated in Figure 7, which indicates a lowest interpolated historical Lake water elevation of some 1434.5 mamsl reached in late-1995 and again in late-2003. The ca. 2003 Lake water elevation observed by Martini et al. (2003) is placed at ~1435 mamsl. This implies that the 6 m interval of corroded speleothems above the Lake water level reported by Martini et al. (2003) extends to an elevation of ~1441 mamsl. Further, that the 2 to 3 m interval devoid of speleothems extends to an upper elevation of ~1438 mamsl.

The interpolated long-term potentiometric data also indicate that the Lake water level fluctuated by ~1 m around a median elevation of 1435.7 mamsl over a period of at least 20 years since the mid-1980s. As the greater positive response of the more recent past (Figure 7) is attributed mainly to the allogenic recharge of mine water, it is postulated that a Lake water elevation of ~1436 mamsl can be taken as representative also of the more recent prehistoric cave environment. The observation by Martini et al. (2003) that the Sterkfontein Cave system was already dewatered at 20 to 25 m above the present water table 3.3 million years (Ma) ago, indicates that the present water table elevation marks a transient position on a downward trajectory spanning millions of years. The Martini et al. (2003) observation suggests a lowering in the range 6.1 to 7.6 m per million years (m/Ma). This is slightly more than the ablation rate of ~5 m/Ma (from Gams, 1989) used by Martini et al. (2003) in their model of cave development. At these rates, the interval reported to be devoid of speleothems might conceivably extend back to the Middle Pleistocene.

If the contemporary Lake depth and that of water bodies in the adjacent P Verhulsel Section of the cave system is restricted to around 4 m (Martini et al., 2003), then these contiguous features would be virtually dry at an elevation of ~1431 mamsl. In this case, the Zwartkrans Spring would also have dried up. Less accessible (or even inaccessible) but deeper water bodies in the cave system such as in the Lincoln-Fault Section to the north (Martini et al., 2003) and, it would seem from later discussion, to the northwest in the vicinity of borehole GP00313 (Figure 2), would continue to represent underground lakes.

The ~1440 mamsl elevation of the Bloubank Spruit channel north of the cave suggests that the Lake water level reaches equilibrium at a similar elevation. This is when the karst water table intersects the streambed ~1500 m upstream of the Zwartkrans Spring as discussed earlier (see also Figure 3). The range of ~7 m between this elevation and that of the spring accommodates the interval identified by Martini et al. (2003) as bracketing the zone of carbonate dissolution that defines the more recent speleogenetic evolution of the cave system. This zone is defined by present-day Lake water elevations between ~1436 and 1439.5 mamsl, which happen to bracket the 2 to 3 m interval of most aggressive recent carbonate dissolution.

The pre-2010 difference of up to 4 m between the streambed and Lake water elevations is a common occurrence observed further upstream. Hobbs and Cobbing (2007) report vertical separations between the Riet Spruit streambed and the water table in the range 12 to 30 m. A rise in groundwater rest levels of 4.9 m on average recorded in 11 monitoring boreholes in the Zwartkrans Compartment (Hobbs, 2014) has reduced this separation since. The separation also reduces toward the downstream end of the karst basin, as is evident in the vicinity of Sterkfontein Cave and at the Zwartkrans Spring. These circumstances partly describe the hydraulic conditions in the vadose zone (including the epikarst) that serves as pathway for surface water ingress to the karst aquifer.

Ancillary groundwater information

Physical hydrogeology

Information obtained during the drilling of monitoring borehole GP00313 (Figure 2) located some 800 m to the northwest across the valley of the Bloubank Spruit from Sterkfontein Cave, provides additional insight into the nature of the ambient subsurface karst environment. Constructed in December 2010 under the supervision of the corresponding author, the borehole intersected highly leached to decomposed dolomite down to its completion depth of 37 mbs, encountering a 15 m cavity spanning the interval 17 to 32 mbs (~1443 to 1428 mamsl). This places the base of the 'cavern' at least 5 m below the elevation of Zwartkrans Spring, and ~11 m below the contemporary Lake water elevation of ~1439 mamsl (Figure 6). The water table depth of 20.6 mbs following completion of the borehole similarly approximates, in absolute terms, the ~1439 mamsl elevation of the contemporary Lake water elevation.

Hobbs et al. (2011) report a groundwater gradient of 0.003 (0.17°) in a northeast direction in the Bloubank Spruit valley passing Sterkfontein Cave. This is in fair agreement with the gradient of 0.005 (0.29°) tangential to the true gradient between the contemporary/current Lake water elevation and the spring. The poorly discernible fall of 0.3 m per 100 m contextualises the 'apparently static' nature of pools in the cave system and the decimetre differences in pool water elevations reported by Martini et al. (2003).

The wetter than average 2008, 2010 and 2014 summers (Figure 4) resulted in exceptional recharge of the karst aquifer in the study area. Compared to the long-term mean annual precipitation (MAP) of 710 mm for the central and southern portion of the COH (Hobbs et al., 2011), these seasons experienced 831, 663 and 721 mm respectively (Figure 4). The observed rise of almost 5 m on average in potentiometric levels since 2007 (also evident in Figure 7) therefore represents a combination of autogenic and allogenic recharge.

Autogenic recharge from rainfall across the COH property is estimated at 17 ± 5% of 710 mm MAP (Hobbs et al., 2011). The error margin of 5% accounts for variation in both the amount and distribution of rainfall. Holland (2007) reports a value of 16% for a similar MAP. Distributed over the ~9800 ha extent of the Zwartkrans Compartment, recharge amounts to 11.9 ± 3.5 Mm^3/a (32.6 ± 9.6 ML/d). Allogenic recharge contributed from the upstream non-karst areas located to the south is quantified on the basis of instream flow measurements carried out at suitable sites in the Tweelopie, Riet and Blougat drainages (Figure 2). Hobbs (2016) reports flow measurements made on 31 occasions between September 2009 and July 2015 at stations F11S12 at the lower end of the Tweelopie Spruit and MRd located ~3.9 km further downstream on the Riet Spruit (Figure 2). The results shown in Figure 8 quantify and elucidate mine water losses to the karst aquifer, indicating a minimum ingress of ~14 ML/d equivalent to ~42 L/s per kilometre (L/s/km). The different slopes of the Period 2 and Period 3 regression lines (Figure 8) indicate a substantial 94% reduction (from ~70 to ~36 L/s/km) in allogenic recharge between these two time periods. Whilst the reduction favours the receiving karst aquifer as the chemistry of this water has a strong mine water character, an explanation for this phenomenon remains elusive. Nevertheless, the spatially concentrated allogenic recharge of mine water in the amount of at least ~14 ML/d (5.1 Mm^3/a), and on occasion >32 ML/d (>11.7 Mm^3/a), is unequivocal. The latter matches that of distributed autogenic recharge. Similar assessments for the Blougat Spruit (Hobbs et al., 2011; Hobbs, 2013; 2014) indicate the ingress of poor quality municipal wastewater effluent in the range 3 to 7 ML/d (1.1 to 2.6 Mm^3/a) where this drainage in its lower reaches traverses carbonate strata.

