The Mushawe meso-alluvial aquifer, Limpopo Basin, Zimbabwe: an example of the development potential of small sand rivers

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Abstract:

The Mushawe river is a right bank tributary of the Mwenezi River in the northern Limpopo Basin. At Maranda, the Mushawe is a sand river with a catchment of 220 km^2. The sand river forms an alluvial aquifer, blocked by a bridge that functions as a sand dam, and supplies water to Maranda No 1 Business Centre. The aquifer is shallow in vertical extent (< 5 m) and responds very rapidly to recharge from upstream rainfall and surface flow. This is typical of alluvial aquifers, which can be considered an extension of surface flow. The aquifer retains water year-round, which can be ascribed in part to the underlying impervious granite geology, which minimises seepage losses. The maximum storage of the aquifer of just under 9,500 m^3 is filled during flood events. Only around half of this remains stored in the aquifer during the year, which is nevertheless sufficient water to double the water supply to No 1 Business Centre. The water balance and model is based on a study site of 500 m of river reach. It should therefore be noted that abstraction further upstream could significantly increase the available water. Furthermore, upstream on-river storage and artificial recharge would allow for maintenance of a greater saturated aquifer volume throughout the year. The scale of aquifers such as Mushawe seems best suited to water supply for domestic and livestock use and sub-hectare scale horticultural irrigation.

Keywords: Alluvial aquifer, Groundwater – surface water interactions, Limpopo Basin, Sand dam

1. Introduction

An alluvial aquifer can be described as a groundwater unit, generally unconfined, that is hosted in horizontally discontinuous layers of sand, silt and clay, deposited by a river in a river channel, banks or flood plain. Because of their shallow depth and vicinity to the streambed, alluvial aquifers have an intimate relationship with surface flow, and indeed it can be argued that groundwater flow in alluvial aquifers is an extension of such flow (Mansell and Hussey, 2005). Most rivers that host alluvial aquifers recharge them annually, although some which are perennial do so continuously (Barker and Molle, 2004). Rivers can be classified as discharge water bodies if they receive a groundwater contribution to baseflow, or as recharge water bodies if they recharge a shallow aquifer below the streambed (Townley, 1998). In semi-arid regions, streams with alluvial aquifers tend to vary from discharge water bodies in the dry season, to recharge water bodies during the rainy season or under a managed release regime (Owen, 1991).

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Intermittent rainfall patterns in semi-arid regions have the potential to impose high temporal variability on alluvial aquifers. In arid and some semi-arid areas, recharge may occurs only after high discharge peaks from heavy rainfall events and be very limited during small to medium flows (Matter *et al.*, 2005). This has been reported from the Kuiseb River in Namibia (Lange, 2005) and the Tabalah Catchment in Saudi Arabia (Lange and Lebinbungut, 2003). In the semi-arid northern Limpopo Basin, work to date on alluvial aquifers has focussed on the major tributaries of the Limpopo River, such as the Shashani and Thuli Rivers (Mansell and Hussey, 2005) and the Mzingwane River (Owen and Dahlin, 2005; Moyce *et al.*, 2006). Little work has been done on small alluvial aquifers – here understood to refer to aquifers on rivers draining a meso-catchment (catchment area of approximately 10^1 – 10^3 km^2; Blöschl and Sivapalan, 1995). Whilst these aquifers will have lower potential storage than larger ones – which are seen as good sources for irrigation water (Owen and Dahlin, 2005; Moyce *et al.*, 2006; Raju *et al.*, 2006) – small alluvial aquifers may be easier to access for poor rural communities. This is because a smaller head difference between the riverbed and the bank can allow for cheaper or manual pumps. Thus accessing small alluvial aquifers for irrigation represents a possibility for development for smallholder farmers. However, little knowledge is available on the hydrogeological characteristics of small alluvial aquifers. In this study, the alluvial aquifer of the Mushawe River is evaluated to determine the development potential it represents in terms of (i) continuity and (ii) quantity of water supply.

