Modelling the water level of the alluvial aquifer of an ephemeral river in south-western Zimbabwe

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Water from the alluvium of ephemeral rivers in Zimbabwe is increasingly being used. These alluvial aquifers are recharged annually from infiltrating floodwater. Nonetheless, the size of this water resource is not without limit and an understanding of the hydrological processes of an alluvial aquifer is required for its sustainable management. This paper presents the development of a water balance model, which estimates the water level in an alluvial aquifer recharged by surface flow and rainfall, while allowing for abstraction, evaporation and other losses. The model is coupled with a watershed model, which generates inflows from upland catchment areas and tributaries. Climate, hydrological, land cover and geomorphological data were collected as inputs to both models as well as observed flow and water levels for model calibration and validation. The sand river model was found to be good at simulating the observed water level and was most sensitive to porosity and seepage.

Keywords: Alluvial aquifer; ephemeral river; hydrological processes; modelling; Shashani River; Zimbabwe

1 Introduction

In large parts of tropical Africa, including Zimbabwe, the groundwater aquifers of the underlying crystalline basement rocks are the main water source for rural populations even though they have limited water supply potential (Davies and Burgess, 2013, Mazvimavi et al., 2007, MacDonald et al., 2008, Chilton and Foster, 1995). Water is typically abstracted from deep wells and boreholes and often from unreliable hand pumps; therefore, water users have to walk lengthy distances and queue for long periods to a functional water source to obtain an adequate supply of water.

As an alternative to the low yielding and unreliable groundwater aquifers and the limited availability of surface water resources, many communities in the semi-arid regions of south-western Zimbabwe have found the alluvial aquifers of ephemeral or episodic rivers to be a viable alternative source of water (de Hamer et al., 2008). The channels of these ephemeral watercourses contain extensive sand deposits (Figure 1). There is usually surface flow only after a rainfall event (Davies et al., 1994), with no surface flow for most of the year (Benito et al., 2009), but there is presence of subsurface water within the sand all year round (Herbert, 1998). These sandy alluvial valley aquifers are frequently referred to as ‘sand rivers’ and they are the most common river type in the arid and semi-arid regions of southern Africa (Davies et al., 1994).

[Figure 1 near here]

The water in the sediments of ephemeral rivers is naturally filtered by the sand and is thus clean enough for safe domestic use. Abstraction of water from such sand river alluvial aquifers is commonly referred to as sand-abstraction. These sand rivers have been exploited by rural communities in many areas of Zimbabwe either by shallow pits dug in the sand or collector wells in the river bank and provide a valuable, readily available water supply for local people (Hussey, 1997, Hussey, 2003). The alluvial aquifers of these ephemeral rivers thus comprise a vast, largely untapped potential for potable water abstraction and they are increasingly being used to supplement or replace the traditional groundwater resources that are becoming depleted. This resource, nonetheless, is not without limit, and an understanding of the hydrological processes of an alluvial aquifer is a basic requirement for its sustainable management.

The sustainable yield of a sand river aquifer depends on the recharge it receives and its distribution in time, the geometry of the sand river deposits, the hydraulic properties of the sand and the amount of water abstraction, evaporation and other losses (Herbert et al., 1997). Recharge refers to the amount of water reaching the saturated zone of the sand riverbed and the resulting increase in water level in the alluvial aquifer (Mpala et al., 2016). Sand rivers are recharged from replenishment by the intermittent surface flow (Horst, 1975), as well as from intermittent rainfall. Rainfall recharge depends on the depth to the saturated zone and the properties of the sand while evaporation only has a significant influence when the water level depth is less than 0.6 m (Neal, 2012, Quinn et al., 2018).

Hydrological models have been used to study the ephemeral rivers of Namibia and Kenya. Morin et al. (2009) developed a flood-routing model with components accounting for channel-bed infiltration to estimate aquifer recharge from the infiltrating floodwater in Namibia. Hut et al. (2008) developed a groundwater-flow model to study the hydrological processes in an aquifer with the presence of a sand dam in Kenya. Sand dams are a form of silted weir and are commonly built across sand rivers to retain more sand as a way to increase the amount of water available. They found that there were significant water losses from the alluvial aquifer to the adjacent banks and from seepage under the sand dam to the downstream alluvial aquifer.

The alluvial aquifers of ephemeral rivers have also been modelled in Zimbabwe (Mansell and Hussey, 2005, Love et al., 2010b, Mpala et al., 2016). Mansell and Hussey (2005) developed a simple single cell model of a sand river aquifer and calibrated it with limited data from four rivers in southwestern Zimbabwe: the Shashani, Huwana, Wenlock and Dongamuzi. The model represented the channel upstream of a site (including tributary channels) by a tank containing sand with water flowing out of the tank at the downstream boundary. Field results indicated that the velocity tends to decrease with time, i.e., in proportion to the depth of the water surface. The model therefore assumed that the velocity was inversely proportional to the depth of the water surface below the sand. The other flows into and out of the tank were due to evaporation from the surface of the sand, seepage and abstraction from wells. There was also periodic recharge from precipitation falling on the surface of the sand as well as from precipitation falling outside the channel and percolating through the banks (Mansell and Hussey, 2005).