The further significance of the reduction in ingress between Periods 2 and 3 (Figure 8) is the associated increase in the volume of water passing station MRd as surface flow. This averaged 13 ML/d (150 L/s) in Period 3, with a measured minimum and maximum of 4.7 ML/d (54 L/s) and 30.3 ML/d (351 L/s) respectively. The progression in surface flow at station MRd from period to period is shown in the inset diagram (Figure 8). This indicates the increasing volume of mine water contributing directly to surface flow downstream past the Sterkfontein Cave and, by extension, its increasing importance as an additional allogenic surface water component in the catchment. This component is not subject to the attenuation (neutralisation and dilution) of the acidic mine water that is observed in the karst aquifer as described hereunder.

As discussed previously, a substantial portion of the recharge returns to the surface as groundwater resurgence in the channel of the Bloubank drainage. It is therefore considered more appropriate to recognise the allogenic recharge component as a transmission flux rather than a transmission loss. Importantly, the effluent groundwater is of better quality than that of the surface water, as shown by respective pH values of >6.5 vs. ~4.1, and sulphate values of <1800 vs. ~2200 milligrams per litre (mg/L) (Hobbs, 2017). This is attributed mainly to two activities in the karst aquifer, namely the neutralising effect of carbonate strata and mixing with much better quality autogenic recharge. The mixing is greatest at the discharge end of the karst basin.

Chemical hydrogeology

If the water level rise in the cave system is due (at least in part) to allogenic recharge driven mainly by mine water drainage, then this would likely also be reflected in the cave water chemistry. A review of available groundwater chemistry data for the karst

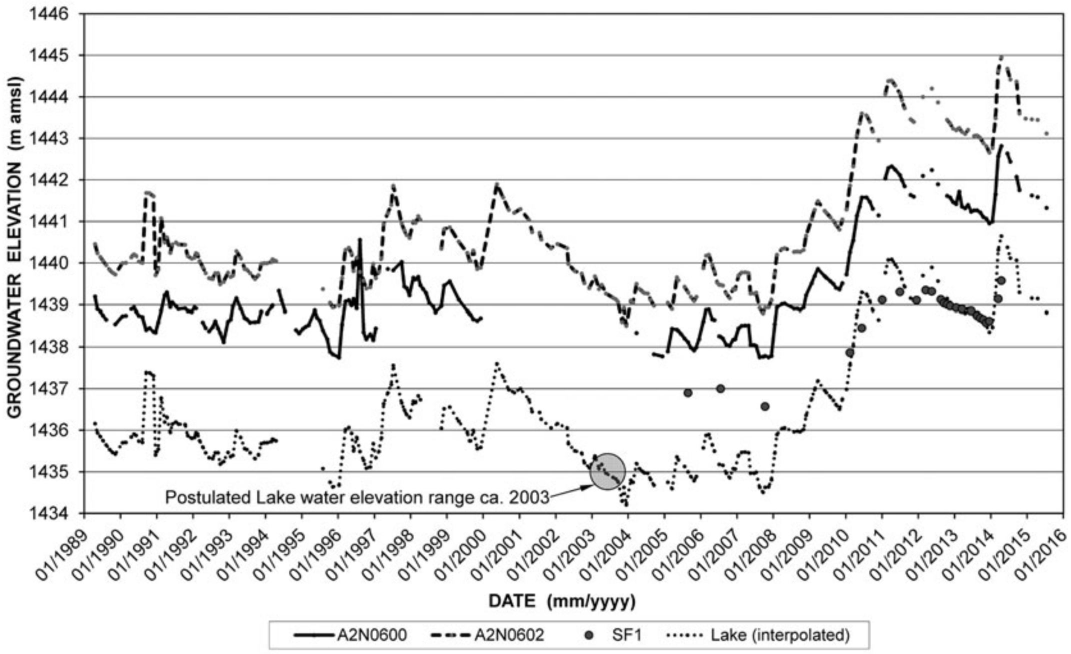

Figure 7. *Long-term potentiometric response in monitoring boreholes A2N0600 and A2N0602 (Figure 2) compared to the measured and interpolated response in borehole SF1 serving as a proxy for the Lake water level.*

hydrosystem shared by the cave system provides the following insight.

An electrical conductivity (EC) data logger installed in the Lake in May 2005 provides historical information on the cave water salinity response before the elevation reached by the Lake water level ca. mid-2010 rendered the data logger inaccessible. The record (Figure 9) reveals an initial EC value of ~47 milliSiemens per metre (mS/m), followed after a break of some 32 weeks by a higher value of ~60 mS/m. A laboratory analysis of contemporary pre-break Lake water gave an EC value of 59 mS/m (Figure 9). This suggests that the initial salinity values in the data logger record are in error by some 12 mS/m. Assuming the post-break record as representative of the entire record, then the salinity of the cave water changed little in the record period, and certainly in the post-break period. The much more recent value of 78 mS/m in December 2015 (Table 1) represents a 32% increase over the April 2005 value of 59 mS/m.

