Modelling the impact of increased pumping on the storage of the Shashane alluvial aquifer in southern Zimbabwe

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Abstract

In the perspective of finding alternative water resources for rural communities to support current needs and future increased crop production, a portion of the Shashane River in the southern part of Zimbabwe is considered as a case study to analyse the potential in sustaining such development schemes.

A 3D transient numerical groundwater model over a stretch of 5.6km was built in order to understand the groundwater flow and the natural variation of the storage. In addition, the model was used to estimate the recharge of the system as well as assess the impact of intensified abstractions on the storage and the different hydrologic components of the aquifer.

The aquifer has an available storage capacity of 0.35 Mm3 per the studied stretch of 5.6km. Such capacity is depleted naturally to its half due to evaporation. Less water is evaporated when abstractions are increased. The system has the potential to fulfil a total demand of 800 m3/day without failure which can irrigate a potential area of 3.5 ha per km. Despite the increased abstractions, the recharge in the following wet season is almost certainly ensured by surface flow except for some dry years that occur once every 20 years. The increased abstractions lead to an increase in recharge reduce runoff in the following year to a certain extent and thus affect other users downstream, including dependent ecosystems, which requires further studies.

The aquifer studied can support substantial socio-economic development plans as it was shown that if coupled with good management practices, it can lead to considerable livelihood improvement. The use of the model can be extended to test different development alternatives and serves as a tool to justify future development decisions.

Key words: Sand rivers, modelling, MODFLOW, transient, alluvial aquifer

1. Introduction

Water is a resource that is mistakenly considered to be both renewable and infinite putting it under the threat of excessive and unsustainable use. Its uneven availability creates zones of high and scarce abundance. While wet areas rigorously thrive in the abundance of this resource in the advancement of a healthy quality of life, dry areas struggle with the management of this scarce resource to promote development and socio-economic prosperity.

The situation in rural sub-Saharan Africa is more critical. The availability of water resources ready for clean drinking water supply to the entire population of Africa is limited. Water scarcity decreases the potential for development in semi-arid to arid regions in the sub-Saharan Africa. Particularly in rural areas, communities lack access to water during the dry season which prevents any perspectives of agricultural development. The Shashane catchment in Southern Zimbabwe is an illustrative example of such a situation. The catchment is characterized by ephemeral river systems overlain by erratic rainfall patterns that make the water availability unreliable resulting in a low agricultural productivity. Groundwater is usually the alternative source, but the rocky geology of the region makes extractions of this resource expensive and the quality not always suitable.

The sand river systems are characterized by perennial groundwater flow that constitute a potential and reliable water resource during the dry season. The Shashane sandy river bed in particular contains significant amounts of water. However, there is a limited understanding of the groundwater dynamics, the volumes of water stored and the sustainable abstraction rates. Hence, it is uncertain how much water available to intensify crop production practices along the river. Therefore, the main objective of the study is to assess the impact of increased pumping on the storage of the aquifer and the potential area that the system can irrigate.

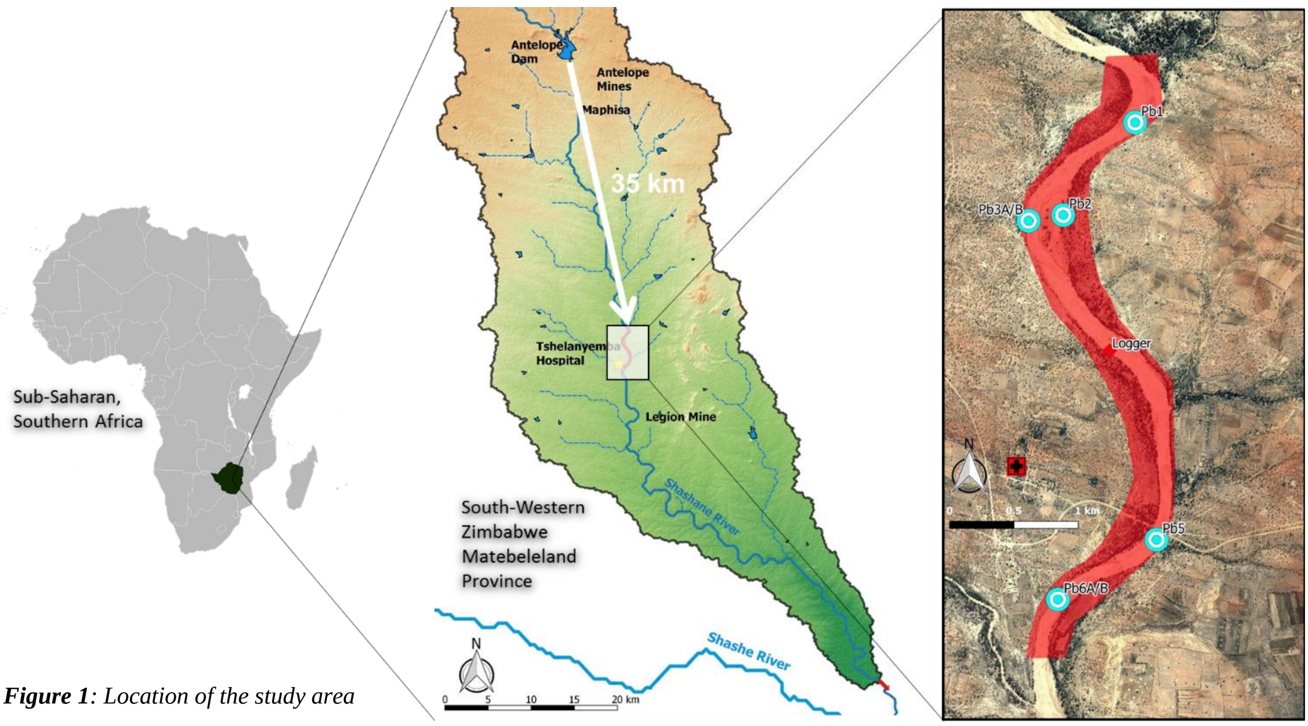

Figure 1*: Location of the study area*

2. Study Area

The area considered for this study is located in the Matabeleland South Province in the southwestern part of Zimbabwe. It corresponds to the catchment of the Shashane river which flows south as a major tributary to the Shashe river, all part of the Limpopo catchment. The Shashane catchment drain to a 200 km long river that flow north to south-east with a meandering course near the flattening reach in the south. The elevation in the catchment varies from 1400 m in the upstream reach to 750 m in the downstream reach. The upstream is characterized by visible mountain ranges and incised deep gorges through the hills (**Figure 1a**). The river uniformly slopes downstream to reach the broad flat-bottomed valleys (Ashton et al. 2001) with slopes between 1:300 and 1:500.

The study area is defined by a dry subtropical climate with precipitation value averaging 450 mm/year making it one of the driest areas in the region (Mansell and Hussey 2005). The area experiences one rainy season per year extending from November to March (Mpala et al. 2016). The potential evapotranspiration attains values of 2000 mm/year as described by the FAO's drought assessment report of the Limpopo river basin

In terms of geology, the bedrock of the region is formed by granite gneiss rock with the presence of localized highly elevated greenstone belt formations (Ashton et al. 2001). The deep aquifer have a very low yields due to secondary porosity. Fluvial deposits on the other hand, consisting primarily of gravels and sands cover the river beds and feature cheap and easy access to good quality water (Chinoda et al. 2009). The soils are shallow and poorly developed ranging from 30 cm to 150 cm, leading to a fast infiltration of the rainfall from the surface to the underlying bedrock. A very important groundwater resources is found in the alluvial deposits. For instance, yields in these formations range between 40 to 5200 m^3/day in the neighbouring Mzingwane catchment (Moyce et al. 2006). These groundwater resources are expected to support large scale irrigation (ZINWA, 2008). The research was focused on a study area located at 35 km downstream of the Antelope dam and continues for 5.6 km. The research site is located nearby the Tshelanyemba village (Figure 1b). The main consumptions come from the nutrition gardens which were designed by Dabane trust, the local hospital and the high school. Surveys conducted in 2017 revealed that the total consumption of the nutrition gardens amount to 1m3/week which is negligible in comparison to the storage capacity of the aquifer. The riparian vegetation covers the river banks and is formed mainly from trees and isolated shrubs as well as small scale gardens which constitute the sole and natural consumer of the water of the riverbed since the riverbed and the banks are connected through a layer.