Mpala et al. (2016) subsequently applied this model to the Shashani and Manzamnyama Rivers, both located in south-western Zimbabwe, and performed a sensitivity analysis to determine the parameters that the model is most sensitive to. As in Mansell and Hussey (2005), they found that the recession of the water level in the alluvial aquifers was mostly sensitive to the area of the channel contributing to the flow and the depth of sediments within the river channel. Love et al. (2010b) used the WAFLEX model together with a water balance module to compute the water balance of alluvial aquifer blocks in the Lower Umzingwane River of southwestern Zimbabwe. They found that average abstraction was of the same order of magnitude as alluvium flow and thus these two parameters were found to be important components of the water balance.

In addition to recharge of an alluvial aquifer vertically from the surface, there will be some horizontal flow from upstream. However, this horizontal flow (measured by Mansell and Hussey (2005) to be between 0.07 to 0.33 metres per day (m/day) depending on the river), is several orders of magnitudes less than the vertical flow (measured at more than 70 m/day) and was not modelled by Mansell and Hussey (2005) nor Mpala et al. (2016), who both used a single cell model. Moreover, Mansell and Hussey (2005) suggested that when the surface flow ceases, the channel is made up of hydraulically isolated sections and recommended that more research be undertaken to investigate the distribution of flows within alluvial channels and in particular to determine whether the assumption that the channel becomes divided into hydraulically separate units is correct. Improving this would likely improve the sensitivity of the model to both the rapid water level changes following a storm and the simulation of the recession curve during the dry season.

This paper seeks to improve our understanding of the hydrology of sand rivers by extending the single cell water balance model developed by Mansell and Hussey (2005) to multiple cells and combining it with the flows generated by an appropriate hydrological model to simulate catchment runoff. A revised version of the model is presented here, which treats the sand river aquifer as a series of interconnected alluvial aquifers and utilising a watershed model to estimate inflows from tributaries and from upstream catchment areas. Previous studies have shown uncertainty in their modelling due to lack of data, notably regarding hydraulic parameters (de Hamer et al., 2008). To address this gap, this paper also presents topographical and geomorphological data collected on the Shashani River to quantify the model parameters and water level data used to calibrate and validate the model. An analysis is also conducted to determine the sensitivity of the model to its different parameters.

2 Study area

The model was developed and calibrated on the Shashani River in southwestern Zimbabwe (Figure 2). The Shashani River was chosen because communities currently exploit the water from its alluvial aquifer and because of the presence of river flow data and water level measurements on that river system. The river is 206 km long with an estimated catchment of 2,826 km^2 and is one of seven major ephemeral rivers that make up the Zimbabwean portion of the Limpopo Basin.

[Figure 2 near here]

The Shashani River catchment is located in the middleveld region, a grassland region of intermediate altitude with a subtropical climate that makes up most of

Zimbabwe. This middleveld region experiences one rainy season per year, beginning in late October and lasting until early April. In the Shashani River catchment, total annual precipitation averages around 600 mm at the headwaters of the river and decreases to less than 450 mm at the outlet of the catchment (Mpala et al., 2016, Mansell and Hussey, 2005). Rainfall in the middleveld is erratic with long dry spells commonly occurring with a few intense storms of short duration contributing to most of total annual precipitation. These climatic conditions are prone to the formation of sand rivers as the incomplete weathering processes result in coarse sediment filling up river channels (Edwards et al., 1983, Mansell and Hussey, 2005).

3 The sand river model

The sand river model simulates both surface and near surface flow. Surface flow refers to water flowing above the alluvium, which in the case of the Shashani River occurs only intermittently following a storm. Near surface flow is the flow within the alluvium.

3.1 Surface flow

The flow of surface water was modelled using Manning’s equation:

$$Q = VA = \left(\frac{1}{n}\right) A R^{\frac{2}{3}} \sqrt{S} \quad (1)$$

where Q = flow rate (m^3/s), V = velocity (m/s), A = flow area (m^2), n = Manning’s roughness coefficient, R = hydraulic radius (m) and S = channel slope (m/m). Rearranging equation 1 to estimate the depth of the surface flow, and assuming that the width of the river is much greater than the depth of the flow, flow depth, *d*, can be estimated as:

$$d = \frac{nQ}{w^{\frac{5}{3}} \sqrt{S}} \quad (2)$$

where *w* refers to the river width and approximates the hydraulic radius *R*.

3.2 Flow within the alluvium

Where there is no surface flow, the alluvium is considered to consist of saturated and unsaturated zones. The horizontal flow within the saturated zone of the alluvium was calculated using Darcy's law and was found to be several orders of magnitude less than the surface flow. This is in agreement with Horst (1975) who mentioned that the flow within the alluvium is relatively small when compared with surface flow. Figure 3a, for instance, shows the flow conditions in five cells of the Shashani River following a major rainfall event (28th March 1980). The surface flow rate following that event was approximately 20.3 m^3/s. At the same time, the flow rate within the alluvium was only 0.001 m^3/s (average of the five alluvial cells depicted in Figure 3). Note that the sections shown are currently uninhabited and, for this reason, abstraction is zero. The horizontal subsurface flow is therefore ignored for simplicity of modelling.

[Figure 3 near here]

Figure 3b depicts flow conditions at the end of the dry season (28th October 1980) when there was no surface flow and further illustrates that the subsurface flow is insignificant. For this reason, it is assumed that during the dry season, the water level in the alluvium drops to an extent that natural rock dykes and the unevenness of the riverbed surface leads to compartmentalisation of the river channel. The alluvial channel was thus represented in the model by a series of separate tanks, which are fed by vertical recharge from the intermittent surface flow as well as from rainfall, with losses consisting of evaporation, seepage and any abstraction (Figure 4). The assumption of isolated compartments in the model does lead to discontinuities in the water surface at the boundaries of the sections. However, in practice, the slope of the water surface is such that the difference in water level between sections is generally less than a few centimetres over lengths of several hundred meters, as the results of the topographic survey will show below.

