Using distributive surface water and groundwater modelling techniques to quantify groundwater recharge and baseflow for the Verlorenvlei estuarine system, west coast, South Africa.

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This is submitted in fulfilment of the requirements of a

DOCTOR OF PHILOSOPHY IN THE FACULTY OF EARTH SCIENCES

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Declaration

I hereby declare that the entirety of the work contained herein is my own, original work, that I am the authorship owner thereof (unless to the extent explicitly otherwise stated) and that I have not previously in its entirety or in part submitted it for obtaining any qualification. This thesis is submitted in fulfilment of a Doctor of philosophy in the department of Earth Science, Stellenbosch University.

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Abstract

For effective management of groundwater resources, recharge rates and baseflow volumes need to be quantified to determine sustainable abstraction regimes and to quantify the ecological reserve, the amount of water needed to maintain the natural environment. While a variety of methods have been used to estimate groundwater recharge, estimates vary due to method, temporal and spatial resolution used. In rainfall/runoff modelling where potential recharge is determined by calculating the amount of water that percolates through the unsaturated zone, aquifer components are usually lumped resulting in over or under estimation of recharge. In contrast, groundwater models which include distributive aquifer components are commonly setup to lump climate and surface variables, thereby neglecting seasonal and climatic variability. In this study, a combined rainfall/runoff and groundwater modelling approach was used to determine the net recharge and baseflow in the RAMSAR-listed Verlorenvlei sub-catchment on the west coast of South Africa. This sub-catchment is an important biodiversity hotspot but is also an important agricultural region, hence there is competition for water resources. To understand the water dynamics within the catchment a four-phase approach was taken to determine the baseflow and ecological reserve requirements. This involved firstly, determining the limits of the sub-catchment boundary. Although the Verlorenvlei lake is supported by the Verlorenvlei sub-catchment which is itself fed by four main tributaries (the Hol, Krom Antonies, Kruismans, and Bergvallei), previous research has indicated that only one of these tributaries (the Krom Antonies) played an important role in the delivery of fresh water to the lake system. Initially the catchment boundary was thus modelled on the Krom Antonies tributary, although the understanding gained by the delineation was applied to the entire sub-catchment. To include spatial and temporal variability in groundwater recharge estimates, a rainfall runoff model was used to determine potential recharge using regionalised climate and assumptions regarding aquifer hydraulic conductivity. The potential recharge estimates within the sub-catchment exceeded previous studies (30 % higher), with the daily timestep nature of the J2000 model (Krause, 2001) assumed to account for this difference. To determine whether aquifer hydraulic conductivity could impact groundwater recharge rates, a groundwater model (MODFLOW) was constructed for the main assumed freshwater source of the Verlorenvlei, the Krom Antonies. The groundwater model included distributive aquifer hydraulic conductivity, although the input recharge was lumped which reduced climate seasonality and daily variability. The resultant output from the groundwater model was net recharge (0.3-11.4 % of rainfall) and average baseflow (14, 000 - 19, 000 m3.d-1), with the model suggesting that the baseflow from the Krom Antonies was not enough to meet evaporation demands (90, 000 m3.d- 1) and that there must be another much larger source. To incorporate daily climatic fluctuations and seasonality in baseflow estimates, the groundwater components of the J2000 were distributed using the net recharge and aquifer hydraulic conductivity from the Krom Antonies. By distributing the groundwater components within the J2000 model, the proportion of interflow to recharge was improved allowing for comprehensive estimates of runoff and baseflow from each tributary. While the model was calibrated using streamflow measurements from the gauging structure on the Kruismans, the measurements were particularly hindered by the DT limit (discharge table) of the station (3.675 m3.s-1), which resulted in reduced confidence in modelling high flow events. To incorporate the limited resolution of the station as well as limited length, an Empirical Mode Decomposition (EMD) was applied to the runoff data and water levels measured at the sub-catchment outlet. The results of the model adaption was that the Krom Antonies was not in fact the main freshwater source, with the Bergvallei supplying the majority of groundwater (49 %) as well as a large contributor of streamflow (29%). While the Hol was initially believed to be a minor contributor, the tributary had the largest ratio of baseflow (0.56), which acted to reduce its flow variability. While the Krom Antonies and Bergvallei is comprised of highly conductive sandstones and quaternary sediments, the Hol which is mainly comprised of shales, resulted in a larger groundwater flow attenuation which reduces its susceptibility to drought and climatic variability. The results of this study highlighted that on average the streamflow (20, 500 m3.d-1) from the feeding tributaries was not able to meet the evaporative demand of the Verlorenvlei and that the lake was mainly supplied by low occurrence high flow events. With the Verlorenvlei under threat due to continued agricultural expansion, it is likely that the lake will dry up more frequently in the future, especially if flows are hampered during wet cycles, when ecosystems regenerate.

List of Publications

Peer Reviewed Journals:

Watson, A., Miller, J., de Clercq, W.P., 2017. Investigating Potential Additional Sources of Groundwater Flow into a Defined Watershed. Procedia Earth Planet. Sci. 17, 546–549. doi:10.1016/j.proeps.2016.12.138 Impact factor: 0.57

Watson, A.P., Miller, J.A., Fleischer, M., de Clercq, W.P., 2018. Estimation of groundwater recharge via percolation outputs from a rainfall/runoff model for the Verlorenvlei estuarine system, west coast, South Africa. J. Hydrol. 558, 238–254. doi:10.1016/S1532-0464(03)00032-7 Impact factor: 3.727

Watson, A.P., Miller, J.A., de Clercq, W.P., Submitted. Combining rainfall/runoff and groundwater modelling to calculate recharge and baseflow in the data scarce Verlorenvlei catchment South Africa. J. Hydrol Impact factor: 3.727

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Sigidi, N.T., Miller, J., **Watson, A**., Clarke, C.E., Butler, M., 2017. Geochemical and Isotopic Tracing of Salt Loads into the Ramsar Listed Verlorenvlei Estuarine Lake, South Africa. Procedia Earth Planet. Sci. 17, 909–912. doi:10.1016/j.proeps.2017.01.015 Impact factor: 0.57

Reports:

Miller, J.A., **Watson, A**., van Rooyen, J., Eilers, A., Nthabeliseni, T., van Gend, J., Petronio, A., Agyare-Dwomoh, Y., Harilall, Z., 2018. Using Isotope Hydrology to Understand Salination Along the West Coast of South Africa. Report No. N/A Impact factor: N/A

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1.Introduction

1.1. Context

Globally, the management and protection of water resources are centered around the provisions for domestic, industrial and agricultural sectors. While each of these sectors are reliant on water to function effectively, each sector considers different priorities and guidelines. These sectors are commonly assigned assurance of supply priorities (e.g Mirrilees et al., 1994), as well as restrictions to the maximum concentration of various elements (WHO, 2004). Water resources within a country are available from two main reserves, namely surface and groundwater. Rainfall that is collected and stored above surface constitutes the surface water reserve, while the groundwater reserve is water held below surface and is only available through abstraction. Water resources within these reserves are subject to natural and anthropogenic activities that affect their quality and quantity (de Andrade et al., 2008). The availability of water in these two reserves is influenced by rainfall and climate-based processes that control the amount of resource available. While the surface water reserves are continuously reduced by evaporation, after rainfall percolates through the vadose zone into the saturated zone, groundwater is not further subject to evaporative reductions (except shallow aquifers). While previously these two reserves have been treated as independent of one another, generally they are interconnected (Winter et al., 1998), making understanding the linkages between these systems critical for effective water management.

Climatic change has had a detrimental impact on the availability and quality of both surface and groundwater resources across many parts of the world. With climate change impacting global average temperatures (Solomon et al., 2009; Walther et al., 2002), hydrological processes such as precipitation, evaporation and surface runoff have changed as a result (Legesse et al., 2003; Middelkoop et al., 2001). Increases in precipitation can be mainly attributed to enhanced evaporation resulting in higher atmospheric moisture conditions in some regions (Trenberth, 2011), while in other regions precipitation has reduced and become more variable (Easterling et al., 2000). Changes in evaporation have impacted precipitation frequency, intensity and duration, as well as soil moisture conditions because of intense drying (Seneviratne et al., 2010). The net effect of these processes is that groundwater recharge is extremely variable both temporally and spatially, due to the percolation of precipitation being impacted by soil moisture. This is problematic because to determine sustainable groundwater abstraction volumes, water managers need to understand groundwater recharge rates, in order to ensure that groundwater abstraction does not over impact low flow conditions required for ecosystem maintenance (e.g Zhou, 2009). In order to determine sustainable groundwater abstraction rates, high spatial and temporal variations of groundwater recharge as well as aquifer properties are required.

The use of hydrological models offers solutions in terms of providing estimates of groundwater recharge and aquifer baseflow provided that enough data is available for calibration. In much of semi-arid sub-Saharan Africa, the availability of data required to build hydrological models is limited, aggravated by many gauging structures being decommissioned due to poor maintenance (e.g. Wessels and Rooseboom, 2009). Although gauging data is a problem within sub-Saharan

Africa, there is usually access to nearby climate and rainfall records. While hydrological models cannot make up for a lack of fundamental field data, climate and rainfall records can be used to understand catchment dynamics. Hydrological models differ in terms of the processes that are simulated as well as their spatial and temporal resolution, with model users selecting appropriate models based on data availability as well as climate and aquifer conditions within study catchments. The majority of models also have the capability to include Global Circulation Model (GCM) scenarios, and hence water managers can use them to prepare for drought and floods in the future.

Critical for drought relief strategies is to investigate both the impact that a reduction in rainfall (e.g Jiménez-Martínez et al., 2009) as well as over abstraction pose to groundwater resources and low flow conditions (e.g Parkin et al., 2007). Hydrological models that simulate both surface water and groundwater components are required to assess the quantity of groundwater that can be exploited, without seriously impacting local groundwater resources and low flow conditions. While a semi-distributed coupling approach (e.g SWAT-MODFLOW) has previously been used for humid conditions (Kim et al., 2008), a fully distributed coupling approach for semi-arid environments is required. The J2000 rainfall/runoff model (Krause, 2001) which is a fully distributed hydrological model has previously been used in semi-arid conditions (e.g Bugan, 2014; Schulz et al., 2013). To be fully distributed, the J2000 model requires validation with groundwater models to estimate baseflow (low flows) from aquifers. MODFLOW (Harbaugh et al., 2000), which is a finite difference groundwater model, has the potential to distribute groundwater components using site specific aquifer properties and observed aquifer water levels. The coupling of the J2000 and the MODFLOW model, can be used to help water managers decide whether to use groundwater or whether more expensive adaption strategies, such as sea-water desalination are required for drought relief strategies in semi-arid environments.

Between 1999-2009, Melbourne, Australia, home to around 4.3 million people, was subjected to one of its worst droughts in recorded history. To counteract the reductions in rainfall, the city of Melbourne implemented water-wise initiatives at a variety of levels which cut water demand in half, as well as investing in alternative supplies such as desalination. During the drought, reservoir levels dropped to an all-time low of 25.6 %, before the drought eased (Patterson, 2015). Australian climate change experts suggested that the front that brings rainfall in from the southern-ocean has shifted by a degree in latitude, thereby reducing rainfall by 15 % across the south-east of Australia (Steffen et al., 2018). Similarly, between 2011 to 2017 the state of California (USA), suffered from the effects of an intense drought that lead to large-scale vegetation destruction (Anderegg et al., 2012) as well as severe declines in fish populations (Sydeman and Thompson, 2014). Reductions in rainfall were suggested to be a consequence of natural variability, namely through the impact of El Ninõ cycles, although human-induced global warming intensified temperatures, making the drought more severe (Seager et al., 2014). Like Melbourne, the government imposed mandatory water restrictions across the state to curtail supply demand (Bernstein, 2015), with the government relying heavily on groundwater to meet demand (Elmore et al., 2006). While the agricultural sector accounted for 80% of the water demand across California (Rothausen and Conway, 2011), restrictions were not imposed due to the economic impacts that would occur if this sector was affected.

The Western Cape recently experienced a severe drought with dam levels reaching a low of 24.7 % in 2017. The drought had serious implications for agricultural activities within the region, with a net loss of around R5.9 billion expected for 2017/2018 (Pienaar, 2018) and job losses of around 30 000 positions. Climate change predictions for the Western Cape suggest that this situation may become the norm, with reduced rainfall subject to increased variability (Lumsden et al., 2009). To deal with the extended periods of low rainfall volumes, the City of Cape Town implemented water-wise initiatives, as well as investing in sea-water desalination endeavours. Like California, agriculture in the Western Cape makes up a large portion of water demand, accounting for 43 % of the total requirements (WWF, 2018). In response to the drought, farms in the Western Cape were forced to used 60 % less water with many farmers opting to remove unprofitable crops. However, unlike the Australian government, the Western Cape Government promoted the use of groundwater reserves to curtail supply deficits while surface water storage schemes recover. Groundwater which was augmented as part of the relief strategy was mainly from the hinterlands of the City of Cape Town, in agricultural regions that are reliant on groundwater during summer. Regions that have been particularly reliant on groundwater for extended periods offer the potential to understand and foresee the impact that the City of Cape Town's drought relief strategy may have on the groundwater reserve sustainability.

The Verlorenvlei is a RAMSAR listed estuarine system located in the Western Cape, where agriculture has been reliant on groundwater for continued expansion. The semi-arid sub-catchment is of particular ecological importance in terms of high biodiversity profiles, which offers the potential to understand the impact that groundwater abstraction can have on low flow conditions, which ecological communities rely on. While understanding the flow regime dynamics of this sub-catchment is important for the preservation of the Verlorenvlei, understanding the potential impact that over abstraction can have on low flow conditions is of importance when considering groundwater as an emergency supply source for the City of Cape Town's drought relief strategy. In this study the J2000 rainfall/runoff model and the MODFLOW groundwater model were used to estimate groundwater recharge and aquifer baseflow for the Verlorenvlei sub-catchment. The modelling approach made use of model parameters that were estimated for a gauged adjacent sub-catchment as well as bulk aquifer properties due to data scarcity. As the sub-catchment has only historical streamflow data, an empirical mode decomposition was used at the sub-catchment outlet to improve the modelled results. The results from this study are relevant to other agricultural regions sitting in the fringes of large urban centres, where competition for groundwater resources to mitigate water supply problems will become more intense as well as climate change acting to reduce groundwater recharge rates.

1.2. Problem statement

In the Western Cape, the reduced availability of surface water resources combined with the need for higher food production, has resulted in increased groundwater abstraction. Effective methods of determining sustainable groundwater abstraction regimes are required that incorporate high spatial and temporal variations of groundwater recharge. Knowledge of the high spatial and temporal variations in groundwater recharge rates is necessary to determine the amount of groundwater resources available for exploitation (Scanlon et al., 2002). This is particularly important because it is now widely accepted that groundwater abstraction rates can no longer equal recharge rates (Alley and Leake, 2004; Sophocleous, 2000), because this approach makes limited provisions for the natural environment and the ecological reserve (Zhou, 2009).