2. Methods

2.1. Study area

The northern Limpopo Basin is a semi-arid area, with rainfall varying from 630 mm a^{-1} at in the north to 360 mm a^{-1} in the south (Love *et al.*, in prep.). Rainfall is seasonal, controlled by the Inter Tropical Convergence Zone and falling between October/November and March/April (Makarau and Jury, 1997). Rainfall occurs over a limited period of time, and often a large portion of the annual rainfall can fall in a small number of events (De Groen and Savenije, 2006). This temporal unreliability of rainfall means there is a need for intra-annual storage to guarantee water supplies for domestic use and for agriculture. However, with potential evaporation exceeding rainfall (Farquharson and Bullock, 1992), the water yield from the developed surface water resource often falls short of the demand, deficits being more evident during droughts (e.g. Nyabeze, 2004). For this reason, groundwater is an attractive water source in the northern Limpopo Basin.

The alluvial aquifers of the Mzingwane Catchments are the most extensive of any tributaries in the Limpopo Basin (Görgens and Boroto, 1997), and are common on the Shashe, Mwenezi, Mzingwane and Thuli Rivers and their tributaries (see figure 1). Alluvial deposits are present in the lower reaches of most of the larger rivers and some of the minor tributaries. They are narrow bands, typically less than 1 km in width on the largest rivers, too several metres on smaller river. The distribution of these aquifers is determined by the river gradient, geometry of channel, fluctuation of stream power as a function of decreasing discharge downstream due to evaporation and infiltration losses, and rates of sediment input due to erosion (Owen, 1991). Recharge of these alluvial aquifers is generally excellent – although annual – and is derived principally from river flow. No river flow occurs until the channel aquifer is saturated and such full recharge normally occurs early in the rainy season (Owen and Dahlin, 2005).

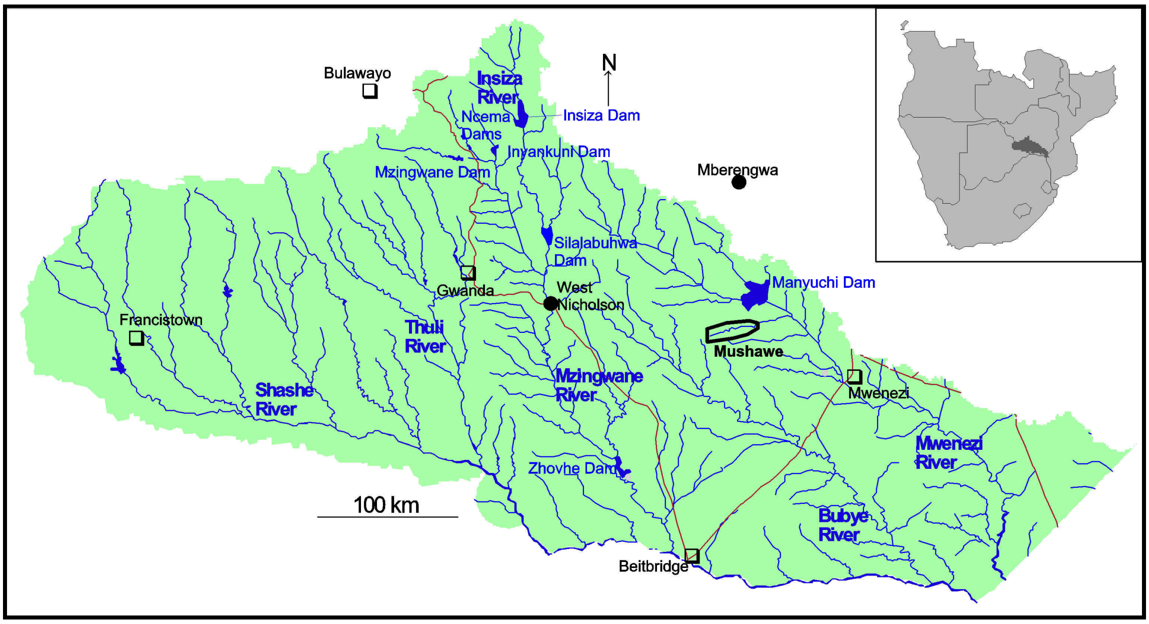

Figure 1. Location of the selected study catchment, shown with thick black border, within the northern portion of the Limpopo Basin. Reference meteorological stations are shown by black circles, other towns by open squares, rivers by blue lines and railways by red lines. Inset: location in southern Africa (shaded).