[Figure 4 near here]

The main processes controlling the water level in the saturated zone are:

(1) Recharge from intermittent surface flow

Surface flow at the upstream end of the river channel was first routed through the Shashani Dam and then through the Gulati Dam (cf. Figure 2) using the level pool routing method, with the outflow from the reservoir of the Gulati Dam used as input to the sand river model. Additional runoff was received into the alluvial aquifer channel from tributaries, with each tributary feeding the cell corresponding to its position along the river channel (Figure 5).

[Figure 5 near here]

The flow on the river channel was first converted to a flow depth using equation 2. When water was present on the riverbed, the alluvium was recharged at a rate governed by the infiltration rate and since the infiltration rate is relatively high for sandy channel beds, recharge normally occurs within one time step (i.e., one day) (Mpala et al., 2016). While the flow within the unsaturated zone of the alluvium could be modelled using Richard’s equation, the high infiltration rates measured on the Shashani River mean that this would be unlikely to result in any significant improvement in the modelling outputs, while increasing the required computational requirements. In order to maintain mass balance, the volume of water contributing to recharging the alluvium was removed from the surface flow, and if the flow depth in one time step was less than the capacity of the alluvium, the alluvium was not completely recharged.

(2) Recharge from intermittent rainfall

The amount of recharge from rainfall (when there is no surface flow) is a function of the water table depth (d_{wt}), the moisture content and nature of the sediments, and the rainfall intensity (Mansell and Hussey, 2005, McDougall and Pyrah, 1998). When the water table is near the surface, the infiltrating water from rainfall passes directly to the saturated zone while for greater water table depths most of the recharge is absorbed by the unsaturated alluvium and does not contribute to recharging the saturated zone. To take this into consideration, parameters d_{wts} and d_{wtd} are introduced, representing the water table depth (d_{wt}) under shallow and deep conditions, respectively. If $d_{wt} < d_{wts}$, water passes directly to the saturated zone and if $d_{wt} > d_{wtd}$, all the infiltrating rainwater is absorbed by the unsaturated zone. For the Shashani River the values of d_{wts} and d_{wtd} were estimated as 1.5 m and 3.0 m, respectively, and were parameters subjected to the sensitivity analysis described below. The actual depth to the water table is normalised with respect to these limiting values by:

$$d_* = 1 - \left(\frac{d_{wt} - d_{wts}}{d_{wtd} - d_{wts}}\right) \quad (3)$$

The moisture content of the soil is also defined in a normalised form θ_*:

$$\theta_* = \frac{\theta - \theta_{dry}}{\theta_{sat} - \theta_{dry}} \quad (4)$$

where θ_{dry} and θ_{sat} are the moisture contents in the air-dry and saturated states, respectively. The actual recharge is a function of θ_*^m where m is a recharge exponent with a typical value of around two.

The model can take account of seepage from the banks of the channel by increasing the rainfall value by an appropriate factor.

(3) Evaporation from the alluvial surface

The amount of evaporation from the alluvial surface depends on the depth to the water surface and the properties of the sand, and decreases with an increase in water table depth. This is estimated by the model for three different sand types using the method described in Mansell and Hussey (2005), which is based on the work of Hellwig (1973).

(4) Abstraction

Abstraction refers to the water pumped from the alluvial aquifer by communities living near the river for domestic and agricultural purposes. Abstraction is based on daily water requirements, which depend on the size of the human and livestock populations and the area of plots irrigated by smallholders living near the river. Hence, a daily household abstraction rate was calculated on the basis of the average number of people living in a household, the type and average number of livestock that a typical household possesses, and the surface area of irrigated plots along the river.

According to the 2012 Zimbabwe Population Census, an average household in the district in which the study area falls (Matobo District) comprises 4.6 people (Zimstats, 2012). Data from the Livestock Production Department were used to estimate the type and average number of livestock per household in the study area (Jele, 2018). The daily per capita domestic water requirement was based on findings from household surveys carried out by Dabane Trust, whose results are in agreement with the water consumption data of the Water for Africa Institute[1], while estimates by the Food and Agriculture Organization (Pallas, 1986) were used to determine the water requirements for the different types of livestock common in the study area. The irrigation water requirements were based on an annual water requirement of 15,000

[1] https://water-for-africa.org/en/water-consumption/articles/water-consumption-in-africa.html

m^3/ha/year, with the general assumption that irrigation would be done for only four months in the year (Moyo et al., 2017). The estimated number of people, type and number of livestock, and the area under irrigation per household were then used to calculate the daily household water requirement (Table 1).

[Table 1 near here]

Determining the total number of households abstracting water from the alluvial aquifer required counting the number of households in the areas of the river that currently use sand water abstraction using high-resolution satellite images. It was assumed that only households located within 3 km of the sand river use its water. The number of households was then multiplied by the daily household water requirement described above to estimate the rate of abstraction per unit length of river and this value was then used as input to the sand river model.