In order to determine spatial and temporal variations in recharge rates, there are several key pieces of information required. Firstly, the boundaries of the catchment or sub-catchment must be defined in order to balance the recharge against the groundwater flow dynamics. Once the catchment boundaries have been determined, the amount of precipitation received in different parts of the catchment must be quantified. Because precipitation is derived from weather stations which are not ubiquitous across a catchment, the weather records (precipitation) must be regionalised across a catchment to account for variations in elevation. Once regionalised precipitation is determined, the amount of precipitation that enters the unsaturated zone must be quantified by accounting for losses due to evapotranspiration and surface runoff. This amount is referred to as the potential recharge as it is the amount of precipitation that has the potential to percolate into the aquifer system. The amount of potential recharge that actually reaches the aquifer system is a function of the hydraulic conductivity of the aquifer and losses due to interflow. Sustainable aquifer recharge allows for unattenuated baseflow generation which in term supports and maintains the ecological reserve to a catchment system. The various steps outlined above require detailed knowledge of the catchment system and detailed records of weather, groundwater recharge rates and groundwater abstraction volumes. If these criteria are met, then net recharge, daily baseflow and the ecological reserve requirements can be defined.

However, in many catchments, particularly in arid and semi-arid environments in sub-Saharan Africa, this information is poorly known or not available any longer. In this context, net recharge, daily baseflow and the ecological reserve are difficult to quantify. However, these types of arid and semi-arid environments are often those that are most threatened by climate change and hence most in need of the predictive capabilities of hydrological modelling. In order to address the data limitation problems associated with these catchments, detailed hydrological and hydrogeological models can be developed to simulate recharge dynamics and allow for estimation of baseflow and the reserve requirements, using historical data. Our ability to do this is limited by the distributive vs lumping properties of most models. Distributive components are where the physical characteristics that impact the water balance are constrained either spatially/temporally or both. In contrast, components are lumped due to limited data and such a single property value is representative of the entire catchment. Typically, rainfall runoff models will contain distributed information regarding the surface water processes, but commonly lump aquifer properties (Fig. 1a,b). In contrast, groundwater models accommodate distributed aquifer properties but are commonly setup to lump surface water properties (Fig. 1c). To accurately constrain daily baseflow, needed to determine ecological reserves, both surface water and groundwater processes must be distributed. This requires "coupling" of surface water models with groundwater models, where "coupling" can take several different forms. In some cases, coupling can refer to the physical joining of rainfall/runoff models with groundwater models. In others, it means the linking of groundwater model results into a rainfall/runoff model. However, the coding for rainfall/runoff models can also be adapted to incorporate distributive groundwater components. The approach taken will depend on the data availability in the catchment, the types of models employed and the technical skills of the modellers.

In this study, this problem has been examined by construction of a J2000 rainfall/runoff model (Krause, 2001) and a groundwater model MODFLOW (Harbaugh et al., 2000) for a particular study catchment. The selected study site, the Verlorenvlei sub-catchment on the west coast of South Africa, is data poor and in particular has very limited gauging data. Weather records were regionalised to determine potential recharge using the J2000 model and converted to net recharge using the MODFLOW model. Aquifer hydraulic properties from MODFLOW were then distributed in the J2000 model such that the net recharge was equivalent to J2000 model recharge. This allowed assessment of the river regime flow dynamics, and thereafter the sustainability of the Verlorenvlei lake system, by incorporation of assumptions regarding the average daily agricultural abstraction volumes.

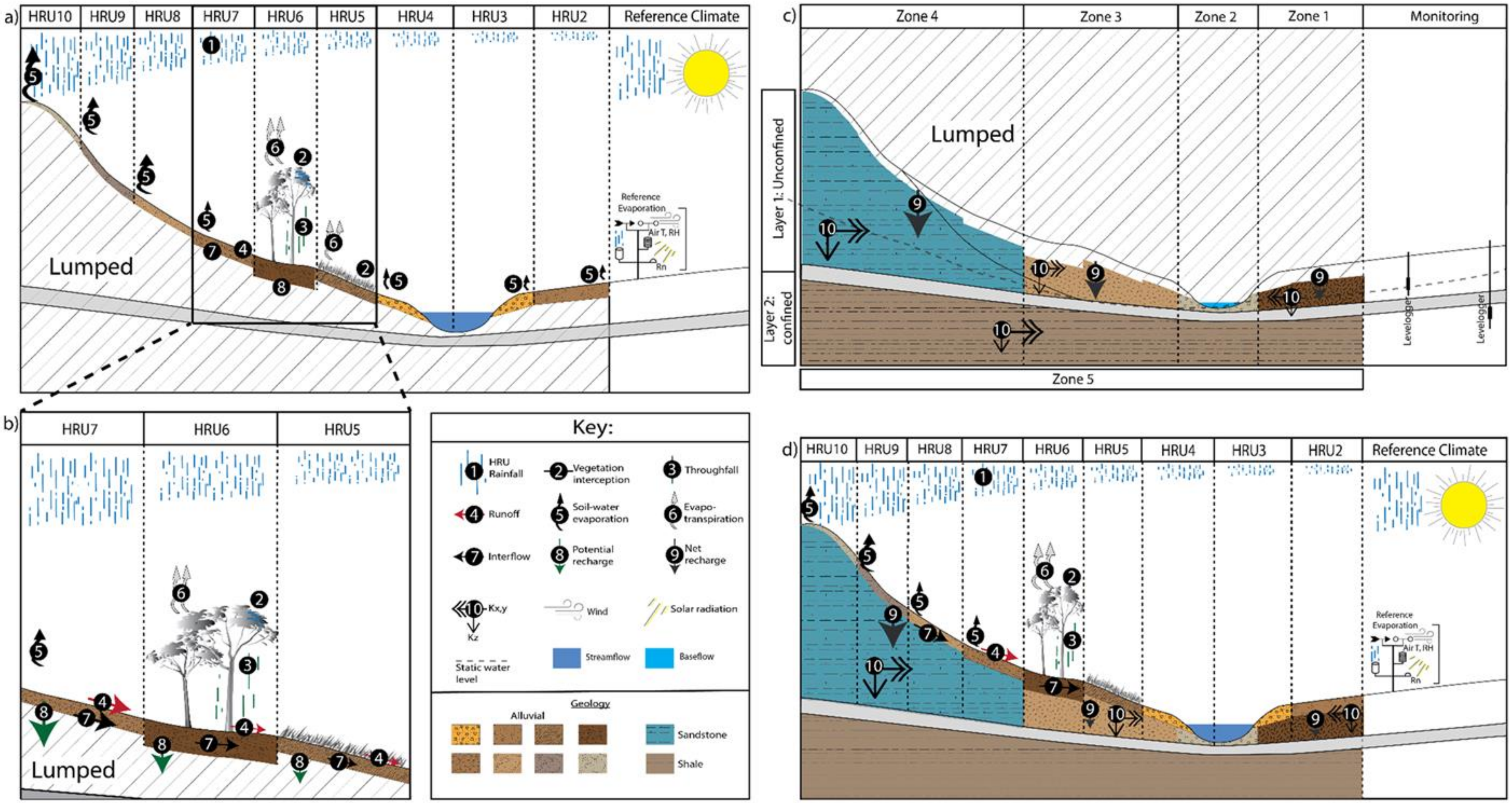

Fig 1. The layout of processes included in the a, b) daily rainfall/runoff modelling of potential recharge, c) groundwater modelling of the net recharge and average baseflow and d) distribution of the groundwater components within the rainfall/runoff model to estimate baseflow and runoff for each tributary.

1.3. Aims and Objectives

To understand the resource base available as well as particular flow dynamics which could impact the Verlorenvlei, a number of aims and objectives have been identified. These aims and objectives address a number of questions that need to be answered before the ecological reserve can be determine and proper management decisions can be made regarding the allocation of surface and groundwater within the sub-catchment.

AIM [1]: To delineate the sub-catchment boundary extent of the Verlorenvlei, by considering the surface water and groundwater contributions of the main feeding tributaries

1.1) To use a tributary, namely the Krom Antonies to understand variables that might influence surface water and groundwater routing within the sub-catchment

1.2) To understand the bearing that the geological dip direction might have on groundwater flow within the Krom Antonies and how this could be tested

1.3) To decide whether the geological boundary has merit, or whether a more conventional surface water boundary should be used for the sub-catchment

1.4) To apply the sub-catchment delineation procedure required for modelling of potential and net recharge, baseflow and runoff

AIM [2]: To estimate potential recharge using a rainfall/runoff model (J2000) constructed for the sub-catchment using climate and surface water variables.

2.1) To compile the relevant input data including climate, monitoring and parameter estimates required to estimate potential recharge within the sub-catchment

2.2) To use the measuring data, namely rainfall and climate data, primary and secondary aquifer water levels to validate the potential recharge outputs and determine whether these estimates are realistic for the sub-catchment

2.3) To compare the potential recharge estimates with previous estimates in the region and account for these difference

2.4) To understand the variables that control the potential recharge estimates within the sub-catchment and whether the model needs adjustment to align with previous recharge estimates

AIM [3]: To apply the potential recharge and previously documented aquifer hydraulic conductivity into a groundwater model (MODFLOW) of the Krom Antonies, to determine net recharge and average daily baseflow from the tributary

3.1) To determine the net recharge for the TMG formations and quaternary sediments within the tributary and understand why potential recharge might not equate to net recharge for these formations

3.2) To quantify the resultant interflow component from the tributary if potential recharge and net recharge are not equivalent.

3.3) To determine whether the Krom Antonies is the major source of groundwater for the Verlorenvlei.

3.4) To determine whether the groundwater baseflow from the Krom Antonies is enough to meet the evaporation demand of the lake.

AIM [4]: To estimate runoff and baseflow for the main feeding tributaries of the Verlorenvlei, required for determining the hydrological components needed for ecological reserve determination.

4.1) To distribute the groundwater components of the rainfall/runoff model of the sub-catchment through the transfer of aquifer hydraulic conductivity and net recharge from the groundwater model of the Krom Antonies.

4.2) To identify the main sources of baseflow and runoff for the Verlorenvlei.

4.3) To use flow exceedance probabilities to determine the hydrological components of the ecological reserve for each tributary

4.4) To use a number of indices (CV/BFI/RD1-RD2/RG1-RG2) to understand the tributary flow contribution as well as factors that may influence its variability

1.4. Thesis layout

The thesis is made up of four papers which logically follow on from each other. Papers 1 and 2 are published, papers 3 and 4 have been submitted and are under review. The first contribution “Investigating Potential Additional Sources of Groundwater Flow into a Defined Watershed” details with the delineation of the sub-catchment and investigates the potential that geology could have on inflows and outflows into the sub-catchment, using a test site the Krom Antonies. The second contribution “Estimation of groundwater recharge via percolation outputs from a rainfall/runoff model for the Verlorenvlei estuarine system, west coast, South Africa” uses the results from the catchment delineation investigation to construct a rainfall/runoff model used to estimate potential recharge within the entire sub-catchment. The third contribution “Combining rainfall/runoff and groundwater modelling to calculate recharge and baseflow in the data scarce Verlorenvlei catchment South Africa” applies the potential recharge with documented aquifer hydraulic conductivity to estimate net recharge and average aquifer baseflow for the Krom Antonies tributary. The fourth contribution “Distributive rainfall/runoff modelling to determine runoff to baseflow proportioning and its impact on the determination of the ecological reserve” distributes the groundwater components within the rainfall/runoff model using the net recharge and aquifer hydraulic conductivity from the Krom Antonies to improve the runoff/baseflow proportioning needed for ecological reserve determination and understanding river flow regime dynamics within the sub- catchment.

1.5. References

Alley, W.M., Leake, S.A., 2004. The journey from safe yield to sustainability. Ground Water 42, 12–16.

Anderegg, W.R.L., Kane, J.M., Anderegg, L.D.L., 2012. Consequences of widespread tree mortality triggered by drought and temperature stress. Nat. Clim. Chang. 3, 30.

Bernstein, S., 2015. California Governor Orders 25% Reduction in Water Usage Statewide. Scientific. America.

Bugan, R.D.H., 2014. Modeling and Regulating Hydrosalinity Dynamics in the Sandspruit River Catchment (Western Cape). Stellenbosch University. Unpublished PhD thesis, 1-216

de Andrade, E.M., Palácio, H.A.Q., Souza, I.H., de Oliveira Leão, R.A., Guerreiro, M.J., 2008. Land use effects in groundwater composition of an alluvial aquifer (Trussu River, Brazil) by multivariate techniques. Environ. Res. 106, 170–177.

Easterling, D.R., Meehl, G.A., Parmesan, C., Changnon, S.A., Karl, T.R., Mearns, L.O., 2000. Climate extremes: observations, modeling, and impacts. Science (80-.). 289, 2068–2074.

Elmore, A.J., Manning, S.J., Mustard, J.F., Craine, J.M., 2006. Decline in alkali meadow vegetation cover in California: the effects of groundwater extraction and drought. J. Appl. Ecol. 43, 770–779.

Foley, J.A., DeFries, R., Asner, G.P., Barford, C., Bonan, G., Carpenter, S.R., Chapin, F.S., Coe, M.T., Daily, G.C., Gibbs, H.K., Helkowski, J.H., Holloway, T., Howard, E.A., Kucharik, C.J., Monfreda, C., Patz, J.A., Prentice, I.C., Ramankutty, N., Snyder, P.K., 2005. Global Consequences of Land Use. Science (80). 309, 570–574.

Harbaugh, B.A.W., Banta, E.R., Hill, M.C., Mcdonald, M.G., 2000. MODFLOW-2000, The U.S Geological Survey modular Groundwater-Water Model — User guide to modularization concepts and the ground-water flow process. U.S. Geological. Survey. 130.

Jiménez-Martínez, J., Skaggs, T.H., van Genuchten, M.T., Candela, L., 2009. A root zone modelling approach to estimating groundwater recharge from irrigated areas. J. Hydrol. 367, 138–149.

Kim, N.W., Chung, I.M., Won, Y.S., Arnold, J.G., 2008. Development and application of the integrated SWAT-MODFLOW model. J. Hydrol. 356, 1–16.

Krause, P., 2001. Das hydrologische Modellsystem J2000. Beschreibung und Anwendung in großen Flussgebieten, in: Umwelt/Environment, Vol. 29. Jülich: Research Centre.

Legesse, D., Vallet-Coulomb, C., Gasse, F., 2003. Hydrological response of a catchment to climate and land use changes in Tropical Africa: case study South Central Ethiopia. J. Hydrol. 275, 67–85.

