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Alluvial aquifer characterisation and resource assessment of the Molototsi sand river, Limpopo, South Africa

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ABSTRACT

Study region: Molototsi sand river, Limpopo, South Africa.
Study focus: Ephemeral sand rivers are common throughout the world's dryland regions, often providing a water source where more conventional sources are unavailable. However, these alluvial aquifers are poorly represented in the literature. Extensive field investigations allowed estimation of stored water volume and characterisation of an alluvial aquifer.
New hydrological insights for the region: Computed alluvial aquifer properties included hydraulic conductivity of 20–300 m/d, porosity of 38–40%, and aquifer thickness of 0–6 m. Dykes and other subcrops commonly compartmentalise the aquifer though do not form barriers to flow. A hydraulic disconnect between deep groundwater (occurring in fractured metamorphic rocks) and the alluvial aquifer was revealed by groundwater levels and contrasting hydrochemistry and stable isotope signatures. The dominant recharge process of the alluvial aquifer is surface runoff occurring from torrential tributaries in the catchment's upper reaches. A fraction of available storage is currently abstracted and there exists potential for greater exploitation for smallholder irrigation and other uses.

1. Introduction

Ephemeral sand choked rivers commonly occur in the world's dryland regions. Such systems experience surface flows only following infrequent torrential rainfall (Tooth, 2000). Where the underlying geology is of low permeability, e.g. African crystalline basement, infrequent torrential flows fully recharge the alluvial aquifer creating an accessible water resource where unfavourable climate and geology create few alternatives. Drylands occupy 41.3% of the world's surface and are home to 2.1 billion people (UN, 2017), many of whom obtain their water supplies from resources contained within "sand rivers" (Seely et al., 2003; Love et al., 2011). Smallholder irrigation often obtains water from sand rivers where minimal head difference and proximity of river channel to riparian fields minimises equipment and energy costs. The relatively low stored water volumes can be sufficient for small-scale farming activities (Hussey, 2007; Love et al., 2007).

South African agricultural policy makers and farmers recognise that subsistence/smallholder agriculture can reduce the

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vulnerability of food-insecure households, improving livelihoods, boosting nutritional intake, and help mitigate food price inflation (Wenhold et al., 2007; Baiphethi and Jacobs, 2009). Most poor and food-insecure households in South Africa are concentrated in former rural homeland areas. Increased food productivity by subsistence and smallholder farmers in these areas would enhance long-term food security reversing the decline of smallholder agriculture (Baiphethi and Jacobs, 2009; Pereira and Drimie, 2016). The decline in smallholder farming is in part due to increased failure of borehole and surface water sources, especially poorly maintained systems installed before 1994 (personal communication, local farmers, October 2016 and May 2017). Identification of alternative cheap water sources to enhance smallholder agriculture should be prioritised. Sand rivers are potentially such alternative water sources.

Sand rivers are defined as shallow unconsolidated alluvial deposits presently accumulating along active stream courses within which thin saturated basal sands form limited aquifers. The depth to water table is less than a two metres, within unconsolidated sands that are regularly recharged. Sand rivers can be considered a renewable resource where long-term groundwater depletion is not expected if used sustainably (Owen, 1989). Moderate yields are possible from permeable alluvial deposits with high initial specific yield. However, the stored volume is often a limiting factor as sand rivers range in width from 10 to 100s m and in thickness from 1 to 30 m. Water qualities are generally good due to frequent recharge and the filtering effect of the sand (Nord, 1985; Owen and Dahlin, 2005; Cobbing et al., 2008). Sand river aquifer recharge is almost entirely from surface water flow during occasional floods generated by torrential rainfall events; little recharge results from normal annual precipitation (Owen and Dahlin, 2005).

Similar landforms elsewhere in the world, such as wadis in the Middle East (e.g. Meirovich et al., 1998; Gheith and Sultan, 2002) and arroyos in southwestern USA (e.g. Reid and Frostick, 1997; Graf, 1988), are fairly well-researched, due to local recognition of their value, local research capacity, and data availability (Tooth, 2000). In contrast, sand rivers in Africa have received little attention in peer-reviewed literature. Much of the research is available only in grey literature. Handbooks by Nissen-Petersen (2006) and Hussey (2007) provide guidance for the exploitation of sand river water resources. Reports describing the exploitation of sand rivers in Botswana are presented by Davies et al. (1998) and Herbert et al. (1997), in Zimbabwe by Owen (1989), and in Southern Africa by Clanahan and Jonck (2004). Regarding aquifer characterisation studies, geophysics was used to assess aquifer geometry by Owen and Dahlin (2005) in Zimbabwe and Arvidsson et al. (2011) in Mozambique. Field tests for aquifer parameters in Botswana are described by Davies et al. (1998), in Zimbabwe by Mansell and Hussey (2005) and Love et al. (2007), and sand river resource assessments in Zimbabwe using modelling are presented by de Hamer et al. (2008) and Love et al. (2011). Moyce et al. (2006) and Mpala et al. (2016) conducted water resource assessments using remote sensing techniques. Investigations on the Kuiseb sand river in Namibia are reported by Dahan et al. (2008); Morin et al. (2009) and Benito et al. (2010). Outside of these areas, there are few published aquifer characteristic data for assessment of the sustainability of sand river exploitation (Mansell and Hussey, 2005). This study contributes data from field investigations and observations of a sand river currently utilised for small-scale irrigation. The objectives include characterisation of aquifer properties, geometry and relationship between surface water, and shallow and deep groundwaters. The study aims to estimate the stored water volume to determine if the resource is being used sustainably and potential for further exploitation. The Molototsi study site is typical of sand rivers found in the wider region, overlying crystalline basement rocks and receiving infrequent surface flows. Increased exploitation of sand rivers in the region could mitigate the impact of borehole and surface water infrastructure deterioration by increasing food security, alleviating poverty and boosting nutritional intake.

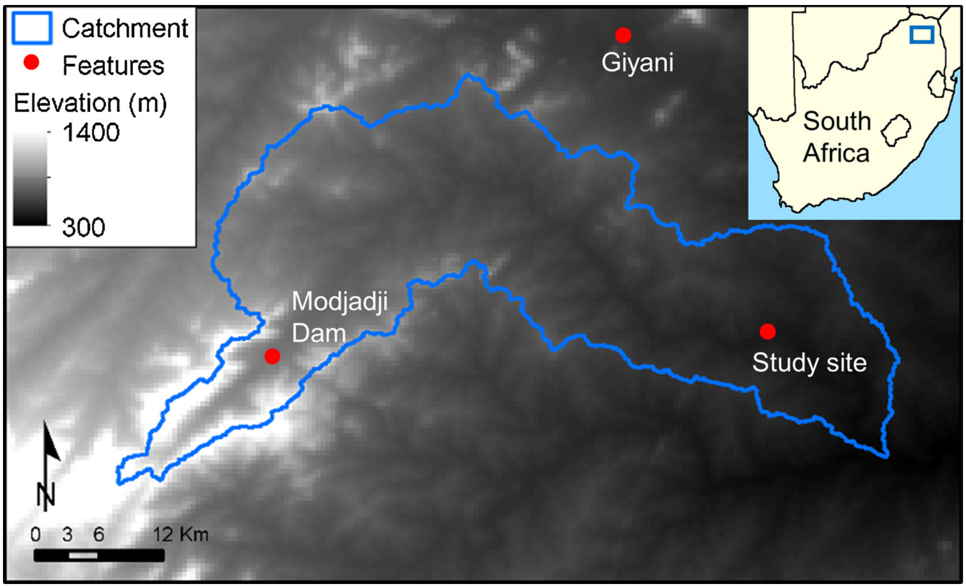

Fig. 1. Study site location.

2. Study site

The Molototsi catchment lies within the Mopani District of Limpopo Province in northeast South Africa (Fig. 1). The catchment covers an area of 1170 km^2 and the river has a length of approximately 120 km. The Molototsi drainage system comprising quaternary catchments B81G and B81H, as defined by the South African Department for Water and Sanitation (DWS), is a torrential or sand river system with infrequent surface flow. The catchment drains west to east where it joins the perennial Groot Letaba River that flows through Kruger National Park into the Olifants River forming part of the transboundary Limpopo Basin.