3. Methods

A groundwater numerical model was built to understand the dynamics of the aquifer system. The main objective of modelling a limited stretch of the aquifer is to assess its capacity to sustain development schemes in the neighbouring region (**Figure 2**). The developed model is transient and aims to provide the following information:

- Estimation of the recharge in all scenarios
- Assess the interaction of surface and groundwater
- Simulate local water management scenarios for agricultural development
- Determine the best spatial locations of abstraction wells
- Optimize the abstraction rates to sustain agricultural development
- Determine the potential area to be irrigated by the aquifer.

The modelling was accomplished using Aquaveo's GMS as a commercial graphical user interface to USGS MODFLOW.

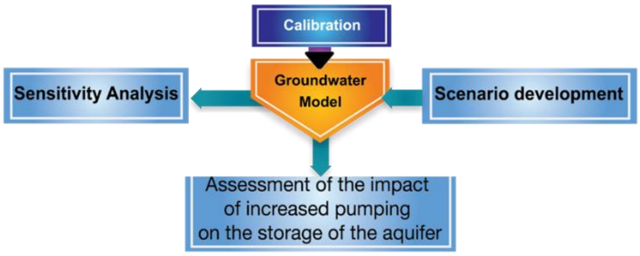

Figure 2: *Framework of the research*

- **Model discretization**

Spatial discretization parameters are summarized in **Table 1**. The vertical dimension was added using one layer of varying bottom level. This was done assuming that the hydraulic conductivity of the clays was too low and no flow conditions were assumed.

Table 1: *Discretization parameters*

Model grid specifications		Model time-discretization parameters	
South-west corner	(0,0)	Stress period length	1 day
No. of layers	1	No. of stress periods	Varying depending on the type of simulation. Min = 365 days; Max = 6000 days
No. of rows	265		
No. of columns	70		
Column width	20 m		
Row width	20 m		
No. of active cells	2565	No. of time steps	1 per stress period
Total cells (active + inactive)	18550		

The model was built in transient mode. The recharge mechanism of the sand river systems as well as the decline in water levels during the dry season is fast and therefore a stress period of one day was considered.

Top level elevation of the riverbed surface were derived combining DEM data with a total station survey. Elevations were considered constant within a cross section profile. Bottom elevations were estimated by delineating the extent of the main channel using cross-sectional and longitudinal probing profiles. This allows the determination of the location of the deepest level of the aquifer. The form of the cross section was determined using the probing and geophysics. Shallow parts do not exceed 1 meter whereas the deepest parts reach up to 5 meters.

The simulation starts at the very beginning of the dry season of 08/04/1986 and the longest period of simulation is 16 years.

At the end of the last runoff event and after recession of the surface water the river bed will be fully saturated and therefore initial heads were assumed to be equal to the top surface elevation.

- **Aquifer parameters**

The material forming the alluvial aquifer was assumed to be uniform and isotropic of fine to coarse sand with expected hydraulic conductivity values varying from 80 to 140 m/day. On the other hand, storage coefficients are assumed to be uniform across the model area. As the aquifer is unconfined, only specific yield needs to be specified with a field measured value of 15%. These values were used as initial estimates for the model which eventually were varied during the calibration for the best fit obtained between measured and estimated values.

- **Boundary conditions**

Model boundaries on the river banks were defined as "no-flow" boundaries since lateral discharge is assumed to be insignificant. Boundaries in the northern (upstream) and southern (downstream) extents of the aquifer were designated as a constant head boundary (CHD) in which head varied with time using a time dependent relationship and the discharge data which define the water level in the aquifer during the dry season. The water level at the boundaries never dries up providing infinite source of out- and in-flows. Imposing time-variable constant head boundary leads to forcing the system to behave accordingly and the inflows and outflows in and from the aquifer are not natural (**Figure 3**).

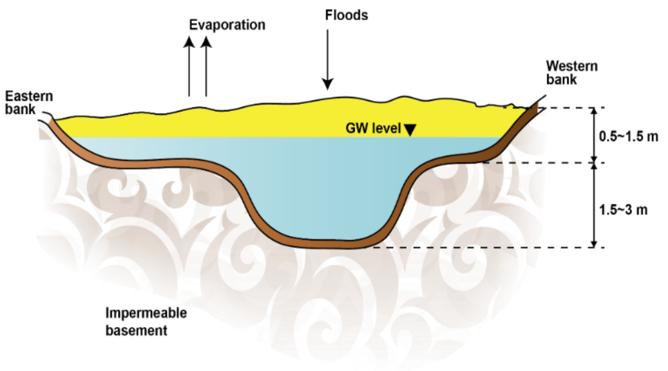

Figure 3: *Conceptual model of the riverbed*

- **Recharge**

In this study, recharge is assumed to be a function of flood events only as direct rainfall recharge is insignificant in comparison to the empty storage of the aquifer. As no runoff is observed until the aquifer is re-saturated entirely (**Figure 4**), runoff occurrence was used as a signal to full saturation. To ensure full saturation, recharge input was given in excess and drains on the surface of the riverbed make sure that at the end of the runoff event the head above the aquifer is at the topographical surface elevation.

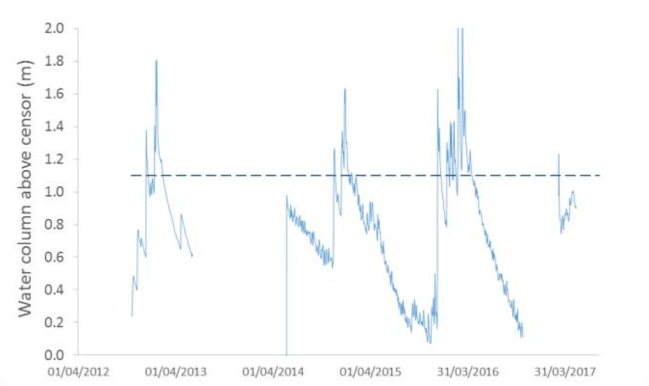

Figure 4: *GW Logger data*

- **Evaporation**

Evaporation is simulated using the EVT package. The evaporation is implemented by specifying a maximum rate of evapotranspiration (ET) and the extinction depth below which zero evapotranspiration occurs. Between the two depths, evapotranspiration varies linearly. Evapotranspiration was chosen between 0.0055 and 0.0025 mm/day which correspond respectively to an average potential evapotranspiration and actual evapotranspiration (FAO). The extinction depth used had 1 meter value for sand materials (Hellwig, 1973).