The Lake water chemistry over time is compared to recent Zwartkrans Spring water chemistry in Figure 10. Further comparison is made in Figure 11, showing that the differences are primarily associated with the proportional sulphate (SO_4^{2-}) concentrations. The SO_4^{2-} concentration of 46 mg/L in cave water ca. 2001 (Table 1) already indicates a measure of impact, as that of pristine karst springs in the COH is typically <5 mg/L (Hobbs, 2015). The presence of a mine water impact on karst groundwater in the Zwartkrans karst basin is evident as early as

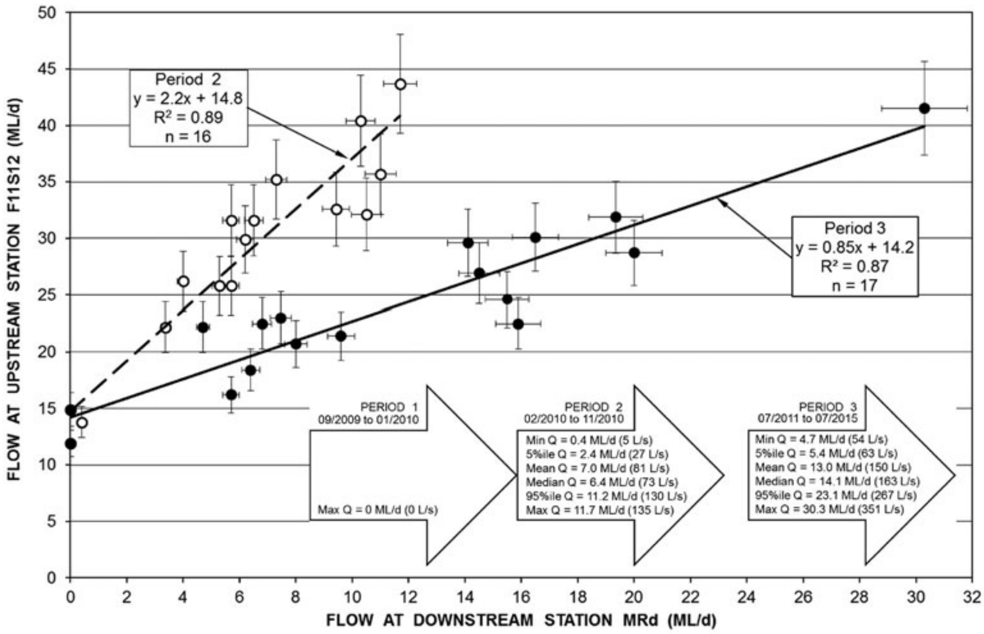

Figure 8. *Correlation of streamflow at stations F11S12 and MRd in the Riet Spruit valley, with vertical error bars denoting ±10% at F11S12 and horizontal bars ±5% at MRd; inset diagram shows periodic progression of surface flow measured at station MRd (from Hobbs, 2016).*

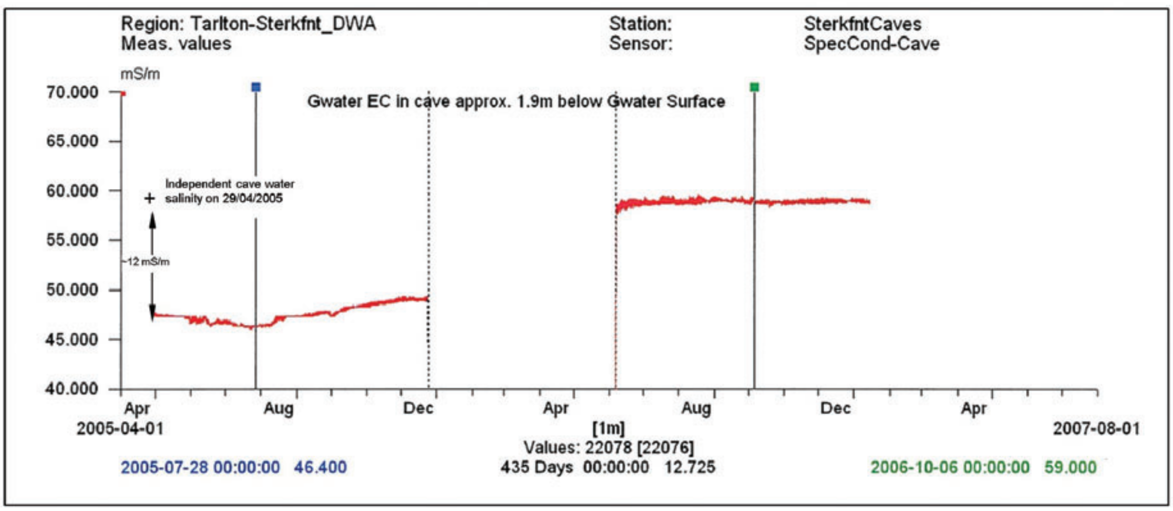

Figure 9. *Semi-continuous electrical conductivity pattern in Lake water captured over a period of 22 months by a data logger installed in Sterkfontein Cave in May 2005 (use of image courtesy of DWS).*

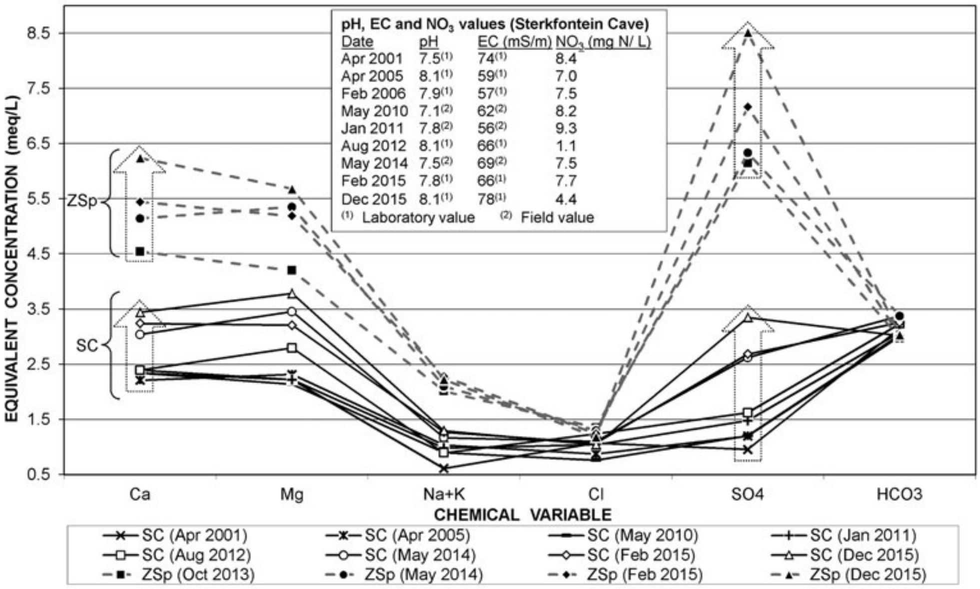

Figure 10. *Schoeller graphical comparison of historical and recent Lake water (SC) chemistry compared to recent Zwartkrans Spring water (ZSp).*

1985, when Bredenkamp et al. (1986) reported a SO_4^{2-} concentration of ~200 mg/L for borehole A2N0586 (Figure 2), and the earliest record for the nearby borehole A2N0584 indicates a value of ~370 mg/L in mid-1985 (Hobbs and Cobbing, 2007). These boreholes are located immediately downstream of surface monitoring station F11S12 and proximate to where allogenic recharge starts in the losing reach of the Riet Spruit. It is common knowledge that mine water pumped from active mines in the Western Basin was discharged to the Tweelopie Spruit, and it is this water that is considered responsible for the observed historical impact. These observations suggest therefore that the Sterkfontein Cave system experienced a historical mine water impact, albeit less apparent, well before the start of mine water decant in 2002.