Lumsden, T.G., Schule, R.E., Hewitson, B.C., 2009. Evaluation of potential changes in hydrolocally relevant statistics of rainfall in Southern Africa under conditions of climate change. Water SA 35, 649–656.

Middelkoop, H., Daamen, K., Gellens, D., Grabs, W., Kwadijk, J.C.J., Lang, H., Parmet, B.W.A.H., Schädler, B., Schulla, J., Wilke, K., 2001. Impact of climate change on hydrological regimes and water resources management in the Rhine basin. Clim. Change 49, 105–128.

Mirrilees, R. I., Forster, S. F., & Williams, C. J. 1994. The application of economics to water management in South Africa. Water Research Commission. No. 415-494

Parkin, G., Birkinshaw, S.J., Younger, P.L., Rao, Z., Kirk, S., 2007. A numerical modelling and neural network approach to estimate the impact of groundwater abstractions on river flows. J. Hydrol. 339, 15–28.

Patterson, B., 2015. What Australia Can Teach the World about Surviving Drought. Scientific. America'

Pinto, I., Jack, C., Hewitson, B., 2018. Process-based model evaluation and projections over southern Africa from Coordinated Regional Climate Downscaling Experiment and Coupled Model Intercomparison Project Phase 5 models. Int. J. Climatol. 1–11.

Rothausen, S.G.S.A., Conway, D., 2011. Greenhouse-gas emissions from energy use in the water sector. Nat. Clim. Chang. 1, 210.

Scanlon, B.R., Healy, R.W., Cook, P.G., 2002. Choosing appropriate techniques for quantifying groundwater recharge. Hydrogeol. J. 10, 18–39.

Scanlon, B.R., Jolly, I., Sophocleous, M., Zhang, L., 2007. Global impacts of conversions from natural to agricultural ecosystems on water resources: Quantity versus quality. Water Resource. Research. 43.

Scanlon, B.R., Longuevergne, L., Long, D., 2012. Ground referencing GRACE satellite estimates of groundwater storage changes in the California Central Valley, USA. Water Resource. Research. 48, 1–9.

Schulz, S., Siebert, C., Rödiger, T., Al-Raggad, M.M., Merz, R., 2013. Application of the water balance model J2000 to estimate groundwater recharge in a semi-arid environment: A case study in the Zarqa River catchment, NW-Jordan. Environ. Earth Sci. 69, 605–615.

Seager, R., Hoerling, M., Schubert, S., Wang, H., Lyon, B., Kumar, A., Nakamura, J., Henderson, N., 2014. Causes and Predictability of the 2011-13 California Drought, NOAA.

Seneviratne, S.I., Corti, T., Davin, E.L., Hirschi, M., Jaeger, E.B., Lehner, I., Orlowsky, B., Teuling, A.J., 2010. Investigating soil moisture–climate interactions in a changing climate: A review. Earth-Science Rev. 99, 125–161.

Solomon, S., Plattner, G.-K., Knutti, R., Friedlingstein, P., 2009. Irreversible climate change due to carbon dioxide emissions. Proceedings of National Academy of Sciences of the United States of America

Sophocleous, M., 2000. From safe yield to sustainable development of water resources—the Kansas experience. J. Hydrol. 235, 27–43.

Steffen, W., Hughes, L., Dean, A., Rice, M., 2018. Climate change and Drought June 2018. Climate Council of Australia

Sydeman, W., Thompson, S.A., 2014. Potential Impacts of Climate Change on California's Fish and Fisheries.

Taylor, R.G., 2012. Ground water and climate change. Nature Climate Change

Trenberth, K.E., 2011. Changes in precipitation with climate change. Clim. Res. 47, 123–138.

Walther, G.-R., Post, E., Convey, P., Menzel, A., Parmesan, C., Beebee, T.J.C., Fromentin, J.-M., Hoegh-Guldberg, O., Bairlein, F., 2002. Ecological responses to recent climate change. Nature 416, 389.

Wessels, P., Rooseboom, A., 2009. Flow-gauging structures in South African rivers Part 2: Calibration. Water SA 35, 11–20.

WHO, 2004. Guidelines for drinking-water quality: recommendations. World Health Organization.

Winter, T.C., Harvey, J.W., Franke, O.L., Alley, W.M., 1998. Ground water and surface water: A single resource. USGS Publ. 79.

WWF, 2018. WWF AGRICULTURAL WATER FILE, farming for a drier future. Office of the Premier of the Western Cape. 4–5.

Zhou, Y., 2009. A critical review of groundwater budget myth, safe yield and sustainability. J. Hydrol. 370, 207–213.

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Investigating potential additional sources of groundwater flow into a defined watershed

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Abstract

With the advancement in digital elevation model accuracies and scales, automated catchment delineation has become more widely used to reduce cost, time and errors that can occur during manual delineation. However, this type of delineation does not consider possible influxes of groundwater due to geological structures, such as dipping of bedding planes. This paper investigates the impact that the dip of bedding planes may have on groundwater flow in a catchment in the Sandveld, South Africa. Reasons for this investigation are that a number of boreholes in the area seem unaffected by pumping, even during the recent drought. To understand the possible contribution that the dip of bedding planes may have on groundwater flow, factors such as runoff, infiltration and recharge need to be considered. The intensity of rainfall can impact the amount of surface runoff or percolation that can occur from an event. Cross sections will be used to determine the dip of the bedding planes, and in conjunction with the relationship between surface runoff, deep drainage and subsurface runoff in high relief areas, will give an indication as to which areas should be excluded or included in the water balance of the catchment.

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1. Introduction

In South Africa, catchment management agencies (CMAs) are being implemented to manage and allocate water in a more equitable, efficient and sustainable manner. To date, only a few of these catchment management agencies have been developed. The division of these CMAs is done around primary river networks and at present the smallest operating unit is quaternary, although catchment sizes of primary, secondary and tertiary exist.

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Due to the large topographic variation in these catchments, there is a need for further refinement to define smaller catchments. Several authors[3] constructed fifth level river networks using the ArcHydro package within ArcMap. Currently, there is a strong emphasis on using GIS in catchment delineation in order to automate catchment mapping. This is done to reduce cost, time and errors that can occur during manual delineation and this method of delineation is effective in producing hydrological and agricultural response zones[3]. However, in delineating catchment boundaries for modelling, it is important to map all water that might be flowing into the watershed. In hydrological modelling, where the area of interest is surface water flow and therefore runoff generation, the catchment delineation based on topography is satisfactory. In groundwater modelling, where the area of interest is subsurface water flow, the dip of the bedding planes and other water bearing structures can influence whether specific areas should be included or excluded from the water balance. For example, an area would be included in the balance where the dip of the bedding planes suggest that water would flow into the catchment, even though topography dictates that water will flow out the catchment and vise visa for areas that would be excluded from the balance.

In groundwater modelling, one of the first steps is to identify the model domain and types of boundary conditions to assign. These boundary conditions are assigned based on the assumption that all water that would affect the simulation is contained within the model domain. The objective of this paper is to investigate the possible affect that the dip of bedding planes could have on the choice of the model boundary conditions. For example, if the user was certain that all water flowing into the catchment was contained within the model domain, the user would specify a “No Flow” boundary surrounding the study area. This would facilitate faster run times for the model and quicker convergence of unknown parameters. If it is suspected that the model domain does not contain all water flowing into the catchment, the user would then specify “Constant Head Cells” to the boundary conditions.

In this study, possible contributions from groundwater flow were considered in a catchment that feeds the Verlorenvlei estuarine lake on the west coast of South Africa. In this catchment a number of boreholes were unaffected by large scale pumping, even though there has been little recharge due to recent drought conditions. These boreholes are situated in areas where dipping bedding planes could contribute groundwater to the catchment.

2. Study site

The Verlorenvlei estuarine lake is situated on the west coast of South Africa in the Sandveld. The lake is located between Redelinghuis and Elands Bay, where the estuary connects the lake to the sea. The estuarine lake is fed by four main tributaries: Kruismans, Bergvallei, Hol and Krom Antonies rivers, of which the Krom Antonies is the most important in terms of fresh water and low salt load. The Krom Antonies River originates in the Piketberg mountains, where mean annual precipitation greatly exceeds that of the lower lying areas of the catchment. The highest amount of rainfall in the area is received in the Piketberg Mountains, which are around 1300 m above sea level. These mountains that are south-east of Verlorenvlei average annual rainfall is approximately 780 mm/year. Mist is also considered to be a significant contributor to soil moisture. Rainfall declines moving north-west from the Piketberg Mountains, reaching a low of 210 mm/year at the mouth of Verlorenvlei[2].

The Neoproterozoic Malmesbury Group (780-750 Ma), which are the oldest rocks in the area, is represented by the Piketberg Formation, and consists of greywache, sercite schist, quartzitie, conglomerate and limestone[4]. The Malmesbury Group has been intruded by the Cambrian Cape Granite Suite, specifically by the Riveria Pluton between Redelinghuis and Piketberg unconformably overlying the Malmesbury Group and Cape Granite Suite, is the Palaeozoic Table Mountain Group (TMG), that is made up of a number of different sedimentary formations. The Piekenierskloof, which is the oldest of the TMG formations, reaches a maximum thickness of 900 m within the Western Cape and consists of coarse grained sandstone, mudrock and conglomerates[5]. The Graafwater Formation consists of sandstone, siltstone and shale and is up to 430 m thick. The Peninsula Formation, which is the youngest of the sandstones, is comprised of minor shale, conglomerates and quartz arenite, and has an average thickness of 2000 m[5].

In terms of the hydrogeology of the area, unconsolidated primary-porosity and fractured-rock secondary-porosity aquifers are found. The primary-porosity alluvial aquifer is normally high yielding and is made of coarse-grained, clean sand. Geological features such as weathering zones, bedding surfaces and fault planes control groundwater flow in the secondary-porosity aquifer. A strong connection exists between the primary and the secondary aquifer as evidenced by the recharge relationship between these two aquifers and by the piezometric head of the secondary aquifer being higher than the water table1.

3. Methodology

The catchment was initially delineated using the GIS software package ArcHydro. The digital elevation model (DEM) that was used was the Shuttle Radar Topography Mission (SRTM) 30 m, due to its accessibility and good resolution. Final outputs from the delineation were compared to elevations from the Stellenbosch University Digital Elevation Model (SUDEM) 5 m elevation6. The following functions in ArcHydro were used in the automated delineation: (1) Fill sinks, (2) Flow direction, (3) Flow accumulation, (4) Stream definition, (5) Stream segmentation, (6) Grid delineation and (7) Polygon features. During the automated catchment delineation, a number of functions were used to prepare the DEM so that dominant stream and topographic highs can be identified. The DEM was initially modified to remove cells that have undefined drainage directions to prevent discontinuity in drainage networks. Each cell in the DEM was assigned a code which was used to identify the flow direction from the cell. These directions were accumulated to determine and to identify stream channels and topographic highs. Threshold functions were used to define the amount of accumulation that would be used to assign stream networks. A grid of stream segments was then created with their own identifiers and these files were then converted to raster format.

4. Discussion

There are a number of factors that need to be considered when assessing the possible impact that the dip of bedding planes might have on aquifer boundaries. The intensity of rainfall can influence whether rainfall that is received on a land surface contributes to runoff or is allowed to infiltrate and therefore contributes to recharge. Understanding the rainfall dynamics of the catchment is important in trying to decide whether geological orientation should be considered and therefore what alterations of the model boundary conditions should be made. Low intensity, long duration rainfall events could have a significant impact on recharge, while there should be little runoff. High intensity, short duration rainfall events could mainly contribute to runoff and therefore will play no part in recharge. What is essential in this study is to quantify surface runoff, deep drainage and subsurface runoff components. This will be done using information from landtype classification. Climatic information will be gathered regarding the number of rainfall events that occur over and above certain threshold. The definition of a threshold events needs to be defined within the respective study area. These thresholds will be defined using analysis of rainfall records to determine the probability of exceedance. The intersection of intensity and duration need to be considered within this threshold.

A number of extra cross-sections will be run through the catchment to get a more detailed assessment of the geological structure, including dip of bedding planes, and allow an assessment of the extent of domains where groundwater input might be derived from beyond the watershed. The mountains south of the catchment are dipping towards the catchment, meaning that groundwater could be flowing into the catchment across the delineated watershed and therefore could contribute to the water budget (Fig 1). The dip of the bedding planes on the north- west side of the cross section suggest that there is an area beyond the watershed that could contribute to groundwater flow (Fig 2).

The difference in salinity between the Krom Antonies River (Fig 2) and Hol River (Fig 2, outside water shed) is notable, in both groundwater and surface water. These two rivers derive water from the same mountain range, although their geochemical concentrations are dissimilar, where the Hol river is high in Electrical Conductivity and the Krom Antonies low. A possible cause of this could be the limited flushing of salts within the Hol river, where the dip of bedding planes is redirecting the majority of groundwater flow into the Krom Antonies catchment. To be able to determine whether these geological features have an influence on the groundwater flow within the catchment, the groundwater model has to be set up using the initial watershed boundary that was constructed within ArcHydro. If the model does not calibrate with measurements that were collected in the field and pump tested boreholes are outside of published transmissivity values', then further investigation is needed to include the dip of bedding planes.

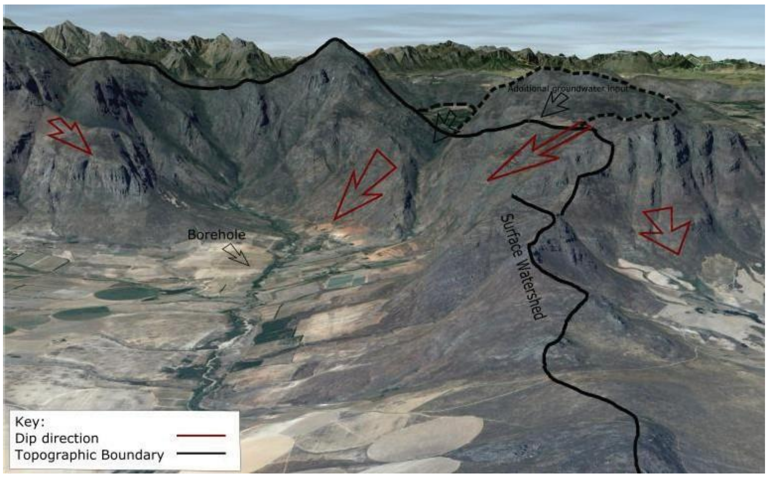

Fig 1: The dip of bedding planes to illustrate possible contributors to groundwater flow (Vertical exaggeration 2:1)

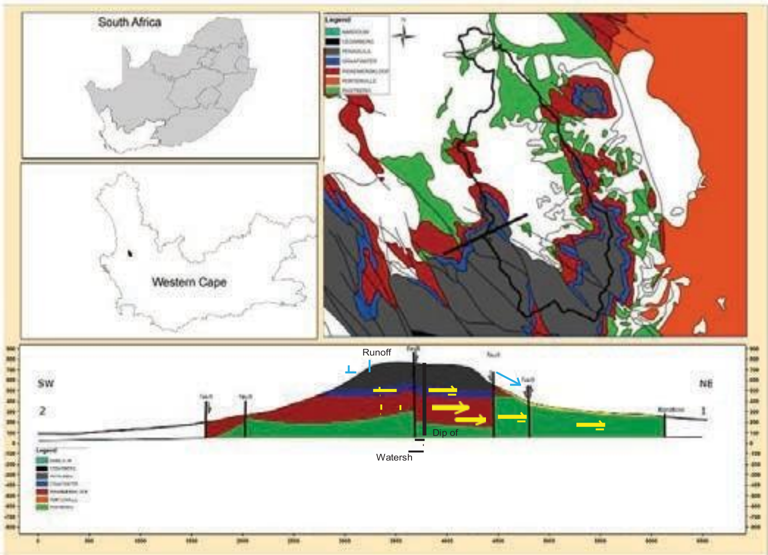

Fig 2: Possible groundwater contributions from areas that are not contained within the watershed (after[4])

5. References

1. Conrad, J, Nel, J, Wentzel, J. The challenges and implications of groundwater recharge: A case study on the Northern Sandveld, Western Cape, South Africa. Proceedings of the 2004 Water Institute of Southern Africa (WISA) Biennial Conference.

