Estimation of groundwater recharge via percolation outputs from a rainfall / runoff model for the Verlorenvlei estuarine system, west coast, South Africa.

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Abstract

Wetlands are conservation priorities worldwide, due to their high biodiversity and productivity, but are under threat from agricultural and climate change stresses. To improve the water management practices and resource allocation in these complex systems, a modelling approach has been developed to estimate potential recharge for data poor catchments using rainfall data and basic assumptions regarding soil and aquifer properties. The Verlorenvlei estuarine lake (RAMSAR #525) on the west coast of South Africa is a data poor catchment where rainfall records have been supplemented with farmer's rainfall records. The catchment has multiple competing users. To determine the ecological reserve for the wetlands, the spatial and temporal distribution of recharge had to be well constrained using the J2000 rainfall/runoff model. The majority of rainfall occurs in the mountains (±650 mm/yr) and considerably less in the valley (±280 mm/yr). Percolation was modelled as ~3.6% of rainfall in the driest parts of the catchment, ~10% of rainfall in the moderately wet parts of the catchment and ~8.4% but up to 28.9% of rainfall in the wettest parts of the catchment. The model results are representative of rainfall and water level measurements in the catchment, and compare well with water table fluctuation technique, although

estimates are dissimilar to previous estimates within the catchment. This is most likely due to the daily timestep nature of the model, in comparison to other yearly average methods. These results go some way in understanding the fact that although most semi-arid catchments have very low yearly recharge estimates, they are still capable of sustaining high biodiversity levels. This demonstrates the importance of incorporating shorter term recharge event modeling for improving recharge estimates.

1. Introduction

Wetlands are systems that are saturated either by surface or groundwater with vegetation that has adapted to periods of saturated soil conditions. These systems are regarded as one of the most productive ecosystems on earth, providing valuable functions in filtering water, collecting sediments and retarding flow during flood events (Barbier et al., 1997; Baron et al., 2002). Due to the highly productive nature of these systems, they have also been the target of often intensive agricultural development (Schuyt, 2005), resulting in competition for water resources. The availability of water is further impacted by climate change (Fay et al., 2016) and high potential evapotranspiration (Přibáň and Ondok, 1985), which exacerbate this competition. Whilst the amount of water needed to sustain different agricultural crops is well constrained (Allen et al., 1998), less constrained is the water needed for the ecology and biodiversity profile of natural wetlands, , often termed the ecological reserve. The ecological reserve is defined by the quantity and quality of water that is required to maintain aquatic ecosystems (Hughes, 2001). These maintenance conditions are identified using ecological, geomorphological, hydraulic and hydrological knowledge of each system. Usually maintenance flow requirements are set for both peak and low flow periods, during average and low rainfall years, although the survival of wetlands is critically dependent on the degree to which the ecological reserve is met during low flow, especially during drought years. During such times, baseflow from aquifers contributes the majority of the ecological reserve, and for this reason baseflow is one of the most important parameters to constrain in a wetland catchment.

While there are many factors that influence baseflow from aquifers, the most important and variable is the rate of groundwater recharge. Various approaches can be used to estimate recharge, but essentially they can be grouped into three methods: 1) physical, for example water table fluctuation (WTF) (Crosbie et al., 2005) or channel water budget (Rantz, 1982); 2) chemical, for example chloride mass balance (Ting et al., 1998) or applied tracers (Forrer et al., 1999); and 3) numerical, for example rainfall/runoff modelling (SWAT, Arnold *et al.*, 2000) or variably saturated flow modelling (HYDRUS: Šimůnek et al., 2012). For the physical and chemical methods, some component of climate is compared to a groundwater component, for example the comparison between precipitation volume and groundwater level. This approach can also be called actual recharge, as it determines the amount of water that reaches the groundwater table (Rushton, 1997), but in doing so it neglects any processes that occur in the unsaturated zone, thereby reducing its spatial and temporal extent. However, for numerical modelling of recharge, it is not possible to neglect what is happening in the unsaturated zone, as most models require information on the physical and chemical pathways of recharge. Therefore, this type of approach is rather defined as potential recharge, which is constrained by the amount of water that has percolated through the unsaturated zone, contributing to the saturated zone (Rushton, 1997), and hence requires knowledge of the percolation rate.

Within numerical modelling, the percolation rate (Scanlon et al., 2002) can be modelled either by looking at variably saturated flow or rainfall/runoff partitioning. Both these methods use a water-balance to determine the percolation volume using input data, such as climate (rainfall, temperature), vegetation (interception) and biosphere (soil texture) to partition water into runoff, infiltration, evaporation and recharge. These two methods differ in their ability to simulate soil moisture. Variably saturated flow models can simulate vertical distributions of soil moisture and estimate recharge by routing water through the soil column using soil hydraulic conductivities. Many rainfall/runoff models partition infiltrated water into storages based on soil type parameters (J2000: Krause, 2001; and ACRU: Schulze, 1995) . This makes variably saturated flow more favourable for estimating recharge for detailed studies due to its ability to simulate soil moisture. However, for larger spatial scales, rainfall/runoff models are able to model representative recharge (Scanlon et al., 2002) and are therefore more commonly used in regional scale studies.

This study looks at evaluating how well the percolation output from a J2000 rainfall/runoff model represents actual recharge and whether this can be used as a valid recharge input to a groundwater model for a wetland catchment. The J2000 model is a distributive hydrological model that can be used to simulate various components of the hydrological cycle by calibration of parameters using streamflow, climate and rainfall data. The validation of the percolation output is done by comparison to physical rainfall and water level data in the Verlorenvlei estuarine lake, a RAMSAR Convention (#525) listed wetland on the west coast of South Africa, north of Cape Town, where the high biodiversity profile is linked to the intermittent connection between fresh and salt water. The catchment is also an important agricultural area, in particular supporting 15% of the South African potato industry (Potatoes South Africa, 2015). Despite the value of the region and lake system, the catchment is relatively data poor, partly because of a lack of operating gauging stations, and in spite of ongoing agricultural monitoring. At present, it is not sufficient to allow groundwater abstraction rates to be in equilibrium with recharge estimates, as this does not consider the requirements of the ecological reserve. Therefore, a groundwater model is needed to assess permissible abstraction rates, of which large spatial (catchment) and high temporal (daily) estimates of recharge are needed. Data poor catchments are a common feature across much of Africa, and this method may provide a mechanism for establishing sustainable groundwater management in other data scarce regions, particularly those that are also semi-arid to arid.