- **Well package**

Table 2: *Pumping scenarios and strategies*

	Spatial Distribution				
	Strategy 1	Strategy 2	Strategy 3	Strategy 4	Total demand for each scenario (m3/day)
# of wells	11	15	21	30	
Spacing between wells (m)	500	350	250	150	
Scenario A: QA (m3/day/well) =	30	22.0	15.7	11.0	330
Scenario B: QB (m3/day/well) =	75	55.0	39.3	27.5	825
Scenario C: QC (m3/day/well) =	172.8	126.7	90.5	63.4	1900.8

The well package is designed to account for the water losses from the cells due to abstractions. For each cell only one value of the net discharge can be specified for each stress period.

Abstraction rates depend on the scenario to be simulated. On the other hand, a simulated pumping well is considered to be screened through the full saturated thickness of the cell. No refinement of the cells around the wells was performed which means that simulations performed will be independent of the abstraction method.

- **Evaluation of the model**

Calibration

Calibration is only necessary during the dry season since the model assumes full saturation of the riverbed during the wet days. Calibration is performed over the dry season of the year 1986 which is representative of an average dry season with 263 dry days and assumes that the water levels drop in a similar way every year depending on the duration of the dry year only. Calibration was done by modifying the evapotranspiration maximum rate, extinction depth of evaporation and hydraulic conductivity. As the model extent is small and the grid is refined, low discrepancies are expected between calculated and measured heads.

Sensitivity analysis

The task was performed for the major input parameters defining the model and the changes on the output variable was assessed. The outputs could be displayed either by showing the variation of the head for the year of 1986 or in terms of the storage at the last day of the dry season as a percentage of the maximum available storage capacity.

The values of the parameters and the ranges of variations are given in **Table 2**. Ranges of the values of all the parameters were obtained from literature.

Table 3: *Sensitivity parameters*

Parameter	Value	Max %	Min %
Hydraulic Conductivity (m/day)	120	+100%	-50%
Specific Yield (%)	0.15	+66%	-50%
Max ET rate (m/day)	0.0025	+50%	-50%
Extinction depth (m)	1.3	+25%	-50%

Abstraction Potential

The scenarios developed in this section have the aim to assess the ability of the Shashane sand river aquifer to sustain agricultural development in the future. The assessment focuses on available water volumes as well as the locations of pumping and the effect of intensive abstractions on the storage of the aquifer.

The scenarios are built upon the baseline scenario (Scenario 0) which simulates the behaviour of the system without abstractions. A total of 4 strategies (1, 2, 3 and 4) were developed to assess the effect of the spatial distribution of abstraction wells on the yield. For each spatial strategy, increased abstraction rates are also simulated (A, B and C). The total demand is maintained constant between the pumping strategies (1, 2, 3&4). The wells were carefully positioned along the deep part of the subsurface (5 m) to avoid drying up due to shallow channel. Finally, Simulations are run on an average dry season of 263 days. The scenarios are summarized in **Table 3.**

- **Optimizing pumping rate**

For Strategies 3 and 4, the abstraction rate was varied to determine the maximum abstraction rate per well (hence total demand) which the system is able to sustain without any failure, for the entire dry season. With this, the potential irrigated area can be estimated.

- **Estimating Recharge**

Recharge was artificially calculated using MODFLOW's flow budget, as the volume necessary to re-saturate the aquifer, therefore it is not simulated as a natural process. For Scenario 0, the recharge was computed for the entire observation period.

4. Results and Discussion

- **Calibration**

The parameters used for final calibration are ET, extinction depth and hydraulic conductivity. ET rate was to 2.5 mm/day while the extinction depth was increased to 1.3 m to compare with the maximum drop observed in the Logger data. The hydraulic conductivity value that gave the best fit was 120 m/day which corresponds to the measured hydraulic conductivity values in the field. **Figure 5**, indicate that the calibration of the model is very satisfactory with heads above 80 cm of depth in the aquifer are simulated accurately. Heads below 80 cm are higher in the model simulations than in the observation. This is a direct result to the evaporation which is the dominating outflow component in the model. Its effect becomes irrelevant around the extinction depth. For the simulated period, the difference between simulated and observed lower heads is less

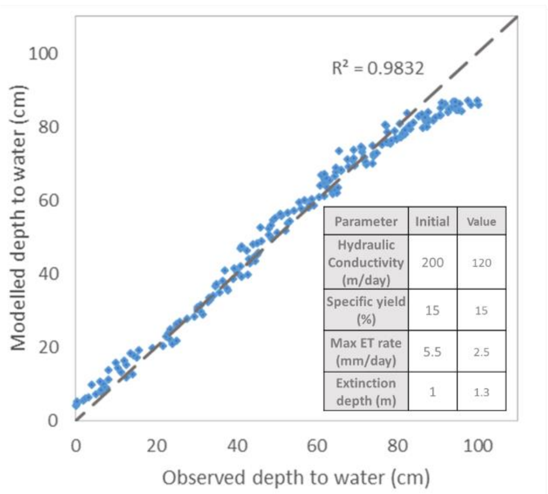

Figure 6: *Calibration results*

than the tolerated error and the impact would only be significant when the duration of the dry seasons are above average. ET rate and extinction depth control the outflows rate whereas hydraulic conductivity controls the groundwater flow rate. Lower hydraulic conductivity values would lead to slower flow of groundwater and therefore more exposure to the effect of Evaporation which leads to a lower drop of GW levels. Such results of calibration are expected for a model of such scale and with such boundary conditions.

- **Sensitivity Analysis**

Figure 6 illustrates a summary of the sensitivity of the storage of the system at the end of the dry season (D263) to a variation in the input parameter. In general, a decrease in the value of any the input parameter leads to an increase in the storage at the end of the dry season, whereas the opposite might not be true.

To start with, the model is insensitive to variations in the

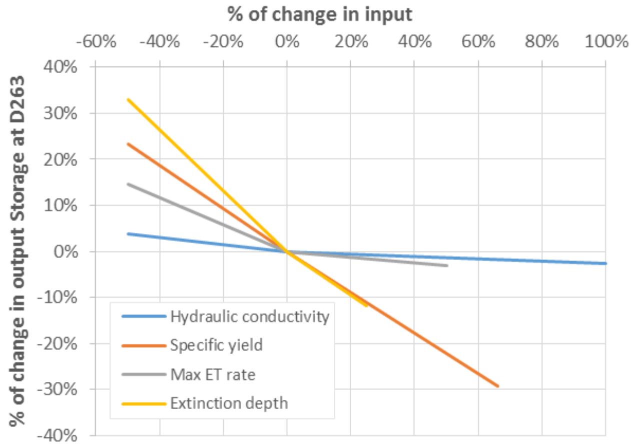

Figure 5: *Sensitivity of the model*

values of conductance of the drains installed to simulate surface water flow. Drains define a head dependent flux boundary conditions, and are conditioned both by the values of the conductance and the head of the water on top. As long as conductance is not too low (representative of silts or clay), the drained flux only depends on the head above the drain which is caused by excess recharge. The higher the recharge is given, the slower it takes the drains to eliminate the excess water out of the model. Nevertheless, to make sure that the effect of the drains does not influence the results, excessively high values of conductance were given. This parameter is purely a modelling hack and does not mimic any natural parameter that could be measured.

On the other hand, doubling the horizontal hydraulic conductivity results only in a 3% decrease in the total storage of the aquifer. The hydraulic conductivity is a parameter that conditions the ability of flow of the water in the porous material. The storage decrease indicates that there is more water lost. This could be explained by the ability of a high conductive material to replenish the system faster and therefore provides more water is available for evaporation loss. Although, hydraulic conductivities of 240 m/day were not observed in the Shashane River therefore lower increase will lead to an even insignificant decrease in the storage. Such a result indicates that any errors associated with the choice of the hydraulic conductivity will not significantly impact the output of the model.