The Mushawe river is a right bank tributary of the middle Mwenezi River (see figure 1). At Maranda, the Mushawe is a sand river with a catchment of 220 km^2 (figure 2). The sand river forms an alluvial aquifer, blocked by a bridge that functions as a sand dam, and supplies water to the adjacent Maranda No 1 Business Centre (approximately 5 m^3day^{-1}). Upstream of No 1, the river is 35 km long and the catchment is composed of fields (32.4 %), medium to dense woodland (13.2 %), mixed grassland and woodland (48.4 %), rocky hills (5.6 %) and wetland (0.4 %), see figure 2. The catchment is underlain by granitic rocks of the Limpopo Belt North Marginal Zone, aged 2.72 - 2.52 Ga (Mineral Resources Centre, 2007). The mean annual rainfall for the nearest reference rainfall stations are 472 mm a^{-1} at Mberengwa (1987-2000) and 421 mm a^{-1} at West Nicholson (1962 – 2008; see figure 1 for locations).

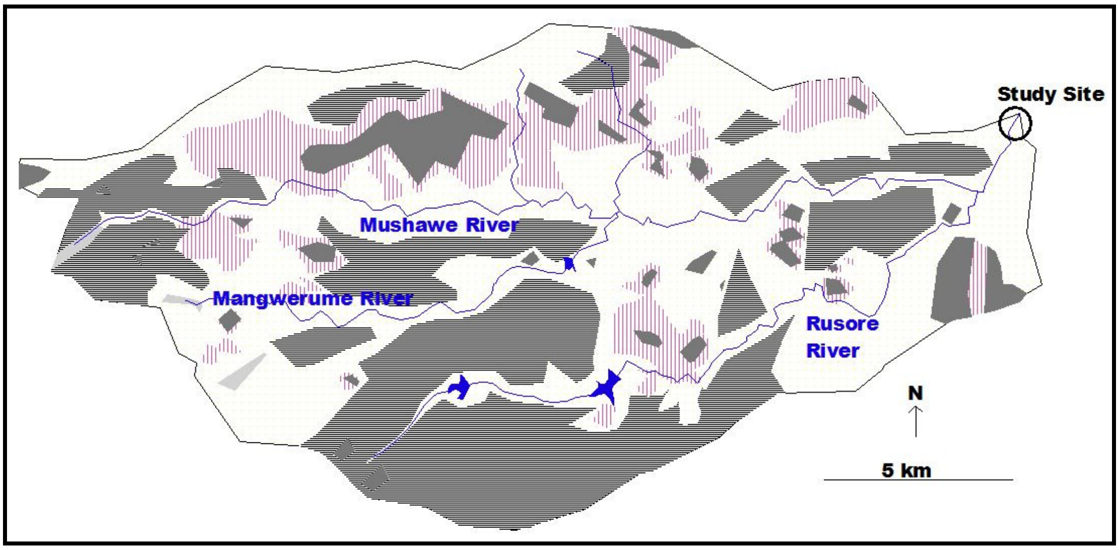

Figure 2. Land cover and drainage in the Mushawe catchment, derived from false colour composite using bands 3, 4 and 5 of a portion of Landsat Scene p170r075, dated 1 June 2000 - Landsat 7 spacecraft, supported by national 1:50,000 topographic mapping and ground

truthing. The hydrogeological study site is marked in the northeast (downstream) corner of the catchment. Key to land cover: blue = water bodies, vertical stripes = forest, horizontal stripes = fields, dots = mixed grassland and woodland, dark grey = bare rock, light grey = wetlands.

2.2. Data collection

The channel width and slope were surveyed in the field and the depth of sand determined by physical probing with a steel probe. The surface of the riverbed was surveyed by GPS. Composite samples of alluvial material were collected from various depths of the aquifer.