2. Environmental Setting

The Verlorenvlei catchment makes up the southern part of the Sandveld, a sub-region along the south-western coastline of South Africa, where the soils are particularly sandy. The catchment consists of the Piketberg Mountains in the east, which form the highest topographic elevation (1446m) and the eastern boundary of the catchment, down to Elandsbaai on the west coast. The dominant feature of the catchment is the Verlorenvlei estuarine lake, which is situated between Redelinghuis and Elandsbaai (Fig. 1), where the estuary transports semi-fresh water into the ocean (Fig. 1). The estuarine lake itself is around 15 km^2 in size, where the catchment has an area of 1832 km^2.

2.1 Hydrology

The estuarine lake is fed by four main rivers, the Kruismans, Bergvallei, Hol and Krom Antonies (Fig. 1). Previously, gauging stations existed along the Kruismans and Hol rivers, but have not been operational since 2009. There is still active water level monitoring within the estuarine lake close to Elandsbaai (Fig. 1). During dry periods, when the water level in the lake is low, stagnant and saline conditions exist, which favours the growth of large algal blooms. During the last seventeen years of monitoring, low water levels of below 0.5 m have been measured for 5 months in 2001, 9 months between 2004 and 2005, and more recently for 4 months between 2015 and 2016. (Fig 2). The likely cause of these low water levels can be attributed to changes in rainfall patterns, although agricultural abstraction has potential in reducing flow in the lake's major feeding rivers. Although no gauging stations currently exist on the Krom Antonies River, it is considered the most significant contributor of both the quantity and quality of flow into the lake, as it receives water from the Piketberg Mountains. The Kruismans River originates from the east side of the Piketberg Mountains, which drains a large, relatively flat agricultural region (Fig. 1). The river passes through a wide neck in the eastern arm of the Piketberg Mountains, and then firstly joins up with the south draining Bergvallei River, and thereafter the north draining Krom Antonies and Hol Rivers (Fig. 1). The point on the Kruismans River after these three rivers have joined is termed the confluence. Below the confluence, the river is variably referred to as the Kruismans River and the Verloren River, but essentially drains westward until the beginning of the actual lake west of Redelinghuis.

2.2 Hydrogeology

The catchment geology is comprised of three major rock units (Fig. 3). The oldest rocks in the area are the Neoproterozoic Malmesbury Group, represented by the Piketberg Formation comprised of greywacke, sercitic schist, quartzite, conglomerate and limestone (Rozendaal and Gresse, 1994). These rocks make up the secondary fractured rock aquifer (Fig. 3). These rocks have been intruded by the Cambrian Cape Granite Suite. Although drilling has indicated their presence at depth, outcrops within the catchment are very poor to non-existent. The youngest rocks in the catchment are the sedimentary rocks of the Cambrian Table Mountain Group (TMG) which overlies both the Malmesbury Group and the Cape Granite Suite. The TMG makes up the Piketberg Mountains, and in this region is dominated by three formations, which are the Peninsula, Graafwater and Piekenierskloof formations (Johnson et al., 2006). The TMG makes up an important fractured rock aquifer in the Western Cape, and the Peninsula and Piekenierskloof formations are two of the most important aquifer units. The primary aquifer, which is located in the valley of the catchment, is made up of quaternary sediments dominated by coarse-grained, clean sand and therefore is high yielding. Previous recharge estimates for the primary aquifer are between 0.2 to 3.4% of rainfall, although majority of recharge is thought to occur primarily within the high lying areas, which are dominated by the TMG aquifer (Conrad *et al*., 2004), similar to other high elevation regions in the Western Cape.

2.3 Climate and Vegetation

In the Piketberg Mountains, where the Krom Antonies originates, the mean annual precipitation is around 537 mm/yr (Lynch, 2004) (Fig. 4). Rainfall decreases moving north-west from the Piketberg Mountains, reaching a low of 210 mm/yr at the mouth of Verlorenvlei, which is around 50 m above sea level (Lynch, 2004). The west coast is subject to a Mediterranean climate, where rainfall is generated by cut-off lows and synoptic scale low-pressure systems during winter (Holloway et al., 2010). Mist and dew are also considered potential contributors to soil moisture but these are not monitored within the catchment. In summer, daily average air temperatures are between 17 and 23 ℃, with mean evaporation rates between 5.5 and 7.35 mm/day (Schulze et al., 2007)

. During winter, daily average air temperatures are between 8 and 13 ℃, with mean evaporation rates between 1.5 and 2.3 mm/day (Schulze et al., 2007). The dominant vegetation types within the study area are Strandveld and coastal Fynbos (Acocks, 1988). Strandveld is present in the western coastal plains, whereas Fynbos grows on sandy soils, which is further inland and closer to the sandstone geology. These vegetation types are adapted to low rainfall environment; therefore, direct soil evaporation is likely to be more important than transpiration although these are currently not well constrained within this catchment.

2.4 Landuse

Agriculture in the Sandveld is the major water user in the area, accounting for 90% of the total water requirements. Potatoes are the main food crop grown, accounting for over 6600 hectares and using around 20% of total recharge (DWAF, 2003). Potatoes in the Sandveld are usually grown in sandy soils, resulting in high yields, but require large amounts of water and fertilisers to grow successfully. Tea is the second most grown crop in the catchment, making up around 5000 hectares, although water is only used during processing. Tea is also planted in sandy soils and is generally rainfed, therefore having limited impact on groundwater resources. Other high water-use agricultural activities include citrus and viticulture. Natural vegetation is also used for livestock grazing.

3. Methodology

3.1 Data Collection Methods

Within the catchment, climate and water level fluctuations within the primary and secondary aquifer were monitored with the installation of weather stations and borehole and piezometer level loggers (Fig. 1). These instruments were positioned throughout the catchment to understand groundwater responses to rainfall, and to validate the potential recharge outputs from the J2000 rainfall/runoff model. During this study rainfall and water level responses were monitored in boreholes between January and December 2016.

3.1.1 Climate and rainfall

Rainfall, windspeed, relative humidity, solar radiation and air temperature were measured by automatic weather stations (AWS) within, and outside the study catchment. These measurements were used as inputs into the Penman Monteith equation to estimate daily reference evaporation for the J2000 model. Climate data was collected from six stations (Fig. 1) of which four (Redelinghuys, Lambertsbaai Nortier (NC), Cape Columbine (CC) and Elandsbaai) are managed by the South African Weather Service (SAWS), and the other three (SV-AWS, Riviera, Piketberg) are managed by the Agriculture Research Council (ARC). The stations located within the study catchment are Redelinghuys, SV-AWS, Piketberg and Elandsbaai (Fig. 1). AWS data was screened to detect any data flags (such as anomalous or negative readings), missing records or short monitoring periods. Two new stations were installed in the catchment (Fig. 1), an Adcon Telemetry system (C-AWS) at the confluence between the Hol, Krom Antonies and Kruismans rivers at an elevation of 209 m, and a Mike Cotton Systems (M-AWS) at the foot of the Piketberg Mountains at an elevation of 237 m. On both systems, rainfall measurements have an accuracy of ±0.2 mm, temperature is ±0.5°C at 20°C and humidity is ±1-3% between 0 and 90% and 3-5% between 90 and 100% humidity. The confluence weather station (C-AWS) was installed to monitor the driest area, while the mountain weather station (M-AWS) was to monitor the wettest accessible area. Both weather stations used telemetry, which allowed for near real-time readings and troubleshooting.