The upper reaches of the Molototsi catchment are hilly with populated valleys and forested slopes. Approximately 70 km^2 of these upper reaches lie upstream of the Modjadji Dam, which was constructed in 1997 with a capacity of 8.4 Mm^3 to supply potable water to the Groot Letaba municipal area (DWAF, 2010). Downstream of the dam are flat, sparsely populated plains supporting a mopaneveld vegetation (scrub and bushland); population centres and agriculture becoming sparser in the drier east away from the mountains. A pipeline is under construction to supply potable water from the Nandoni Dam (in a catchment to the north) to the sparse villages (Ringwood, 2016). These villages have no sewage systems and may represent a significant source of groundwater pollution (Mopani, 2017a). The catchment is ungauged (with the exception of an 18-month period during dam construction) though data are available from DWS recording Modjadji Dam releases. Dam releases are via a spillway when the reservoir exceeds full capacity and have occurred in nine of the 21 years from 1997 to 2018. No release has occurred since April 2014 (to June 2018).

Agriculture in South Africa is dominated by a corporate-industrial structure with large-scale commercial agriculture producing over 90% of marketed food (Greenberg, 2013). Recent agricultural policies focus on the importance of subsistence agriculture recognising its contribution to food security and poverty alleviation (Baiphethi and Jacobs, 2009). Within Mopani District, only 6.7% of land is designated as arable of which 43% is under irrigation (Mopani, 2017b). The most important crops in terms of monetary value are citrus and subtropical fruits (e.g. mango and banana) with several irrigation schemes in the region utilising water supplied by dams on the Letaba River. There are no such large irrigation schemes within the Molototsi catchment; rather the farms are small and largely non-mechanised. Physical measurement of a sample of 142 farms within the catchment using Google earth reveals a median size of 4.8 ha and a 10th to 90th percentile range of 0.9–17.7 ha. Most of these farms are run by cooperatives with permission to farm communal lands governed under the Traditional Leadership and Governance Framework (Act 6, 2005). Discussions with local farmers revealed that most of these cooperative farms irrigate crops with water obtained from boreholes drilled by DWS. However, riparian farms often utilise water from sand rivers even if they have a borehole. Sand river abstraction points include simple hand-dug pits, open brick-lined wells (Fig. 2) and buried metal tanks. Furrow and drip irrigation systems dominate, using diesel pumps. Use of electric submersible pumps is rare due to a general lack of rural electrification. Crops cultivated for local market in addition to domestic subsistence use include okra, courgettes, tomatoes, maize, chili peppers, *amaranthus* and fruits such as mango and watermelon.

The climate of this region of Southern Africa is categorised as hot semi-arid according to the Köppen-Geiger climate classification

Fig. 2. Riverbed well at the study site in October 2016 during a drought.

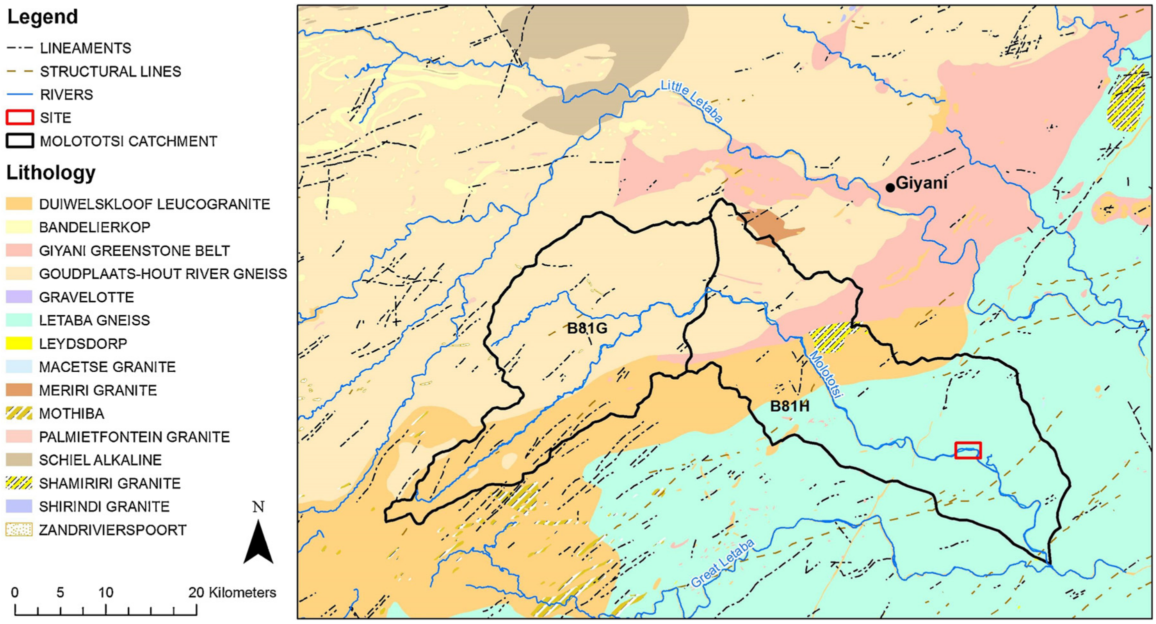

Fig. 3. Geological map of the Molototsi catchment including delineation of the DWS quaternary catchments and the study site location.

system (Peel et al., 2007). The region experiences a single rainfall season during the summer months of November to March that is characterised by high interannual variability and occasional torrential rainfall. Potential evapotranspiration greatly exceeds precipitation, droughts are regular phenomena, often coinciding with El Niño – Southern Oscillation (ENSO) periods (Lindesay, 1988; Cook et al., 2004; Reason et al., 2005). Severe droughts in the past in Limpopo have resulted in crop failure and economic loss (Trambauer et al., 2014). Limpopo suffered a period of reduced rainfall from around 2012 culminating in the major drought in 2016 (Maponya and Mpandeli, 2016; Sunday Times, 2016); this is affirmed by DWS rainfall and dam status data.

The regional air circulation is governed by high pressure cells located across the northern part of Botswana and South Africa that interact with low pressure cells from the Antarctic besides warm air masses generated from the southern Indian Ocean (Abiye, 2016). A distinct rainfall gradient is observed between the catchment headwaters in the west and the lowveld to the east. In the west, a DWS rain-gauge at Modjadji Dam measured mean annual rainfall for 1998–2017 of 1219 mm. No formal rain-gauges are sited in the east of the catchment though a farmer-monitored rain-gauge close to the study site recorded a mean annual rainfall from 2006 to 2017 of 430 mm.

The geology of study site is described by Holland (2011) and Witthüser et al. (2011): Archaean crystalline basement rocks dominate the regional geology (Fig. 3), these being some of the oldest rocks on Earth with ages up to 3600 Ma (Kramers et al., 2007). The hydrogeologically significant Ventersdorp-related mafic dykes, represented as structural lines in Fig. 3, were intruded around 2700 Ma ago (Ernst, 2014). The NE-SW trending structures, clearly visible on satellite imagery and on aeromagnetic maps, relate to the position of this region within the metamorphic Limpopo fold Belt. The Limpopo Belt is considered to be the earliest identified Himalaya-type orogeny, formed when the Zimbabwe and Kaapvaal cratons collided around 2600 Ma ago (Zhao et al., 2000). A geological cross-section of the study site (Fig. 4) shows boreholes drilled where geophysical surveys identified dykes to exploit groundwater within zones of enhanced fracturing at the dyke country rock margins.

The Molototsi is typical of ephemeral sand rivers throughout the Limpopo region. A 1300 m long reach of its course was selected for study. This focus study site (Fig. 5), was chosen due to the presence of a large number of boreholes and a riverbed well, the latter used for small-scale irrigation. The Molototsi sand river, at the study site, is considered typical of the landform in terms of geometry, gradient, and geology.

3. Methodology

3.1. Groundwater levels

Groundwater levels were monitored to assess the relationship between deep and shallow groundwater. Three boreholes were drilled by DWS at the study site to 120 mbgl (metres below ground level) in July/August 2016. Two of these boreholes (H14/1702 and H14/1703) were equipped with Solinst pressure transducers to monitor groundwater levels at hourly intervals. Water levels in the riverbed well and the other boreholes were monitored using a manual dip-meter approximately every three months to May 2017.