To accommodate for seasonal variations in water usage as a result of changes in mean daily temperature and precipitation, monthly abstraction factors, *f*, were calculated using the following relationship:

$$f = \frac{\left(1 - \frac{P_{mi}}{P_{max}}\right) * \frac{T_{mi}}{T_{max}}}{2} \quad (5)$$

where P_{mi} = monthly precipitation for month i, P_{max} = mean maximum precipitation for the month with the highest mean precipitation, T_{mi} = mean monthly temperature for month i and T_{max} = mean maximum temperature for the month with the highest mean temperature. T_{max} was set at 22˚C and P_{max} at 120 mm based on rainfall and temperature data for the study area.

(5) Seepage

It is assumed that the amount of seepage to the underlying bedrock, *seep*, is a function of the water table depth, i.e.:

$$seep = ks * (seddep - d_{wt}) \quad (6)$$

where *ks* is a seepage coefficient and *seddep* = depth of sediments.

Since three out of the above five processes controlling the water level in the saturated zone of the sand river aquifer are functions of the water table depth, i.e., recharge from intermittent rainfall, evaporation from soils and seepage, Newton's method was used to solve iteratively for the water level at the end of each time period.

3.3 Influence of upstream reservoirs on the sand river model

The reservoirs of the two dams on the Shashani River are operated as a coupled system with the reservoir of the Shashani Dam in the upper reaches of the catchment used to replenish the reservoir of the Gulati Dam situated just upstream of the research site (Figure 2). In consequence, and for simplicity of modelling, the reservoirs were modelled as one hypothetical reservoir whose capacity and surface area were the sum of the capacity and surface area of each individual reservoir, respectively. The level pool routing method was used to calculate the outflow hydrograph through the following relationship (Chow et al., 1988).

$$\left(\frac{2S_{t+1}}{\Delta t} + Q_{t+1}\right) = (I_t + I_{t+1}) + \left(\frac{2S_t}{\Delta t} - Q_t\right) \quad (7)$$

where I_t and I_{t+1} are the inflow values at time t and $t+1$, respectively, Q represent the outflow, Δt represents the time step and S is the value for storage.

The sand river model included a module that routed the outputs of the R-R model through the coupled reservoir system using the level pool routing method described above before being used as the upstream input to the sand river model. This approach required knowledge of the initial volume of water in the reservoirs and their storage volume. The routing also considered abstraction from the reservoir for irrigation purposes and losses through evaporation and seepage.

4 The rainfall-runoff model

This study uses the *Hydrologiska Byråns Vattenbalansavdelning* (HBV) Rainfall-Runoff (R-R) model to generate flows from upstream catchment areas and tributaries into the above sand river model. The HBV model is widely used to simulate catchment runoff in Zimbabwe. It was first applied in the humid subtropical climate of eastern and northern Zimbabwe (Liden et al., 2001, Andersson et al., 2006). Love (2013) and Love et al (2010a) used it to simulate the runoff of two catchment in southern Zimbabwe. The HBV model remains more popular than other commonly used R-R models such as SWAT because it requires fewer parameters to run it. SWAT is a complex physically based model that requires daily rainfall, maximum and minimum temperature, solar radiation, relative humidity and wind speed data as inputs (Devia et al., 2015). The HBV model requires only temperature, evaporation and precipitation as climatic parameters (Devia et al., 2015), which are available for the study catchment.

The HBV model is a semi distributed conceptual model (Lindström et al., 1997) with the catchment divided into sub catchments, which are themselves also subdivided into different elevation zones, with a maximum of 20 elevation zones allowed per sub catchment. Moreover, each elevation zone can be further subdivided into a maximum of three vegetation zones or land cover types (Devia et al., 2015). The model has three subroutines: snow accumulation and melt, response and routing, and soil moisture accounting (Lindström et al., 1997), and follows a water balance approach:

$$P - E - Q = \frac{d}{dt}(SP + SM + UZ + LZ + lakes) \quad (8)$$

where P = precipitation, E = evaporation, Q = runoff, SP = snow pack, SM = soil moisture, and UZ and LZ are the upper and lower groundwater zones, respectively, while *lakes* represent the volume of the lakes in the sub basin (Devia et al., 2015).

This study uses he HBV-light version of the model. The catchment, whose flow discharges into the sand river under study, was classified into three vegetation zones in order of increasing field capacity (FC), namely grassy woodland or row crops, wooded meadow or pasture and bare soil with crop residue cover. The catchment was also subdivided into 18 elevation zones, with the elevation of the catchment varying between 1428 m and 1030 m from the headwater to the catchment outlet, respectively. The proportion of each vegetation zone for each elevation zone was calculated.

5 Data collection

Climatic, hydrological, land cover and geomorphological data were obtained as described below, because they were required as inputs to the R-R and/or sand river models and, together with observed flow and water levels, for model calibration and validation.

5.1 Climatic data

Daily rainfall and temperature (mean, max and min) data were obtained from October 1976 to October 1983 from two weather stations in Zimbabwe (West Nicholson and Bulawayo) and from a weather station in neighbouring Botswana (Francistown) through Climate Data Online (CDO). Rainfall and temperature over the catchment were then estimated through interpolation using Thiessen Polygons. Estimates of rainfall and temperature were also obtained from interpretations of radar images, which were sourced from World Weather Online (WWO). These rainfall and temperature estimates were downloaded for a grid cell covering most of the study area during the period January 2012 to June 2017. The CDO dataset was used primarily to calibrate and validate the HBV model. Data for the period October 1976 to October 1977 were used to warm up the model, while the calibration and validation were done using data from October 1977 to October 1980 and from October 1980 to October 1983, respectively.