Due to the limited AWS coverage and therefore limited rainfall measurements within the catchment, rainfall records were collected from nearby farmers to increase the network coverage (Fig. 1). The farm rainfall records used were those that were measured continuously, and where the rain gauges were located away from trees or other infrastructure. Record SD-R is on the Hol River beneath the Piketberg Mountains and so has a similar setting to record M-AWS. Record KK-R is in the middle of the Krom Antonies drainage, a sub-section of the catchment. Record FF-R is actually from outside of the catchment but is the only rainfall record from the top of the Piketberg Mountains and shows significantly higher rainfall than any other rainfall station. Daily rainfall was recorded at 8am in the morning, measuring rain that had fallen in the previous 24 hours. The rainfall records of the farmers were validated by comparison to the AWS data. The rainfall measurements from VL-R, which is approximately 400 m from C-AWS, agreed with the record from the C-AWS to within ± 8mm. Climate and rainfall records presented are from 1 January to the 31 December 2016, although M-AWS only started on the 1st of March. Farmers records were used to assess how dry 2016 was in comparison to previous years.

3.1.2 Groundwater Levels

In this study, shallow groundwater is defined as water that is held in the primary aquifer within the Quaternary sediments (Fig 3, B1). The depth of the shallow groundwater was monitored in 26 piezometers that were installed into the banks of the Krom Antonies, Hol and Kruismans rivers between 1 and 2 meters from the edge of each river (Fig. 1). The piezometers were screened near the bottom to allow for lateral water flow, and a geotextile filter was used to reduce sediment build up. Where it was necessary, clay was used to seal the casing from above. Caps were fitted to the tops of all the piezometers, although only four piezometers, one for each stream, were selected for continuous water level monitoring. Water levels were monitored using Heron levelogger Nano 10 m pressure transducers, which have an accuracy of ±5 mm for water level and ± 0.5 °C for the temperature. These sensors were installed at the maximum possible depth in each piezometer, to allow for the longest measurement period, as it was expected that in the dry season the water level would drop below the piezometer. The installed piezometer depth varied between 2.5 and 3 m, due to presence of an impervious clay layer. Primary aquifer piezometers were monitored from 1 January to the 31 December 2016.

Groundwater within the secondary aquifer of the catchment (Fig 3, B2) was monitored at six existing boreholes (Fig. 1). EC profiling in these boreholes suggests that they are screened below 15 m, but borehole installation records are not available. Only boreholes that did not contain pumps were used for these installations. Water level fluctuations were measured with Heron Levelogger Nano pressure transducers, which have an accuracy of 0.05% FS and ± 0.5 ℃ (where FS is defined as the maximum water level fluctuation range). Because of this, the maximum drawdown in each borehole was determined and matched to an appropriate depth range (10 m, 30 m and 60 m FS). Water levels from transducers in both piezometers and boreholes were pressure compensated using weather stations that were no more than 20 km from any of the monitoring points. Water levels in secondary aquifer boreholes were monitored from 1 January to the 31 December 2016, although sensor failure (KKB03), incorrect sensor positioning (NFB05) and sensor removal (KVB06) reduced record length.

3.1.3 Water Table Fluctuation (WTF) method

The WTF method is one of the most common and simplest methods that can be used to calculate net recharge from shallow unconfined aquifers (Healy and Cook, 2002). The main assumption in the method is that the rise in groundwater level in an unconfined aquifer is due to recharge water arriving at the water table and can be expressed as:

$$R = \Delta h \times Sy \quad (1)$$

where Sy is specific yield and Δh is the change in water table height.. Mechanisms that can influence water table fluctuations are: 1) near surface evapotranspiration; 2) changes in atmospheric pressure which can be overcome using vented pressure transducers or by atmospheric correction of pressure transducers; and 3) entrapped air between the wetting front and the water table caused by a saturated soil surface which is impervious to air; 4) pumping from nearby wells 5) natural or induced changes in surface water elevation; and 6) oceanic tides (Healy and Cook, 2002). The WTF method requires the identification of water table rises that are solely attributed to precipitation to estimate recharge (Healy and Cook, 2002) but with aquifers that are hydraulically connected to streams this can be difficult (e.g Brookfield et al., 2017). The removal of river response functions (RRF) (Spane and Mackley, 2011) using multiple regressions allows streamflow responses to be filtered out, although accurate streamflow records are required to do this. Within fractured rock aquifers with low porosities, water level responses to recharge are typically very large (e.g. Bidaux and Drogue, 1993) and while these responses can be measured, determining the specific yield is difficult. Consequently the WTF method is difficult to apply to this aquifer type.

3.1.4 Soil Types

Nine different soil types have been identified within the catchment and include Arensols, Leptosols, Solonetzs, Fluvisols, Planosols, Regosols, Lixisols, Cambisols, and Luvisols (Batjies et al., 2012). These largely reflect poorly formed, young soils, which are variably saline and are, or were, occasionally water logged. The Harmonized World Soil Database v1.2 (HWSD) (Batjes et al., 2012) was used to extract soil type information, including water storage capacity, average soil depth, depth of each horizon, texture and granulometry, which was then fed into the J2000 model (Table 1). For each soil type, two horizons were defined at a depth of 300 and 700 mm, where the proportion of sand to silt to clay in each was set. This allowed for groupings based on water holding capacity, which is necessary for defining the properties of medium pore storage (MPS) and large pore storage (LPS). MPS and LPS essentially represent two types of soil structure that differ in their pore size where LPS has a larger pore size than MPS.

3.2 Percolation Model Setup

Percolation modelling was conducted using the JAMS/J2000 hydrological modelling package (Krause, 2001). The processes that have the largest impact on modelled percolation, and therefore included in this study, are interception, infiltration, evapotranspiration, soil-water storage, and lateral water transport (Fig. 5). The model involves three main steps: (1) allocate how much rainfall goes to interception and how much to infiltration, based on vegetation cover types and rainfall patterns; (2) of the rainfall that infiltrates, allocate how much is lost to evapotranspiration, how much is lost to surface runoff, and how much actually infiltrates further; and (3) of the amount that actually infiltrates further, assign how much contributes to interflow into the river system, and how much becomes modelled recharge calculated as percolation into the aquifer. In this study, percolation rate is calculated per hydrological response unit (HRU: Flügel, 1995).