Grain size distribution was determined by the sieve shaker method. The used sieves are US standard sieves with sizes 4000, 2800, 2000, 1000, 500, 250, 180, 125 and 32 µm. Porosity (n) was then derived indirectly using the coefficient of grain uniformity (Vukovic and Soro, 1992):

$$n = 0.255(1 + 0.83^{U}) \qquad U = \frac{d_{60}}{d_{10}} \tag{1}$$

Where n = porosity (-), U = coefficient of grain uniformity (-), d_{60} = sieve size for which 60 % of the sample passed (mm) , d_{10} = sieve size for which 10 % of the sample passed (mm).

Specific yield was determined from the volume of water that drained under gravity from a known volume of saturated aquifer material:

$$S_y = \frac{V_{wd}}{V_{tot}} \tag{2}$$

Where S_y = specific yield (-), V_{wd} = volume of water drained under gravity (cm^3), V_{tot} = total volume of saturated sand (cm^3).

Hydraulic conductivity was determined by using a permeameter: a 60 dm^3 bucket with a 25 mm outlet at the base was completely filled with aquifer material and a constant head maintained by continuous inflow. The time taken for water flowing out of the outlet to fill a 5 dm^3 bucket was recorded and the hydraulic conductivity computed as follows:

$$K = \frac{V}{tA} \tag{3}$$

Where K = hydraulic conductivity (m day^{-1}), V = volume of water flowing out of the outlet (m^3), t = time (day), A = cross-sectional area of bucket (m^2).

Eight piezometers were driven through the aquifer to bedrock, placed in pairs at four locations upstream of the bridge (figure 3), which forms the downstream edge of the study site (see figure 2). Discharge was measured using a limnigraph located on the bridge and the surveyed river cross-section at the bridge. Daily observations were made of the water level in each piezometer and at the limnigraph. Rainfall was measured daily, using 17 catch-gauges, spread throughout the catchment of the study site, and catchment rainfall calculated using Thiessen polygons.

Figure 3. The field study site, showing the riverbed and location of instrumentation. L = limnigraph, P1-P8 = piezometers no. 1 – no. 8, A = abstraction point. Satellite image from Google Earth.

2.3. Aquifer Water Balance

The riverbed was divided into eight cells, each allocated to a piezometer (figure 4). The volume of water stored in aquifer on a given date was determined from the specific yield, the daily piezometer readings, the depth measurements and the riverbed surface cell area (equation 4). Potential storage was calculated for a unit length of river from the river cross-section and specific yield (equation 5).

$$V_{gw} = S_y \sum_{i=Pz1}^{i=Pz8} A_i h_i \tag{4}$$

$$V_{pot} = S_y \sum_{i=Pz1}^{i=Pz8} A_i h_i\_max \tag{5}$$

Where V_{gw} = volume of stored groundwater (m^3), V_{pot} = potential volume of stored groundwater (m^3), A_i = area of riverbed cell of piezometer i (m^2), h_i = height of water table in piezometer i (m), h_{i_max} = length of piezometer i (m), S_y = specific yield (-).

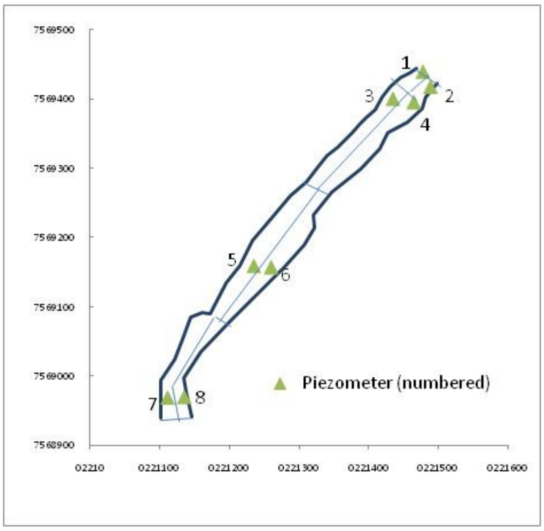

Figure 4. Cells for the aquifer water balance determination

2.4. Long-term Variability

The results of the rainfall-runoff models were used to show the relationship between rainfall and frequency of flow in the Mushawe River at the study site. Since the aquifer must be saturated before flow occurs at the study site, the frequency of flow events was used as an input to the aquifer models, in order to determine the change in volume of water stored. By comparison with the long time series of rainfall data from West Nicholson (1962 – 2008), the long term variability of the alluvial aquifer was predicted.