A reduction in the maximum ET rate and the extinction depth lead to a significant increase in storage, hence less water is evaporated. The extinction depth effect on storage is more pronounced that ET. In addition, while a deeper extinction depth depletes the storage more, a higher ET rate surprisingly does not significantly affect the storage. This is a direct result to the choice of the model representative output parameter. A higher ET rate will lead to a faster drop in the water table, however, the final level will be similar as it is already conditioned by extinction depth. Potential evapotranspiration was used as the ET rate which is already an over-estimation of the actual ET calculated in the study area by ETo, therefore the uncertainty in the output caused by ET is small. However, there significant uncertainty associated with the extinction depth as these values are withdrawn from literature of other sediments of similar nature.

Finally, the model output presents a high sensitivity to the specific yield. Higher specific yield values result in a decrease in the storage observed at the last day of the dry season and vice versa. An increase in the specific yield means that there is more water available to flow in or outside the cell when the water balance for each cell is solved and therefore any increase in the specific yield allows more water to flow out of the model. There is a lot of uncertainty associated with determining the specific yield for the sand river aquifers. While the best method to obtain the values is through field tests, these might amount to be expensive. In this research only values of specific yield that were measured were considered, although the measurement methods were unreliable and need to be improved.

- **Storage variation**

After assessing and calibrating the model, the model was run on a period of 16 years. The water budget was calculated from the output of the model. The available storage variation was computed and presented in **Figure 7** as available storage percentage. The storage variation occurs in a series of wet and dry seasons varying in length. The storage drops gradually to reach on average 45 to 50% of the maximum storage capacity at the last day of the dry season. This is observed for instance in 1985, 1990 and 1993. Years 1991-1992 show a more intense storage drop where storage at the end of the dry season drops to reach 40% of storage. These periods have a drought duration typically above average which could in some cases last for two years in a row. Moreover, the spatial irregularities in the bedrock, create local ponds at which point evaporation losses are higher. After 300 days storage capacity will not vary much even if the entire following year is dry as observed. This is a direct consequence of the assumptions made the outflow components. As the evaporation is the main outflow component of the model, the effect of evaporation ceases when the water level drops below the extinction depth. As a consequence, the storage decline stops after the evaporation effects ceases, and is maintained by the upstream head boundary condition only. The relatively large depletion leaves a small volume available for abstraction, which could be increased if abstractions are initiated early in the dry season, as less volume water would be available for evaporation from the riverbed.

During the wet season, the majority of the days are dry since the discharge events are usually, short. During these dry days defined as dryspells, the water table decays with the same patterns as in the dry season until the next surface flow event occurs. As seen in in section 5.2.2, the maximum dry spell duration could reach up to 63 days during which time the storage could drop down to 70% of its maximum capacity. Rainfed crops could be severely damaged when such long dryspells persist, however the aquifer storage at that level provides significant amounts of water that can sustain irrigation during that time.

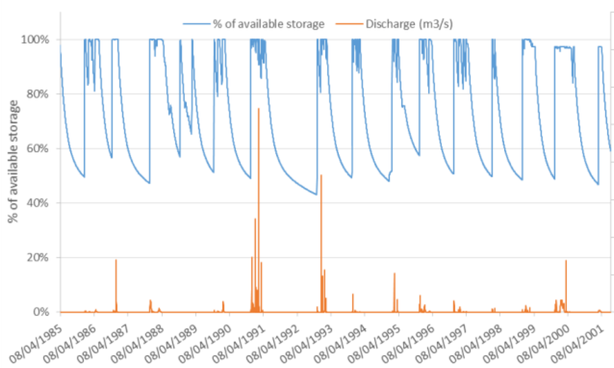

Figure 7: *Variation of available storage*

- **Recharge estimation**

The simulation of the variation of the storage for the period of 16 years show that the aquifer system recharged almost certainly every year except for the 1992 where no discharge occurred indicating extreme drought conditions. A beneficial application of the model would be to use the storage variation results to estimate the recharge of the aquifer in the boundaries of the study area. Recharge volumes vary between 150000 m3 to 380000 m3 per year. The magnitude depends on the length of the dry season of the previous year in addition to the frequency and duration of dry spells during the wet year. HY1987 observed the lowest recharge since the dry season was very short and no occurrence of dry spells. The highest recharge occurs during HY1991 and HY1997 where the volumes reach a cumulative of 380000 m3. These volumes exceed the maximum available capacity of the aquifer and suggest that recharge is cumulative and a function of the occurrence and duration of dry spells. Indeed, both years had a cumulative duration of dry spells between 50 and 70 days with shorter but more frequent dry spells. During the dry spells, the storage naturally depletes as an effect to evaporation losses, which is immediately re-saturated in the next discharge event. The depletion-recharge cycles increase the cumulative amount of recharge for a given hydrological year.

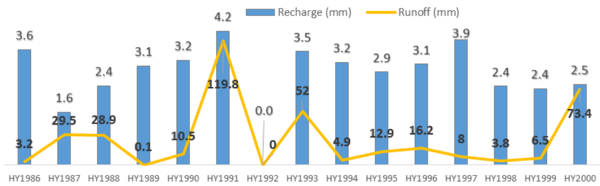

Figure 8: *Recharge calculated for the entire catchment*

These estimations are only valid for the study area stretch of 5.6 km. It is therefore difficult to generalise these calculations as diffuse areal recharge from rainfall if the mechanism is actually concentrated recharge from infiltration of runoff water. If we assume however, that recharge can be uniform for the entire stretch of the river in the drainage area, the recharge values computed above can be rescaled to the entire area of the drainage so as to compare its magnitude to other components of the hydrological cycle (**Figure 8**). This would lead to recharge values varying between 1.6 and 4.2 mm per year. This is considerably lower than the average annual runoff of 33 mm. On a yearly basis, sometimes, recharge observed was higher than runoff of that year (example of HY1986, 1989, 1994, 1998). This results from the model assumption that any discharge events was considered an indication of the saturation of the riverbed, regardless of the magnitude of the discharge measurements which sometimes was measured at 0.001 m3/s. A threshold value could have been applied to account for a reliable detection threshold, however, in the absence of stage data this was not possible. Stage data would have been more appropriate to estimate the relevance of the flood event in recharging the aquifer (Gustar & Demuth, 2008; Loucks & Beek, 2005).

- **Scenarios simulations:**

Effect of the strategies storage and on meeting the daily demand

Figure 9, illustrates the percentage of meeting an assigned demand defined by the number of wells and the pumping rate per well for scenarios 1, 2, 3 and 4 and for both cases of scenario B and C. The total demand is fully met for the entire dry season for all A scenarios, therefore the results were not presented. For the B scenarios where the total demand is 825 m3/day, all distributions fail at meeting the daily demand at different durations after the beginning of the dry season. In strategy 1B, failure is observed after 130 days of the beginning of the dry season and continues until the last dry day where the total daily demand is met at 40%. For the remaining strategies 2B, 3B and 4B, failure occurs around 40 days after the failure of 1B, with a higher percentage of meeting the daily demand as compared to 1B, however, failure is increased rapidly in the remaining days. This shows a clear advantage of using sparse pumping strategy midway the dry season. A strategy with distributed wells and decreased pumping rates per well should cause less local over-exploitation and therefore is expected to perform better than concentrated pumping. This is dependent on the location of wells with regard to the depth of the aquifer as well as any other obstacles in its upstream or downstream. In these simulations of sparse located wells, wells positioned downstream of rock sills dry up quickly as the water level drops below the obstacle, preventing any replenishment of the downstream. The closer the well to the upstream obstacle, the quicker it will dry and therefore the total daily demand decreases at higher rate than a scenario where the wells are located a bit downstream of the obstacle.