2. Lynch, S.D. 2003. Development of a raster database of annual, monthly and daily rainfall for Southern Africa. School of Bioresources Engineering and Environmental Hydrology. University of Natal. Pietermaritzburg, South Africa.

3. Maherry, A.M, Horan, M.J.C, Smith-Adao, L.B, van Deventer, H, Nel, J.L, Schulze, R.E, Kunz, R.P. 2013. Delineating River Network Quinary Catchments for South Africa and allocating Associated Daily Hydrological information. WRC report No.2020/1/12.

4. Rozendaal, A, P.G, Gresse, Scheepers, R and de Beer, C.H. 1994. Structural setting of the Riviera W-Mo deposit, Western Cape, South Africa. *Journal of Geology*. 97, 184-195.

5. Rust, I.C. 1967. On the sedimentation in the Table Mountain Group in the western Cape province. D.Sc. thesis (unplubl.), Univ. Stellenbosch, 110 pp.

6. van Nieker, 2005. Stellenbosch University Digital Elevation Model (SUDEM). Centre for Geographical Analysis. Stellenbosch University.

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Research papers

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 spa- tial 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 cap- able 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 (Prˇibánˇ and Ondok,1985), which exacerbate this competition. Whilst the amount of water needed to sustain different agricultural crops is well con- strained (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 of water that is required to maintain aquatic ecosystems (Hughes,

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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: Šimu˚nek et al., 2006). 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. How- ever, for larger spatial scales, rainfall/runoff models are able to model representative recharge (Scanlon et al., 2002) and are there- fore 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 sup- porting 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 con- sists of the Piketberg Mountains in the east, which form the highest topographic elevation (1446 m) 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 km2 in size, where the catch- ment has an area of 1832 km2.

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

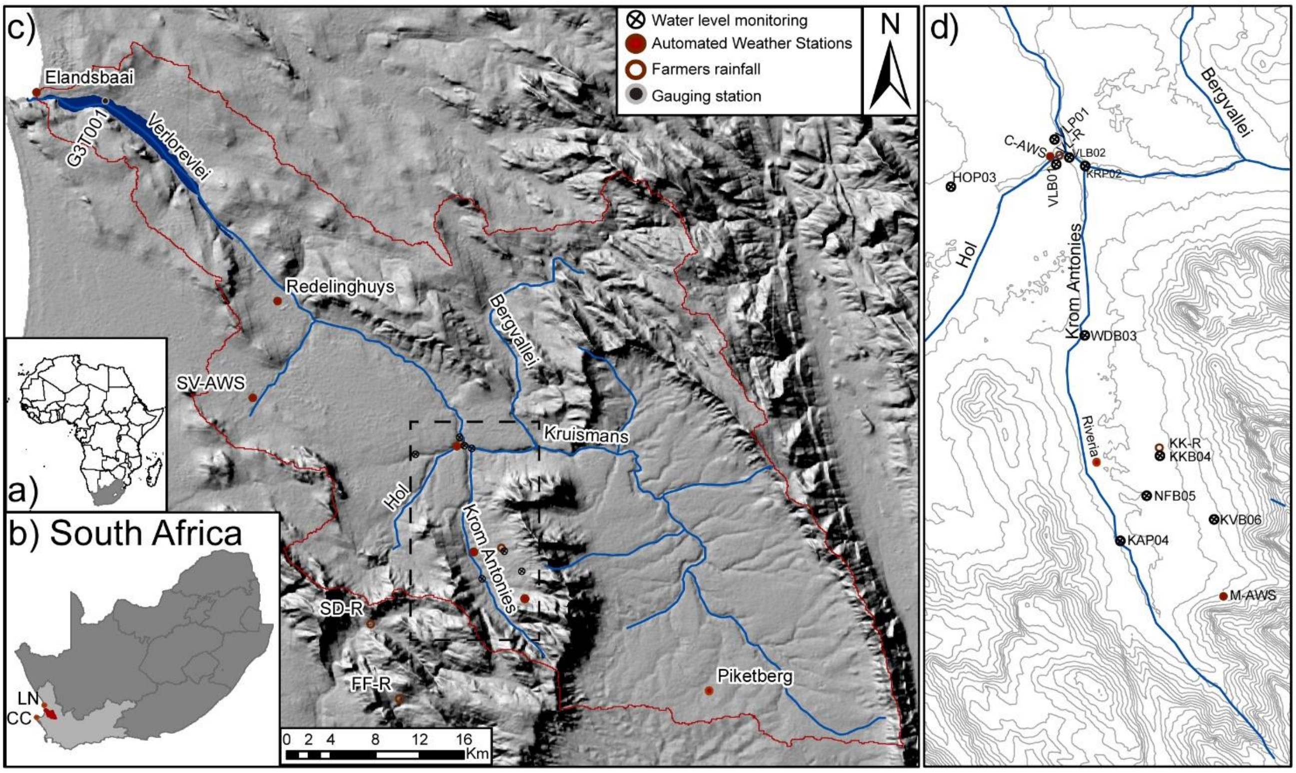

Fig. 1. (a) Location of South Africa within Africa; (b) location of Western Cape (WC) within South Africa showing the study catchment and data collection points outside the catchment (LN, CC); (c) extend of the study catchment and the data collection points within the catchment; (d) dominant data collection area with collection points.

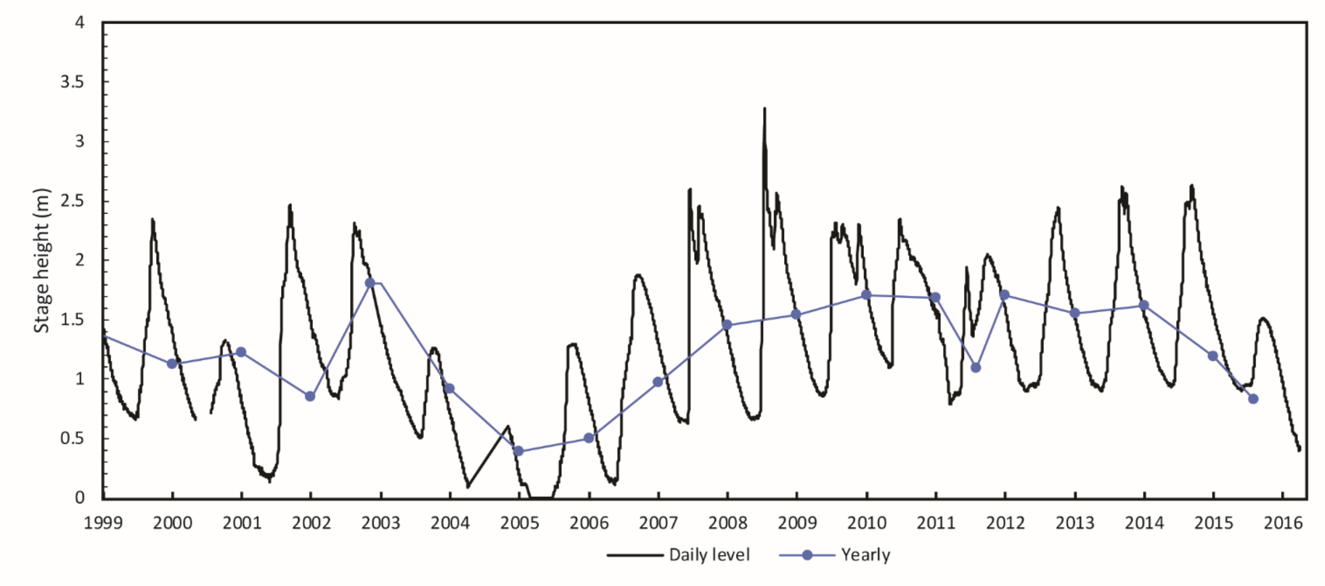

Fig. 2. Daily and yearly average stage height monitored between 1999 and 2016 within the Verlorenvlei estuarine lake at Department of Water Affairs (DWA) G3T001 station. Stage height is equivalent to water depth, where a stage height of 0 indicates the lake is dry at the monitoring point.

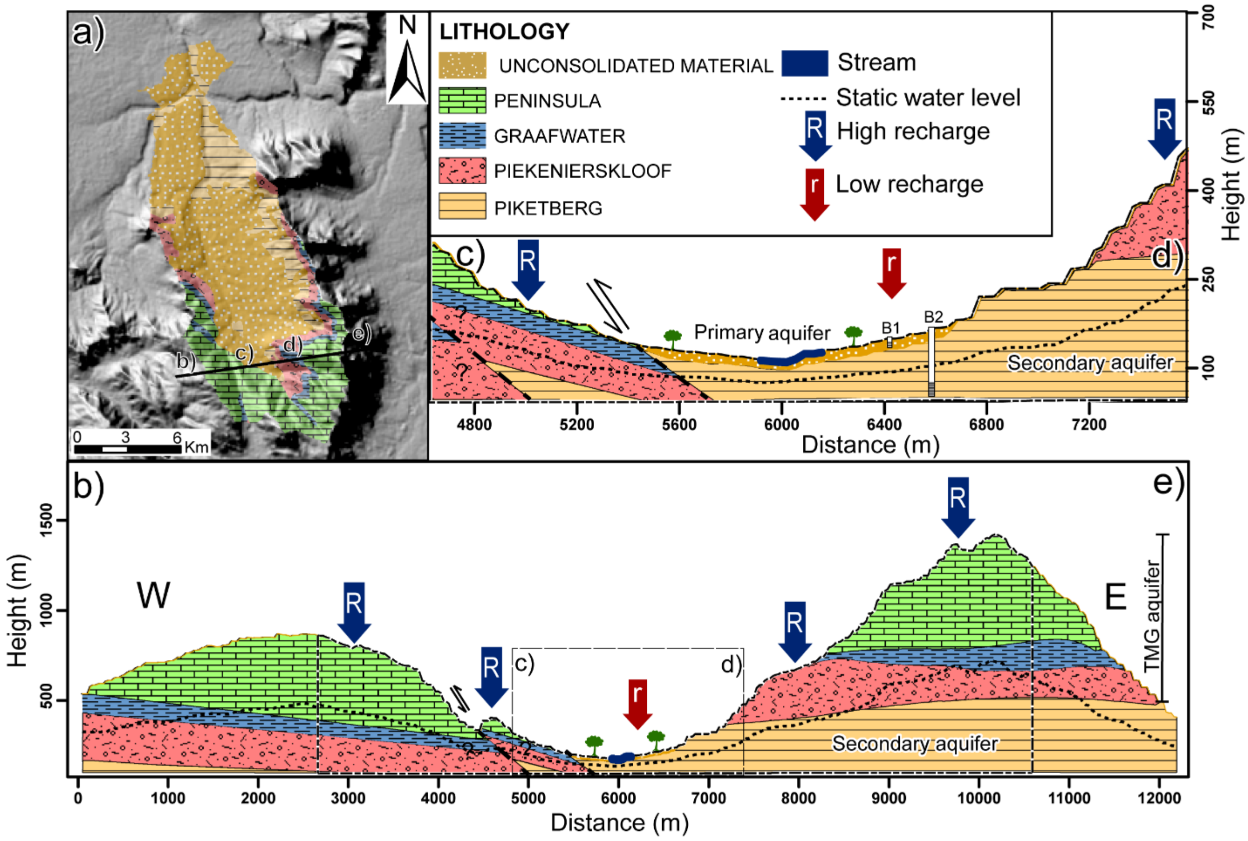

Fig. 3. Cross section of the dominant recharge locations with respect to TMG, secondary (b–e) and primary aquifers (c–d) as well as primary (B1) and secondary boreholes (B2) (Vertical exaggeration 2:1).