An older climatic dataset was required because the only reliable and available hydrological records for the study catchment, which are required to calibrate and validate the R-R model, were from 1977 to 1990. The radar dataset, because of its more recent and continuous data, was used for calibrating and validating the sand river model as well as for the sensitivity analysis. Dabane Trust provided daily evaporation data for the period November 1999 to June 2003, which were measured using an evaporation pan set up on the bank of the Shashani River.

5.2 Hydrological data

Daily flow data covering the period January 1969 - December 2015 for three gauging stations on the Shashani River (Figure 2) were obtained from the Zimbabwe National Water Authority, albeit data quality issues prevented the use of the entire dataset. These hydrological data were used to calibrate and validate the R-R model. In addition to river flow data, the water level in the alluvial aquifer was collected using an automatic pressure transducer positioned in a piezometer installed at Tshelanyemba on the Shashani River. This digital logger was installed in 2012 and was set to record water level on an hourly basis during the rainy season. During the dry season, the recording interval was reduced to once a day, as the change in water level is usually slow and gradual during that season. The logger was installed to record not only the depth to which the water level drops within the river sand, but also the height of the river flow above the surface.

Weekly water level and reservoir volume data were acquired from the Zimbabwe National Water Authority from October 1994 to January 2018 for the Shashani Dam and from March 1980 to April 2017 for the Gulati Dam.

5.3 Land cover data

In addition to climatic variables, the hydrological model requires information about land cover, which was determined from high-resolution satellite images from Google Earth. The process consisted of using an Iterative Self Organizing (ISO) cluster unsupervised classification (Dhodhi et al., 1999). The number of classes was specified and then an algorithm generated the initial cluster centres (ESRI, 2017). Initially five classes were used, and this was subsequently reduced to four and finally to three classes, which is the maximum number of land cover types allowed by the HBV model. This required combining similar land cover types, and estimating their field capacity through calibration.

Being a semi-distributed model, HBV is designed to simplify the modelling process and makes it easier for users with limited data. This means that the model output might not be as accurate as those from a fully distributed model such as SWAT. The use of three land cover types was thus a limitation of the model, as five land cover types would better represent the catchment. Nonetheless, having fewer land cover classes works particularly well in places such as the study area where there is limited land cover data, and where the land cover types have to be estimated from satellite images. In any case, the model was able to simulate very well the observed river flow, as described below. The results of the classification are shown in Table 2 and Figure 6.

[Table 2 near here]

[Figure 6 near here]

5.4 Topographical and geomorphological data

Topographical and geomorphological data encompassing channel width, depth and porosity of the sediments in the river channel, as well as infiltration were collected on the Shashani River to quantify the parameters required to run the sand river model. The survey was conducted in August 2016 on three non-connected sections of the river measuring 4.7 km, 4.9 km and 9.9 km (19.5 km in total), representing 50% of the length of the river channel (Figure 5).

Topographical measurements were collected using a Total Station Theodolite (TST). The measurements were taken along the length of the river at intervals of 400-700 m. Measurements across the width of the river were taken at 5 m intervals in the upper sections of the river where it is less than 50 m wide, while at the lower end of the Shashani River, where the river width increases to well over 100 m, the measurement were taken at every 10-20 m. The bedrock profile of the river channel was also established through physical probing to determine the depth of rock or clay layers from the sediment surface.

Sediment samples were collected in each of the surveyed river sections, with a total of ten sampling points taking over the length of the river. The grain size distribution was determined using the dry sieving method with sieves conformed with the American Standard Test Sieve Series of the American Society for Testing and Materials International. Using this technique, the coefficient of grain uniformity (U) was determined. The sediment porosity (n) was then determined using that coefficient through the following equations developed by Vuković and Soro (1992) and previously adopted in southern Zimbabwe by Love et al. (2008):

$$U = \frac{d_{60}}{d_{10}} \quad (9)$$

$$n = 0.255(1 + 0.83^{U}) \quad (10)$$

where d_{60} is sieve size for which 60% of the sample passed (mm) and d_{10} is the sieve size for which 10% of the sample passed (mm).

Porosity was also measured by taking sediment core samples. The samples were obtained below the sediment surface by digging a 1.2 m deep pit and then inserting a metre long uPVC pipe at a one-metre depth to take a horizontal sediment core in each of the three river sections. Porosity was calculated using the following equation:

$$n = \frac{V_v}{V_t} \quad (11)$$

where V_v = volume of voids (determined by measuring the amount of water required to saturate the sample), V_t = total volume of the sample (determined by calculating the geometric volume of the bulk sample). The porosities determined using equations 10 and 11 were found to be similar, with an average value of the two used for the purpose of this study.

The infiltration rates were determined using a single ring infiltrometer of one metre long and a diameter of 110 mm. Forty centimetres of the infiltrometer was inserted into the sand. Water was then poured into the 60 cm of the infiltrometer remaining above ground and times were recorded at every 10 cm depth of infiltration. One set of infiltration measurements was carried out in each of the three river sections.

6 Methods

6.1 Calibration and validation of the hydrological model

The R-R model was calibrated using a sequence of over 100,000 runs with randomly generated values of the model parameters. The Nash-Sutcliffe Efficiency (NSE) coefficient (Nash and Sutcliffe, 1970) was used as an indicator of the accuracy of the resulting model. The model was calibrated using observed hydrological data from gauging station B77D (Number 5 in Figure 2) covering the period October 1 1977 to

September 30 1978, while the validation period extended from 1 October 1 1980 to September 30 1981. These periods were selected as they had relatively good quality data from the gauging station. More recent data, especially from the 1990s onwards, showed that the stations were slowly degrading in data quality possibly due to siltation of the weirs.