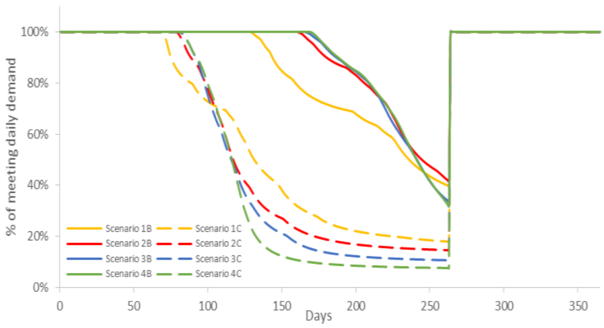

***Figure 8**: Percentage of meeting the daily demand*

For scenarios C, the system fails at delivering the total daily demand of 1900 m3/day, after the first 72 days of the dry season. Even though strategies 2C, 3C and 4C fail after 97 days, the decrease is more pronounced and the system provides less water per day for these sparse strategies as compared to concentrated pumping strategy. This is an expected result for high pumping rates per well, which will induce more drawdown that leads to interference between the neighbouring wells.

The total demand associated with Scenarios A, reduces the storage of the aquifer at the last day of the dry season to 35% of

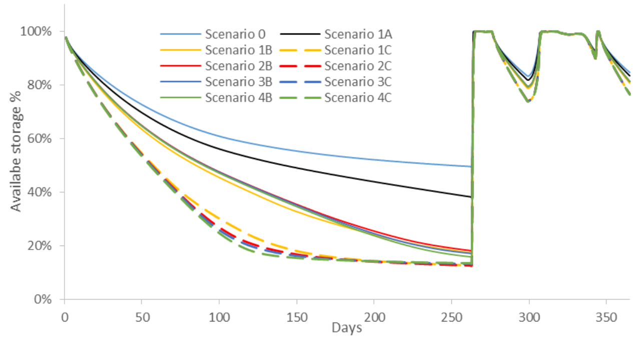

***Figure 7:** Variation of storage for different scenarios*

the maximum storage capacity as compared to 45% in the no-abstraction scenario (**Figure 10**). In scenario B, and at the same dry day, storage reaches 10% whereas in Scenarios C, the storage is almost entirely depleted. The pumping strategy influences the total storage of the aquifer. The sparser the distribution of the wells is, the more depleted is the aquifer, which indicates clearly the advantage associated with a more distributed strategy. The wells when strategically and sparsely positioned allow abstraction of more water from the aquifer.

Effect of abstraction on evaporation losses

Figure 11, illustrates the relationship between the supplied volumes to meet the demand and the total evaporation from the aquifer as a cumulative flux for the entire dry season for a specific daily demand. With increasing daily demand, the total cumulative evaporation during the dry season decreases. When no abstraction are applied, evaporation losses are equivalent to 53% of the available storage. This percentage decreases as the abstraction increase. For instance, an abstraction of 800 m3 per day lead to a 20% reduction in the evaporation rate and more water of the storage is available for use. For management strategies, more abstractions is favoured as it allows to maximize the use of the available water in the storage instead of losing to natural evaporation.

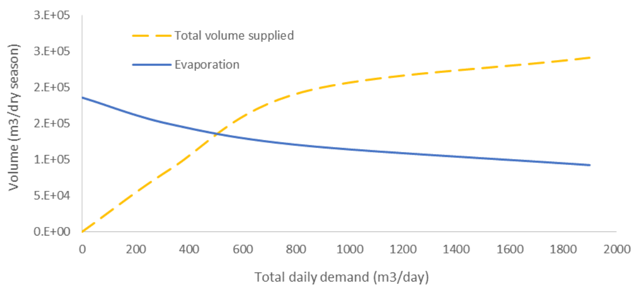

***Figure 9:** Effect of increased abstractions on evaporation*

Optimization of pumping rate and potential irrigated area

If we consider that meeting the total demand of the entire dry season as the criteria of the success of the system then determining the maximum abstraction rate per well which meets this criteria helps determine the maximum extractable volumes and thus the degree of agricultural development in the region. For this, the best abstraction strategy would be either strategy 3 or 4 even if strategy 2 showed better results. The reason behind this choice, stems in the distribution of gardens in the region which are distributed on both sides of the river banks. A sparse well network minimizes the distances of the pipes and

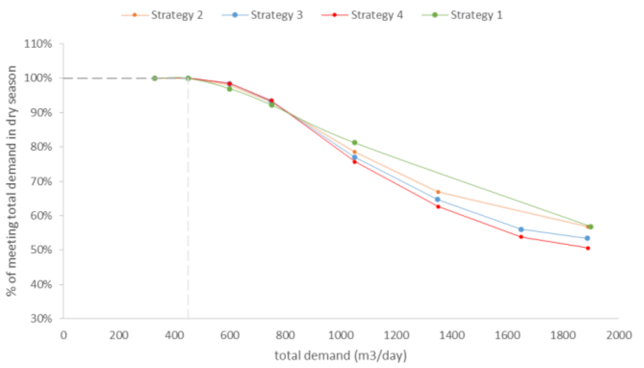

Figure 10: *percentage of meeting the dry season demand*

installations and could possibly reduce the associated costs. Since the optimized daily pumping rate is intended to meet the total demand at 100%, this means that the strategy is not relevant especially since no significant differences between the strategies in terms of the percentage of meeting the total demand was recorded (**Figure 12**).

The maximum total demand that could be entirely satisfied by either one of the abstraction strategies 3 or 4 without any drying up of the wells is 450 to 800 m3/day. There is not a significant difference between the two scenarios for the abstraction rate. Such volume will be used in the majority for irrigation of the neighbouring gardens, as abstractions for drinking water consumption are assessed to be negligible. This estimation is based on the assumption that irrigation starts immediately at the beginning of the dry season. Any delays in the start of the irrigation season results in more available water for abstraction at the end of the dry season, although evaporation losses might reduce the amount. **Table 4** summarizes the potential area that the determined a bstraction rate can irrigate.

Effect of pumping on recharge and runoff

Recharge was calculated for a selected year's representative of average, medium and extreme drought duration, defined by the duration of the dry seasons (see section 5.2.4). The year selected area, HY1986, HY1988, and 1991/1992, and HY1995. The pumping strategy followed is Strategy 4 with a pumping rate of 26.6 m3/day per well and total demand of 800 m3/day.

Figure 13, illustrates the annual recharge, computed for no pumping scenario and a pumping scenario corresponding to 800 m3 per day. The annual recharge is higher in the pumping scenario as the storage is depleted, and therefore there is more available empty volume to be recharged in the next surface water runoff event. As surface runoff is the main recharge source, increasing abstractions might lead to a reduction in surface water flow produced. For years with high runoff, the increase of the recharge will not disturb the surface flow regime as much as it

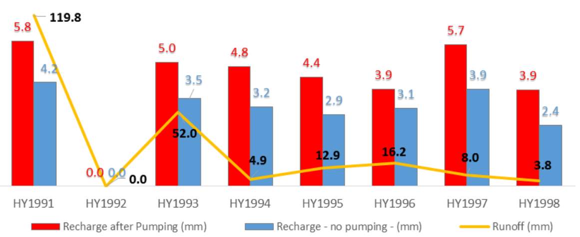

Figure 11: *Increased recharge due to pumping*

could do for years with low runoff. For example, in HY1994, recharge in the pumping scenario increases by 1.6 mm more than the recharge in the no-pumping scenario. If this amount is solely taken from runoff, the yearly runoff of that year could then be reduced by roughly the same amount. Runoff could be as low as 3.8 mm for HY1998. Increased pumping might significantly decrease the discharge if a low runoff is to be recorded in the next year in the case of such large scale irrigation.