Also evident in Figures 10 and 11 is the shift of the Lake water composition toward that of the Zwartkrans Spring water, with bicarbonate (HCO^{3-}) gradually losing its anionic dominance to sulphate. The springwater chemistry itself shows an unambiguous shift toward a more dominant Ca-SO_4 composition reflecting the mine water impact on groundwater draining the Zwartkrans karst basin.

Closer inspection of the data indicate that the SO_4^{2-} concentration in the Lake water increased quasi-linearly by 70% in the period April 2001 to April 2012, followed by a quasi-exponential increase by 106% from 78 mg/L in April 2012 to 161 mg/L in December 2015 (Figure 12). It is reported by Hobbs et al. (2011) and Hobbs (2013) that SO_4^{2-} typically comprises 62% (~3700 mg/L) of the median total dissolved solids (TDS) concentration of ~5200 mg/L in Western Basin raw mine water. This proportion reduces to ~20% in ambient surface water and <2% in pristine karst groundwater in the study area. A value of much >20% in surface water and very much >2% in groundwater therefore serves as an indicator of a mine water impact on a respective receiving water resource. The ratio value for Zwartkrans Spring water in December 2015 was 46% (Table 1). By comparison, the value of 31% for the coeval Lake water composition reflects a lesser influence of mine water on the cave water chemistry. Nevertheless, the progression of this ratio and the SO_4^{2-} concentration in Lake water since April 2001 (and especially since 2010) shown in Figure 12 validates concern for the future quality of this water.

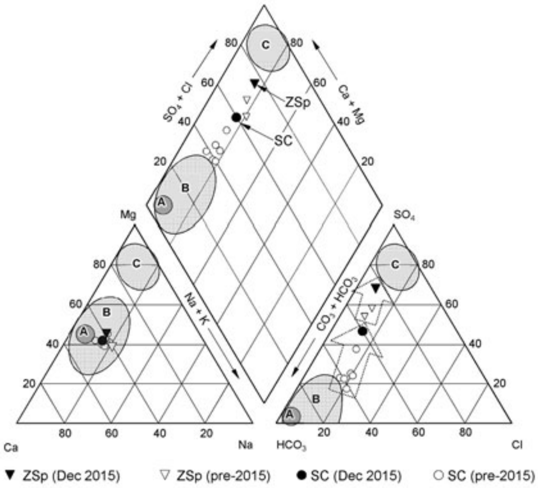

Figure 11. *Piper diagram of historical (open circles) and recent (arrowed) Lake water (SC) chemistry compared to historical (open triangles) and recent (arrowed) Zwartkrans Spring (ZSp) water, as well as the typical positions of other relevant water type plotting fields (**A**) = 'pristine' karst groundwater, (**B**) = general natural to weakly impacted karst groundwater, (**C**) = mine water and associated sources).*

Seen in context, the impact of mine water on the karst groundwater in the broader Zwartkrans basin is already problematic under circumstances where numerous 'private' water supply boreholes reflect SO_4^{2-} concentrations of up to ~2000 mg/L (Hobbs, 2016). This compares poorly to the 500 mg/L limit set in the SANS (2015a; 2015b) national standard for drinking water.

Table 1. Analytical results for Sterkfontein Cave Lake and Zwartkrans Spring water chemistry.

Variable	**Unit**	**Sterkfontein Cave Lake**									
Date	**mm/yyyy**	**04/2001**	**04/2005**	**02/2006**	**05/2009**	**05/2010**	**01/2011**	**07/2012**	**05/2014**	**02/2015**	**12/2015**
pH	$-\log_{10}\alpha_{H+}$	7.5	8.1	7.9		7.90	7.6	8.1	7.5	7.8	8.1
EC	mS/m	74.0	59.0	57.0		59.00	45.5	66.4	68.6	66.4	78.0
TDS	mg/L	365.0	369.0	406.0		364.00	394.0	423.0	502.0	500.0	528.0
Ca	mg/L	47.0	44.0	46.0		47.00	48.0	48.0	61.0	65.0	69.0
Mg	mg/L	27.0	28.0	37.0		26.00	27.0	34.0	42.0	39.0	46.0
Na	mg/L	14.0	23.0	20.0		20.00	22.0	20.0	26.0	28.0	29.0
K	mg/L	0.2	1.0			1.10	1.0	1.0	1.5	2.1	1.4
SO_4	mg/L	46.0	58.0	61.0		58.00	71.0	78.0	126.0	129.0	161.0
Cl	mg/L	38.0	31.0	46.0		27.00	37.0	44.0	39.0	38.0	37.0
T.Alk.	mg/L	158.0	151.0	161.0		152.00	154.0	162.0	169.0	163.0	151.0
NO_3-N	mg/L	8.4	7.0	7.5		8.20	9.3	1.0	7.5	7.7	4.4
^{2}H	‰				−19.80	−20.30					
^{18}O	‰				−3.53	−3.55					
^{3}H	TU				1.3 ± 0.30						
Error	% diff.	5.5	0.7	0.9		1.90	5.6	0.7	0.2	1.0	4.9
So_4 /TDS	%	13.0	16.0	15.0		16.00	18.0	18.0	25.0	26.0	30.0