The calibration and validation of the HBV model was successful (NSE coefficient = 0.86). The NSE coefficient values can range from $-\infty$ to 1 with a value of one corresponding to a perfect match between the modelled river flow and the observations. A NSE coefficient of zero indicates that the modelled outputs are as accurate as the mean of the observed data while a negative value means that the model is a worse predictor than the average of the observations. As a general classification, a model is considered good if $0.65 < \text{NSE} < 0.75$ and very good if $\text{NSE} > 0.75$ (Moriasi et al., 2007).

6.2 Calibration and validation of the sand river model

The sand river model was calibrated and validated using observed water level data collected between October 2014 and October 2016 and between October 2016 and June 2017, respectively. The calibration of the model consisted of adjusting manually the following model parameters: Manning’s roughness coefficient, evaporation rate, the moisture content, moisture exponent, the dry moisture content and the saturated moisture content, as well as the deep and shallow water depths, with the NSE coefficient used as an indicator of the accuracy of the resulting model. The calibration followed a three-step iterative procedure involving macro-level calibration, a sensitivity analysis and micro-level calibration, i.e., an approach adapted from Ormsbee and Lingireddy (1997) and used in Mpala et al. (2016).

6.3 Sensitivity analysis of the sand river model to its parameters

A sensitivity analysis was conducted on the above eight parameters of the sand river model as well as abstraction, porosity and the seepage coefficient to determine the variables influencing the most the sand river model outputs. For this, the value of each model parameter was increased and decreased by 10%, 20%, 30% and 40% and noting the resulting change in water level.

7 Results

7.1 Characteristics of the Shashani River

Figure 7 shows the topography of parts of the three surveyed sections of the Shashani River. There is a general decrease in river gradient in the downstream direction, although the presence of artificial sand dams can alter the gradient.

[Figure 7 near here]

The width of the Shashani River increases from 22 m at the upstream end of the research site (yellow arrow in Figure 2), to 125 m at the outlet of the catchment, but reaching over 200 m in width in parts of the river section located the furthest downstream. This is illustrated in Figure 8, which shows results of the topographical survey at different locations along the length of the river. As sand rivers get wider, they develop more extensive sedimentation and thus become more suitable for water abstraction. The average sediment depth was found to gradually increase from around 1 m at the upper end of the alluvial aquifer to approximately 3 m a few kilometres before the end of the alluvial aquifer zone. Sediment tests were carried at eight sampling points on the Shashani River, with infiltration rates of 3.10 m/hr., 3.13 m/hr. and 3.60 m/hr. measured for each of the three river sections, while the average porosity ranged from

0.375 to 0.430 with a median value of 0.405. Coarser sediments higher up in the catchment have higher rates of infiltration while in the lower section of the river where there is a higher proportion of finer sediments the infiltration rate is smaller.

7.2 Calibration and validation of the sand river model

The calibration and validation procedure resulted in an average NSE coefficient of 0.70, which means that the developed sand river model is good on the basis of the classification presented in section 6.1. It should also be noted that this coefficient incorporates the calibration of both the sand river model and the HBV model. This is also an improvement on the work of Mansell and Hussey (2005) and Mpala et al. (2016) whose single cell model did not reach a NSE coefficient higher than 0.65. Figure 9 shows a plot of the observed water levels at the research site together with the water levels produced by the hydrological model over 852 days extending from September 2013 to January 2016, covering two complete hydrological years. Although water level data were collected at the research site from October 2012 until August 2017, this particular period was chosen because it was a period where there was a complete time series of water level observations with no missing values. The water level logger installed on the Shashani River by the team malfunctioned on a few occasions due to flood damage and this resulted in some periods being unusable.

The model simulates relatively well the recession curve following the first and second rainy season depicted in Figure 9 (Days 1186 – 1526 and Days 1636 – 1850). The model is also very sensitive to sudden flooding of the river channel at the beginning of the rainy season (Day 1137 and Day 1530). Aquifer recharge is relatively rapid due to the high infiltration rates experienced in medium to coarse river sand. As a result, surface water reaches the subsurface water within an hour, resulting in an almost instant rise in the water table, with the water table capable of rising rapidly from an annual low to fully saturated conditions within a day, for instance see days 1532 – 1533. The major model limitations were in modelling subtle variations in water level especially during the rainy season as a result of the several storm events occurring during that season. The model understated both the sharp rise in water level and the subsequent sharp drop as a result of these sporadic events (days 1145 – 1185 and days 1537 – 1635). This could largely be due to reliance on Manning's equation for surface flow routing, which was chosen for this model due to its simplicity. Although more complex surface routing functions such as diffusion wave and kinematic wave could have been used, they would have added complexity to the model without improving overall model accuracy as surface flow occurs for a very few days in the year.