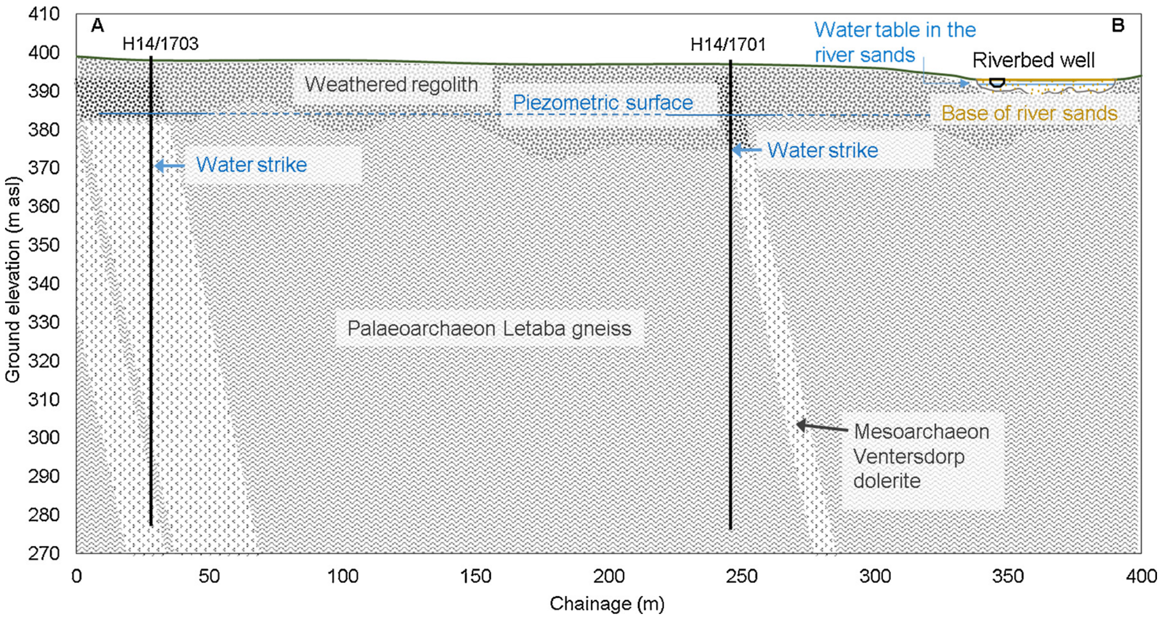

Fig. 4. Geological cross section at the study site based on DWS drillers' logs (see Fig. 5 for cross section location).

Fig. 5. Plan of the study site showing borehole information, testing and sampling locations (SI = stable isotope, FHP = falling head permeameter) and the geophysical survey lines (VES = vertical electrical sounding) along the Molototsi sand river. The elevations are taken from a 20 m DEM developed from 20 m topographic contour data and spot heights by the Department of Geography and Environmental Studies at the University of Stellenbosch (m asl = metres above sea level). Groundwater levels are as measured in October 2016. Depths and water strikes are from DWS drillers' logs; DUV1 is an old borehole for which no logs could be found. Image source: Google earth; Imagery ©2018 DigitalGlobe.

3.2. Aquifer properties

Four methods of hydraulic conductivity estimation were applied to the sand river in May 2017; test-pumping of the riverbed well, falling head permeameter tests, grain size analysis (also for porosity), and salt dilution tests.

Two pumping tests were undertaken by measuring drawdown and recovery with a pressure transducer logging water levels at 30 s intervals in the riverbed well (Fig. 2) during daily abstraction periods (~3–4 h) for irrigation. The well, of nominal diameter of 4 m, is lined with brick lining to 1.2 m, with a cone-shaped base below the brick lining where the pump lies 1 m below the rest water level located at 0.5 mbgl. An abstraction rate of 1 l/s was determined using the farmer's flow meter. Drawdown data were analysed using the Papadopulos and Cooper (1967) method with AquiferWIN32 software (ESI, 2007), while recovery was analysed manually by the Barker and Herbert (1989) method using generated nomograms. These methods were specifically developed for application to data from large diameter wells where well bore storage is considered.

Falling head permeameter tests were conducted at 12 sites in the sand river. Test sites were installed by excavating pits into the sand to the water table, then inserting a plain, open-ended tube 510 mm long and 43 mm diameter halfway into the riverbed. The test involved filling the tube with water so that the falling water level rate could be measured by a pressure transducer until equilibrium is achieved. Testing sites were chosen to measure falling head rates within a range of alluvial grain sizes and positions within the channel, including coarse clean sand in active central channel, gravel bars, and silty lee of mid-channel outcrop. The vertical hydraulic conductivity was estimated using the Hvorslev (1951) method and application of the Darcy equation as described by Chen (2000).

Five unconsolidated sand samples from the sand river were collected and analysed in the laboratory for porosity and grain size distribution through mechanical sieving. There are multiple methods available to estimate hydraulic conductivity from grain size distributions. The HydrogeoSieveXL program presented by Devlin (2015) computes 15 of these methods and was applied to the five grain size analyses.

Six salt dilution tests were conducted to assess groundwater flow velocities. Each test consisted of excavating a pit in the sands to the water table, installing a slotted tube with a bottom end cap (piezometer, 510 mm long and 43 mm diameter) into the saturated river sand and adding ~150 g of table salt. Following solution of salt through agitation, the rate of decline in electrical conductivity (EC) with dissolution due to groundwater throughflow was monitored for an hour with an EC sonde within the piezometer. Four more pits were excavated around each testing site where the background salinity was measured. Testing and analysis followed the methodology presented by Davies et al. (1998) who applied this test to sand rivers in Botswana.

3.3. Aquifer thickness

A crucial aquifer characteristic of sand rivers is aquifer thickness (i.e. saturated sand depth) for estimation of groundwater storage. Methods of sand depth assessment applied included trial pits, probing and geophysical surveys.

Trial pits were manually excavated in the sand river to determine depth to the water table and the nature of the sediment infill while conducting falling head permeameter and salt dilution tests, and while sampling alluvial sands and groundwater. Additionally, an excavator mining sand for road construction was utilised to excavate trial pits to the base of the alluvial sands.

Probing involves inserting, usually by physically hammering, a metal rod into the sands and is the most accurate method of determining depth to bedrock. Probing in this case employed a dynamic cone penetrometer (DCP); such testing was conducted as part of a separate civil engineering project to assess the strength and thickness of subsurface layers.

Geophysical investigations were conducted by DWS in May 2017. DWS initially conducted a simultaneous magnetometer and EM34 survey in the centre of the channel along a 1300 m reach, followed by longitudinal and vertical resistivity surveys (Fig. 5). Magnetometer and EM34 surveys are conducted by DWS when attempting to locate dykes within deep bedrock geology and, while being unable to estimate sand thickness, could identify dykes subcropping within the alluvial sands. A Model G5 Proton Memory Magnetometer (Geotron, South Africa) was used to measure ground magnetic flux density at 10 m intervals. An EM34 (Geonics Ltd, Ontario, Canada) was used to measure subsurface apparent conductivity with a coil spacing of 20 m. The equipment was operated in two dipole orientations: horizontal (vertical coils) for identifying vertical structures at shallower depths and vertical (horizontal coils) for horizontal structures at deeper depths. The effective investigation depths for the two dipole orientations are approximately 0.75 and 1.5 times the intercoil spacing respectively (McNeill, 1983). EM34 measures the bulk conductivity of the ground, the measurements representing a weighted mean (related to depth) of the formations within the range of investigation (MacDonald et al., 2001). The resistivity system (Geotech Environmental Equipment Inc., Colorado, USA) was operated using a Schlumberger array with a half-distance between current electrodes (AB/2) of 30 m for the longitudinal survey. At this electrode spacing, each reading gives the bulk resistivity averaged, though biased towards shallower depths, over a depth of approximately 40 m (Van Nostrand and Cook, 1966). This longitudinal resistivity survey could indicate the consistency of sand aquifer thickness and identify the locations of subcropping dykes. Five cross-channel vertical electrical sounding (VES) surveys were conducted using a Schlumberger array. This expanding array system is used to determine the thickness of the upper dry zone by recognising the resistivity contrast between it and the underlying saturated zone. Similarly, the depth to bedrock is recognised by the marked resistivity contrast between solid rock and saturated sands. Such a survey gives a 1D resistivity profile centred on the centre of the river channel. The VES data (electrode spacing and apparent resistivity) was entered into RES1D, an open source computer program developed by Geotomo Software, Malaysia (Loke, 2001). RES1D computes an estimate of the number of layers, their thickness and resistivity for each profile giving the lowest root mean square (RMS) error.