Variable	**Unit**	**Zwartkrans Spring**							
Date	**mm/yyyy**	**05/2006**	**07/2009**	**05/2010**	**10/2012**	**10/2013**	**05/2014**	**02/2015**	**12/2015**
pH	$-\log_{10}\alpha_{H+}$			7.60	7.2	7.1	7.1	7.7	8.0
EC	mS/m			79.00	96.1	109.0	105.5	106.0	124.0
TDS	mg/L	492.0		547.00	643.0	714.0	773.0	797.0	882.0
Ca	mg/L	60.0		66.00	77.0	91.0	103.0	109.0	125.0
Mg	mg/L	38.0		37.00	44.0	51.0	65.0	63.0	69.0
Na	mg/L	41.0		36.00	44.0	45.0	47.0	51.0	50.0
K	mg/L	1.4		1.50	1.8	2.3	2.2	2.3	2.0
SO_4	mg/L	123.0		154.00	240.0	295.0	304.0	344.0	409.0
Cl	mg/L	55.0		57.00	36.0	48.0	46.0	44.0	42.0
T.Alk.	mg/L	142.0		160.00	165.0	149.0	169.0	151.0	152.0
NO_3-N	mg/L	12.0		12.00	8.6	12.0	9.1	9.2	7.5
^{2}H	‰		−16.80	−17.60					
^{18}O	‰		−3.17	−3.11					
^{3}H	TU	0.8	2.4 ± 0.30						
Error	% diff.	25.0		5.50	2.6	2.6	3.9	3.3	3.2
So_4 /TDS	%			28.00	37.0	41.0	39.0	43.0	46.0

Note: The 2009 isotope data are from Hobbs et al. (2011), and are presented for validation of the May 2010 data, as well as for reporting of the informative tritium (^{3}H) data.

Included in the Lake water analyses is nitrate (NO^{3-} as N), which is not found in mine water. The concentration in Lake water (4.4 mg/L in December 2015) and contemporary springwater (7.5 mg/L) indicates a measure of additional contamination by nutrient-rich water. Possible sources of nitrate include the following:

- upstream agricultural land use practices, e.g. the Oaktree area supports extensive irrigated agriculture that includes maize, vegetables and cut-flowers requiring fertilisation;
- municipal wastewater effluent discharged from a wastewater treatment works into the Blougat Spruit at locality 'B' (Figures 1 and 2), resulting in a median concentration of ~11 mg N/L in the receiving watercourse (Hobbs et al., 2011);
- on-site sanitation facilities on the numerous smallholdings in the Oaktree area, as this area is not served by the local municipality sewerage system;
- the pre-2007 operation of a piggery on a property located ~1150 m upstream (southwest) of the cave (Figure 2) on the right bank of the Bloubank drainage, anecdotal evidence suggesting that this facility routinely washed its accumulation of piggery waste into the watercourse as a means of disposal; and
- the sanitation facility serving the Sterkfontein Cave Visitors Centre which, until its replacement in mid-2011 with an environment-friendly self-contained bioreactor system, comprised a septic tank system.

Although still within the SANS (2015a; 2015b) limit of 11 mg N/L set for drinking water, the presence of nutrients in the cave water raises concern for the impact on stygofauna colonies in the cave system. Graening and Brown (2000) and Boulton et al. (2003) recognise the negative impacts of organic pollution on cave ecosystems as being related to (a) the alteration of the community assemblage, (b) impoverishment of biodiversity,

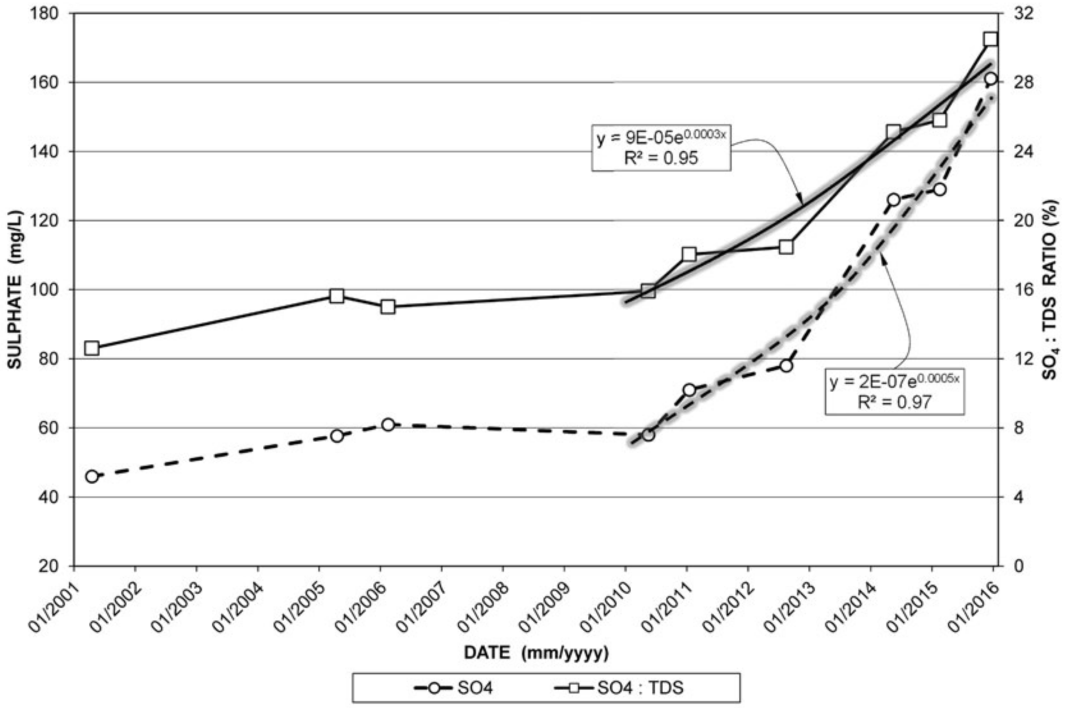

Figure 12. *Pattern and trend of SO_4^{2-} concentration and the SO_4:TDS ratio in Lake water composition since 2001, showing the exponential trend since 2010.*

and (c) increased risk of predation from surface fauna. Routine biomonitoring activities, particularly those that target stygofauna, offer an early warning of potentially catastrophic longer-term impacts on stygobitic fauna (Culver and Sket, 2002). It is recognised that cave ecosystem monitoring extends well beyond the water resource environment alone, and includes disciplines such as geomicrobiology, geobiology, biogeochemistry and biogeography (Engel et al., 2008). The authors therefore wholly support the monitoring requirements put forward by Durand and Peinke (2010) in developing a holistic understanding of the karst ecology of the COH within the paradigm advanced by Williams (2008).