Table 4: *Potential irrigated area*

	Dabane plan		Zimbabwe, Ministry of agriculture		Based on plant demand	
	High deficit irrigation				Full irrigation	
m3/day/ha	30		41		57.6	
Daily available water (m3/day)	450	800	450	800	450	800
Potential irrigated area (ha)	15	26.7	11.0	19.5	7.8	13.9
Potential irrigated area (ha/km)	2.7	4.8	2.0	3.5	1.4	2.5
Surface of irrigation scheme (ha)	3		3		3	
Potential number of irrigation schemes	5	9	4	7	3	5

• Limitations of the model

As the famous saying goes: "All models are wrong, however, some are useful". The model developed for this research was intentioned to provide a tool to understand the riverbed aquifer behaviour in natural conditions, and assess the extent at which agricultural development in the region can rely on for water provision. There were many simplifying assumptions followed and their justifications were presented. The assumptions adopted for the modelling have direct impact on the validity of the results and their implications will be discussed.

As the modelled study site is only a portion of a 102 km river stretch, head boundary conditions needed to be defined in the upstream and downstream of the study area to simulate the extent of the aquifer. Constant head with time variation was chosen for this exercise. The variations of the head was build using the discharge data and a simple assumption that any occurrence of the discharge brings the head to the saturation level. This might be a valid assumption considering the used time step, however, imposing the head in the upstream and downstream, will be forcing the system in the middle to react accordingly. In fact when simulations were run without any evaporation losses, the water table still decreased as a reaction to the drop in the heads at the boundary conditions. This could explain the low calibrated evaporation rate compared to the potential evaporation rates.

The constant head boundary conditions have no significant impact on the storage as inputs/outputs are negligible compared to the storage of the aquifer. In fact, simulations were performed were pumping stopped during the dry season and after the water level dropped below the extinction depth. The storage remained constant and was as expected not replenished by the constant head boundary conditions. Still, an alternative to these conditions would be to simulate a larger extent of the area (20 km for example) in a free draining condition by installing drains at the downstream extreme. This will mimic the natural behaviour of the system and allow the boundaries to react freely to the stresses within the study area.

No flow boundaries were considered on the riverbanks and the underlying bedrock. Such assumptions were justified because of the low conductive fine material which limits the losses. Since the processes in the aquifer are rapid, it is safe to make such an assumption. However, in the presence of undetected faults, the losses might be exacerbated.

During the flood event and after saturation of the riverbed, the water head exceeded the sand surface level. It was unable to use MODFLOW to simulate surface water flow. A modelling hack was adopted by installing drains at the level of the sand surface and therefore anytime the water head exceeds the sand surface level, the excess is eliminated out of the model. The use of drains for such a purpose was not found in literature before, perhaps due to the lack of small scale groundwater model. There are still some issues observed with their use. At the beginning of each simulation the drains seem to drain out water is not in excess even in the dry season. This creases extra losses in the storage and underestimate the real values of the storage. Therefore correction on the flow budget need to be done, or an improvement of the drains performance is suggested.

Modelling the abstractions was done without consideration of the method of abstraction and its feasibility. It assumed that the wells operate at a high efficiency and no issues are to be expected with the pumps. Besides, no refinement of the cells was applied around the wells and therefore drawdowns are expected to be more severe in the vicinity of the wells.

Finally, the optimized pumping rate using the numerical model is higher than that obtained by the tank model. The simple explanation is that the losses in the tank model were predetermined and conditioned by the logger data, therefore are not influenced by the hydrology of the system. Whereas in the numerical model, increased pumping allow more water to be pumped that would have otherwise be lost to evaporation. If the tank model adopts a linear model for the evaporation, similar results could be obtained from both.

5. General Discussion

The studied portion of the riverbed alluvial aquifer provides a maximum storage capacity of almost 700000 m3 of which only 50% is available for abstractions. There is however a high uncertainty regarding this estimation related to the values of the hydraulic and geometrical parameters of the aquifer. The deepest point of the channel was chosen from the geophysics to be uniformly 5m, with the exception of locations of subsurface obstacles. This assumption cannot be generalized as the geophysical assessment was limited to one cross section. Besides, the cross section was at the level of a tributary which generally favours erosion and therefore thicker riverbeds are expected compared to cross-sections away for the tributaries. In addition, the 3D subsurface model relied on many sparse interpolations between the probed cross sections despite the extreme irregularity in the bedrock. Considering 3.85 m as the average thickest depth of the main channel decreases the estimated total storage capacity by 12%. On the other hand, disparities with regard to the specific yield values which conditions the available volume for abstraction, will give gross differences in the estimations of groundwater. This work may be overestimating the potential of the aquifer.

The results of scenario 0 showed that the system's water losses are dominated by evaporation as illustrated in the groundwater dynamic analysis (see section 5.2.1.). Gonzalez-Carballo (2018) following the same approach, concluded that water level drops could be explained to a high extent only by evaporation. In opposite, Wekesa (2017) pointed out the existence of an additional loss to evaporation which he attributed to seepage into the underlying fractured bedrock. While vertical and lateral seepages might affect the water table decrease, their impact is localised as the weathering degree of the rock is irregular and cannot be generalized. In another hand, at locations where the water is trapped behind subsurface rock obstacles, volumetric evaporation losses tend to increase. As the water cannot flow downstream, the only way out for it is through evaporation which is enhanced by the shallow depth of the water table and the upstream constant replenishment. Finally, losses to evapotranspiration of the riparian vegetation were clearly observed through the diurnal fluctuations of the water table (see section 5.2.1.) and a clear indication that the riverbed aquifer system is connected to the banks. Gonzalez-Carballo (2018) modelled this connection with aquifer system and the riverbanks and quantified that these losses are insignificant in comparison to the direct losses to evaporation. This supports the assumption of no lateral flow in the model.

The natural depletion of the aquifer is recharged almost certainly at the beginning of the first or second surface flow event of the next wet season. This is highly dependent on the water content in the aquifer, the level of the water table, the hydraulic conductivity of the sediment and the intensity of the rainfall and the size of the system. Recharge from direct rainfall is generally almost negligible for small to medium systems, and runoff is considered by many authors as the main recharge mechanism. Wekesa (2017) for instance showed that indeed the surface water is only recorded once the groundwater level entirely saturated. Although the system he studied is shallow of 2 m depth and therefore bigger systems will require more intense events for their saturation. Studies on ephemeral rivers in Australia estimated that 25% of the discharge are infiltrated in the riverbeds, and the total recharge to the sand rivers amount to 4% of the total catchment recharge. While the ratio of infiltrated water from the discharge was not considered in this study, estimated recharge values from the model varied between 2 to 4 mm per year which account between 2 and 5% of the total catchment recharge (Nyagwambo, 2006; Sangwe, 2001). Note that the upscaling of the results was performed for indicative purposes only and recharge cannot be considered as a uniformly diffusive mechanism along the entire river. A more thorough estimation of recharge for the catchment's alluvial aquifer is recommended.