3.4. Hydrochemistry

Groundwater samples were collected on three occasions in the Molototsi area in February 2012, October 2015 and January 2017. On site analysis involved measurement of pH, temperature and electrical conductivity (EC) while samples were filtered and acidified prior to transport to a laboratory for major ion analysis utilising Dionex ion chromatography (anions) and atomic absorption spectroscopy (cations). Groundwater samples were collected after purging boreholes with a portable water pump or the irrigation pump. Additionally, twenty water samples for stable isotope analysis (δ18O and δ2H) were collected in May 2017 from sites in the sand river, a borehole, surface water and rainwater. Samples were collected in nalgene vials filled to the rim with caps sealed with tape.

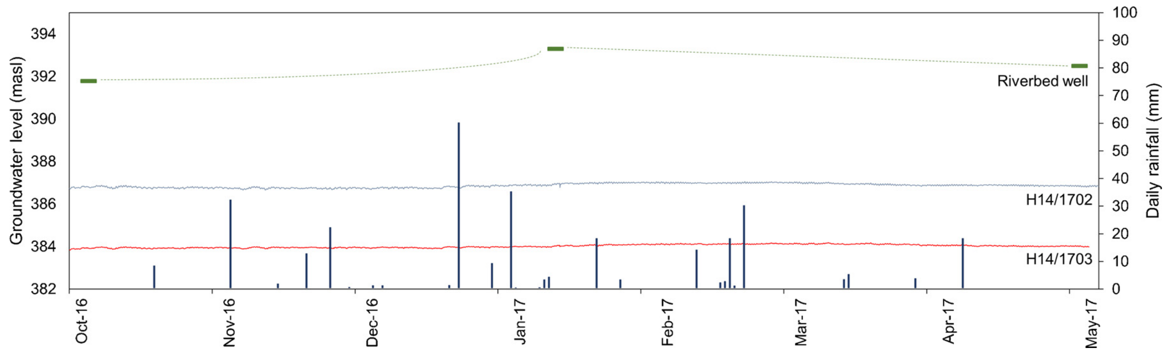

Fig. 6. Borehole and sand river groundwater levels and daily rainfall at the study site. The dashed line for the riverbed well is interpolated between three measurements whereas the borehole groundwater levels are continually monitored.

4. Results and discussion

4.1. Groundwater levels

Fig. 6 shows that logged borehole groundwater levels are stable with little to no identifiable response to rainfall; this is despite the occurrence of four rainfall events with daily totals > 30 mm, one of which had a daily total of 60 mm. Nor does there appear to be a correlation between the borehole groundwater levels and the water level in the sand river. The groundwater level in boreholes H14/1701 and DUV1 also showed little seasonal variation with depths to groundwater stable at 13 and 15 mbgl. These levels are 5–8 m deeper than the observed water table depth in the sand river. Therefore, there is no groundwater discharge into the sand river, i.e. it is not a gaining stream. The Molototsi is also very unlikely to be a losing stream or the sand river water table would not settle at the observed depth of ~1 m and a response would likely be seen in the borehole groundwater levels to surface water flows. The groundwater levels are suggestive of a disconnected stream, where an unsaturated zone exists between the stream and deeper groundwater. It is noted that on the cross-section in Fig. 4, the groundwater level in the boreholes is interpolated as a piezometric surface rather than a water table. The drillers' logs reported in Fig. 5 and presented in the Supplementary material (Fig. S1) stated that water strikes occurred at greater depths than the rest water level, usually at the locations of fractured layers. The initial water strikes were observed at depths of 16–31 mbgl and rest water levels are 7–18 mbgl.

Weathered and fractured layers commonly occur above the water strikes, within the level of the resting groundwater, although water was not encountered at the time of drilling through these materials. It seems likely that groundwater is present only in fractures and is, therefore, much deeper (approximately 20 m) below the base of the sand river than the observed water levels in the boreholes suggests. The cause of the driving head leading groundwater to rise up the boreholes may be a connection with the NE-SW trending lineaments shown in Fig. 3 that extend to the wetter mountains and recharge zone to the southwest, thus creating a semi-confined aquifer system.

The groundwater level in the sand river varied from a maximum depth of 1.2 mbgl in October 2016 following two years of drought to above surface during occasional flow events. The proximity of the water table to the surface in October 2016 following one of the worst droughts on record (Baudoin et al., 2017) illustrates the potential of the sand river aquifer for productive use. Wipplinger (1958) and Nord (1985) are the commonly cited sources of the maximum evaporation extinction depth for river sands, who showed with laboratory evaporation tests on cylinders of sand (Wipplinger, 1958) and observations in sand rivers in Botswana (Nord, 1985) that the rate of evaporation decreased asymptotically until no further evaporation occurs at a depth of 0.9 m. This depth generally matched multiple observations in the Molototsi sand river at the end of the drought. A depth of 0.9 m is not the maximum depth that the water table would reach because of the influence of drainage, i.e. down-gradient groundwater flow and/or seepage through the base of the sand river, which can reduce the level further. Phreatophytes on the channel banks and occasionally within the channel could also reduce the groundwater level to below the maximum evaporation extinction depth. This was observed in hyperarid environments such as the Kuiseb River in Namibia where the water table drops to depths of 4.5–7 m (Benito et al., 2010). During excavation of 50+ small pits in May 2017 for falling head permeameter testing, water sampling, and for conducting salt dilution tests (locations shown in Fig. 5), water table depth was measured and varied from 0 to 1 m depending on the position within the channel. Low spots where surface water flow last occurred and narrow gaps between in-channel outcrops (Fig. 7) had a water table closer to the surface than beneath protruding sand and gravel bars.

4.2. Aquifer properties

The results from the aquifer properties tests are presented in Table 1. The calculated hydraulic conductivity of 170–290 m/d from the riverbed well pumping tests are provided with uncertainty due to: a) the non-cylindrical shape of the well with its conical base and b) pumping was halted prior to achievement of steady state giving little curve to match during analysis. These factors, evidenced by the discrepancy in hydraulic conductivity estimates between tests in the same well, at least from the Papadopulos and Cooper (1967) method, indicate that these results should be regarded as tenuous at best. Falling head permeameter tests gave vertical

Fig. 7. In-channel outcrops at chainage 1200 m that almost block the sand river.

Table 1
Aquifer properties derived from field investigations. K_h/K_v = anisotropy ratio (horizontal hydraulic conductivity / vertical hydraulic conductivity). HydrogeoSieveXL hydraulic conductivity result reported is the median of the 15 methods.

Analysis method	Test number											
	1	2	3	4	5	6	7	8	9	10	11	12
Pumping tests – hydraulic conductivity (m/d)												
Papadopulos and Cooper (1967)	170[a]	293[a]										
Barker and Herbert (1989)	183[a]	175[a]										
Falling head permeameter – hydraulic conductivity (m/d)												
Chen (2000)	58	318	140	22	125	112	216	49	58	244	91	118
Hvorslev (1951): Isotropic	61	333	147	23	131	118	226	52	61	257	96	124
Hvorslev (1951): K_h/K_v = 5	59	325	143	22	128	115	220	50	59	250	93	121
Grain size analysis – hydraulic conductivity (m/d) and porosity (%)												
HydrogeoSieveXL	28[b]	79	72	74	61							
Water volume test	39%											
Vukovic and Soro (1992)	[b]	40%	39%	39%	38%							
Salt dilution tests – groundwater flow velocity (m/d)												
Davies et al. (1998)	12[a]	12[a]	22[a]	31[a]	14[a]	12[a]						

[a] Uncertain for reasons described in the text.
[b] Uncertain as the finest sieve used was 0.425 mm (other samples analysed down to the < 0.075 mm fraction).