The installation of an EC data logger in a well adjacent to the Zwartkrans Spring in July 2013 provides the record of springwater salinity shown in Figure 13. This reflects a gradual rise at an initial rate of 0.8 to 1 mS/m per month, reducing to 0.5 mS/m per month by August 2016. The anomalously low values recorded at point 'A' on the timeline mark the passing of a flood event between late-January and mid-February 2014 that inundated the spring and well by up to some 2 m. The much broader anomaly at point 'B' is less readily explained, perhaps marking the passage of karst groundwater bearing a mine water impact diluted by natural recharge associated with the preceding flood-generating conditions in the catchment. The observations demonstrate the utility of spring measurements (e.g. flow behaviour, turbidity, chemistry) to (a) reflect a composite of 'everything that has happened upstream' (White, 2002), and (b) provide information on the functioning of the whole hydrosystem drained by the spring (Bakalowicz, 2005).

The environmental (stable) isotopes of hydrogen (^{2}H) and oxygen (^{18}O) together provide a further indicator that sheds light on the impact of mine water on the cave water chemistry. Holland (2007), Hobbs and Cobbing (2007) and Hobbs et al. (2010; 2011) have shown the utility of these isotopes in distinguishing pristine karst groundwater from that impacted by an evaporative allogenic component such as mine water. The isotopic signature of a Lake water sample is superimposed on those for a range of local surface and groundwater sources in Figure 14. The comparatively close grouping of springwater in the 'pristine' karst groundwater field is apparent. Further up the local evaporation line is the May 2010 impacted Sterkfontein Cave water, and beyond that the coeval more impacted Zwartkrans Spring water. This trend reflects the transition from isotopically 'heavier' pristine karst groundwater to 'lighter' increasingly impacted water as the evaporative signature assumes dominance.

Discussion

Groundwater flow

The hydrostatic response of the Sterkfontein Cave water level reflects that observed elsewhere in the Zwartkrans Compartment. Further, the recent driver of this response is a combination of exceptional autogenic and allogenic recharge in the 2008, 2010, 2011 and 2014 hydrological years. The implementation of mine water control and management interventions in the Western Basin (Coetzee et al., 2011) forecasts a greater and sustained discharge of treated/neutralised mine water into the Zwartkrans Compartment karst basin via the Tweelopie Spruit. The current ~34 ML/d capacity of the mine water treatment plant is earmarked for expansion to ~53 ML/d (DEA, 2014) sometime in the future.

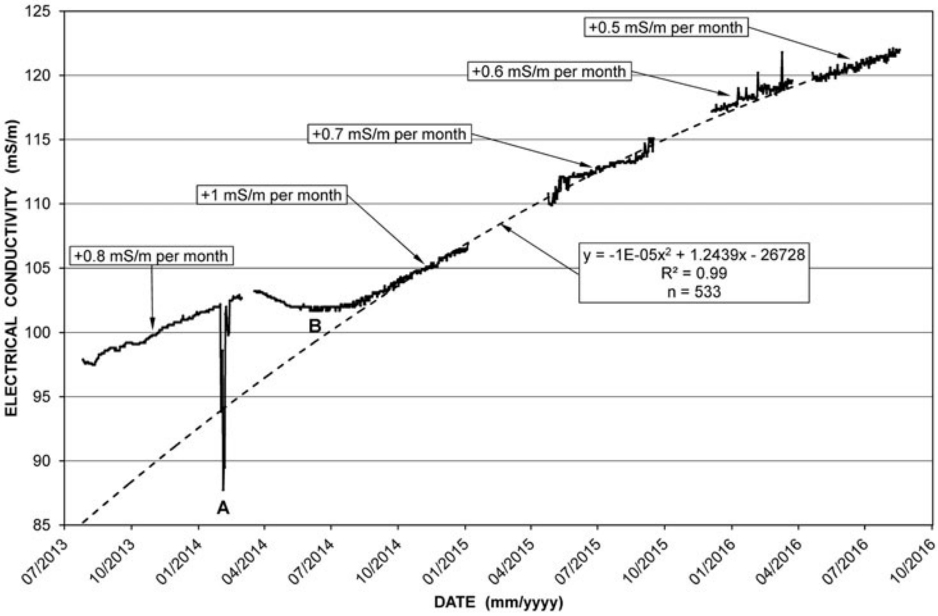

Figure 13. *Chemograph of recent electrical conductivity pattern and trend of Zwartkrans Spring water as reflected in daily data logger recordings; the sharp anomaly labelled **(A)** is explained in the text, and a hypothesis put forward to explain broader anomaly **(B)**.*

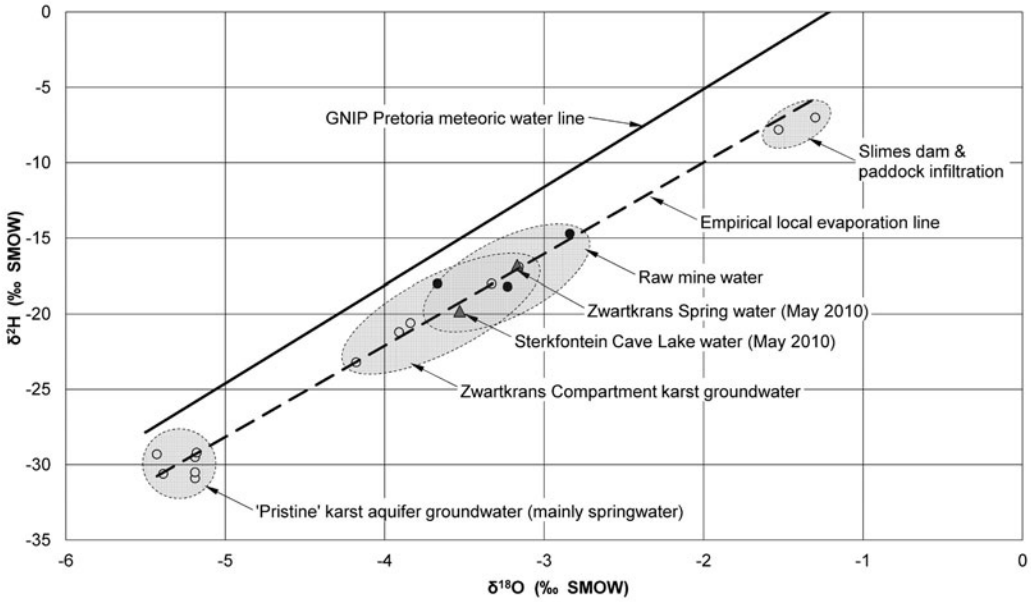

Figure 14. *Stable isotope composition of water associated with various sources relevant to the study area, presented as characteristic plotting fields that distinguish between sources on the basis of the relative evaporative signature associated with primarily a mine water impact; solid symbols represent raw mine water point sources (modified from Hobbs et al., 2011).*

The fluctuating potentiometric response in the lower reaches of the Zwartkrans basin immediately upstream of the Zwartkrans Spring dictates the magnitude of groundwater resurgence in the channel of the Bloubank Spruit. This groundwater discharge is additional to that of the spring, and increases as the water table intersection with the stream channel migrates upstream following a rising water table. These circumstances are reversed with a declining water table, the response benefitting from a highly permeable epikarst.