Despite the assumption that the recharge occurs in tandem with discharge events, there are still some years that go without a single drop of water to recharge the aquifer. According to the data, this occurred once every 20 years, although the length of the data series was too short for a strong affirmation. In the case of these long durations of the dry season, one of the advantages of the sand aquifer is that evaporation does not affect the losses once the water level drops below the extinction depth, and only slow losses to deeper aquifer or the river banks persist. This allows for water to be available for drinking purposes at specific locations where water tends to accumulate behind subsurface obstacles. However, the quality of the water for drinking purposes might deteriorate due to increased salinity (Gonzalez-Carballo, 2018) and large scale irrigation cannot be sustained during such prolonged dry seasons since the storage will be depleted through pumping in the previous year. Dependent vegetation on the other hand, should not be affected significantly as the plants can still access water beyond the retained fraction until the moisture content reaches the wilting point.

The dominance of evaporation results in a natural depletion of the aquifer to half of its available storage. This is particularly an issue with wide and shallow systems. Love et. al. (2007) indicates that shallow aquifers less than a meter in depth can dry up due to evaporation losses within 24 hours of a river flow. Nord (1985) estimated that evaporation losses account for a total of 25% of the available water in the aquifer in Botswana. This water should then be utilized immediately after the river flow as this will reduce the available water to evaporation and allocates it for other uses. Following this, if the water is utilized entirely during the dry spells or the beginning of the dry season, the water level will decline creating more available empty volume to be filled in the subsequent wet event, and thus maximizing the use of the aquifer water.

Maximizing the amount of water used from the aquifer can be achieved by intensively pumping the entire available storage and waiting for it to be recharged with the first flood event of the wet season, however, this might have implications on the riparian vegetation and the generated runoff if the pumping is intense. The yearly runoff is highly variable where some years could observe 110 mm whereas others observe runoff as low as 3 mm. On the other hand, pumping directly increases the recharge volume of the following year as it creates more empty storage for the runoff to refill. If runoff in the following year happens to be low, an increase of recharge of 1.8 mm because of pumping (see figure 5-31) might reduce the generation of the runoff to an extent that it affects other users downstream. In the absence of any particular study that supports this finding on the scale of sand river, Sapriza-Azuri, Jódar, Carrera, and Gupta (2015), using a global model in the semi-arid southern part of Spain, indeed found that intensive pumping reduces runoff generation by up to 50% throughout the entire year.

On an average dry season of 255 dry days, the studied stretch of 5.6 km of the aquifer has the potential of providing

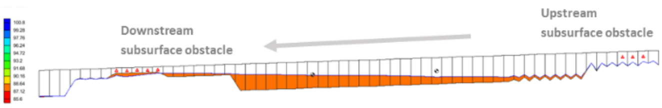

Figure 12: *Longitudinal cross section of the riverbed*

between 450 and 800 m3 per day without failure of the system in the entire dry season. Higher volumes will lead to drying up of some wells before others. When the water level drops below the level of a subsurface obstacle, the system will be separated into discontinuous micro-reservoirs and the micro-reservoirs downstream of the obstacle will not be replenished from the upstream. Therefore, the wells located at the immediate downstream of such an obstacle are most likely to dry up first. In contrast, wells located upstream of an obstacle are likely to remain productive for longer periods, as these obstacles work as subsurface dams that prevent downstream losses. At these levels, the water depth is shallow and easily accessible and therefore well installations should be targeted in the vicinity of such areas. Determining these localities, groundwater can be used to support local scale commercial irrigation.

The spatial pumping strategies were evaluated with the model. A strategy with sparsely located wells and reduced pumping rates was expected to be more reliable. It was revealed that this might be true when the abstraction rates are not relatively high. When pumping rates are low (i.e. meeting the total demand without failure), there is no observed difference between the strategies. For medium pumping rates (i.e. when failure is minimized) the sparse strategies are indeed more reliable than concentrated pumping, although the difference is within 3% (see figure 5-23). Higher pumping rates are better achieved using fewer wells because of the interference between wells closely spaced. Mulder (1973) reported a radius of influence of wells of 750 m for an optimum pumping rate of 1900 m3/day in the Limpopo riverbed aquifer system. Mulder's (1973) findings are for a system that is three times larger than the study area. Since for low pumping rates, the spatial strategy is not important, it is however suggested to opt for the distributed pumping network if small and more irrigation schemes are planned.

The optimized pumping rate could be allocated mainly to agricultural uses since domestic use demand was estimated to be negligible (Mansell & Hussey, 2005). Such abstracted volumes will be able to irrigate a potential area between 8 and 27 ha per 5km stretch or 1.4 to 4.8 ha/km depending on the followed irrigation strategy. This compares with the estimations made by Maspovo (2008) which approximated a potential irrigated area of 11.8 ha/km for an aquifer that is twice as big as the Tshelanyemba study site. Dabane's perspective is to develop gardens of 3ha of area using an irrigation scheme of 3 mm every 3 days. This is a very high deficit irrigation that stresses crops and therefore yields are expected to be low. On the other hand, the Ministry of Agriculture suggests a daily irrigation of 41 m3/day with which an average of 2.5 ha can be covered per km of river. A lower area can be covered by full irrigation, as this strategy aims to meet the entire demand of the crop which is higher than the potential evapotranspiration for crops like wheat, citrus, vegetables and sugarcanes, commonly planted in communal areas in Zimbabwe (Acquah and Masanzu, 1997; Mhlope 2017; Hussey 2005).

Sand river abstraction methods vary from traditional to technologically advanced methods. For limited abstractions, the local dwellers rely on hand dug wells in the sediment which are deepened depending on the drop of the next water level. These wells are seasonal and need to be reinstalled in the next season. Modern methods on the other hand are either humanized or motorized and vary in scale depending on the purpose of the abstraction. In the study area, users rely on scoop holes, whereas garden owners either use rower pumps, diesel pumps or solar pumps. Rower pumps allow for a low pumping rate that does not suit large irrigation schemes. Diesel pumps are the commonly used as they are portable and easily operated. Solar pumps installed by Dabane in the recent years are the favourites as they reduce the energy costs. However, their operation still creates a maintenance dependency on the supplier. On many occasions installed solar pumps were disconnected as a consequence to flooding events and require the intervention of Dabane for maintenance which sees the farmers switching back to diesel pumps as they are more reliable for them for the time being. Conveyance structures are more suitable to divert water from the river channel to the arable lands (Owen, 1989).

6. Conclusion

This research aimed at evaluating the potential of shallow sand riverbed aquifer for use in agricultural development projects in the neighbouring areas.

The riverbed aquifer is formed by homogenous highly conductive fine to coarse sand, however, the gentle slope slow down the groundwater flow to 100 m per year. The sand layers are very irregular all over the river and tend to deepen around the main channel and outer bends where thicknesses can reach up to 5m. The irregularities in subsurface due to the presence of natural bedrock barriers, divide the subsurface system into disconnected pools every 600 to 1000 m. The disconnection between the pools limits the groundwater replenishment and reduces their exploitation potential. However, the subsurface obstacle create location with a shallow water level easily abstracted.

The studied portion of the aquifer has a maximum storage capacity reaching around 0.7 Mm3 of which only 0.35 Mm3 per 5.6km or 3500 m3 per 1ha of the channel are available for abstractions. Naturally, evaporation is the dominant outflow mechanism, which accounts for almost 50% of the available water in an average dry season. It is conditioned by the extinction depth which is as deep as 1m for sands. Evaporation losses persist through the dry season in location where the water table is high, generally where the water accumulates behind subsurface aquifers. Another loss mechanism is the downstream groundwater flow which is only observed locally at the level of the created pools. In the entirety of the system, the upstream extent of the aquifer, allows for replenishment of the groundwater although its contribution is almost entirely negligible.