hydraulic conductivity estimates generally varying from 20 to 330 m/d, depending on the grain size distribution at the testing location, with a median of 118 m/d. The grain size analyses (Fig. 8) revealed moderately well sorted gravelly sands low in fines (particles < 0.075 mm). The hydraulic conductivity range given by HydrogeoSieveXL for all methods and all samples (excluding sample 1; see note on Table 1) was 10–260 m/d with a median of 71 m/d. A water volume test was conducted in the laboratory on the sand samples, which indicated porosity of 39%. This result was supported by application of the empirical relationship between coefficient of grain uniformity and porosity presented by Vukovic and Soro (1992) that gave results of 38–40%. The computed groundwater flow velocities from the salt dilution tests were extremely high: 12–31 m/d. With a hydraulic gradient of around 0.002, based on analysis of high resolution (20 m) DEM data, these flow velocities equate to hydraulic conductivities in the 1000s m/d. It is suspected that the disturbance of the sands while inserting the piezometer – some excavation and backfilling was required – created loose and consequently very high hydraulic conductivity materials in the vicinity of the tests. However, the testing was not in vain, as it confirmed that groundwater flow is occurring in the sands, i.e. groundwater is not stagnant between bedrock outcrops and subcrops (shallow bedrock close to the sand surface that does not outcrop). This is an important finding as it means abstraction from a riverbed well should not drain an aquifer compartment as replenishment is occurring. What's more, test sites in narrow gaps between in-channel outcrops (Fig. 7) gave higher (relative) velocities than open channel areas and sites close to riverbanks, which is further

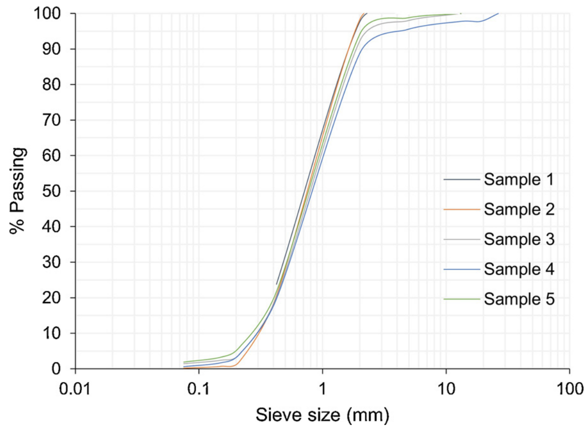

Fig. 8. Grain size distribution curves. Note that sample 1 was analysed with a finest sieve of 0.425 mm.

evidence for the connectedness of aquifer compartments.

Many African sand river studies have measured hydraulic conductivity by various means, and the results from an extensive literature review are presented in Table 2. It can be seen that the results from this study are typical of values for the region. However, the difficulty experienced in obtaining reliable aquifer property data from this study site, particularly regarding the riverbed well pumping tests and salt dilution tests, should be emphasised. Table 2 may also demonstrate that the simple grain size analysis method of hydraulic conductivity estimation performs adequately for this sediment type.

Analysis of the sand profile within trial pits generally showed near surface layers characterised by poorly sorted thinly bedded sediments deposited rapidly by ephemeral floods, the sediment being derived locally from the rapid erosion of soils and regolith by surface runoff from areas denuded of vegetation. Below this zone, there was commonly a zone of well graded coarse sands and gravels, likely deposited by a perennially flowing river.

4.3. Aquifer thickness

Trial pits excavated in the sand river struck bedrock at 1.5–2 mbgl. DCP testing (essentially probing) also occasionally reached the

Table 2
Comparison of hydraulic conductivity measurements of sand rivers from this study with studies from elsewhere in Africa.

Sand river location	Hydraulic conductivity	Testing method	Reference
Molototsi River, Limpopo, South Africa	170–290 m/d[a] 20–330 m/d median:118 m/d 10–260 m/d median: 71 m/d	Riverbed well pumping/recovery test Falling head permeameter HydrogeoSieveXL based on grain size analysis	This study
Musina, Limpopo River, NE South Africa	120 m/d	?	Mulder (1973) cited in Cobbing et al. (2008)
Shashe River, NE Botswana	148-207 m/d (horizontal) 91–216 m/d (vertical)	Salt dilution tests Falling head permeameter	Davies et al. (1998)
Tati River, NE Botswana	~700 m/d	Sand river borehole pumping tests	Herbert et al. (1997)
Nhandugue River, Mozambique	32-639 m/d	Grain size analysis	Arvidsson et al. (2011)
River Gash, E Sudan	36-105 m/d	Pumping tests	Elsheikh et al. (2011)
Shashane, Huwana, Wenlock and Dongamuzi, Zimbabwe	14-181 m/d	Constant head permeameter	Mansell and Hussey (2005)
Mzingwane, Zimbabwe	200 m/d	Grain size analysis	Moyce et al. (2006)
Mzingwane, Zimbabwe	32-46 m/d	Laboratory tests	Love et al. (2011)
Mnyabezi, Zimbabwe	69 m/d	?	de Hamer et al. (2008)
Runde and Sessami Rivers, Zimbabwe	20-60 m/d	Grain size analysis	Owen (1989)
The following studies are all discussed and cited in (Owen, 1989):			
Mahalapshwe River, E Botswana	20-250 m/d median: 80 m/d	Grain size analysis	Thomas and Hyde (1972)
E Botswana	average: 260 m/d	Pumping tests	Nord (1985)
Grootvlei, Zimbabwe	150-264 m/d	Pumping tests	Hineson (1970)
Bauchi state, Nigeria	Crystalline bedrock: 43–620 m/d average: 182 m/d Sedimentary bedrock: 13–640 m/d average: 230 m/d	Pumping tests	Water Surveys Group (1986)

[a] Uncertain for reasons described in the text.

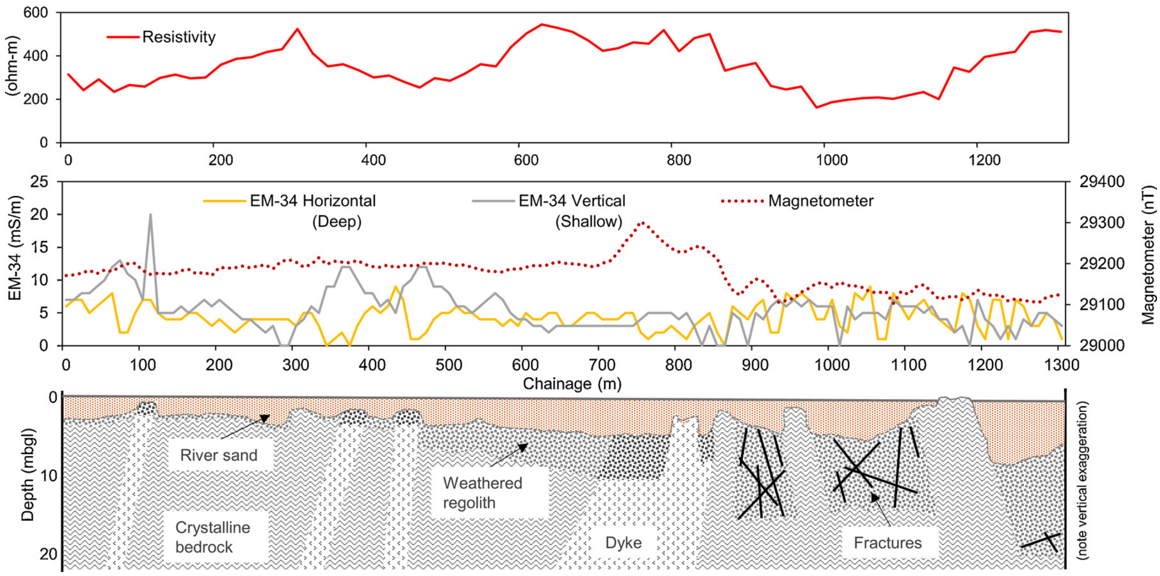

Fig. 9. Plots showing the longitudinal resistivity survey measurements (top),the magnetometer and EM34 electromagnetic readings, and the interpreted geology based on all geophysics (including VES), trial pits, probing and field observations. The orientation is west to east, or C to D, on Fig. 5.

base of the sands at depths of 0.9–2.0 m. Several other locations gave clues as to sand depth, including as-dug blocky regolith material remaining beside river pits excavated for irrigation water or for cattle watering, and weathered regolith material exposed by sand mining, again suggesting a sand depth of around 2 m.

The longitudinal geophysical survey results are shown in Fig. 9. Magnetic anomalies are caused by the anomalous distribution of magnetic minerals. Generally, and applicable to this study, basic rocks have a higher magnetic susceptibility than acid rocks, and greater metamorphism tends to decrease magnetic susceptibility (Clark and Emerson, 1991; Hunt et al., 1995). Therefore, younger dolerite dykes within granitic gneisses may register as positive anomalies. Such a feature is clearly visible on the magnetometer plot at chainage 725–860 m.