Wilkinson (1973) also recognised resurgence of groundwater in the valley of the Bloubank Spruit as an important hydraulic element of the cave system. The mechanism suggested by Wilkinson (1973), however, considered that resurgence occurred as underflow on the buried contact between the alluvial valley fill deposit and the underlying dolomitic bedrock located 12 m below the streambed. Further, that the resurgence was driven by a hydraulic gradient of 0.1 (6°) directed to the north roughly normal to the stream channel. Current understanding presented in this paper indicates a much more gentle hydraulic gradient of 0.003 to 0.005 (0.17 to 0.29°) directed to the northeast.

Exploration borehole information presented by Hobbs et al. (2011) suggests that the main flowpath (thalweg) of groundwater discharge through the Zwartkrans Compartment toward the spring coincides with the Bloubank Spruit drainage and its ephemeral upper reach (Figures 1 and 2), and also the Riet Spruit tributary. In the vicinity of Sterkfontein Cave, this flowpath lies ~600 to 700 m to the northwest of the cave. It is hypothesised, therefore, that the location of the cave system on the southeast periphery of the thalweg offers an explanation for the lesser mine water impact on the Lake water chemistry. Reference by Groenewald (2010) to the cave as occupying a low energy groundwater system, and by Martini et al. (2003) to the apparently static pools in the cave system, support a 'backwater' location on the margin of a postulated subsurface thalweg.

The tritium (^{3}H) results reported in Table 1 for Lake and springwater provide further support for a 'backwater' location of the cave system. Under circumstances where ^{3}H values <2 tritium units (TU) are generally associated with groundwater residence times in the order of several decades, and higher values are indicative of more recent recharge and faster circulation (see for e.g. Hershey et al., 2010), the difference between the Lake and springwater ^{3}H values indicates differing hydrodynamic conditions between the cave and spring settings.

The thalweg hypothesis finds support in the 'flood induced pseudoplume' described by Ewers et al. (2012), which postulates the intrusion of contaminated groundwater from a flooded major conduit into peripheral porous karst strata during storm/flood events. Although the intruded water returns to the major conduit as groundwater levels recede following passage of the flood, it leaves behind contaminants in various settings in the intruded portion of the karst aquifer. These circumstances are more likely to prevail in the lower (downstream) reaches of a karst basin where conduits are convergent (Ewers et al., 2012), a hydrogeologic setting that suitably describes the locations of the Sterkfontein Cave system and the nearby Zwartkrans Spring. The association of raw mine water with flood conditions in the receiving water resources environment provides the 'contaminants' (e.g. SO_4^{2-}) in the flood induced pseudoplume. The concomitant rise in groundwater levels similarly provides the mechanism for the lateral intrusion of 'contaminated' groundwater into the cave system.

Finally, it is also worth noting that the cave straddles the north-northwest-dipping contact between the chert-poor Oaktree Formation that forms the southeast (and oldest) lithostratigraphic margin of the Malmani Subgroup, and the overlying chert-rich strata of the Monte Christo Formation (Martini et al., 2003). The latter is generally associated with a more productive aquifer, in contrast to the typically more barren character of the Oaktree Formation. These circumstances underscore the distal location of the cave Lake relative to a postulated subsurface thalweg. This is illustrated in Figure 15, which reflects a combination of the hydrophysical and hydrochemical dynamics that describe the cave system hydrogeology in relation to the broader karst environment along two transects labelled Profile 1 and Profile 2 respectively.

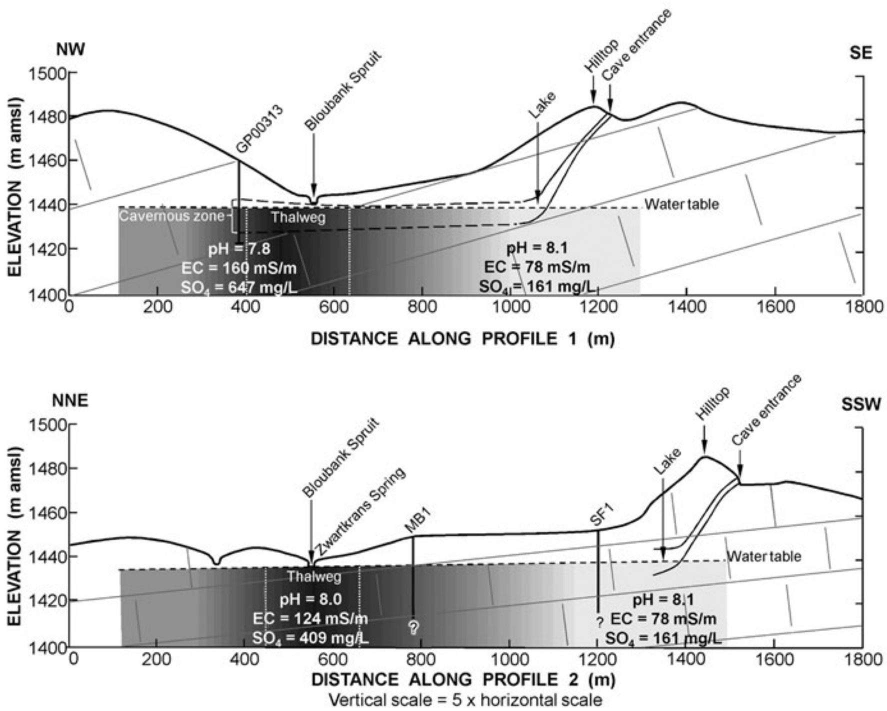

Figure 15. *Schematic hydrophysical and hydrochemical profiles (see Figure 2 for positions) through the Bloubank Spruit valley and Sterkfontein Cave illustrating the relationship between the groundwater and Lake water elevations, and the stream channel and subsurface thalweg; the lateral gradational shading reflects the relative intensity of a mine water impact on ambient karst groundwater as shown by the December 2015 pH, EC and SO_4^{2-} values for representative thalweg and cave sampling locations.*

Physico-chemical groundwater flux

The semi-quantitative assessment of allogenic and autogenic recharge contributions to the Zwartkrans karst basin together with the hydrochemical characterisation of the input-output components, provides a means to explore the physico-chemical groundwater budget for this basin. The mine water impact adds relevance to such an exercise.