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 com- prised of greywacke, sericitic 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

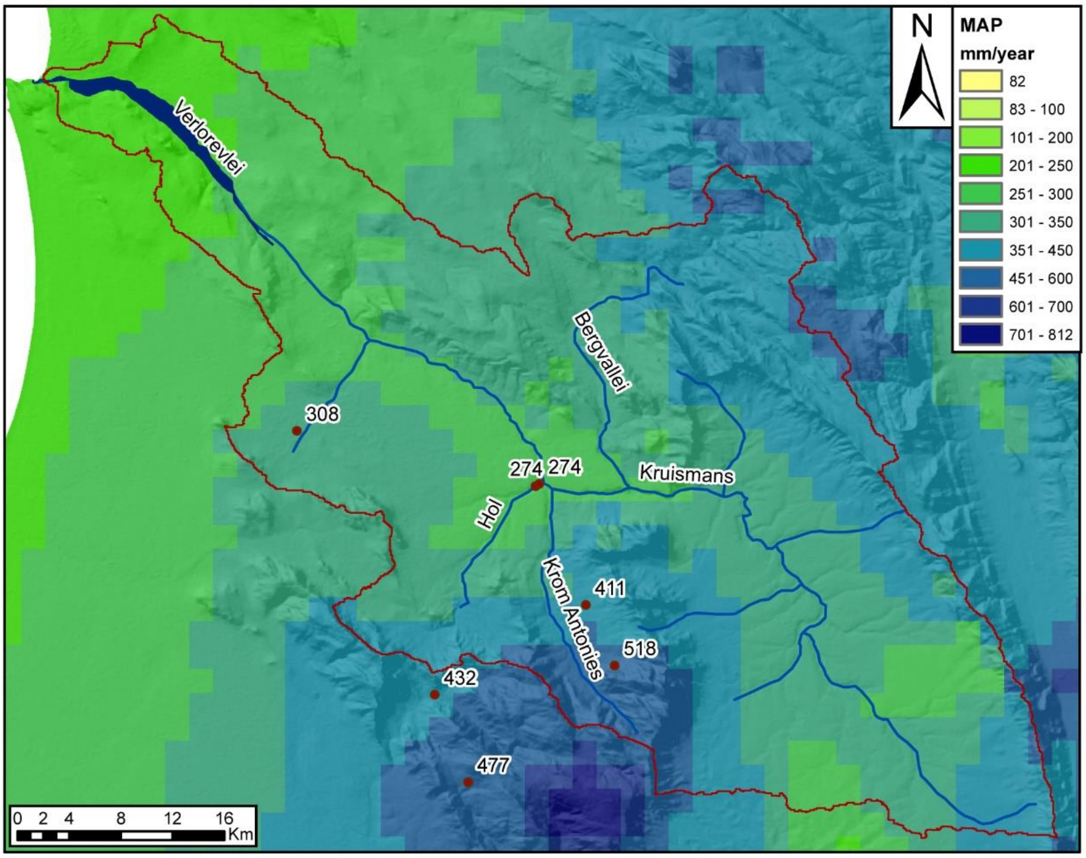

Fig. 4. Mean Annual precipitation (MAP) across the study catchment and at rainfall monitoring points (after Lynch, 2004).

coarse-grained, clean sand and therefore is high yielding. Previous recharge estimates for the primary aquifer are between 0.2 and 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 cli- mate, 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 °C, 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 °C, 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 rain- fall 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 ha and using around 20% of total recharge (DWAF, 2003). Potatoes in the Sand- veld 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 ha, 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 through- out the catchment to understand groundwater responses to rain- fall, 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 h. The farmers' rainfall records 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 ±8 mm. 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 m 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 pro- filing 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 °C (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:

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

where Sy is specific yield and Dh 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 speci- fic 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 catch- ment and include Arensols, Leptosols, Solonetzs, Fluvisols, Plano- sols, Regosols, Lixisols, Cambisols, and Luvisols (Batjes 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*

An HRU is an area with homogenous physiological and topographical features, used for distributive hydrological modelling 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 and 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 weight- ing (IDW) method for the regionalization of the input climate data, which is derived from the climate stations. Due to the scarce net- work of the climate stations within the catchment, and the signif-icant 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 eleva- tion, which

Table 1
Estimated waterholding capacity of MPS and LPS using pedotransfer functions from the HYDRUS model using average soil depth and texture from the Harmonized World Soil Database v1.2 (HWSD).

				Pressure (mbar)			60-15,000	60-0	Waterholding	
Soil type	Horizon	Depth (mm)	Sand, Silt, Clay (%)	0	60	15,000	MPS	LPS	MPS	LPS
Arenosol	A	300	89,6,5	0.3781	0.1602	0.0501	0.1101	0.2179	33.03	65.37
	B	700	90,5,5	0.3769	0.1563	0.0515	0.1048	0.2206	73.36	154.42
								Total	106.39	219.79
Leptosol	A	100	43,29,28	0.4129	0.3542	0.074	0.2802	0.0587	28.02	5.87
								Total	28.02	5.87
Solonetz	A	300	35,37,28	0.426	0.3806	0.0757	0.3049	0.0454	91.47	13.62
	B	700	27,37,36	0.4506	0.4021	0.086	0.3161	0.0485	221.27	33.95
								Total	312.74	47.57
Fluvisol	A	300	44,33,23	0.4066	0.352	0.0671	0.2849	0.0546	85.47	16.38
	B	700	45,31,24	0.4064	0.3485	0.0864	0.2621	0.0579	183.47	40.53
								Total	268.94	56.91
Planosol	A	300	56,25,19	0.3908	0.3166	0.0584	0.2582	0.0742	77.46	22.26
	B	700	44,23,33	0.414	0.3526	0.0785	0.2741	0.0614	191.87	42.98
								Total	269.33	65.24
Regosol	A	300	69,19,12	0.3834	0.2781	0.046	0.2321	0.1053	69.63	31.59
	B	700	70,17,13	0.3821	0.278	0.0481	0.2299	0.1041	160.93	72.87
								Total	230.56	104.46
Lixisols	A	300	63,15,22	0.3831	0.3067	0.062	0.2447	0.0764	73.41	22.92
	B	700	53,13,34	0.3979	0.3361	0.0766	0.2595	0.0618	181.65	43.26
								Total	255.06	66.18
Cambisol	A	300	42,26,32	0.4173	0.3575	0.0782	0.2793	0.0598	83.79	17.94
	B	700	41,25,34	0.4205	0.3609	0.0802	0.2807	0.0596	196.49	41.72
								Total	280.28	59.66
Luvisol	A	300	51,22,27	0.3994	0.3329	0.0706	0.2623	0.0665	78.69	19.95
	B	700	45,21,34	0.4126	0.3506	0.0789	0.2717	0.062	190.19	43.4
								Total	268.88	63.35

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 Eq. (2):

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

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 rain- fall amount and elevation. The regression line created between the elevation and rainfall correlation should have a r2 value greater than a specified limit, which in this study was set as 0.75. The calculation is then made according to Eq. (3):

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

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.

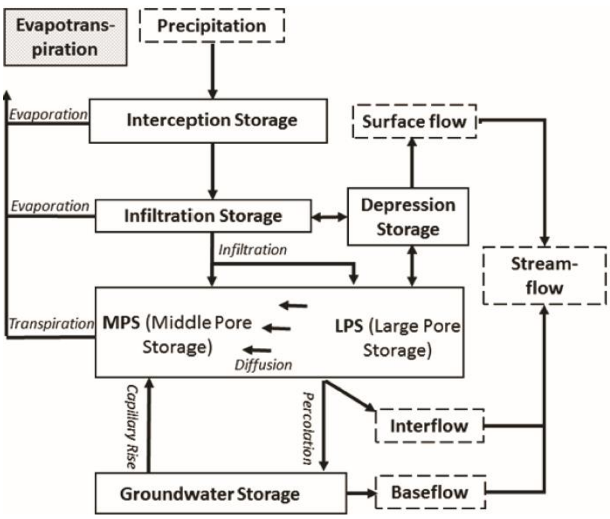

Fig. 5. Schematic showing the processes that are simulated within the J2000 model to allow for the estimation of percolation (after Krause 2001).

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 maxi- mum 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 (Intmax) is calculated according to Eq. (4):

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

where α is the storage capacity per m^2 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 (Šimů nek et al., 2006) incorporating soil textures from the HWSD. A pedo- transfer function was applied to three hypothetical pressure scenarios namely: 0 mbar, 60 mbar and 15,000 mbar. The storage capacity of MPS, water held at field capacity, was calculated by the difference in water content between 60 mbar and 15,000 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 maxi- mum 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 Eq. (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 *soilLatVertDist* 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 Eq. (6):

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

where *SoilOutLPS* is the calibration factor for the definition of LPS outflow (values range from 0 to 10) (Nepal, 2012). During this study, the *SoilOutLPS* calibration factor was determined using the Kruismans gauging station that was operational from 1970 to 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 affected 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.

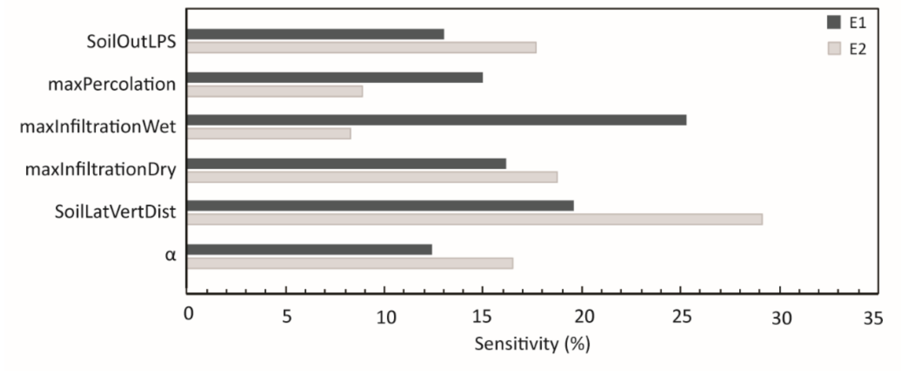

Fig. 6. The sensitivity of the parameters based objective functions for Nash Sutcliffe Efficiency with squared differences (e2) and absolute values (e1).

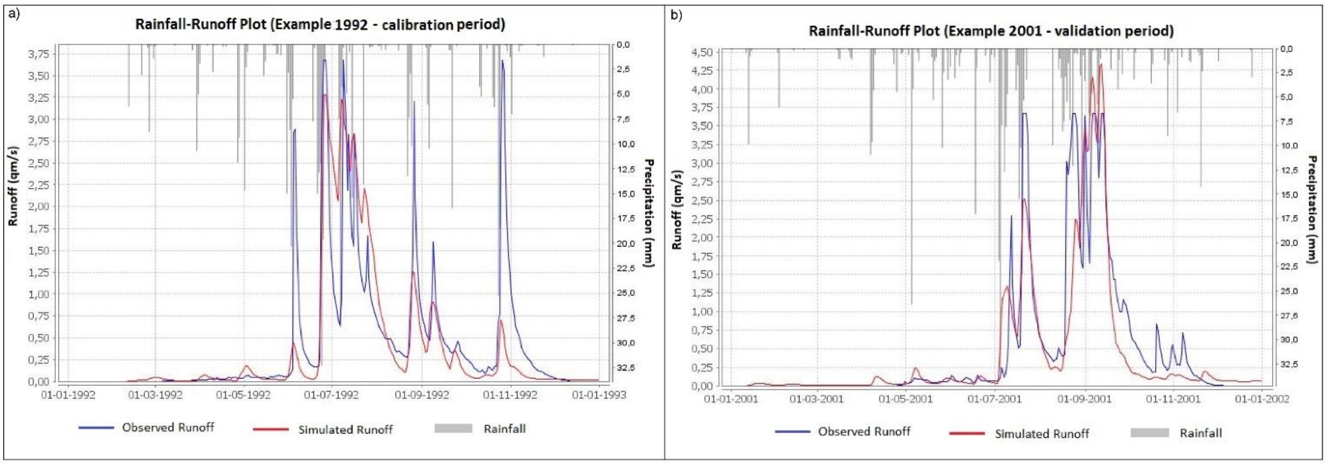

Fig. 7. Simulated runoff (red line) for the J2000 model for an example calibration (1992) and validation (2001) period showing measured runoff (blue line) and rainfall (grey bar) (y-axis scale for periods are not identical). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Table 2

Global and calibrated values (OPTAS) for parameters that influenced J2000 modelled percolation.

Parameter	Description	Global Range	Unit	Calibrated value
SoilOutLPS	Outflow coefficient for Large Pore Storage	0–10	Non-dimensional	0.2
MaxPercolation	Maximum percolation rate to the groundwater storage	0–100	mm	2
MaxInfiltrationWet	Maximum infiltration during wet season	0–200	mm	40
MaxInfiltrationDry	Maximum infiltration during dry season	0–200	mm	100
SoilLatVertDist	Lateral-vertical distribution coefficient	0–10	Non-dimensional	0.7
α	Storage capacity of particular land cover for rain	0–5	mm	0.1

Based on this data, 2 mm/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 (Fig. 7a and b). 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 to 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 and 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.

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.

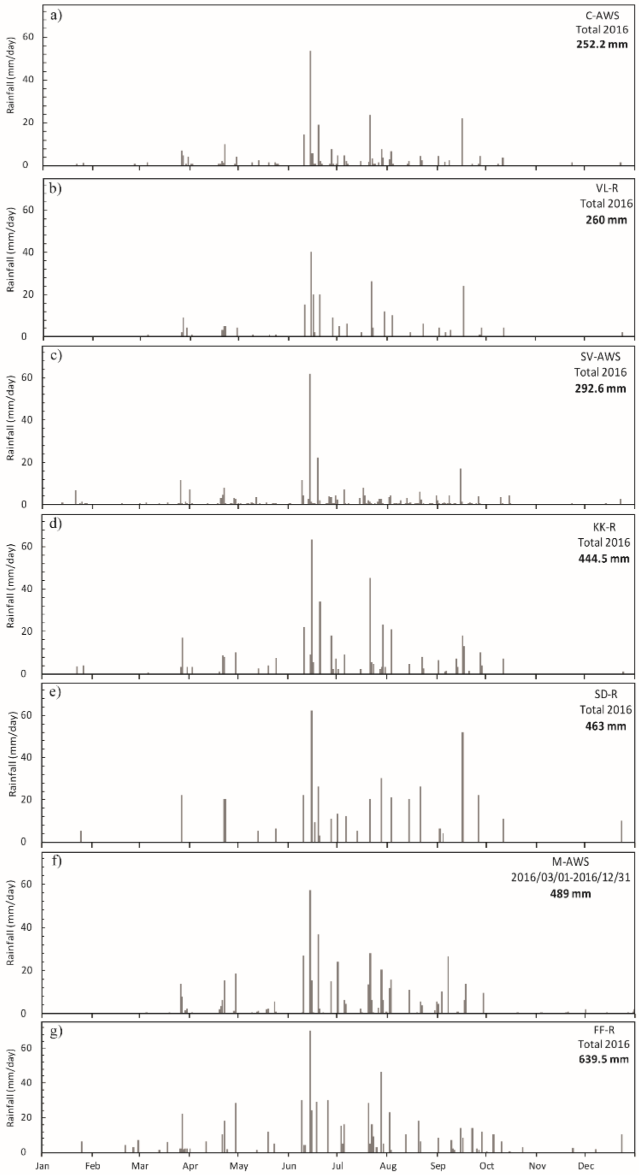

Fig. 8. Daily rainfall measured for 2016 at: (a) C-AWS, (b) VL-R, (c) SV-AWS, (d) KK- R and (f) M-AWS within the study catchment and (g) FF-R (e) SD-R outside the catchment.

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. 9d).

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).