3.2.1 Definition and setup of HRUs

A HRU is an area with homogenous physiological and topographical features, used for distributive hydrological modeling in the J2000 modelling system. The SRTM-DGM (90 m) was used as the input Digital Elevation Model, where data gaps were filled using the standard fill algorithm from ArcInfo (Jenson and Domingue, 1988) after which flow direction, flow accumulation, slope, aspect, solar radiation index, mass balance index, and topographic wetness index were derived. HRU's were thereafter delineated using an AML (ArcMarkupLanguage) based automated tool (Pfennig et al., 2009). Finally, each HRU is assigned a file containing model parameters for each dominant soil, land use and geology class, and these remain constant throughout the modelling period (Flügel, 1995). The number of recommended HRUs is between 13-14 HRUs/km^2 (Pfannschmidt, 2008). However, the AML tool delineated 7008 HRUs within the modelled catchment giving a ratio of ~ 4 HRUs/km^2. As flow paths rely on slope, the HRU delineation tool increases the number of HRU's across uniform topography and decreases the number of HRUs in areas of high topography such as the Verlorenvlei catchment.

3.2.2 Assignment of HRU Climate Properties

The J2000 modelling system uses the inverse distance weighting (IDW) method for the regionalization of the input climate data, which is derived from the climate stations. Due to the scarce network of the climate stations within the catchment, and the significant differences in rainfall between the valley and the mountains, two farmers' rainfall records, FF-R and KK-R, were included in the study. FF-R was particularly important as it is at the highest elevation, which allowed for more representative estimations, due to better corrected rainfall in higher relief HRUs. Rainfall data was regionalised by defining *n* weather records available (in this case eight) and estimating the influence of each on the rainfall amount for each HRU by assigning a weighting *(Wi)* to each rainfall record using Eqn 2:

$$W(i) = \frac{\left(\frac{\sum_{i=1}^{n} wDist(i)}{wDist(i)}\right)}{\sum_{i=1}^{n}\left(\frac{\sum_{i=1}^{n} wDist(i)}{wDist(i)}\right)} \qquad 2$$

where *W(i)* is the weight of each weather station and *Dist(i)* is the distance of each weather station to the area of interest. In the case of data that is impacted by elevation such as rainfall, an elevation correction is carried out by examining the correlation between rainfall amount and elevation. The regression line created between the elevation and rainfall correlation should have a r^2 value greater than a specified limit, which in this study was set as 0.75. The calculation is then made according to Eqn 3:

$$MV_C = \sum_{i=1}^{n}\left(\left(\Delta H(i) * b_H + MV(i)\right) * W\ (i)\right) \qquad 3$$

where MV_C is the corrected rainfall value, $\Delta H(i)$ is the elevation difference between the station (i) and the HRU, b_H is the slope of the regression line and *MV(i)* is the measured rainfall value.

3.2.3 Setting of Interception vs Infiltration Amounts

The J2000 model makes use of land use classes to determine the influence that vegetation has on the water balance. These classes are defined according to wetlands, waterbodies, cultivated (temporary/permanent, commercial, dryland/irrigated), shrub land and low Fynbos (thicket, bushveld, bush clumps, high Fynbos). The model calculates throughfall by reducing net rainfall by the vegetational interception capacity (Krause, 2001).The interception module uses a simple storage approach, which calculates a maximum interception storage capacity based on the Leaf Area Index (LAI) of the particular land use class. Seasonal changes have an impact on vegetation LAI and therefore, the model incorporates variations in LAI based on season. When the maximum interception storage is reached, the surplus is passed as throughfall to the soil module. Interception storage is exclusively emptied by evapotranspiration. The maximum interception capacity (Int_{max}) is calculated according to Eqn.4:

$$Int_{max} = \alpha * LAI \quad (4)$$

where α is the storage capacity per m² and set to 0.1 mm based on previous work in the region (Steudel et al., 2015), and LAI is set for the season of the land use class.

3.2.4 Proportioning of Water into Different Soil Components

Throughfall is then passed onto the soil module, where the amount that infiltrates is calculated and the remainder is lost to surface runoff (Krause, 2001). The amount of infiltrated water is empirically determined by the model, using the maximum soil infiltration rate and the relative soil saturation deficit. The relative soil saturation deficit is determined using a relationship between the actual MPS to LPS, the maximum MPS to LPS and their water storage capacity. The water storage capacity for MPS and LPS was determined using the Rosetta, HYDRUS 1-D model (Šimůnek et al., 2006) incorporating soil textures from the HWSD. A pedotransfer function was applied to three hypothetical pressure scenarios namely: 0 mbar, 60 mbar and 15000 mbar. The storage capacity of MPS, water held at field capacity, was calculated by the difference in water content between 60 mbar and 15000 mbar, while LPS, which is water held against gravity, was calculated by the difference in water content between 0 and 60 mbar.

Within the J2000 model, the maximum soil infiltration rate is set for different seasons, where during dry conditions the maximum soil infiltration rate is higher than in wet conditions. The maximum infiltration rate of the soil was set as 100 mm/day during the dry season and 40 mm/day during the wet season, based on previous models constructed in the area (Steudel et al., 2015). If throughfall exceeds this maximum rate, the surplus water is fed to the depression storage. Depression storage is the ability of an area to retain water in pits and depressions, and once the depression storage capacity is exceeded, horizontal overland flow is simulated. Infiltrated water is then subdivided into MPS and LPS. Water can move from MPS to LPS, based on the saturation deficit of MPS where the remaining water is routed to LPS. Water can also move from LPS to MPS via diffusion. The total routed to LPS, calculated as a function of the relative soil saturation and the actual storage capacity, is then divided between percolation and interflow based on the slope. The slope weight is calculated using Eqn 5, based on the actual slope determined from the DEM and a user specified calibration factor *soilLatVertDist*, which represents the distribution of the LPS outflow between lateral (interflow) and vertical (percolation) components:

$$Slope_W = \left(1 - \tan\left(slope * \frac{\pi}{180}\right)\right) * soilLatVertDist \qquad 5$$

where $Slope_W$ is the slope weight and s*oilLATVertDist* is set as 0.7, based on the results of multiple simulations.