3. Results and Discussion

3.1. Field Data

The hydrogeological characteristics of the aquifer are presented in table 2 and show properties that are reasonable for an aquifer composed of fine sand. The porosity value is somewhat higher than the 30 % derived by Moyce *et al.* (2006) and 35 % derived by Owen (1991) for the Mzingwane River and also by Nord (1985) for rivers in neighbouring Botswana. The specific yield is within the expected range for fine sand (Johnson, 1967), which fits with the observed grain size distribution. It is also within the range for alluvial aquifers reported elsewhere (e.g. Barlow *et al.*, 2003; Rodríguez *et al.*, 2006), and close to that reported by Nord (1985) and Owen (1991) from neighbouring river basins. The hydraulic conductivity is on the border between that expected for fine sand and that expected for silt (Bear, 1972), which also fits with the grain size distribution. Note that some authors assume that specific yield is equal to the aquifer effective porosity and thence to the total porosity - see for example the discussion of this problem in Laslett and Davis (1998). In this case, such an assumption would result in a 300 % increase in the volume of stored water calculated.

Table 2. Hydrogeological characteristics of the Mushawe alluvial aquifer

Channel slope	0.17 %
Width of river at limnigraph	50.0 m
Depth of alluvial aquifer material (range)	1.60 to 2.45 m
Depth of alluvial aquifer material (average)	2.10 m
Average cross-sectional area of aquifer	106 m^2

Grain size distribution	Fine sand to silt 54 %, sand 43 %
Porosity (Vukovic and Soro method)	43 %
Specific yield	14.4 %
Hydraulic conductivity	26.8 m day^{-1}

When comparing discharge and depth to water table (figure 5), it can be seen that a small volume of water flows out of the aquifer at the bridge throughout the year – discharge in March to April 2008 was non-zero, but too low to reflect on the graph. During the rainy season the water table initially rises (early December 2007), and then reaches the surface allowing for discharge events. Thus the Mushawe River changes from a discharge water body in the dry season, to a recharge water body during the rainy season, in terms of the definition of Townley (1998).