Regarding the EM34 survey, electrical conductivity of the ground is controlled by clay content, presence of ferromagnesian minerals, porosity, and the presence and nature of pore fluids (e.g. salinity). The crystalline basement rocks of the study site would have low conductivity due to their low primary porosity, whereas weathered regolith would have higher conductivity due to its higher clay and water content. Spikes in the vertical coil profile above dips in the horizontal coil profile, e.g. at chainage 350 and 460 on Fig. 9, are interpreted as weathered mafic material (relatively high conductivity) at shallow depth above a dyke (relatively low conductivity).

Prior to the resistivity survey, it was suspected that the channel bed and banks were mostly weathered regolith, which should contrast well with the alluvial sands, i.e. higher-resistivity channel within low-resistivity background, or, where unweathered gneiss and dolerite underlie the sand river, lower-resistivity channel within high-resistivity background. The resistivity peak visible in the profile in Fig. 9 at chainage 300 is indicative of high-resistivity bedrock close to the surface.

The apparent resistivity curves from the cross-channel VES surveys are shown in Fig. 10 and the interpreted layers are shown in Fig. S2 of the Supplementary material. Processing with RES1D led to estimates of sand aquifer thickness from 1.5 to 5.5 m, though there is some uncertainty due to the presence of either weathered regolith (with lower resistivity than sand) or solid crystalline bedrock (with higher resistivity than sand), or both, immediately beneath the sand.

Accurate determination of sand aquifer thickness proved to be a difficult procedure. Trial pitting is problematic below the water table due to pit collapse (shuttering and pumping is a possible solution), probing involves uncertainty if the sands overlie weathered regolith, and, similarly, VES provides thickness estimates only where layers are well defined. Unfortunately, the geophysical survey did not provide sand thickness estimations of which we have high confidence. Based on the geophysics, the sand thickness is estimated to be around 0–6 m, although we can only confidently assert a sand thickness of up to around 2 m based on the trial pits and probing. The discrepancy in estimates is due to: a) heterogeneities in the level of the bedrock between testing sites and, b) uncertainty in the geophysical interpretations of the presence and thickness of weathered regolith between the sands and the solid bedrock. What the geophysics did reveal are abundant cross-channel dykes and subcrops that compartmentalise (though do not entirely separate due to discontinuities) the aquifer where they reduce its thickness to a metre or less at subsurface or outcropping barriers, as shown in Fig. 7. Subsurface barriers that divide the alluvial aquifer into units are frequently reported in sand river literature, e.g. Morin et al. (2009), Love et al. (2011), Hussey (2007).

Many African sand river studies have measured aquifer thickness by various means and the results are presented in Table 3 for comparison. This list represents all the studies identified from an extensive literature review, which give the depth of alluvial sands in ephemeral sand rivers. The results from this study are typical of sand river aquifer thicknesses in the region though Table 3 shows that

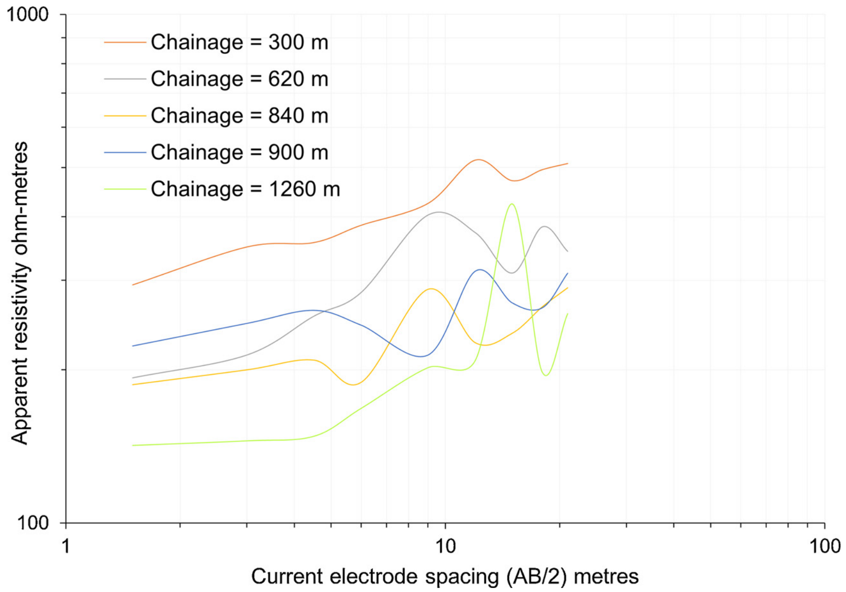

Fig. 10. Vertical electrical sounding (VES) curves.

Table 3
Comparison of sand river aquifer thickness from this study with studies from elsewhere in Africa. GPR = ground penetrating radar.

Sand river location	Sand thickness	Measuring method	Reference
Molototsi River, Limpopo, South Africa	0-6 m (0–2 m)[a]	Trial pits, probing, VES	This study
Shashane, Huwana, Wenlock and Dongamuzi, Zimbabwe	1.48-2.45 m	Probing	Mansell and Hussey (2005)
Umzingwane River, Zimbabwe	5-15 m maximum: 40 m downstream of in-channel outcrops	Continuous VES, GPR	Owen and Dahlin (2005)
Shangani River, Zimbabwe	6-14 m	Continuous VES, GPR, hammer seismics, hand drilling	Dahlin and Owen (1998)
Shashani and Manzamnyama Rivers, Zimbabwe	1.5 m	Probing	Mpala et al. (2016)
Lower Mzingwane River, Zimbabwe	10-20 m	VES	Masvopo (2008)
Kuiseb River, Namibia	average: 10 m	Channel borehole depths	Bate and Walker (1993)
Shashe, Tati, Motloutsi and Ramokgwebana Rivers, NE Botswana	1.8–10 m	Hammer seismics, probing, shaft excavation	Davies et al. (1998)
Tati River, NE Botswana	4-6 m	Hammer seismics, probing, shaft excavation	Herbert et al. (1997)
Nhandugue River, Mozambique	5-10 m	Continuous VES	Arvidsson et al. (2011)
River Gash, E Sudan	8.0–19.5 m	?	Elsheikh et al. (2011)
Mzingwane, Zimbabwe	4-20 m mean: 12.6 m	Resistivity, dam construction drawings, local farmers opinions	Love et al. (2011)
Mnyabezi, Zimbabwe	1.4 m	Probing	de Hamer et al. (2008)
Runde and Sessami Rivers, Zimbabwe	4-8 m	Hammer seismics	Owen (1989)
The following studies are all discussed and cited in (Owen, 1989):			
E Botswana	maximum: 7 m	Probing	Nord (1985)
Grootvlei, Zimbabwe	19-38 m	?	Hineson (1970)
Mahalapye River, Botswana	2.5–12 m average: 6.7 m	Channel borehole drilling	Molosiwa (1987)
The following studies are all discussed and cited in Cobbing et al. (2008):			
Musina, Limpopo River, NE South Africa	mean: 3.5 m	?	Mulder (1973)
Limpopo and Shashe River confluence, NE South Africa	mean: 6 m maximum: 24 m	?	du Toit et al. (2000)
Upper Limpopo River, NE South Africa	10-12 m maximum: 30 m	?	Hobbs and Esterhuyse (1983)

[a] We can confidently assert thickness of 0–2 m based on trial pitting, probing and observations though the geophysical survey suggested (with uncertainty) sand thickness up to approximately 6 m.

Table 4

Groundwater quality results. Turb. = turbidity, EC = electrical conductivity, TDS = total dissolved solids, Alk. = alkalinity, nt = not tested, nd = not detected.