3.2.5 Separation of Percolation from Interflow

The amount of water that is available for actual percolation is then calculated according to Eqn 6:

$$Percolation = (1 - slope_W) * SoilOutLPS \qquad 6$$

where *SoilOutLPS* is the calibration factor for the definition of LPS outflow (values range from 0-10) (Nepal, 2012). During this study, the *SoilOutLPS* calibration factor was determined using the Kruismans gauging station that was operational from 1970-2009 and estimated as 0.2. This low value implies that most of the water that infiltrates is rather lost to evapotranspiration rather than contributing to recharge. However, the actual percolation rate cannot exceed a maximum percolation rate (vertical hydraulic conductivity), the value for which is specified by the user. Maximum percolation was estimated by analysis of groundwater level fluctuations in two boreholes in the secondary aquifer, which were not impacted by drawdown from nearby pumping, WDB03 and KVB06 (Fig. 1). While recharge in these borehole is likely received via groundwater flow from the TMG, they are not affetced by streamflow fluctuation, thereby providing the only means of estimating daily maximum soil percolation. For WDB03 the average daily fluctuation was 2.3 mm and the median 1.1 mm, whilst for KVB06 the average daily fluctuation was 2.9 mm and the median 2.1 mm. Based on this data, 2mm/day was used as the maximum soil percolation rate. If this rate is exceeded, the extra water is fed to interflow. Potential percolation is therefore the sum of actual percolation (percolation simulated by the model) and interflow.

3.3 Model Calibration and Sensitivity Analysis

During model calibration, the aim is to reduce the difference between simulated and measured dependent variables at each time step by modifying the model parameters, to predict the best measured outflow level. To ensure both quantitative and objective estimates of results during model calibration, a validation was used after each model run for both relative and absolute quality criteria (Wheater et al., 2007). As part of the model calibration, a sensitivity analysis (Fig. 6) is used to determine how sensitive estimated input values for different parameters are, with regard to the outputs (Krause et al., 2006; Nepal, 2012). The fully distributed HRU based JAMS/J2000 model was applied to a number of semi-arid catchments, as well as the nearby Berg River catchment (Steudel et al., 2015).

3.3.1 Model calibration and parameter estimations

In this study, calibration was completed by comparison of model outputs to gauging data from the Kruismans sub-catchment, using station G3H001 with records from 1989-2006. The model calibration was split into three periods: 1989-1991 for model initialisation, 1992-1998 for calibration and 1999-2006 for validation (for testing calibration parameter values). Thereafter the calibration parameters were used for modelling between 2013-2016 where a two-year initialisation (2013-2014) was incorporated. Before the automated calibration was conducted, the initial parameterization of the J2000 model was carried out by adapting and transferring model parameter values from the neighbouring Berg River catchment (Steudel et al., 2015). These parameter values were then integrated into the automated optimization tool, OPTAS (Fischer, 2013), which identifies optimal parameter value sets based on multi-criteria analysis (MCA) (Table 2). The automatic calibration makes use of the Nash-Sutcliffe efficiency and the Index of Agreement to describe efficiencies. The Nash-Sutcliffe efficiency (e2) considers variability of the measured outflow, and integrates the sum of the difference squared between measured and modelled outflow, taking into account peak outflow squared residuals (Nash and Sutcliffe, 1970; Pfannschmidt, 2008). For low flow, a modification of the Nash-Sutcliff efficiency, which incorporates unsquared residuals (e1), is used (Pfannschmidt, 2008). Higher e1 and e2 values suggest a better correspondence between observed and modelled discharge. The Index of Agreement (Willmott, 1981), was used to relate the ratio of the mean square error to potential error. This form of criteria for standardized square error is used for estimating the temporal representation of modelled runoff (Giertz et al., 2006). This MCA not only considers the effect of a single parameter on the quality of the output, but also the combined effect of all the parameters on the model.

3.3.2 Parameter sensitivity

The objective of a sensitivity analyses is to determine the influence that various independent variables have on a specific dependent variable, based on a given set of assumptions (Nepal, 2012). Sensitivity analysis can be conducted during construction, calibration and verification of a model (McCuen, 1973), using a variety of different techiques. In this study a Regional Sensitivity Analysis (RSA), also called Monte Carlo filtering (Hornberger and Spear, 1981), was used. RSA aims at identifying regions of input variability that produce extreme output values (Pianosi et al., 2016). During typical RSAs, model parameters are split up into behavioural (good) and non-behavioural (bad) populations depending on whether the variables behave as expected based on the model setup (Pianosi et al., 2016). However, this study an objective function, which makes use of observations against model accuracy, was used. During this type of RSA, splitting criteria are based on the minimum model performance requirements (Pianosi et al., 2016), where given thresholds were taken from previous studies (Nepal, 2012; Steudel et al., 2015).

4 Results

4.1 Monitoring Results

4.1.1 Rainfall Patterns

Rainfall was measured at monitoring locations within the catchment between May and October 2016. Records from C-AWS and VL-R have yearly totals of 252.2 and 260 mm respectively, representing the lowest rainfall recorded in the catchment for 2016 (Fig. 7a and Fig. 7b). The largest rainfall event measured at C-AWS was 54 mm on the July 14, while 40 mm was recorded at VL-R for the same day. Average daily rainfall for C-AWS was 0.64 mm/day, while VL-R was 0.75 mm/day. Of the last five years that were measured, 2015 and 2016 were the two driest years for VL-R (Table 3).

SV-AWS received 292.2 mm rainfall for 2016 (Fig. 7c), which was slightly higher than C-AWS and VL-R. The largest rainfall event measured at SV-AWS was 61.7 mm on the July 14, which is slightly more than C-AWS and VL-R for the same event. The average daily rainfall for SV-AWS was 0.77 mm/day. KK-R received 356 mm of rainfall in 2016, which was higher than C-AWS, VLR and SV-AWS. Rainfall records for KK-R date back to 1965, where in the last 12 years 2015 and 2016 are the two driest consecutive years, although rainfall in 2003 was lower (303 mm) than both 2015 and 2016 (Table3). The largest rainfall event measured at KK-R during 2016 was 63 mm on the July 15. This appears to be the same event albeit recorded a day later than that at C-AWS, VL-R, and SV-AWS. The daily average for KK-R was 0.97 mm/day.

Precipitation gauges at SD-R and M-AWS (Moutonshoek AWS) measured rainfall at the foot of the Piketberg Mountains. SD-R, which is located near the Hol River, received slightly less rainfall (463 mm) (Fig. 7e) than M-AWS (489 mm) (Fig. 7f) which is located near the Krom Antonies River, even though M-AWS had a shortened record (2016/03/01-2016/12/31). Rainfall records for SD-R date back to 1999, and indicate that 2015 (254 mm) was the driest year recorded (Table 3). The largest event measured during 2016 at SD-R was 62 mm on the July 15, while at M-AWS 57.2 mm was recorded for the previous day. The daily average for SD-R was 1.27 mm/day, while for M-AWS it was 1.55 mm/day.

Rainfall measured at FF-R in the Piketberg Mountains (Fig. 7g) for 2016 was the highest (639 mm) in the catchment. Rainfall records for FF-R date back to 2010 and indicate that 2015 was the driest year (398 mm) (Table 3). The largest measured event during 2016 at FF-R was 70 mm for the July 14. The daily average for this location was 1.75 mm/day.