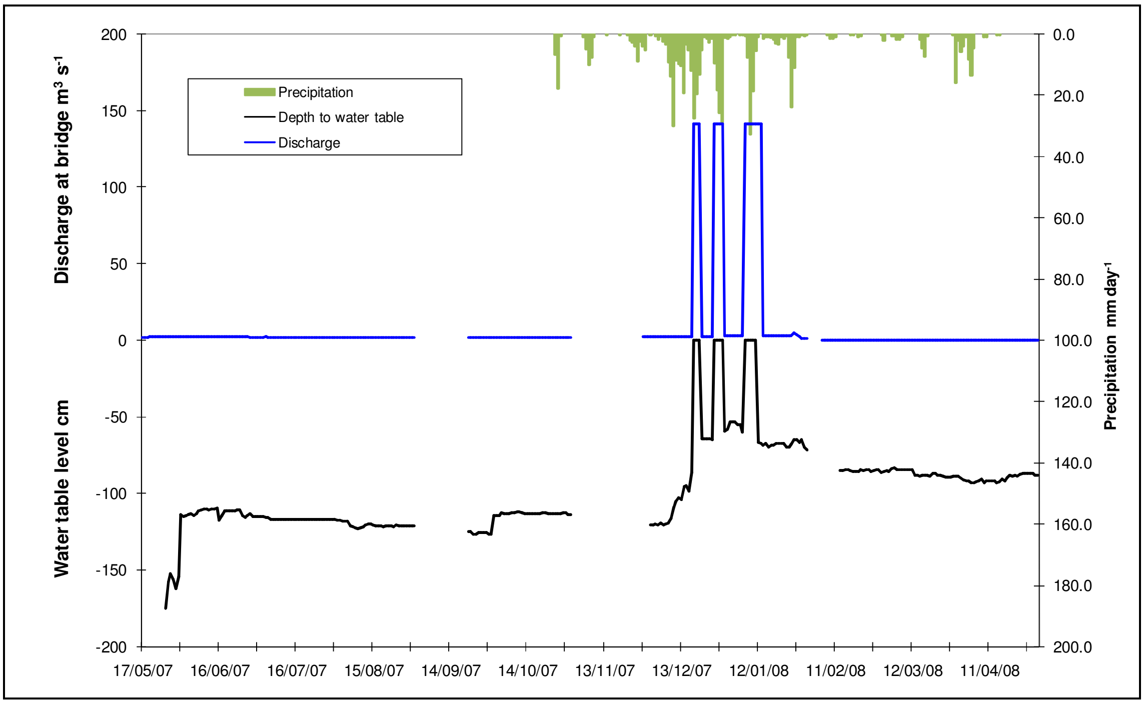

Figure 5. Observed discharge, rainfall and water table level in the study area. Note the extremely rapid and simultaneous rise in the two hydrographs during the three large flow events of late December 2007 to early January 2008. Discharge readings go to a maximum of 141.6 m^3s^{-1}, above which the limnigraph and bridge were overtopped and no measurements could be made.

The discharge hydrograph does not present conventional rise and recession behaviour. This could be explained by considering the aquifer and river as a single unit: water flowing towards the study site, whether surface flow in the river upstream or overland flow, must first saturate the aquifer before surface flow occurs (excluding the small discharge from the aquifer at the bridge). This effect can be seen in early December 2007, where the groundwater hydrograph rises – and can be seen to relate to upstream rainfall events – but the surface water hydrograph does not respond until late December 2007.

During the three peak surface flow events of 18-20 December 2007, 26-29 December 2007 and 7-13 January 2008, the surface water and groundwater hydrographs rise and fall precipitately and virtually simultaneously. It is possible that the aquifer material became fluidised during these flood flows, allowing for flows several orders of magnitude faster than the hydraulic conductivity of 26.8 m day^{-1}. Fluidisation of saturated unconsolidated media typically occurs due to an extremely rapid build up in pore-pressure (Vucetic, 1994; Wang, 2005), for example during floods (Colomer *et al*, 2002). The three floods that produced this response at Mushawe are all large in comparison to the volume of the aquifer: a minimum flow of 142 m^{3}s^{-1} on these events can be compared with the average volume of 1 m length of the aquifer at 106 m^3. The much smaller flow event of 26-27 January 2008 shows a more gradual rise and fall in both the surface water and the groundwater hydrographs.

3.2. Aquifer Water Balance

The volume of water stored in the aquifer varies from a 2007 dry season low of under 2,500 m^3 to a wet season range of 5,000 to 7,000 m^3 (figure 6). The 2008 dry season volume of around 5,000 m^3 is much higher than that for 2007, probably due to the rainy season of the year preceding the 2007 dry season being subject to an El Niño event (Logan *et al.*, 2008).

Each of the three flood events resulted in the (temporary) filling of the aquifer's full potential storage, calculated from equation (5) as 9,428 m^3. This is sufficient water to irrigate 1 ha for a year (based on 10,000 m^{3}ha^{-1}a^{-1}, Ministry of Local Government Rural and Urban Development, 1996). However, this maximum storage declines within a month to 5,000 m^3. This amount is still more than twice the current usage of around 1,825 m^{3}a^{-1}, suggesting the potential for limited expansion of the water supply for No 1 Business Centre.