Table 3
Yearly rainfall totals measured at selected farms within and outside the study catchments from 1999 to 2015 Year (mm)

	Year (mm)																
Farm name	2015	2014	2013	2012	2011	2010	2009	2008	2007	2006	2005	2004	2003	2002	2001	2000	1999
VL-R	169	344	332	275	264	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
KK-R	326	536	538	405	326	516	545	736	671	583	474	355	303	516	568	359	388
SD-R	254	439	605	589	463	573	768	630	730	581	572	367	364	488	736	429	433
FF-R	398	655	774	798	616	720	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A

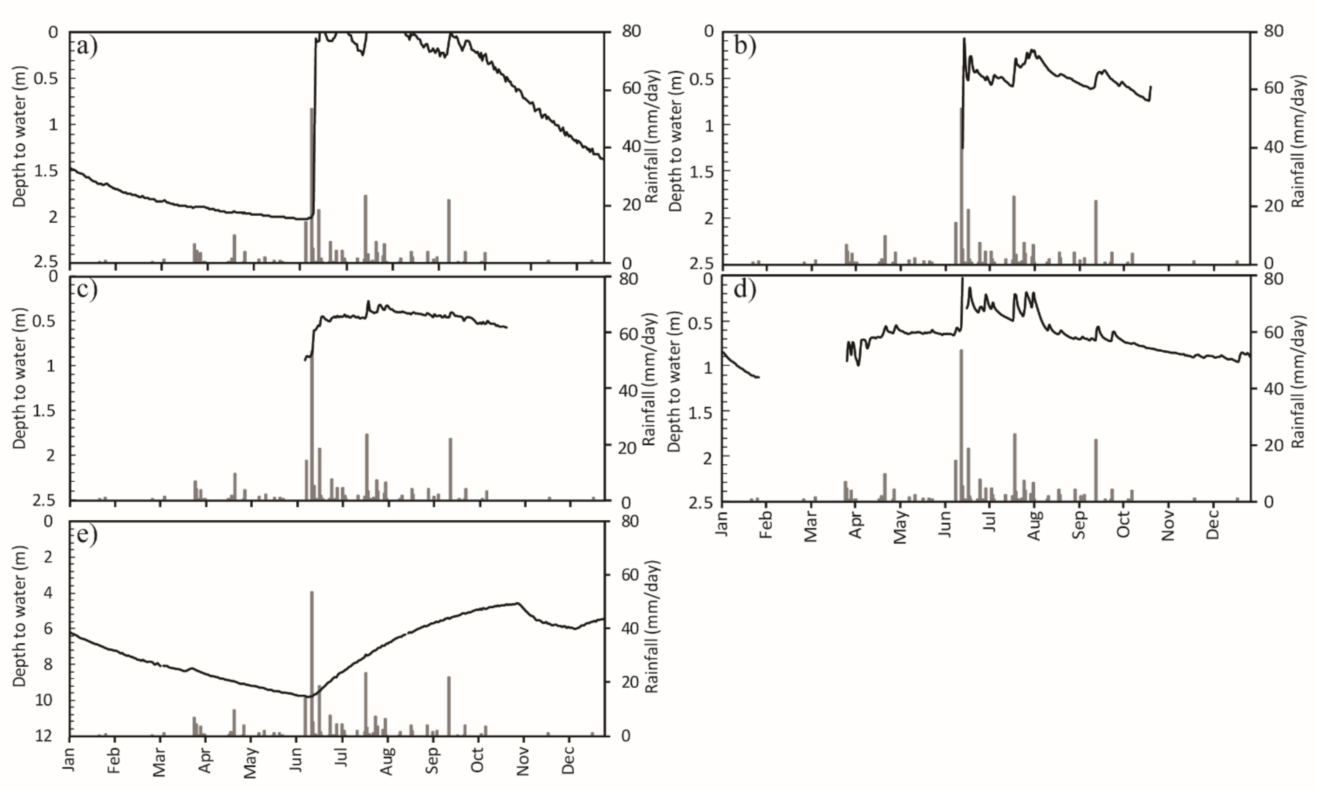

Fig. 9. Shallow groundwater monitoring and rainfall (station C-AWS) within the primary aquifer for 2016 at locations: (a) VLP01, (b) KRP02, (c) HOP02, (d) KAP04 and in borehole (e) VLB02

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.

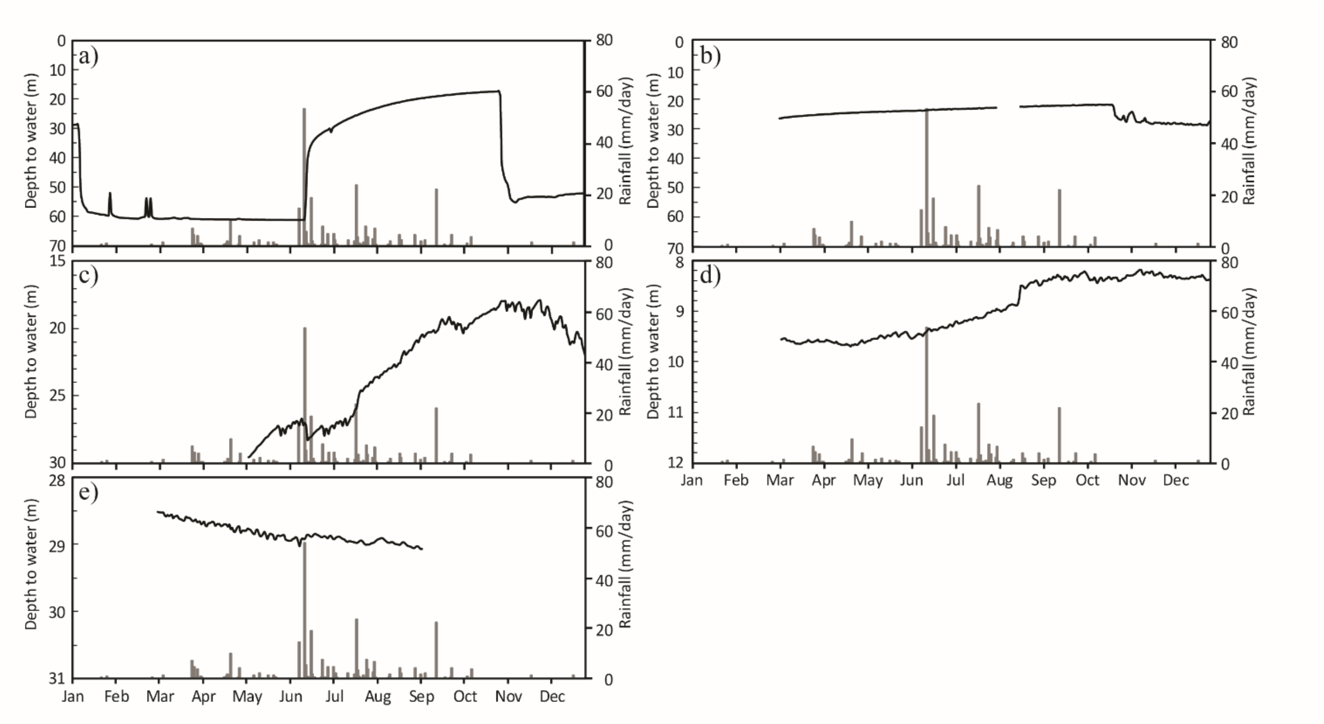

Fig. 10. Groundwater monitoring and rainfall (station C-AWS) within the secondary aquifer for 2016 at locations: (a) VLB01, (b) KKB04, (c) NFB05, (d) WDB03 and (e) KVB06.

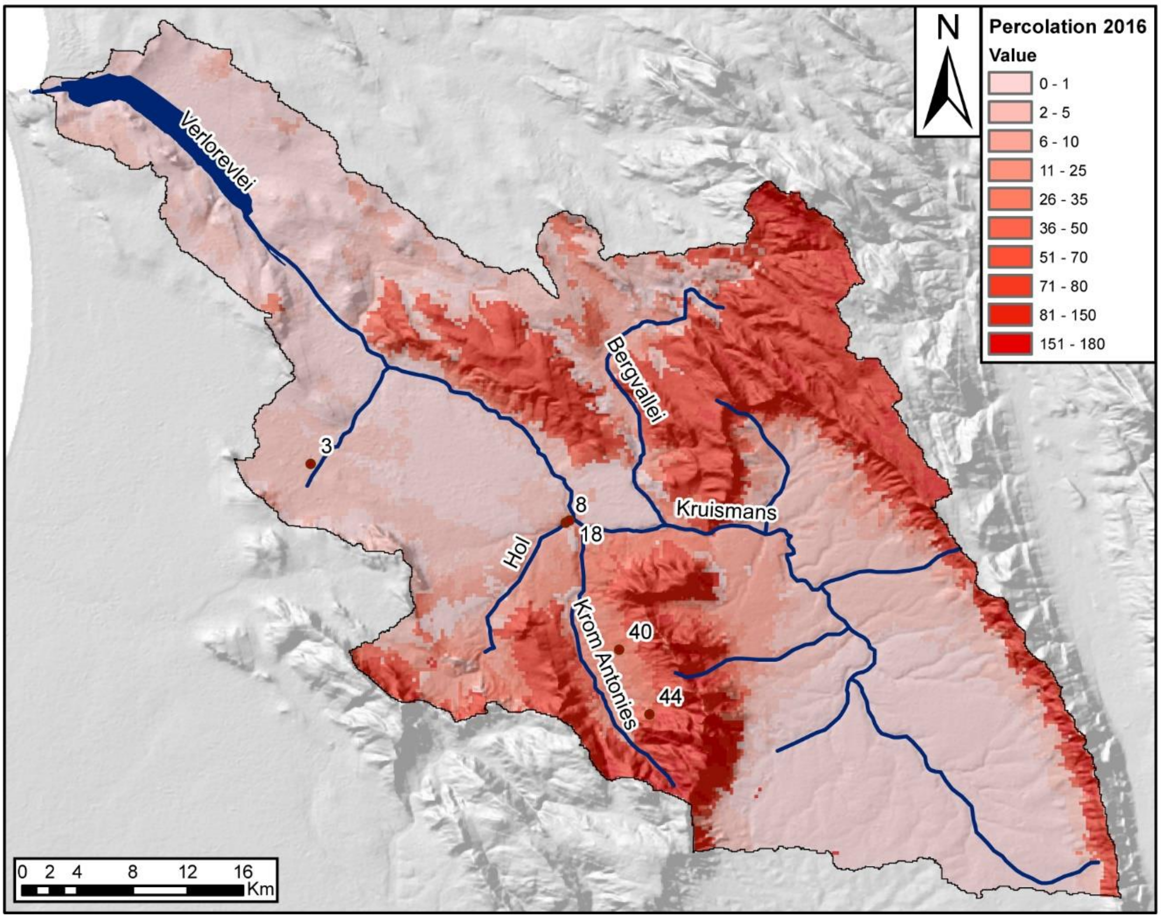

Fig. 11. The simulated percolation estimates from the J2000 model for 2016 across the Verlorenvlei catchment and at rainfall monitoring points.

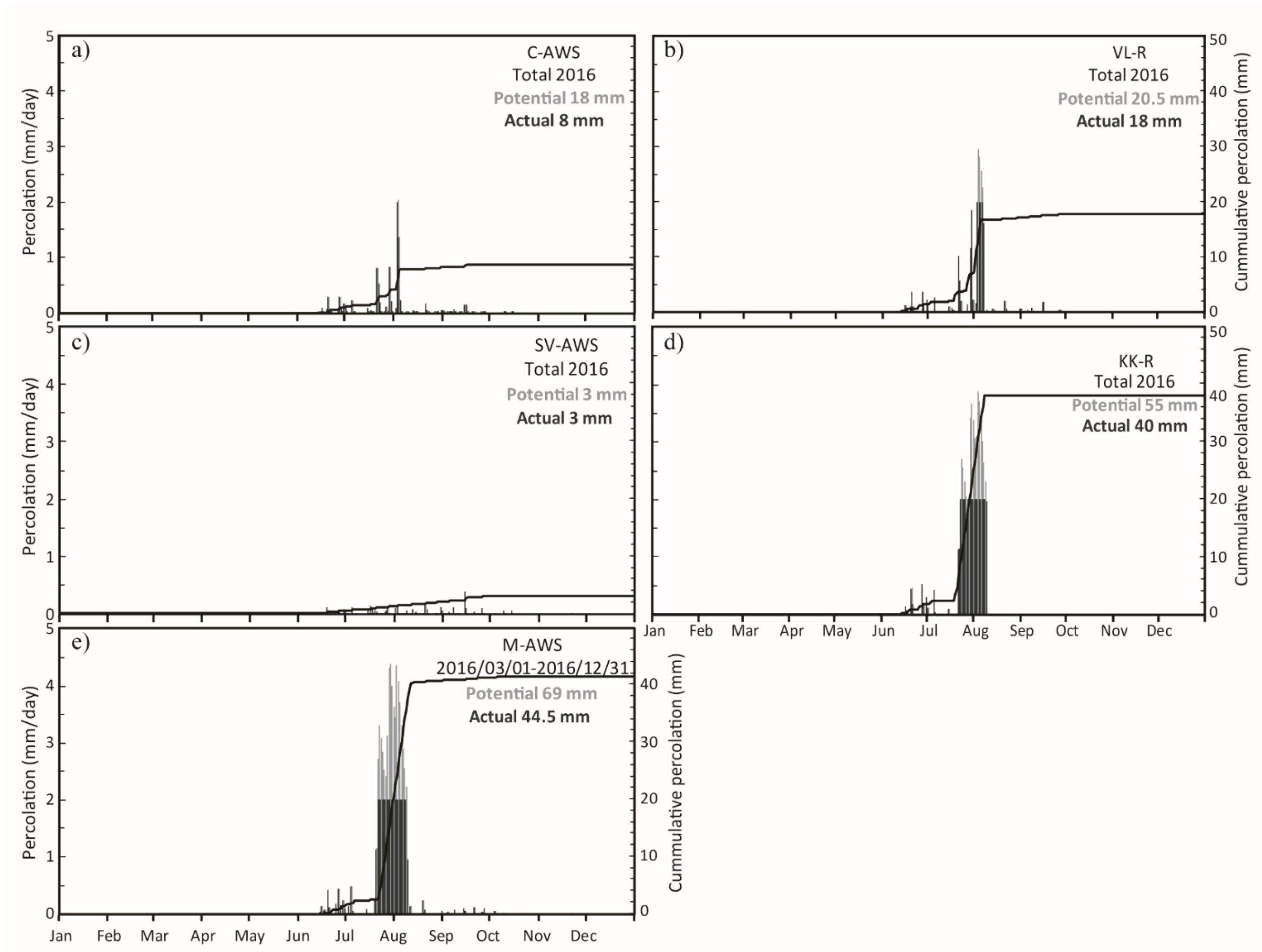

Fig. 12. Simulated potential and actual percolation for 2016 simulated at locations: (a) C-AWS, (b) VL-R, (c) SV-AWS, (d) KK-R and (e) M-AWS.

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.

4.2.4. *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.