4.1.2 Primary Aquifer Groundwater Levels

VLP01, which is the piezometer monitoring sub-surface flow below the confluence, showed a steady water level of around 1.5 m below surface between January 1 to June 14, 2016. Thereafter, due to rainfall received on the June 15, the water level rose 1.5 m to above the piezometer (Fig. 8a). The water level fluctuated around this point from June 15 to September 22. Thereafter a steady drop in water level was measured, reaching a low of 1.2 m below the surface at the end of December. Water level spikes throughout the measuring period were rapid and steep.

Piezometer KRP02, which was installed on the Kruismans River, had a short monitoring length during the dry season, between January 1 to June 15, 2016, due to the water level dropping below the sensor (Fig. 8b). The water level in the piezometer rose to 0.5 m below surface on the June 15, fluctuating between 0.3 to 0.5 m until the October 24. Water level responses at this sensor were rapid, although the occurrence of responses was less frequent than in VLP01. Similarly, piezometer HOLP03 was dry from the January 1 until June 9, thereafter fluctuating from 0.9 to 0.3 m during the wet season (Fig. 8c). At this piezometer, water level responses to rainfall events were slower, where peaks were relatively small.

Piezometer KAP04 showed a steady decline in water level from January 1 until the January 26, 2016, thereafter was dry until the March 27, 2016 (Fig. 8d). Between March and December the water level rose to 0.95 m below surface, fluctuating between 0.8 and 0.6 m from April to June. On the June 15, the water level rose to 0.1 m below surface, fluctuating around 0.5 m until August. Thereafter a steady decline in the water level was observed between the August 13 and the end of December, where the water level was around 0.9 m below the surface. This location showed more rapid responses to rainfall events, which can be observed by the steep spikes in water levels (Fig. 8d).

Shallow groundwater was monitored in borehole VLB02 within the primary aquifer, near the confluence (Fig. 1). The water level in this borehole dropped from 6 to 9 m below surface from January 1 to June 14, 2016. Thereafter, the water level rose above the measured static water level of 4.82 m to 4.88 m in November, with a month rainfall lag. A steady decline in water level was observed from November until December, dropping below 5.5 m below surface.

4.1.3 Secondary Aquifer Groundwater Levels

Secondary aquifer groundwater levels were monitored in five existing boreholes none of which were actively pumped. However, three of the five monitored boreholes were close to boreholes that were pumped. These three include VLB01, KKB04 and NFB05. VLB01 was near three pumped boreholes where significant drawdown was observed. Minor water level recovery occurred when pumping ceased (pump failure) during February and March 2016. However, when pumping recommenced, the water level dropped more than 40 meters between the June 15 and November 1, in 2016 (Fig. 9a). Water level recovery was monitored between the November 1 until the November 15, rising from 60 to 25 m due to the halting of pumping. The water levels monitored at KKB04 recorded limited fluctuations until the stress of pumping was added, where the water level dropped from 26 to 30 m between the October 24 and end of December 2016 (Fig. 9b). KKB04 showed minor drawdown due to the small volume of water being abstracted. Borehole NFB05 has incomplete records, due to groundwater abstraction nearby resulting in drawdown below the sensor position from January 1 to the May 6. Thereafter, NFB05 showed minor fluctuations in water levels around 28 m, recovering to 22 m in late October (Fig. 9c).

Monitoring boreholes WDB03 and KVB06 where away from abstraction points, hence water level fluctuations were minor over the course of the monitoring period. At WDB03 minor fluctuations were recorded throughout the year, persisting at around 9 m and dropping to a low high of 8.1 m in September (Fig. 9d). A slight recovery of 0.2 m was recorded towards the end of December. KVB06 showed limited fluctuations in water levels, persisting at around 28.5 m during the monitoring period (Fig. 9e).

4.2 J2000 Modelling Results

4.2.1 Actual Percolation Results

Actual percolation simulated for 2016 within the catchment ranged from 0 to 250 mm. The highest simulated actual percolation were in the higher relief regions, dominated by the TMG aquifer, which ranged from 80 to 210 mm (Fig. 10). In the valley, which is dominated by the primary aquifer but underlain by the secondary aquifer, simulated percolation ranged from 0 to 80 mm. In the driest part of the catchment at locations C-AWS, VL-R and SV-AWS (Fig. 11a-c), yearly simulated actual percolation corresponded to 8 mm, 18 mm and 3 mm for 2016. Actual percolation was simulated from the June 20 to the September 15 at these locations. Maximum soil percolation was reached (2 mm/day) for one day on the August 3 for C-AWS and for three days between August 3-5 for VL-R. In the moderately wet regions of the catchment (KK-R), simulated actual percolation for 2016 was 40 mm (Fig.11d). Actual percolation was simulated from the June 20 to the September 9 at KK-R. Maximum soil percolation was reached for 18 days between July 23 to August 9, in 2016. In the wettest regions of the catchment (M-AWS) simulated actual percolation for 2016 was 44.5 mm (Fig. 11e). Actual percolation was simulated from the June 20 to the August 20 with maximum soil percolation being reached for 19 days between the July 22 and the August 9 at M-AWS for 2016.

4.2.2 Potential Percolation Results

Potential percolation from the J2000 model includes actual percolation and interflow, and represents the amount of water that has passed through the vadose zone and can potentially contribute to recharge. Yearly potential percolation at locations C-AWS, VL-R and SV-AWS, was 18, 20.5 and 3 mm respectively (Fig. 11a-c), where interflow contributed a total of 10, 2.5 and 0 mm for 2016. Potential percolation was simulated between the June 20 to the September 15, where a maximum interflow of 1 mm was simulated on the August 3 at location VL-R. At KK-R, 55 mm of potential percolation was simulated (Fig. 11d), where interflow contributed 15 mm for 2016. Potential percolation was simulated from the June 20 to the September 9 at KK-R, where a maximum interflow of 1.8 mm on the August 3. At M-AWS, 69 mm of potential percolation was simulated (Fig. 11e), where interflow contributed 24.5 mm for 2016. Potential percolation was simulated from the June 20 to the August 20 at M-AWS, where a maximum interflow of 2.4 mm on the August 3.

4.2.3 Potential Evaporation

Potential evaporation for 2016 at C-AWS, VL-R and SV-AWS, the driest regions in the catchment, was 1454 mm, 1466 mm and 1662 mm (Fig. 12a-c). Potential evaporation at these locations during January was 10 mm/day, decreasing to 2 mm/day for May in 2016. Thereafter, potential evaporation was 2 mm/day until September, rising to 6 mm/day at the end of December. Potential evaporation for 2016 in the moderately wet regions of the catchment at KK-R, was 1363 mm (Fig. 12d). Daily potential evaporation of 10 mm/day was simulated for January, decreasing to 2 mm/day for May in 2016. Thereafter, a potential evaporation of 2 mm/day was simulated from May until October, rising to 5 mm/day at the end of December in 2016. Potential evaporation for 2016 (Mar – Dec) in the wettest region of the catchment at M-AWS, was 942 mm (Fig. 12e). At this location, daily evaporation was 6 mm/day in March until the end of April. Thereafter, potential evaporation was 2 mm/day until September, reaching 6 mm/day at the end of December in 2016.