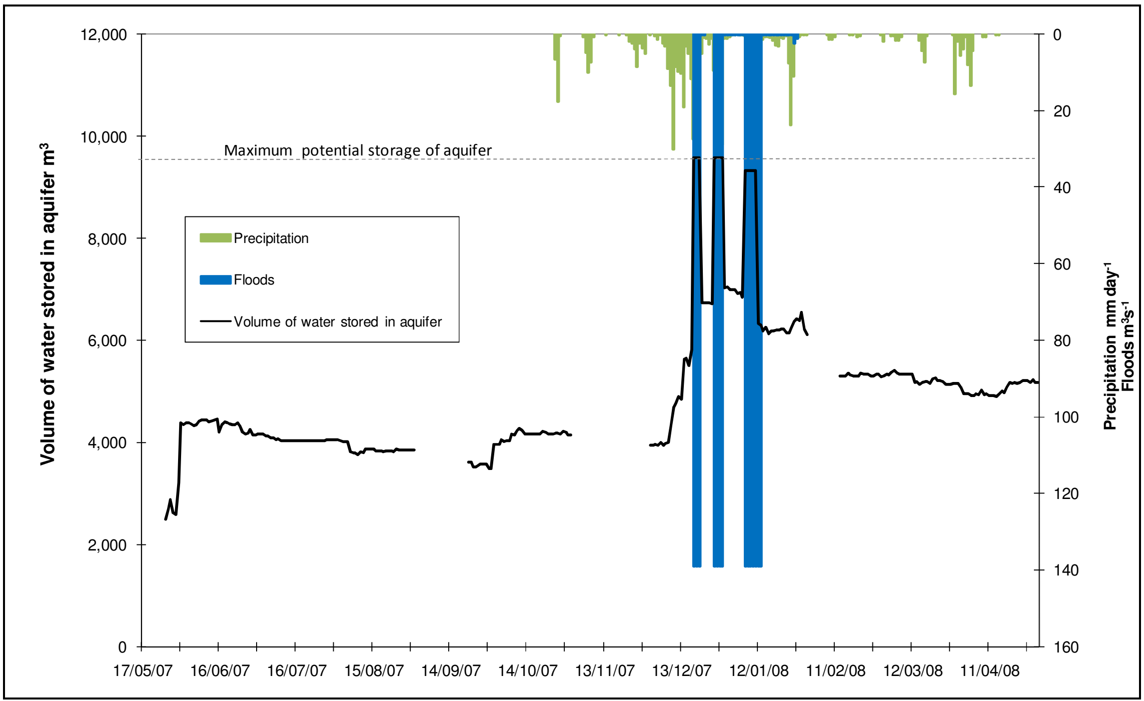

Figure 6. Variation in storage in the Mushawe alluvial aquifer, with rainfall and flood events over 2.5 m^{3}s^{-1} for reference. Note that even during a drought year (2006-7), more than 25 % of the aquifer is still saturated (2,500 m^3).

4. Conclusions

The maximum storage of the aquifer of just under 9,500 m^3 is filled during flood events. Only around half of this remains stored in the aquifer during the year, but this residual volume is nevertheless sufficient water to double the water supply to No 1 Business Centre. The aquifer water balance is based on a study site of 500 m of river reach. It should therefore be noted that abstraction immediately upstream of the study site (reach) could significantly increase the available water. Furthermore, upstream on-river storage and artificial recharge would allow for maintenance of a greater saturated aquifer volume throughout the year.

Alluvial aquifers of a similar scale with greater specific yield would store a lot more water: for example aquifers composed of coarse sands or gravels, with specific yields of 0.25 to 0.35 (Johnson, 1967), would be expected to store twice as much abstractable water as Mushawe. However, the occurrence of such aquifers at meso-scale might be limited by scale factors in erosion and sediment transport dynamics.

In conclusion, it can be said that alluvial aquifers of the scale of Mushawe are likely to be suitable for use at the scale of domestic and livestock water supply and the irrigation of small gardens (sub-hectare). Such gardens would be best suited for horticulture, to maximise the value obtained from the irrigation water. This contrasts with the alluvial aquifers on some major rivers, for example the lower Mzingwane alluvial aquifer which can supply enough water for 20 ha of irrigation from a similar 500 m of river reach (Masvopo *et al.*, this volume).

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