Location	pH	Turb. (%)	EC (mS/m)	All mg/l													SAR
				TDS (mg/l)	DOC	Alk. as $CaCO_3$	Ca	Mg	Na	K	NH_4	F	Cl	N as NO_2 + NO_3	P as PO^4	SO_4	
February 2012																	
Deep borehole	6.9	0.204	284	nt	nt	nt	147	131	170	12	nt	nd	633	14	nd	38	2.4
Deep borehole	7.0	0.105	128	nt	nt	nt	66	76	45	4	nt	nd	87	32	nd	44	0.9
Sand river	7.8	0.060	21	nt	nt	nt	11	5	16	9	nt	nd	16	nd	nd	3	1.0
October 2015																	
Deep borehole	7.4	nt	108	691	2.3	539	47	32	151	5	0.1	0.9	37	1.1	0.1	14	4.2
Deep borehole	7.4	nt	215	1376	3.3	490	89	103	159	3	9	0.8	391	0.1	0.5	16	2.7
Deep borehole	7.3	nt	240	1536	2.9	607	91	99	241	12	0.2	1	387	6.2	0.1	39	4.2
Sand river	7.5	nt	41	262	2.9	127	34	11	29	4	0.6	0.3	29	0.1	0.1	35	1.1
Sand river	8.2	nt	41	262	3.1	129	30	11	35	4	0.1	0.2	44	0.2	0.1	12	1.4
January 2017																	
Deep borehole	7.3	nt	112	717	2.3	521	50	57	101	2	4.6	0.7	58	0.1	0.3	7	2.3
Deep borehole	8.5	nt	68	435	2.4	356	10	7	145	1	0.3	0.8	11	0.2	3.7	11	8.6
Deep borehole	7.7	nt	130	832	0.9	303	57	46	147	8	0.1	1	235	0.1	0.1	30	3.5
Sand river	7.9	nt	17	109	3.9	61	13	5	13	3	0.1	0.2	9	0.3	0.1	4	0.8
Surface water	7.3	nt	11	70	4.2	51	11	3	5	4	0.6	0.1	3	0.1	0.1	2	0.3

sand river aquifers can be substantially thicker where the rivers are wider and more mature, e.g. Mzingwane River in Zimbabwe and Shashe River in Botswana.

4.4. Hydrochemistry

Apparent from the groundwater quality analyses shown in Table 4 (and the Piper plot in Fig. S3 of the Supplementary material), is the difference in deep and shallow groundwater hydrochemistry. Samples from deep boreholes show much higher ionic concentrations (higher salinity), which are characteristic of long water-rock interaction time, whereas the shallow groundwater samples from the sand river have hydrochemical signatures closer to surface water. What is more, nitrate and phosphate levels are generally lower in the sand river samples meaning that current farm activities, through fertilizer and pesticide applications, have little effect on the shallow groundwater quality. The evidence suggests that at the study site there is little to no connectivity between deep groundwater and the sand river, nor between cultivated fields and the sand river. Deep groundwater contributing to baseflow and irrigation water flowing through the surface or subsurface to the sand river could influence the alluvial aquifer hydrochemistry, but that is not apparent. The quality of shallow groundwater sampled in the sand river is very good and the water is fit for human consumption based on the inorganic and dissolved organic carbon (DOC) analyses, unlike the groundwater samples collected from deep boreholes (WHO, 2011). The sand river groundwater also has better sodium adsorption ratios (SAR) whereas the deep groundwater is occasionally at concentrations (SAR > 3.0) that may have adverse effects to crops under irrigation (Olson, 2012).

Fig. 11 shows the plot of stable isotopes δ18O and δ2H in ‰ and the local meteoric water lines (LMWL) for Pretoria and Harare (the study site lies between the two). The main cluster containing the majority of samples are from the sand river. The two results that plot above the LMWLs showing condensation represent a rainfall sample and a sample taken from a rare surface water flow. The similarity in these results indicated that the surface water was derived from recent rain (there had been little mixing with soil or groundwater), as was observed when the river responded quite rapidly to a thunderstorm. The deep borehole sample has a clearly different signature to the sand river samples, showing the greatest depletion. The sand river sample that plots closest to the deep groundwater sample was collected from the centre of the channel. This result may be suggestive of some localised connectivity. However, such high depletion was not measured elsewhere; this sample being the only anomaly from the studied reach. Note that all shallow groundwater samples were collected from the top 100 mm of the groundwater reservoir following more than two weeks with zero rainfall recorded anywhere within the catchment. The result furthest right on Fig. 11 was sampled from a low point in the sand river with stagnant surface water; consequently, it shows the highest evaporation effect. Sand river samples were deliberately collected next to riverbanks and in the channel centre with the latter samples showing a slightly different isotopic signature: Mid-channel samples show greater oxygen-18 depletion whereas the tight cluster (Cluster A on Fig. 11) showing lower oxygen-18 depletion includes the three samples taken immediately adjacent to high regolith banks. Therefore, there may be some groundwater entering the channel through the banks, from bank storage or interflow. Several pairs of samples were collected from immediately upstream and downstream of in-channel outcrops to assess if sand river groundwater was originating from fractures within the outcrops (pairs labelled B, C and D on Fig. 11). It can be seen that there is little difference between results for each pair. The stable isotope results show that deep and shallow groundwater, and surface water, all have different isotopic signatures suggesting there is no leakage to the deeper aquifer, though this is based on just one deep groundwater sample.

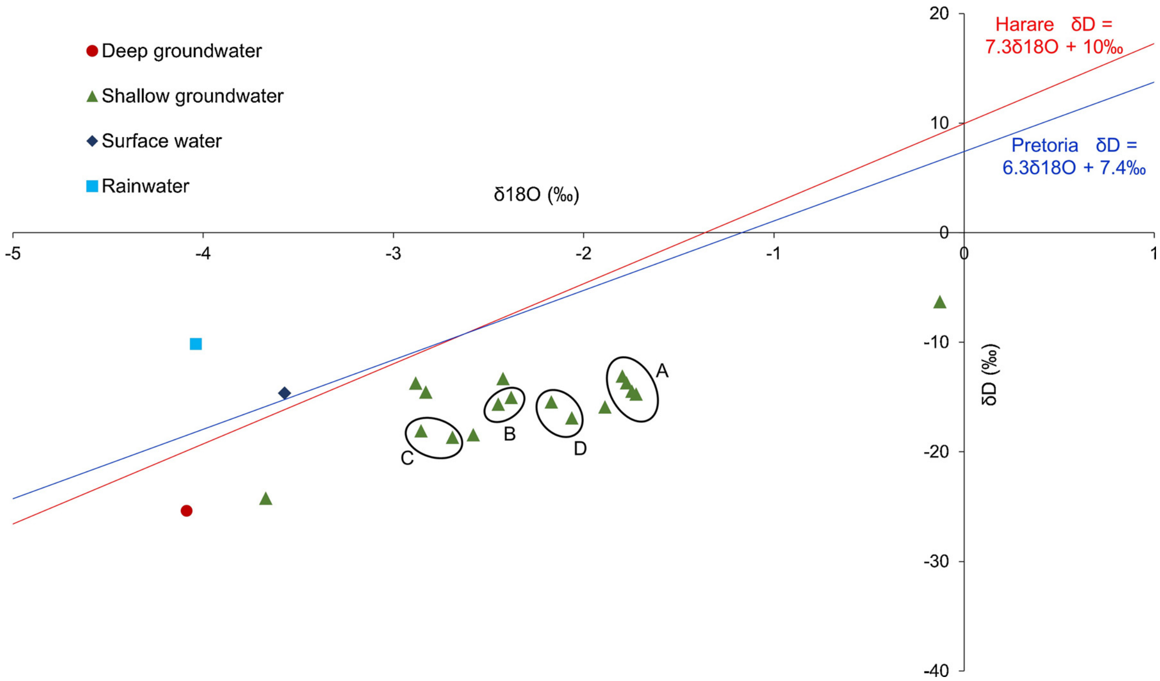

Fig. 11. Plot of stable isotopes ($\delta 18O$ and δD ($\delta 2H$) in ‰) analyses with the local meteoric water lines (LMWL) for Pretoria and Harare. Labels are referred to in the text.

5. Alluvial aquifer recharge

The Molototsi flowed at the onset of rains in December 2016. Nord (1985) and Owen (1989) stated that surface water flow only occurs in sand rivers when the alluvial aquifer is fully saturated. Therefore, following two years of one of the worst droughts on record (Baudoin et al., 2017), the alluvial aquifer was still able to completely refill at the onset of rains to enable surface flow. There was no legacy of the drought, i.e. excessive depletion, preventing complete alluvial aquifer groundwater replenishment and surface water flow. Observations and DWS data showed that the Modjadji Dam was well below capacity in December 2016 and did not release water; dam releases are via a spillway thus occur when the reservoir is at 100% capacity. This indicated that Molototsi surface water flow and alluvial aquifer recharge did not originate from dam releases.