4.2.3 Actual Evaporation

Actual evaporation simulated within the catchment was based on the availability of soil moisture so that evaporation and transpiration can take place. At C-AWS, VL-R and SV-AWS, simulated actual evaporation was 326, 319 and 317 mm respectively for 2016 (Fig. 11a-c). At these locations, little evaporation was simulated between January and March (less than 1 mm/day). Thereafter, 2 mm/day of actual evaporation was simulated from July until the end of December in 2016. Actual evaporation at KK-R was 375 mm for 2016 (Fig. 11d). At KK-R, simulated evaporation from January until March was less than 1 mm/day, although on the April 1 and October 1, 3 mm of actual evaporation was simulated. Actual evaporation simulated at M-AWS was 321 mm for 2016 (Fig. 11e). At M-AWS, little actual evaporation was simulated (less than 1 mm/day) until August where simulated actual evaporation reached 2 mm/day, continuing until the beginning of October in 2016.

4.2.3 Model Sensitivity

The model sensitivity was assessed using an RSA with objective functions for specific variables (Fig 6). For low flow criteria (E1) SoilOutLPS, maxPercolation, MaxInfiltrationDry and α, the sensitivity analysis showed moderate sensitivity (12-16%). Model parameters MaxinfiltrationWet and SoilLatVertDist showed moderate to high sensitivity (19-25%). During peak flow criteria (E2), MaxPercolation, MaxInfilitrationWet and α showed moderate sensitivity (8-16%), while model parameters SoilOutLPS, MaxInfiltrationDry and SoilLatVertDist showing moderate to high sensitivity (18-29%).

4.3 Water Table Fluctuation Results

Monitoring within the primary aquifer showed that the aquifer is hydraulically connected to the stream system, and streamflow contributes to water table rises (Fig. 8). Most of the piezometers and boreholes into the primary aquifer show very erratic fluctuations in the water table making it difficult to separate out direct recharge from streamflow. However, borehole VLB02, which is around 100 m from river shows a steady decline in water level from 6 m to 9 m below surface in mid-June 2016 (Fig 13a), before steadily recovering to 4.82 m in October 2016. The change from decline to recovery is marked by a relatively sharp inflection point and this inflection point is mimicked in piezometers VLP01, KRP02, HOP03 and KAP04. This inflection point appears to be in sync with measured rainfall at C-AWS. The current interpretation of this pattern is that the water level rise in the piezometers and boreholes is from streamflow due to the large change in water levels within the primary aquifer as reflected in the piezometer. Although it is likely that rainfall would also have an impact on this water level rise, streamflow filtering techniques are required in order to estimate recharge via the WTF method. Although borehole, VLB02 seems un-influenced by streamflow, towards the end of October where the water level rises from 4.82 to 4.88 m, without high resolution gauging data to allow for RFF filtering, it is not certain that this rise is attributed solely to rainfall.

5. Discussion

The monitoring of rainfall and groundwater levels within a catchment are important in hydrological studies where the prime objective is estimating groundwater recharge and baseflow, as in the case here. Within the Verlorenvlei catchment, water level fluctuations within the primary unconfined and secondary confined aquifer were measured in the valley that receives lower rainfall than the high recharge mountains. Although, the boreholes are in areas that receive little recharge, they are subject to local groundwater flow that is generated from the high hydraulic gradient created by the mountains on the boundaries of the catchment. The groundwater level monitoring has shown that the primary aquifer responds directly to rainfall but that the secondary aquifer does not, suggesting that it is receiving recharge from somewhere else via a different pathway. The most logical explanation for this is that the TMG aquifer, which makes up the mountainous region of the catchment and therefore has the highest recharge potential, is recharging the secondary aquifer by groundwater flow that bypasses the primary aquifer. Below we assess how representative the data is across the catchment and use this as a basis for evaluating the validity of the recharge estimates.

5.1 Data Evaluation and Representativeness

The two most important output parameters from the J2000 percolation model are simulated rainfall and simulated evapotranspiration. To evaluate the data and its representativeness across the catchment, simulated percolation and evapotranspiration have been compared to potential percolation and potential evaporation at locations C-AWS, SV-AWS, VL-R, KK-R, and M-AWS.

5.1.1 Percolation

C-AWS, VL-R and SV-AWS are in the drier regions of the catchment, where little actual percolation was simulated: 3% of rainfall at C-AWS (Fig. 11a), 7% of rainfall at VL-R (Fig. 11b) and 1% of rainfall at SV-AWS (Fig. 11c). Although C-AWS and VL-R are near each other, and hence would be expected to generate similar percolations, they are in different HRUs and therefore corrected rainfall most likely accounts for this difference. SV-AWS is located at Redelinghuis, which is considerably closer to the coast, where higher evapotranspiration reduces the amount of simulated percolation. In the moderately wet region of the catchment, location KK-R, simulated percolation corresponded to 10% of rainfall during 2016 (Fig. 11d). In the wettest region of the catchment, simulated percolation at M-AWS corresponded to 8.4% of rainfall (Fig. 11e), although surrounding HRU's suggest that a much higher percolation of up to 28.9% of rainfall is possible. Based on these results, actual simulated percolation from the J2000 model resembles the distribution of rainfall across the catchment.

5.1.2 Evapotranspiration

The atmospheric demand for water, which was modelled as potential evaporation, was much greater than simulated evapotranspiration. Simulated evapotranspiration was: 22% of potential evaporation at both C-AWS (Fig. 12a) and VL-R (Fig. 12b) and 19% of potential evaporation at SV-AWS (Fig. 12c). Simulated evapotranspiration was 28 % of potential evaporation in the moderately wet regions of the catchment at KK-R (Fig. 12d) and 34% of potential evapotranspiration at M-AWS (Fig. 12e). Essentially the higher the simulated evapotranspiration, the less water is available for percolation. If these figures are compared to actual rainfall received at different stations in the driest parts of the catchment, simulated potential evaporation is 24.4 mm greater than rainfall. This implies that overall there is very little available for percolation, although on individual days rainfall can exceed potential evaporation. In the middle parts of the catchment which are moderately wet, simulated evapotranspiration was roughly equivalent to rainfall, while in the wettest parts of the catchment in the mountains, rainfall exceeded simulated evapotranspiration by 69.5 mm for 2016. The excess is then portioned into surface runoff, interflow and percolation.