[Figure 9 near here]

7.3 Sensitivity analysis of the sand river model

The sand river model was found to be most sensitive to porosity, moisture content, seepage coefficient and abstraction, while the other parameters did not influence significantly the model outputs (Figure 10). Porosity is the ratio of the fraction of pore space or voids to the volume of material of the sediment. It thus determines the amount of water that can be retained in a given volume of sediments. Seepage from the channel into the surrounding soil and groundwater increase the water level recession rate. The seepage rate was estimated on the basis of water balance calculations, an approach suggested by Love et al. (2010b) who recommended that seepage be estimated by monitoring the recession of the water level when the surface flow is absent and no abstraction is taking place. Seepage, evaporation (to a depth of 0.9 m) and abstraction were found to have the same effect on the water level by continuously withdrawing water from the aquifer and the three accounted for most of the water loss in the aquifer. As abstraction and evaporation on the upper portion of the sediment could be estimated on the basis of measurements (see section 3.2), estimates of the magnitude of seepage were made and refined through calibration.

[Figure 10 near here]

During the site investigations, several dykes and sills across the river channel were detected, which are also visible on satellite images from Google Earth on the upper stretches of the river channel and near its outlet where there is less sediment (Figure 11). Physical probing into the sediment also revealed the presence of the same in the middle sections of the river, although they are mostly covered by extensive sediment, making them difficult to identify from satellite imagery. These dykes and sills act as natural barriers to the flow, resulting in reduced subsurface flow within the sediment and the splitting of the subsurface aquifer into compartments (Figure 12).

[Figure 11 near here]

[Figure 12 near here]

The results of the sensitivity analysis also showed that the model is not sensitive to Manning’s coefficient and the depth parameters (deep water depth, shallow water depth). This suggests that depletion of the aquifer is therefore largely influenced by porosity, moisture content, abstraction and seepage, the later occurs as the water percolates through the semi-permeable clay layer underlying the alluvial aquifer.

Furthermore, the model was, as expected, found to be very sensitive to the input flow data derived from the HBV model, as it sets the boundary conditions at any given time. Rainfall episodes that resulted in even small amounts of surface flow were enough to trigger marked increases in water level within the alluvium as the alluvium rapidly became saturated. The initial water depth, another initial condition that the model requires, was also very important and during calibration it was set to the initial water level data collected in the field.

8 Discussion and conclusions

The saturated alluvium of the ephemeral rivers of the arid and semi-arid regions of south-western Zimbabwe is increasingly being used to supplement or replace the traditional groundwater resources that are feeling the shocks of climate change and failing to meet the requirements of an increasing population against decreasing recharge. Nonetheless, the size of this water resource is not without limit and an understanding of the hydrological processes of an alluvial aquifer is a basic requirement for its sustainable management.

This paper presents the development of a two-dimensional multiple cell water balance model, which estimates the water level in an alluvial aquifer recharged by surface flow and intermittent rainfall, while allowing for abstraction, evaporation and other losses. The model is coupled with a watershed model, which generates inflows from upland catchment areas and tributaries. Topographical and morphological data were collected across a significant length of the Shashani River to quantify the parameters required for the model. The water balance model was calibrated and validated using observed water level data and a sensitivity analysis was performed to determine the influence of different model parameters on model performance, thus helping to better understand flow mechanisms within the alluvial aquifer system.

The model presented in this paper provided a good representation of the hydrological processes of the sand river system resulting in an NSE of 0.7. Similar to the model developed by Mansell and Hussey (2005), the developed model is semi-distributed, but with geometric and geomorphological parameters fully distributed, while the climatic data, initial moisture content, and catchment ratio are lumped. A further development of the model is the use of an R-R model to produce hydrographs that are used as inputs on the upstream boundary and for tributary inflows. This modelling is similar but is also believed to be an improvement on the HBVx-Waflex model used by Love (2013), as it is more accurate in predicting alluvial water levels, probably because it incorporates a daily time step rather than a 10-day time step. The model also uses fully distributed infiltration values, which is different from Morin et al. (2009) who determined a constant infiltration rate across the whole riverbed of the studied river in Namibia.

Surface flow within the channel of sand rivers was found to be short lived, lasting only a few days per year. At a daily time step, infiltration into the sand was found to be almost instantaneous, with full saturation of the alluvium occurring within an hour of the river channel being submerged with floodwater. This is supported by observations, with the infiltrated water providing recharge for the alluvial aquifer immediately below. It has been shown that HBV, which was used as the input model into the sand river model, is a very simple semi-distributed model that does not require a lot of parameters to run, while providing very good results. The output from HBV has been validated with an NSE coefficient of 0.86.

The movement of water within the alluvial aquifer system has also been explored. It has been established that the groundwater flow follows Darcy’s law, but once the surface flow ceases the subsurface flow within the sediments becomes so low that it can be ignored, and the alluvial aquifer was modelled as a series of discrete compartments independent of one another, which are fed by vertical recharge from the intermittent surface flow as well as from rainfall and they lose flow by evaporation, seepage and any abstraction. The presence of impervious dykes and rock sills divides the sediments into separate hydrogeological units, resulting in very little subsurface flow. This is in agreement with Mansell and Hussey (2005) and Benito et al. (2009), who noted that the flow within the alluvium for sections of sand rivers in Zimbabwe and South Africa is minor and that the aquifer should be represented as separate groundwater sections given the presence of rock sills or any other geological barrier as such that prevents groundwater outflow. This was also observed in Kenya, and in areas where there is no geological barrier impeding the subsurface flow, communities living alongside the river have constructed sub-surface soil dams that trap water (Nissen-Petersen 1998).