The aquifer systems underlays an ephemeral river system that serves as its main recharge input. It flows only a limited number of days during the short wet season. It is characterized by highly variable flood pulses that are interrupted by dry spells of almost 63 days of length. Generally, each flood pulse transports a volume much superior to the storage capacity of the entire riverbed aquifer of the catchment. Typically, the aquifer is re-saturated almost entirely with the first or second flood event of the wet season. Recharge estimates established by an upscaling of the local results varied between 2 and 4 mm. Such values were artificially calculated as the volume required to refill the aquifer and were up-scaled over the entire aquifer considering the recharge mechanism as diffusive. Recharge however, is not ensured on a year to year basis. Some years pass by entirely dry without a single recorded discharge. This occurred once only in the 20 years of data available, which provides a significant certainty in recharge occurrence. Increase pumping might significantly reduce runoff generation if the following years have a low runoff. This could impact the dependent ecosystems, the local dwellers and the downstream users.

Simulation of abstractions to determine the best spatial abstraction strategy as well as the sustainable abstraction rates were performed on an average dry year. It was hypothesised initially that a distributed network with lower abstraction rates per wells would maximize the exploitation potential of the aquifer storage. However, it was revealed that the spatial distribution does not seem to have a big influence for relatively low abstraction rates, whereas for high abstraction rates, fewer wells operate for longer before drying up. This occurs as a direct consequence to the discontinuity of the aquifer into micro-reservoirs where wells at the upstream of these reservoirs are more likely to dry up, than those at the downstream.

The portion of the studied aquifer is able to cover a demand of 450 to 800 m3 of water without any registered failure of the system over the dry season. Such volumes are able to irrigate an area between 1.5 and 5ha per km of stretch depending on the considered water demand of the crops.

The research demonstrates that the riverbed aquifers have good groundwater development potential as it showed that significant quantities of water for irrigation and water supply are available and their distribution meets the rural settlements in economically deprived communities in arid to semi-arid regions.

7. References

A4labs. (2016). *DUPC2 Integrated Research for Development - A4Labs full proposal*. Retrieved from Delft: http://a4labs.un-ihe.org/a4labs

Ashton, P. J., Love, D., Mahachi, H., & Dirks, P. H. G. M. (2001). *An Overview of the Impact of Mining and Mineral Processing Operations on Water Resources and Water Quality in the Zambezi , Limpopo and Olifants Catchments in Southern Africa* (ENV-P-V 2001-042). Retrieved from Harare, Zimbabwe:

Chinoda, G., Matura, N., Moyce, W., & Owen, R. (2009). *Baseline Report on the Geology of the Limpopo Basin Area, a contribution to the Challenge Program on Water and Food Project 17 "Integrated Water Resource Management for Improved Rural Livelihoods: Managing risk, mitigating drought and improving water productivity in the water scarce Limpopo Basin"*. Retrieved from WaterNet Harare:

Gonzalez-Carballo, A. (2018). *Assessing Groundwater Recharge And Storage Enhancements In The Alluvial Aquifers Of The Tigray Highlands (Ethiopia).* (MSc thesis), IHE-Delft Institute for Water Education.,

Gustar, A., & Demuth, S. (2008). *Manual on Low-flow Estimation and Prediction*. Retrieved from Geneva, Switzerland:

Hellwig, D. H. R. (1973). Evaporation of water from sand. *Journal of Hydrology, 18*, 317-327.

Hussey, S. W. (2007). *WATER FROM SAND RIVERS: Guidelines for abstraction*: Water, Engineering and Development Centre, Loughborough University.

Loucks, D. P., & Beek, E. V. (2005). *Water Resources Systems Planning and Management: An Introduction to Methods, Models and Applications*. Paris: UNESCO.

Love, D., Uhlenbrook, S., Twomlow, S., & van der Zaag, P. (2010). Changing hydroclimatic and discharge patterns in the northern Limpopo Basin, Zimbabwe. *Water SA, 36*(3).

Love, D., van der Zaag, P., Uhlenbrook, S., & Owen, R. J. S. (2011). A water balance modelling approach to optimising the use of water resources in ephemeral sand rivers. *River Research and Applications, 27*(7), 908-925. doi:10.1002/rra.1408

Mansell, M. G., & Hussey, S. W. (2005). An investigation of flows and losses within the alluvial sands of ephemeral rivers in Zimbabwe. *Journal of Hydrology, 314*, 192-203.

Maspovo, T. H. (2008). *Evaluation of the groundwater potential of the malala alluvial aquifer, lower mzingwane river, zimbabwe.* (MSc thesis), University of Zimbabwe,

Moyce, W., Mangeya, P., Owen, R., & Love, D. (2006). Alluvial aquifers in the Mzingwane Catchment: their distribution, properties, current usage and potential expansion. *Physics and Chemistry of the Earth, 31*, 988-994.

Mpala, S. C., Gagnon, A. S., Mansell, M. G., & Hussey, S. W. (2016). The hydrology of sand rivers in Zimbabwe and the use of remote sensing to assess their level of saturation. *Physics and Chemistry of the Earth, 93*, 24-36. doi:dx.doi.org/10.1016/j.pce.2016.03.004

Mulder, M. P. (1973). *WATER SUPPLY FROM RIVERSAND LIMPOPO RIVER AT MESSINA*. Retrieved from

Ncube, B., Manzungu, E., Love, D., Magombeyi, M., Gumbo, B., & Lupankw, K. (2010). *The challenge of integrated water resource management for improved rural livelihoods: Managing risk, mitigating drought and improving water productivity in the water scarce Limpopo Basin.* Retrieved from

Nord, M. (1985). *Sand rivers of Botswana, Phase 2.* Retrieved from

Nyagwambo, N. L. (2006). *Groundwater Recharge Estimation and Water Resources Assessment in a Tropical Crystalline Basement Aquifer.* (PhD), TU Delft and UNESCO-IHE, Delft.

Owen. (2000). *Conceptual Models for the Evolution of Groundwater Flow Paths in Shallow Aquifers in Zimbabwe.* (Unpublished PhD thesis), University of Zimbabwe,

Owen, R., & Dahlin, T. (2005). *Alluvial aquifers at geological boundaries: geophysical investigations and groundwater resources.* Rotterdam: AA Balkema Publishers.

Owen, R., & Rydzeski, J. R. (1991). *Shallow groundwater as a resource for small-scale irrigation development, in techniques for environmentally sound water resources development*. London: Pentech Press.

Owen, R. J. (1989). *The use of shallow alluvial aquifers for small scale irrigation with reference to Zimbabwe*. Retrieved from

Sangwe, K. M. (2001). *Recharge map of Zimbabwe and validation of rain gauges for country wide total atmospheric chloride deposition studies.* (MSc thesis), University of Zimbabwe,

Sapriza-Azuri, G., Jódar, J., Carrera, J., & Gupta, H. V. (2015). Toward a comprehensive assessment of the combined impacts of climate change and groundwater pumping on catchment dynamics. *Journal of Hydrology, 529*, 1701-1712. doi:http://dx.doi.org/10.1016/j.jhydrol.2015.08.015

Wekesa, S. S. (2017). *Analysis of Groundwater Potential and Interactions with Surface Water in the Sand River Region, Mara Basin Kenya.* (MSc), UNESCO-IHE,

WHO. (2016). *Safely managed drinking water - thematic report on drinking water* (CC BY-NC-SA 3.0 IGO). Retrieved from Geneva Switzerland:

World Bank. (2017). The World Bank in Zimbabwe: Overview. Retrieved from http://www.worldbank.org/en/country/zimbabwe/overview