5.1.3 Recharge Estimates

Percolation simulated using the J2000 model for rainfall/runoff modelling is water that has passed through the vadose zone into an aquifer. The model is unable to consider stacked aquifers, and thus routes water to the upper most aquifer at each location. In the mountains, this will be the TMG aquifer, whereas in the valley it will be the primary aquifer. Water level data measured in the catchment suggests that the secondary aquifer is recharged by the TMG aquifer, while the primary is likely recharged by streamflow and surface runoff that originates in the Piketberg Mountains. The majority of recharge simulated by the J2000 model occurs in the TMG aquifer, whilst considerably less recharge occurs in the primary aquifer. This is consistent with water level data in piezometers and boreholes throughout the catchment. However, the model does not consider recharge that could have occurred by streamflow into the primary aquifer, as the only recharge input that the model considers is rainfall. Within the J2000 model runoff is routed to depression storage after interception is complete, and therefore partitions runoff from infiltration as two separate processes. However, these processes are likely not independent of one another, as runoff water influences primary aquifer recharge. Although the model does not account for the influence of streamflow on recharge to the primary aquifer, during the dry season it is likely that the secondary and TMG aquifers are the only contributors of baseflow, and therefore the quantification of their recharge is the most important.

5.2 Comparison of Recharge Estimates

Previous recharge estimates made by Conrad *et al.* (2004), within the Sandveld used a GIS approach that involved assigning literature estimates of recharge percentages based on MAP across the catchment. In the J2000 method, physical measurements of rainfall from nearby stations are considered, and elevation correction factors are used to assign rainfall to each HRU. While MAP is satisfactory for large scale studies, for targeted studies in smaller catchments such as the Sandveld, these estimates do not provide enough spatial resolution. The resultant net position is that the J2000 model simulates ~30 % more recharge than Conrad et al. (2004). The timestep nature of the J2000 model is producing a higher recharge value than a yearly average approach would. This is because the net yearly total evaporation exceeds the net yearly total rainfall, but daily there will be a higher probability that rainfall may exceed evaporation during the wet season. Furthermore, the spatial resolution (cell-size) of the J2000 (~0.25-1.2 km) and Conrad et al. (2004) are different (~1.5-5 km), therefore for comparison and to produce net yearly recharge estimates, J2000 estimates need to be included in a groundwater model and calibrated using literature estimates of rock and soil hydraulic conductivity. The use of water level fluctuations measured within the catchment are another possible way of estimating recharge, via the Water Table Fluctuation (WTF) method. This method however, only works for fluctuations in the water table in shallow unconfined aquifers, where estimates of specific yield exist. Although, borehole VLB02 meets the criteria specified within the WTF method, during 2016 results showed that this borehole was influenced by streamflow and therefore would require RFF filtering if recharge is to be calculated. In the future for this catchment, RFF could be used to filter out streamflow and provide an additional measure of recharge, when gauging data becomes available.

5.3 Model Evaluation

Rainfall/runoff models have been used and validated in various studies to estimate groundwater recharge (Arnold and Allen, 1999; Hughes, 2004). While these approaches are well documented, it is important to highlight the limitations of these models. The J2000 sensitivity analysis suggests that soilLatVertDist (distribution of the LPS outflow between lateral (interflow) and vertical (percolation) components) is the most sensitive parameter based upon peak flow efficiency criteria (e2) with 28 % variation in model results (Fig. 6). With e2, maximum infiltration rate for dry conditions (19%), SoilOutLPS (calibration factor for the definition of LPS outflow) (17%), α (canopy storage) (16%) are moderately sensitive. Soil maximum percolation (8%) and the maximum infiltration rate for wet conditions (9%) have low sensitivity in e2. For e1, which emphasizes sensitivity for low flow conditions, the maximum infiltration rate for wet conditions shows the highest sensitivity (25%), with all other parameters showing moderate sensitivity (13-18%).

For rainfall/runoff models to produce reliable results, estimates of streamflow from gauging stations are traditionally used for model calibration. However, gauging stations are usually not positioned at the headwaters of the catchment area, where most of the runoff water is typically generated. The J2000 model indicates that a dense network of climate data, including the use of informal rainfall records such as farm records, can be used as a substitute for limited rainfall/runoff data from gauging stations. Records obtained at high elevations were especially important to allow the model to correct rainfall for each HRU based on elevation. Water level monitoring data can be used to determine the direction of groundwater flow, and these measurements, along with a suitable DEM, should be used to determine if there is a large influence of hydraulic gradient on waterflow. Hydraulic gradient is accounted for by the slope function when partitioning water between interflow and percolation. In this model, the slope threshold was set to 0.7 (soilLatVertDist), meaning that if exceeded, all water was directed to interflow. The initial slope threshold used in this study was lower and caused all water to be diverted to interflow. Selection of the "correct" value is largely done on the basis of multiple simulations, by selecting the value that gave the most "reasonable" result, but the definition of "reasonable" varies based on the user. The sensitivity results here suggest that the slope threshold parameter is likely to be one of the most important variables in determining recharge wherever the minimum and maximum elevation in a catchment is significantly different. Despite these issues, the model results in this study are consistent with observation data in this area and known variations in recharge rates for semi-arid regions elsewhere in the world, suggesting that the modelling approach used here could be reproduced in other similar catchments worldwide.

6. Conclusions

Recharge is one of the most important parameters to quantify for addressing sustainable groundwater usage, but groundwater recharge estimates differ widely for different calculation methods even for a particular data set and catchment. In semi-arid and arid environments in particular, these estimates appear to be too low to sustain sufficient ecosystem functioning. In this study, a different approach was taken by using a model that incorporated daily timestep estimates. In spite of the catchment being partially gauged, simulated daily rainfall, evapotranspiration and the proportioning of interflow to percolation were consistent with available climate and water level data. The most sensitive parameter in the model is the terrain slope which directly controls the proportioning between interflow and percolation. However, whilst the model would likely be transferable to other semi-arid to arid catchments, it remains to be tested as to whether the model can cope with humid climates where runoff is likely to be a more significant component. A critical component of this study was to get the densest network of rainfall data possible, where weather station data was supplemented with farmer's rainfall records to improve the modelling results. Farmer's rainfall records thus provide an important additional resource when considering data poor catchments. The daily timestep function of the model yielded a recharge estimate that is ~30% higher than previous estimates. This is because daily fluctuations, which are accounted for in the model, result in lower yearly ET, as ET potentials are lower during the wet season, although further modelling is required to determine net yearly recharge estimates. The results greatly reduce the apparent discrepancy between the very low calculated recharge rates in semi-arid catchments, and the apparent sustainability of most semi-arid catchments.

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