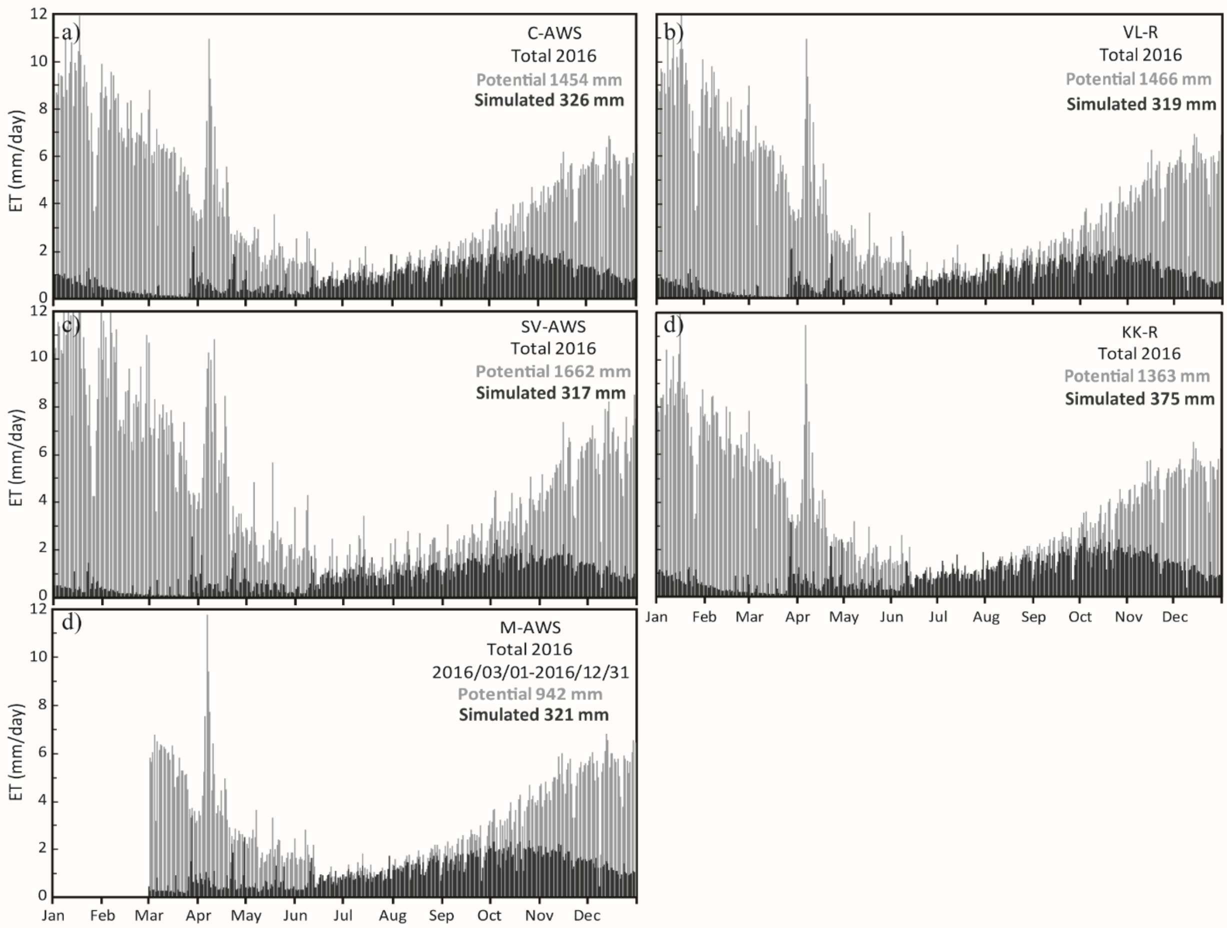

Fig. 13. Simulated potential and actual evapotranspiration for 2016 at locations: (a) C-AWS, (b) VL-R, (c) SV-AWS, (d) KK-R and (e) M-AWS.

4.2.5.*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 showedmoderate 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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8. References

Acocks, J., 1988. Veld Types of South Africa, 3rd editio. ed. Memoirs of Botanical Survey of South Africa. 57.

Allen, R.G., Pereira, L.S., Raes, D., Smith, M., 1998. Crop evapotranspiration - Guidelines for computing crop water requirements - FAO Irrigation and drainage paper 56. Rome, Italy.

Arnold, J.., Muttiah, R.., Srinivasan, R., Allen, P.., 2000. Regional estimation of base flow and groundwater recharge in the Upper Mississippi river basin. J. Hydrol. 227, 21–40.

Arnold, J.G., Allen, P.M., 1999. Automated methods for estimating baseflow and ground water recharge from streamflow records1.

Barbier, E., Acreman, M., Knowler, D., 1997. Economic Valuation of Wetlands. Gland, Switzerland.

Baron, J.S., Poff, N.L., Angermeier, P.L., Dahm, C.N., Gleick, P.H., Hairston, N.G., Jackson, R.B., Johnston, C.A., Richter, B.D., Steinman, Alan, D., 2002. Meeting Ecological and Societal Needs for Freshwater. Ecol. Appl. 12, 1247–1260.

Batjes, N., Dijkshoorn, K., Van Engelen, V., Fischer, G., Jones, A., Montanarella, L., Petri, M., Prieler, S., Teixeira, E., Wiberg, D., 2012. Harmonized World Soil Database (version 1.2). Tech. rep., FAO and IIASA, Rome, Italy and Laxenburg, Austria.

Bidaux, P., Drogue, C., 1993. Calculation of Low-Range Flow Velocities in Fractured Carbonate Media from Borehole Hydrochemical Logging Data Comparison with Thermometric Results. Groundwater 31, 19–26.

Brookfield, A.E., Stotler, R.L., Reboulet, E.C., 2017. Interpreting temporal variations in river response functions: an example from the Arkansas River, Kansas, USA. Hydrogeol. J. 25, 1271–1282. doi:10.1007/s10040-017-1545-9

Conrad, J., Nel, J., Wentzel, J., 2004. The challenges and implications of assessing groundwater recharge: A case study-northern Sandveld , Western Cape, South Africa. Water SA 30, 75–81.

Crosbie, R.S., Binning, P., Kalma, J.D., 2005. A time series approach to inferring groundwater recharge using the water table fluctuation method. Water Resour. Res. 41.

DWAF, 2003. Sandveld Preliminary (Rapid) Reserve Determinations. Langvlei, Jakkals and Verlorenvlei Rivers. Olifants-Doorn WMA G30. Surface Volume 1: Final Report Reserve Specifications. DWAF Project Number: 2002-227.

Fay, P.A., Guntenspergen, G.R., Olker, J.H., Johnson, W.C., 2016. Climate change impacts on freshwater wetland hydrology and vegetation cover cycling along a regional aridity gradient. Ecosphere 7.

Fischer, C., 2013. Automatische Kalibrierung hydrologischer Modelle: Entwicklung und Anwendung des Kalibrierugnssystems OPTAS.

Flügel, W., 1995. Delineating hydrological response units by geographical information system analyses for regional hydrological modelling using PRMS/MMS in the drainage basin of the River Bröl, Germany. Hydrol. Process. 9, 423–436.

Forrer, I., Kasteel, R., Flury, M., Flühler, H., 1999. Longitudinal and lateral dispersion in an unsaturated field soil. Water Resour. Res. 35, 3049–3060.

Giertz, S., Diekkrüger, B., Steup, G., 2006. Physically-based modelling of hydrological processes in a tropical headwater catchment (West Africa)? process representation and multi-criteria validation. Hydrol. Earth Syst. Sci. Discuss. 10, 829–847.

Healy, R.W., Cook, P.G., 2002. Using groundwater levels to estimate recharge. Hydrogeol. J. 10, 91–109. doi:10.1007/s10040-001-0178-0

Holloway, A., Fortune, G., Chasi, V., 2010. RADAR Western Cape risk and development annual review disaster mitigation for sustainable livelihoods programme. PeriPeri Publications, Rondebosch.

Hornberger, G.M., Spear, R.C., 1981. Approach to the preliminary analysis of environmental systems. J. Environ. Mgmt. 12, 7–18.

Hughes, D.A., 2004. Incorporating groundwater recharge and discharge functions into an existing monthly rainfall–runoff model/Incorporation de fonctions de recharge et de vidange superficielle de nappes au sein d'un modèle pluie-débit mensuel existant. Hydrol. Sci. J. 49.

Hughes, D.A., 2001. Providing hydrological information and data analysis tools for the determination of ecological instream flow requirements for South African rivers. J. Hydrol. 241, 140–151. doi:10.1016/S0022-1694(00)00378-4

Jenson, S.K., Domingue, J.O., 1988. Extracting topographic structure from digital elevation data for geographic information system analysis. Photogramm. Eng. Remote Sensing 54, 1593–1600.

Johnson, A.I., 1967. Specific yield: compilation of specific yields for various materials. US Government Printing Office.

Johnson, M.R., Anhauesser, C.R., Thomas, R.J., 2006. The Geology of South Africa. Geological Society of South Africa.

Krause, P., 2001. Das hydrologische Modellsystem J2000. Beschreibung und Anwendung in großen Flussgebieten, in: Umwelt/Environment, Vol. 29. Jülich: Research Centre.

Krause, P., Bäse, F., Bende-Michl, U., Fink, M., Flügel, W., Pfennig, B., 2006. Multiscale investigations in a mesoscale catchment? hydrological modelling in the Gera catchment. Adv. Geosci. 9, 53–61.

Lynch, S., 2004. Development of a raster database of annula, monthly and daily rainfall for southern Africa. Water Res. Comm. 108. doi:10.1024/0301-1526.32.1.54

Meinzer, O.E., Stearns, N.D., 1929. A study of groundwater in the Pomperaug Basin, Conn. with special reference to intake and discharge. US Geol. Surv. Water-Supply Pap. 73–146.

Nash, J.E., Sutcliffe, J. V, 1970. River flow forecasting through conceptual models part I—A discussion of principles. J. Hydrol. 10, 282–290.

Nepal, S., 2012. Evaluating Upstream-Downstream Linkages of Hydrological Dynamics in the Himalayan Region submitted to the Faculty of Chemical and Earth Sciences of the. 2012 262.

Pfannschmidt, K., 2008. Optimierungsmethoden zur HRU-basierten N/A-Modellierung für eine operationelle Hochwasservorhersage auf Basis prognostischer Klimadaten des Deutschen Wetterdienstes: Untersuchungen in einem mesoskaligen Einzugsgebiet im Thüringer Wald.

Pfennig, B., Kipka, H., Wolf, M., Fink, M., Krause, P., Flügel, W. a., 2009. Development of an extended routing scheme in reference to consideration of multi-dimensional flow relations between hydrological model entities. 18th World IMACS Congr. MODSIM09 Int. Congr. Model. Simulation, Cairns, Aust. 1972–1978.

Pianosi, F., Beven, K., Freer, J., Hall, J.W., Rougier, J., Stephenson, D.B., Wagener, T., 2016. Sensitivity analysis of environmental models: A systematic review with practical workflow. Environ. Model. Softw. 79, 214–232.

Pike, A., Schulze, R.E., 1995. AUTOSOILS: A program to convert ISCW soils attributes to variables usable in hydrological models. Pietermaritzburg, South Africa Univ. KwaZulu-Natal, Sch. Bioresour. Eng. Environ. Hydrol.

Potatoes South Africa, 2015. 2014/2015 Potato Industry Report.

Přibáň, K., Ondok, J.P., 1985. Heat balance components and evapotranspiration from a sedge-grass marsh. Folia Geobot. 20, 41–56.

Rantz, S.E., 1982. Measurement and computation of streamflow: Volume 1, Measurement of stage and discharge, Water Supply Paper.

Rozendaal, A., Gresse, P.G., 1994. Structural setting of the Riviera W-Mo deposit, western Cape, South Africa. South African J. Geol. 97, 184.

Rushton, K., 1997. Recharge from permanent water bodies., in: Recharge of Phreatic Aquifers in (Semi)arid Ar- Eas. AA Balkema, Rotterdam, pp. 215–255.

Scanlon, B.R., Healy, R.W., Cook, P.G., 2002. Choosing appropriate techniques for quantifying groundwater recharge. Hydrogeol. J. 10, 18–39. doi:10.1007/s10040-0010176-2

Schulze, R.E., 1995. Hydrology and agrohydrology: A text to accompany the ACRU 3.00 agrohydrological modelling system. Water Research Commission.

Schulze, R.E., Maharaj, M., Warburton, M.L., Gers, C.J., Horan, M.J.C., Kunz, R.P., Clark, D.J., 2007. South African atlas of climatology and agrohydrology. Water Res. Comm. Pretoria, RSA, WRC Rep. 1489, 6.

Schuyt, K.D., 2005. Economic consequences of wetland degradation for local populations in Africa. Ecol. Econ. 53, 177–190. doi:10.1016/j.ecolecon.2004.08.003

Šimůnek, J., van Genuchten, M.T., Sejna, M., 2006. The HYDRUS Software Package for Simulating the Two- and Three-Dimensional Movement of Water, Heat, and Multiple Solutes in Variably-Saturated Porous Media.

Spane, F.A., Mackley, R.D., 2011. Removal of River-Stage Fluctuations from Well Response Using Multiple Regression.

Ground Water 49, 794–807. doi:10.1111/j.1745-6584.2010.00780.x

Steudel, T., Bugan, R., Kipka, H., Pfennig, B., Fink, M., de Clercq, W., Flügel, W.-A., Helmschrot, J., 2015. Implementing contour bank farming practices into the J2000 model to improve hydrological and erosion modelling in semi-arid Western Cape Province of South Africa. Hydrol. Res. 46, 192. doi:10.2166/nh.2013.164

Ting, C.-S., Kerh, T., Liao, C.-J., 1998. Estimation of groundwater recharge using the chloride mass-balance method, Pingtung Plain, Taiwan. Hydrogeol. J. 6, 282–292.

Wheater, H., Sorooshian, S., Sharma, K.D., 2007. Hydrological modelling in arid and semi-arid areas. Cambridge University Press.

Willmott, C.J., 1981. On the validation of models. Phys. Geogr. 2, 184–194.

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Combining rainfall/runoff and groundwater modelling to calculate recharge and baseflow in the data scarce Verlorenvlei catchment South Africa.

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ABSTRACT

Exploitation of groundwater resources has the potential to reduce baseflow from aquifers and impact ecosystem functioning. Areas with high biodiversity hotspots such as wetlands and estuarine lakes are particularly vulnerable. The Verlorenvlei estuarine lake (RAMSAR #525), north of Cape Town, South Africa, is a sensitive biodiversity hotspot but agricultural production within the sub-catchment impacts baseflow generation that sustains the lake during the dry season. Potential recharge from a J2000 rainfall/runoff model for the sub-catchment was used to calibrate net recharge within a MODFLOW model. The model transfer highlighted the need for the inclusion of distributive aquifer properties for groundwater recharge determination. Net recharge was estimated between 0.3-7.0 % in the valley and 7.0-11.7% in the mountains from 2010-2016. The abstraction scenarios used resulted in 40.0 % of recharge being utilised, with more than half the recharge coming via the Table Mountain Group aquifer. The changes in groundwater recharge affected aquifer baseflow by as much as 22.0 %, although without more water level data baseflow reductions cannot be quantified. This study demonstrated that changes in rainfall and net recharge resulted in significant baseflow variability, which could be detrimental to ecosystems that rely on a constant groundwater baseflow supply to maintain sufficient biodiversity levels..