The sand river model was found to be sensitive to porosity, seepage and abstraction. Unconsolidated sediments, such as those found in sand rivers, tend to have higher porosity than consolidated sediments. Porosities of a number of sand river alluvial aquifers in southern Africa have been measured from 37.5 – 43% (Mansell and Hussey, 2005, Walker et al., 2018, Love et al., 2008, Wipplinger, 1958), and those results agree with the measurements obtained during the survey undertaken on the Shashani River.

Through a water balance approach it was found that seepage into the bedrock is an important flux in the alluvial aquifer system of the Shashani River, and it was found to be greater than Darcy’s flow. This is in contrast to Love et al. (2010b) and Love (2013). The major difference could be attributable to differences in geological formations, with the granite/gneiss complex on the Shashani alluvial aquifer system of the current study being more deeply weathered than that of the Umzingwane River system studied by Love. De Hamer (2008) argued that in older terrains that are more deeply weathered, seepage can be a substantial flux. This is because the riverbed under the sand may have seepage lines along boulders and fractured rocks that allow the water to penetrate to greater depths (Nissen-Petersen, 1998).

In a previous study and using a single cell model without using a R-R model to simulate runoff from upstream catchment areas, Mpala et al. (2016) noted that the channel length and the depth of sediments were the two main parameters affecting the accuracy of the modelled water level when compared with actual measurements. The channel length, which is a ratio of the channel area to the actual width of the channel at the point being observed, is an indication of the length of the channel contributing to the flow (Mansell and Hussey, 2005). Mpala et al. (2016) also observed that their model better represented the water level measurements when the sediment depth in the model was equated to the difference between the highest water level and the lowest water level as opposed to the full sediment depth. The results of the single cell model of Mpala et al. (2016) could not be compared with those obtained using the current model, as the geomorphological properties were fully distributed in the current model; therefore all geometric features of the model were included on the basis of field observations.

Despite the improvements, the model's surface flow routing can still be further improved by incorporating either the diffusion or kinematic wave equations. This will be particularly important on sand rivers with extended surface flow, i.e., surface flow that lasts for weeks or months as is common with sand rivers originating in wet regions, such as the Juba and Shabelle rivers in Somalia, which both originate in the Ethiopian Highlands (SWALIM, 2016). The model would also be more robust if field data on seepage from the alluvium into the underlying soil or bedrock were available. Nonetheless, the current model is applicable across all sand river systems in the prediction of subsurface water level, although for sand rivers with perennial surface flow, coupling hydrological models with hydraulic models such as HEC-RAS may need to be explored.

The model currently estimates water level within the aquifer of a sand river system. However, with sufficient topographical data, calculations could be done to determine the amount of water in storage in any sand river. Furthermore, by combining the coupled HBV and sand river model with outputs from General Circulation Models (GCMs), it should be possible to simulate future water level conditions within the aquifer system and thereby be aware of the sustainability of this water resource under changing climatic conditions. In addition, using present abstraction data and population projections, it is also possible to estimate the sustainability of the sand river system as an alternate water source.

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Table captions

Table 1. Estimation of the daily household water abstraction rate.

Table 2. Results of unsupervised classification of the land cover types in the study area.

Figure captions

Figure 1. The Shashani River towards the end of the rainy season in April 2010 at Tshelanyemba where the water logger described below was installed (top) and in the middle of the dry season in August 2016 (bottom). During the dry season, there is no water flowing on the surface of the river but digging up to a certain depth reveals the presence of water within the sediments.

Figure 2. The Shashani River catchment in south-western Zimbabwe together with the location of the gauging and weather stations, water level logger, major dams and the section of the river channel that was surveyed, which is identified as ‘research site’.

Figure 3. Comparisons of the magnitude of the surface flow with the subsurface flow and seepage on the Shashani River between longitudinal cross sections 217 and 222 on a day following a storm during the wet season (a) and on a day during the dry season (b).

Figure 4. Schematic representation of the fluxes within an alluvial channel as modelled by the sand river model. Note that the vertical scale is exaggerated, as the differences in water level are only a few centimetres over lengths of several km.

Figure 5. Schematic representation of the surveyed river channel with the grey polygons representing sections of the river where geomorphological data were collected while no data were collected in the white sections.

Figure 6. The three land cover types (vegetation zones) of the studied catchment following an unsupervised classification procedure using Google Earth. Left: original optical image of the study area with catchment boundary outlined in red. Centre: Initial results of unsupervised classification using ArcGIS. Right: final classified and clipped image of the study area clearly showing the three vegetation zones.

Figure 7. The longitudinal profiles of the Shashani River for parts of the three surveyed sections depicted in Figure 5, with (a) referring to the section of the river closest to the Gulati Dam and (c) the section of the river that is the furthest downstream.

Figure 8. Cross sectional profiles at various points along the river downstream of the Gulati Dam. The yellow line represents the sand river bed, the blue line the water level and the black line the bedrock.

Figure 9. Plot of the observed water levels at the research site together with the water levels produced by the hydrological model over 852 days (two hydrological years) extending from September 2013 to January 2016.

Figure 10. Results of the sensitivity analysis on eight parameters.

Figure 11. Location of dykes on a 1.2 km stretch of the Shashani River as identified using a Google Earth image taken on July 26 2016 (a) and location of dykes and sills mapped using across the length of the river channel mapped dykes and sills along the river channel (b).

Figure 12. Longitudinal profile of a sand river channel showing the influence of the presence of dykes and rock sills on the flow within the alluvium shortly after a storm (a), a few weeks following the rainy season, and during the dry season (c).