Allogenic mine water recharge since mid-2011 amounted to 12.4 ML/d on average. The path followed by this recharge encompasses monitoring boreholes A2N0584 and A2N0586 at the influent (upstream) end, and borehole A2N0600 to the outflow at the Zwartkrans Spring to the northeast. The coarse dimensions of the flowpath from the upstream end to borehole A2N0600 are 4500 m (L) and 2200 m (W). The potentiometric difference of ~30 m defines a hydraulic gradient (i) of 0.0067. Applying Darcy's law in the form $Q = T \cdot i \cdot w$, and assuming an aquifer transmissivity (T) value of ~1000 metres squared per day (m^2/d) [roughly 50% of the T-value determined for the aquifer intersected in borehole A2N0586 (Bredenkamp et al., 1986)], yields an estimated flow of 14.7 ML/d. This is similar to that associated with allogenic mine water recharge as per Figure 8. The estimate does not, however, include the contribution of autogenic recharge accruing to the basin. A mass balance calculation is applied to account for this contribution.

Using the relationship $Q_U C_U = Q_D C_D$ where:

Q_U = 14.7 ML/d (allogenic recharge contribution),
C_U = 2170 mg/L (SO_4^{2-} concentration in allogenic recharge),
C_D = 720 mg/L (SO_4^{2-} concentration at borehole A2N0600), and
Q_D (? ML/d) is the unknown combined contribution of allogenic and autogenic recharge at borehole A2N0600,
yields a Q_D value of 44.3 ML/d.

This indicates an additional contribution of 29.6 ML/d at borehole A2N0600, of which at least 3 ML/d derives from municipal wastewater recharge. The balance (26.6 ML/d) is attributed to autogenic recharge from rainfall. This equates to a water depth of 99 mm or 14% of an MAP of 710 mm on the ~9800 ha catchment, placing it at the lower end of the recharge rate of 17 ± 5% of MAP that finds general application in the COH.

The reasonable agreement between the derived values generates cautious confidence in the groundwater fluxes that broadly characterise the Zwartkrans karst basin since January 2010, when the mine water impact has been more severe. Caution is indicated by the use of SO_4^{2-} as a 'tracer' because of its propensity to participate in any number of bio(geo)chemical reactions such as bacterial sulphate reduction that might influence the mass balance calculation. Further caution is indicated by non-recognition of factors such as longitudinal dispersion in the aquifer and 3-component mixing. Nevertheless, Hartmann et al. (2017) have demonstrated the combined utility of SO_4^{2-} and flow states in karst modelling.

Groundwater balance

The groundwater flux values provide a means to derive an approximate groundwater balance for the Zwartkrans basin. The semi-quantitative basic input values for such a calculation are natural (autogenic) recharge of ~27 ML/d (9.9 Mm^3/a) and allogenic recharge of ~18 ML/d (6.6 Mm^3/a as mine water and municipal wastewater) for a total of ~45 ML/d (16.4 Mm^3/a). Basic output values are groundwater discharge of ~29 ML/d (10.6 Mm^3/a as spring discharge and instream groundwater resurgence) and groundwater use which, in the amount of ~16 ML/d (5.8 Mm^3/a), would balance the groundwater budget. Holland and Cobbing (2008) report a total groundwater use of 8.7 Mm^3/a as being registered with the Department of Water and Sanitation (DWS) for mainly agricultural use in the basin. This figure contrasts sharply with those estimated by Schoeman and Associates (2006) of 14.1 Mm^3/a (reported in Holland and Cobbing, 2008) based on an interpretation of satellite imagery, and of 25.7 Mm^3/a estimated by JFA (2006) from a water budget calculation. A further measure of the uncertainty in groundwater use figures is provided by the 15 Mm^3/a reported earlier by Bredenkamp et al. (1986), and the 18 Mm^3/a by Van Biljon (2006). The uncertainty in this budget component will need to be resolved if a more accurate groundwater balance is to be obtained.

Conclusions

The Zwartkrans karst basin which hosts the Sterkfontein Cave system exemplifies a karst environment that is subjected to a mine water impact. The characteristics of this impact on the Sterkfontein Cave system in particular, and on the broader water resources environment in general, are described on the basis of mutually supportive hydrophysical and hydrochemical information.

A rise of ~3 m in water level in the cave system since late-2009 is attributed to the combined impact of allogenic recharge mainly from mine water, and autogenic recharge from rainfall, both driven by the wet 2010, 2011 and 2014 summer rainfall seasons. The high water level (currently at an elevation of ~1439 mamsl) will be maintained into the future, given the greater sustained discharge of treated/neutralised mine water associated with the mine water control and management interventions in the Western Basin. It is, however, unlikely to exceed an elevation of ~1440 mamsl because of the hydrogeological control exercised on natural discharge from the karst basin.

The prognosis for the Lake water chemistry is that continued deterioration is inevitable, but is unlikely to reach the extent observed in the broader groundwater environment, and even less likely that of surface water. This is premised on the distal location of the cave hydrosystem relative to the main groundwater flowpath through the Zwartkrans karst basin toward the Zwartkrans Spring. These conclusions might temper short- to medium-term concerns for the future of this UNESCO-inscribed fossil site. The dynamic response of the water resources environment to a variety of hydrological and hydrogeological drivers, however, cautions against complacency in this regard, and reinforces the need for monitoring vigilance across a wide spectrum of disciplines such as hydrology, hydrogeology, geo(micro)biology and biogeochemistry.

A greater concern exists for the broader water resources environment, which reflects a more severe but spatially constrained mine water impact in the Zwartkrans karst basin. For example, sulphate levels of up to ~2000 mg/L in a portion of the karst aquifer severely compromise the potability of this water for affected land owners dependent on this resource for water supply. Further, fossil sites such as Bolt's Farm, Rising Star and Swartkrans are located within the thalweg that describes the principal groundwater drainage and coincident poorer water quality pathway through the lower portion of the Zwartkrans basin. These circumstances underwrite the importance of the water resources monitoring programme that is supported by the COH WHS Management Authority in conjunction with the Department of Water and Sanitation.

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