A field visit in May 2017 coincided with a surface water flow event. Tributaries in the lowveld portion of the catchment remained dry while contributing flow was only observed in streams from the mountains, of which only a few are downstream of the dam catchment. Notably, the flow receded within hours and a few days later only minor flow remained. The lowveld tributaries flow infrequently because the sandy soils, typically water-starved vegetation, and deep water tables, mean that flow in these torrential streams is likely to occur only when rainfall is sufficiently intense to exceed the infiltration capacity of the ground, leading to Hortonian overland flow into the incised gullies. Conversely, high intensity rainfall in the mountains caused the tributaries there to flow where steep slopes and thin soils above solid bedrock encourage runoff. Anecdotal, observational and remote sensing analysis indicate the Molototsi River flows and the alluvial aquifer fully recharges a few times per year (Walker et al., 2018).

6. Conceptual model

The information so far described was used to develop a conceptual model of the Molototsi River catchment. Data collected suggested that there is a hydraulic disconnect between the regional groundwater and the alluvial aquifer: Groundwater level monitoring in boreholes located barely 100 m from the Molototsi indicated that the regional groundwater level is below the water table in the sand river. In addition, borehole groundwater levels responded very little to rainfall. Perhaps the clearest evidence of this disconnect was given by the difference between the regional groundwater hydrochemistry and that of the alluvial aquifer. Throughout the measuring period, the hydrochemistry of the sand river resembled the quality of rainwater, whilst higher salinity and ionic concentrations were measured in the regional groundwater.

Stable isotope analyses also resulted in clear differentiation of water samples, although they gave (possibly) some indication of lateral inflow from bank storage. The main pathway of sand river recharge is therefore surface runoff, which occurs predominantly through erosion gullies and torrential tributaries high in the catchment as observed in May 2017. The geophysical surveys revealed the presence of abundant dykes and other subcrops that compartmentalise the alluvial aquifer. However, the lack of decline of the sand river water table to > 1 mbgl in October 2016, even following two years of drought, indicated that these subsurface structures are not barriers to flow. Salt dilution tests confirmed that the narrow passages between outcrops forming aquifer compartments are sites of greater groundwater flow velocity.

An uncertainty in the conceptual model is the regolith underlying the alluvial sands; its presence, thickness, and properties. Observations and geophysics suggest that the sands directly overlie crystalline bedrock in some areas and regolith in others. The perched nature of the sand river aquifer indicate that the regolith must be of very low hydraulic conductivity. Therefore, the base of the aquifer is considered to be the base of the saturated sands.

7. Stored water volume available for productive use

We will consider the study site sand river reach beginning at chainage 0 m and ending at the mid-channel outcrop at chainage 1200 m (visible in Figs. 5 and 7). The width of the sand channel (there is no floodplain), based on Google earth measurements every 100 m, ranges from 46 to 75 m with an average of 60 m. The aquifer thickness measurements ranged from 0-6 m; we can take a conservative average of 2 m because we have greatest confidence in the trial pitting and probing methods to obtain sand depth.

Laboratory porosity analysis of the sands gave a value of 39% and this is conservatively converted to a specific yield of 30% based on porosity/specific yield relationships at various grain sizes plotted by Robson (1993). Multiplying this length, average width, average thickness and specific yield gives the total drainable water volume of the aquifer. However, following surface flow and aquifer recharge, evaporation from the sands reduces the water table to an observed depth of around 1 mbgl. Subtracting 1 m from the aquifer thickness produces a stored volume of about 21,650 m^3.

The purpose of this calculation is to highlight the potential for abstraction for small-scale irrigation, livestock, and water supply, etc. The only farm abstracting groundwater from this reach monitored abstraction with a flow meter from September 2016 onwards, during the drought. The average abstraction from the riverbed well was 7 m^3/d to irrigate approximately 0.2 ha of tomatoes with drip irrigation. Even if abstraction were to occur every day of the year, the total abstracted volume would be 2555 m^3, less than 12% of the stored volume. What's more, this simple but conservative calculation neglects to consider sand river groundwater flow from upstream and occasional surface water flow recharging the aquifer and increasing saturated thickness, which would increase the volume of the shallow groundwater resource. Additionally, neglected are wet periods, days off, or other downtime when no irrigation occurs, all of which would reduce the abstracted volume.

Taking this simple calculation a step further: if we consider the aquifer geometry of the reach to be typical, scale up the irrigated area, assume the same irrigation rate, and irrigate for 9-months of the year; this equates to the potential for two 1 ha farms per kilometre of sand river. Within a 10 km reach centred on the study site, only six farms currently irrigate from the sand river revealing the potential for a three-fold increase in number of farms and a much greater increase in irrigated area.

8. Conclusions

This study aimed to contribute findings from field investigations and observations of a sand river currently utilised for small-scale irrigation to the limited pool of data available in published literature. Field investigations meant the aquifer could be characterised in terms of its properties, geometry and relationship with surface and groundwater. It is important to reiterate the difficulty in obtaining results in which we had a high degree of confidence, therefore, applying multiple methods is recommended. Pumping tests in a riverbed well, falling head permeameter testing, and grain size analysis indicated hydraulic conductivity of 20–300 m/d, depending on how silty or gravelly were the sands at the precise testing location. Porosity was measured at 38–40%. Aquifer thickness was estimated with trial pits, probing, observations and geophysics (magnetometer, EM34, resistivity and VES) and gave thickness varying from zero at in-channel outcrops to about 6 m. Dykes and other subcrops that compartmentalise the aquifer are abundant, though there remain channels through which groundwater can flow (confirmed by salt dilution testing). There is a hydraulic disconnect between the deep groundwater and the sand river, revealed by borehole groundwater levels many metres below the sand river water table and showing no response to rainfall. Contrasting hydrochemistry and stable isotope signatures are further evidence for the deep and shallow groundwater disconnect. The dominant recharge process of the alluvial aquifer is surface runoff, which occurs predominantly through erosion gullies and torrential tributaries in the upper reaches of the catchment.

Even when groundwater inflow from upstream and surface water flows recharging the aquifer are neglected, the Molototsi alluvial aquifer resource is plentiful and abstraction is sustainable when considering the current demand. Opportunities exist to increase abstraction of sand river groundwater at the study site reach more than eight-fold (based on the current 12% usage of stored volume) for smallholder irrigation and other uses. A resource assessment of the Lower Mzingwane sand river in Zimbabwe by Love et al. (2011) indicated that water usage there could be tripled. Similarly, the transboundary Limpopo alluvial aquifer is abstracted at only around 20% of its potential (Cobbing et al., 2008), the Umzingwane sand river in Zimbabwe is described as "highly underutilised" (Owen and Dahlin, 2005), and sand rivers in Botswana are considered "major sources of under-utilised groundwater" (Herbert et al., 1997). Further exploitation of these alluvial aquifers could aid in halting or reversing the declining trend in subsistence and smallholder agriculture in the region boosting food security and nutrition, and alleviating poverty. However, additional factors must be considered, such as the availability of arable land and of irrigation infrastructure.

All of the methodologies applied in this study are transferrable to sand rivers elsewhere.

To conduct a resource assessment, knowledge of alluvial aquifer thickness is important, though Table 3 shows that sufficient thickness for a substantial stored water volume is generally found. More important is to estimate the degree of compartmentalisation of the alluvial aquifer. If outcrops or subcrops create barriers to flow (e.g. Morin et al. (2009); Love et al. (2011); Hussey (2007)), there is the possibility of quickly depleting a small alluvial aquifer unit through abstraction. In such cases, monitoring of regional groundwater levels would indicate whether lateral inflows (i.e. gaining conditions) could sustain alluvial aquifer groundwater levels.

Such monitoring would also suggest whether the sand river is losing or disconnected; hydrochemical investigations would aid in this determination. Investigation of aquifer properties indicates the suitability for abstraction; Table 2 shows that high hydraulic conductivities are typical in sand rivers and that simple grain size analysis is probably sufficient for hydraulic conductivity estimation. To assess the sustainability of abstraction, recharge frequency must be estimated. This is a more straightforward assessment than often in hydrogeological studies because a single river flow equates to complete alluvial aquifer recharge (Owen and Dahlin, 2005). However, as rainfall in dryland regions is highly spatially variable and tributary contributions to flow are consequently highly spatially variable (Hughes and Sami, 1992), traditional means of estimating flow frequency through rainfall or flow monitoring may be unsuitable unless a high density monitoring network is available. Remote sensing proves to be a more appropriate flow detection technique (Walker et al., 2018). Future work would assess the varying abstraction systems observed in sand rivers for efficiency and yield, as well as modelling of increased abstraction to assess the impact on the shallow groundwater resource.

Declarations of interest

None.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at https://doi.org/10.1016/j.ejrh.2018.09.002.

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