1. Introduction

Climate change models for southern Africa, indicate that rainfall will be reduced and subject to more variability in the future (Haarsma et al., 2005; McMichael et al., 2006; Rotstayn et al., 2010; Abiodun et al., 2017). Reduced rainfall impacts the replenishment of surface water reservoirs, leading to potential supply curtailments in urban centres. Such a situation is currently occuring in the city of Cape Town, South Africa, where several years of below average rainfall have resulted in extreme water shortages to about 4 million people. Groundwater has long been considered an alternative water source when surface water resources are limited, but sustainability of the groundwater resource must be taken into account when determining abstraction volumes.

Groundwater abstraction volumes are considered sustainable when abstractions do not exceed recharge and thereby negatively impact the ecological reserve (the amount of water needed to sustain the natural environment) (Zhou, 2009). Over-abstraction of groundwater can lead to significant groundwater depletion (Konikow and Kendy, 2005; Wada et al., 2010), which impacts baseflow to streams, and hence the ecological reserve. Over-abstraction has long lasting, adverse impacts on the health of river and lake systems and the ecological communities they support (e.g Weber and Perry, 2006). Consequently, before groundwater can be used as a supplementary water source during times of drought, sustainable groundwater abstractions need to be quantified. Whilst a range of hydrological and ecological factors contribute to the assessment of sustainability, one of the most important to quantify is baseflow under both high and low flow conditions.

The most common method for estimating baseflow is to use streamflow-gauging measurements, to separate a streamflow hydrograph into runoff and baseflow components. Although there are a variety of hydrograph separation methods, they all require accurate daily streamflow measurements (Arnold and Allen, 1999). In arid and semi-arid regions of sub-Saharan Africa, although reliable rainfall and climate records are usually available, continuous streamflow records are generally limited (e.g Wessels and Rooseboom, 2009). This means that traditional hydrograph separation methods are difficult to apply and instead hydrological modelling approaches, which include estimates of recharge, are needed. As a result, computational modelling to predict and understand groundwater recharge and baseflow has become common practice, especially for data scarce catchments (e.g Bulygina et al., 2009; Ndomba et al., 2008).

In data scarce catchments, a variety of assumptions must be made for models to produce sensible outputs. The two main types of models used to simulate baseflow are rainfall/runoff and groundwater models. Rainfall/runoff models can use available climate records to estimate recharge (eg Watson et al., 2018) and thereafter baseflow (eg Arnold et al., 2000). These distributive models use hydrological response units (HRUs) for surface water routing, to simulate daily water balances (Flügel, 1995). However, groundwater processes are occasionally lumped, with a single hydraulic conductivity value to control the proportion of interflow to recharge throughout the modelling domain. With this modelling approach, recharge is defined as potential recharge (Scanlon et al., 2001) as opposed to net recharge which incorporates aquifer hydraulic properties (Kim et al., 2008). In reality, multiple hydraulic conductivity rates are usually contained within a catchment, and potential recharge from rainfall/runoff models requires further processing before net recharge estimates can be quantified. In comparison, while groundwater models are distributive with the way aquifer and hydraulic properties are treated, they occasionally lump net recharge, which neglects surface water variability (Kim et al., 2008).

Combining rainfall/runoff models with groundwater models, offers the potential for improved modelling outputs in data scarce catchments. The transfer of potential recharge from rainfall/runoff models to groundwater models allows simulation of both climate and groundwater components in a distributive manner, thereby reducing model output uncertainties and equifinality. While a variety of different distributive groundwater models are available (eg. finite difference and finite element), MODFLOW, which is the groundwater model used in this study, is a three-dimensional groundwater model that uses a finite-difference approach to calculate flow through porous media by matching modelled to observed water levels (Harbaugh et al., 2000).

This study used a distributive modelling approach to calculate baseflow between 2010 and 2016 for the Krom Antonies tributary, one of the main tributaries feeding the Verlorenvlei RAMSAR listed estuarine lake on the west coast of South Africa. The Krom Antonies is sustained by baseflow during the dry season, but Verlorenvlei is also an important agricultural area supported in part by groundwater abstraction. The modelling results will be used to constrain the baseflow requirements needed to ensure sustainable resource usage required for the long-term survival of the Verlorenvlei estuarine system. This approach took the potential recharge, modelled using the J2000 rainfall/runoff model (Watson et al., 2018) and converted it to net recharge using a MODFLOW groundwater model and PEST autocalibration. Thereafter, baseflow for the Krom Antonies tributary was modelled. The results help to understand the impact of changing climate on baseflow generation, especially when many ecosystems are on the tipping point of becoming severely altered by over allocation for socio-economic development and drought remediation.

2. Study site

The Verlorenvlei estuarine lake is located along the west coast of South Africa (Fig. 1). The lake itself begins west of the town of Redelinghuys, but drains a sub-catchment of 1832 km^2, extending inland to the Piketberg Mountains. The lake is fed by four tributaries, the Krom Antonies, Hol, Kruismans and Bergvallei. The Hol, Krusimans and Bergvallei tributaries are more saline in comparison to the Krom Antonies (Sigidi, 2018). The salt concentrations of surface and groundwater become increasingly elevated with proximity to the confluence of the Hol, Krom Antonies and Kruismans tributaries (Sigidi, 2018). The Krom Antonies sub-catchment drains an area of around 140 km^2 with the majority of water derived from the Piketberg Mountains.

2.1. *Climate*

The west coast of South Africa is subject to a Mediterranean climate that receives most rainfall during the winter months (May to September). The Piketberg Mountains, with the highest elevation (1446 masl), receive the highest amount of rainfall within the sub-catchment, with an average of 650 mm.yr-1 between 2010 and 2016 (Supplementary Table 1). In comparison, the valley receives significantly less rainfall, due to its lower elevation (120 masl), with an average of 350 mm between 2010 and 2016 (Supplementary Table 1). The annual rainfall within the sub-catchment varied by as much as 50 % between 2010-2016. Particularly dry years were 2011, 2015 and 2016, with wet years being 2010, 2013 and 2014.

The average daily temperatures within the sub-catchment in summer are between 20-30°C with a relative humidity of 50-60%. In comparison, winter temperatures are between 12-20°C with relative humidity of 60-80% (Muche et al., 2018). As a result of the high atmospheric demand for water in summer daily evaporation is between 4 to 6 mm.d-1, while in winter moderate atmospheric demand results in daily evaporation of between 1 to 3 mm.d-1 (Watson et al., 2018).

2.2. *Hydrogeology*

Within the region there are essentially four geological packages (from youngest to oldest): (1) Quaternary alluvial sediments along the valley floors; (2) fractured sandstones and associated rocks of the Cambrian Table Mountain Group (TMG) that make up the high elevation parts of the sub-catchment; (3) the Cambrian Cape Granite Suite, which intrudes (4) the Neoproterozoic Malmesbury Group (MG) rocks, which consist of greywacke, sericitic schist, quartzite, conglomerate and limestone (Rozendaal and Gresse, 1994). In terms of the hydrogeology of the area, the TMG consists of a variety of fractured sandstone and shales belonging to three main formations, namely: 1) the Peninsula Formation, 2) the Graafwater Formation and 3) the Piekernierskloof Formation. While the Peninsula and Piekernierskloof sandstones have well developed interconnected fractures and are considered aquifers,

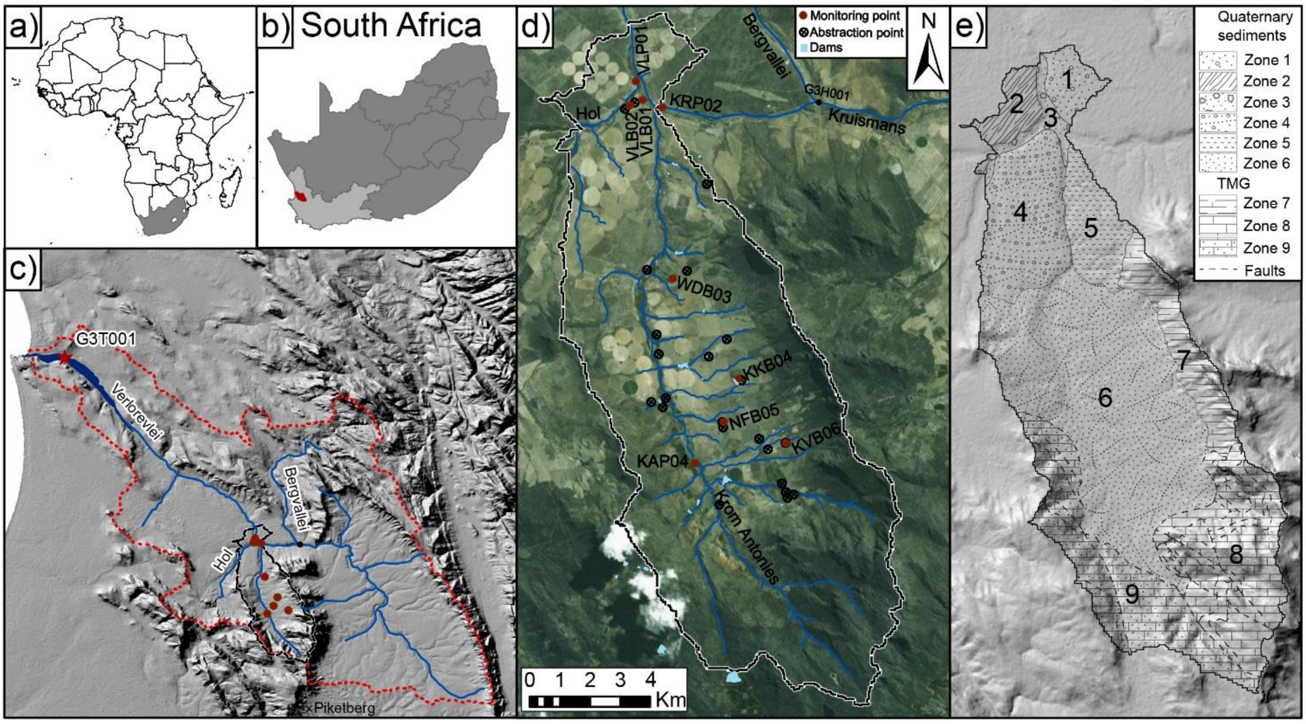

Fig. 1. a) Location of South Africa within Africa, b) Location of Western Cape and Verlorenvlei within South Africa, c) Extent of the Verlorenvlei catchment showing weir G3T001 as well as the main feeding rivers, d) Extent of the Krom Antonies catchment showing monitoring boreholes, abstraction points and dams (after Watson et al., 2018) and e) model grid showing the boundary of the modelled catchment including the primary aquifer model zones (1-6) and TMG aquifer zones (7-9).

the Graafwater as a shale dominated unit, is more aquitard in nature. In comparison, the floor of the valley is comprised of Quaternary sediments and clean sands that make up a primary aquifer. The primary aquifer is relatively high yielding and although it is commonly saline (Sigidi, 2018), it is frequently used for irrigation purposes. Underlying the TMG aquifer and the primary aquifer is the Neoproterozoic Malmesbury Group, referred to as a secondary aquifer because of the dominance of shale horizons and the channelization of groundwater along cleavage planes and fractures. For further hydrological, hydrogeological, climate and vegetation information regarding the Verlorenvlei, refer to Watson et al. (2018).

2.3. *Landuse and agriculture*

The major water user in the area is agriculture, accounting for 90 % of total water requirements (DWAF, 2003). The region is a well-known potato production area, where around 6600 hectares of produce are planted annually. Potato production has been estimated to use up to 20 % of total recharge (DWAF, 2003). The favoured irrigation system for potatoes is centre pivots, where the crops are irrigated with streamflow from the estuarine lake and the main feeding tributaries during winter. During summer when there is reduced streamflow, groundwater is abstracted to support the agricultural development. Other important agricultural users include viticulture, citrus, and tea. The detailed overall water consumption of these crops in the catchment has not been properly documented. Borehole pumping volumes estimated by the farmers throughout 2016 range from around 20 m3/day to 1680 m3/day between January and May, with a maximum sub-catchment draw of around ±7000 m3/day from the secondary aquifer.

3. Conceptual model

The conceptual model includes the TMG aquifer, the primary aquifer (alluvial) and the secondary aquifer (MG) (Fig. 2 and Fig. 3). The primary and TMG aquifers are both treated as unconfined aquifers as groundwater flow is not held between confining units. Although the TMG aquifer has been called semi-confined as a result of differences in the hydraulic conductivity of the different formations, the Peninsula Formation is dominant in the field area and is regarded as largely unconfined (Lin, 2007). The secondary aquifer is treated as a semi-confined aquifer with groundwater flow being restricted by the shale-dominated character of host lithologies.

Recharge is predominantly received in the mountains, which feed into the TMG aquifer. The valley which feeds the alluvial aquifer, is subject to significantly less recharge as yearly evapotranspiration is equivalent to rainfall. Watson et al. (2018) modelled 28.9 % of potential recharge received in the wettest region in the mountains and 1-7% of potential recharge received in the valleys for 2016. The secondary aquifer is not directly recharged by rainfall but rather via groundwater flow from the TMG and primary aquifers (Fig. 2).

The spatial extent of groundwater flow is influenced by the presence of the Piketberg Mountains to the south-west and south-east of the Krom Antonies which form the boundaries of the sub-catchment (Fig. 3). The Krom Antonies was modelled as a gaining system, and where it exits it represents an accumulation of all flow within the sub-catchment. Surface runoff in the wet season is generated from the valley and mountains. Rainfall that does not percolate through to the TMG or the secondary aquifer contributes to streamflow but may also be evaporated (Fig. 2b). During the dry season, baseflow is responsible for supplying water to the Krom Antonies. In the smaller tributaries of the Krom Antonies there is little to no flow during the dry season, with the main tributary drying up for a few months of the year.

The ability of the aquifer to supply baseflow to the Krom Antonies tributary is hindered when pumping stresses are applied to abstraction points into both the primary and secondary aquifers. Theoretically this could induce groundwater flow from the TMG aquifer to the secondary or primary aquifers, thereby lowering the water table and reducing the stream stage (Fig. 2c). Dams and reservoirs constructed along flow networks can reduce baseflow to the Krom Antonies tributary by limiting groundwater flow beneath the reservoirs due to saturated conditions. Dam spillways are also responsible for retarding surface water flow until reservoir levels are sufficient to allow for overflow. In summary, the conceptual model of the sub-catchment includes the spatial distribution of recharge across the different aquifer types, the baseflow generation mechanism for the Krom Antonies tributary and factors that may reduce aquifer baseflow such as groundwater abstraction and baseflow barriers such as dams.

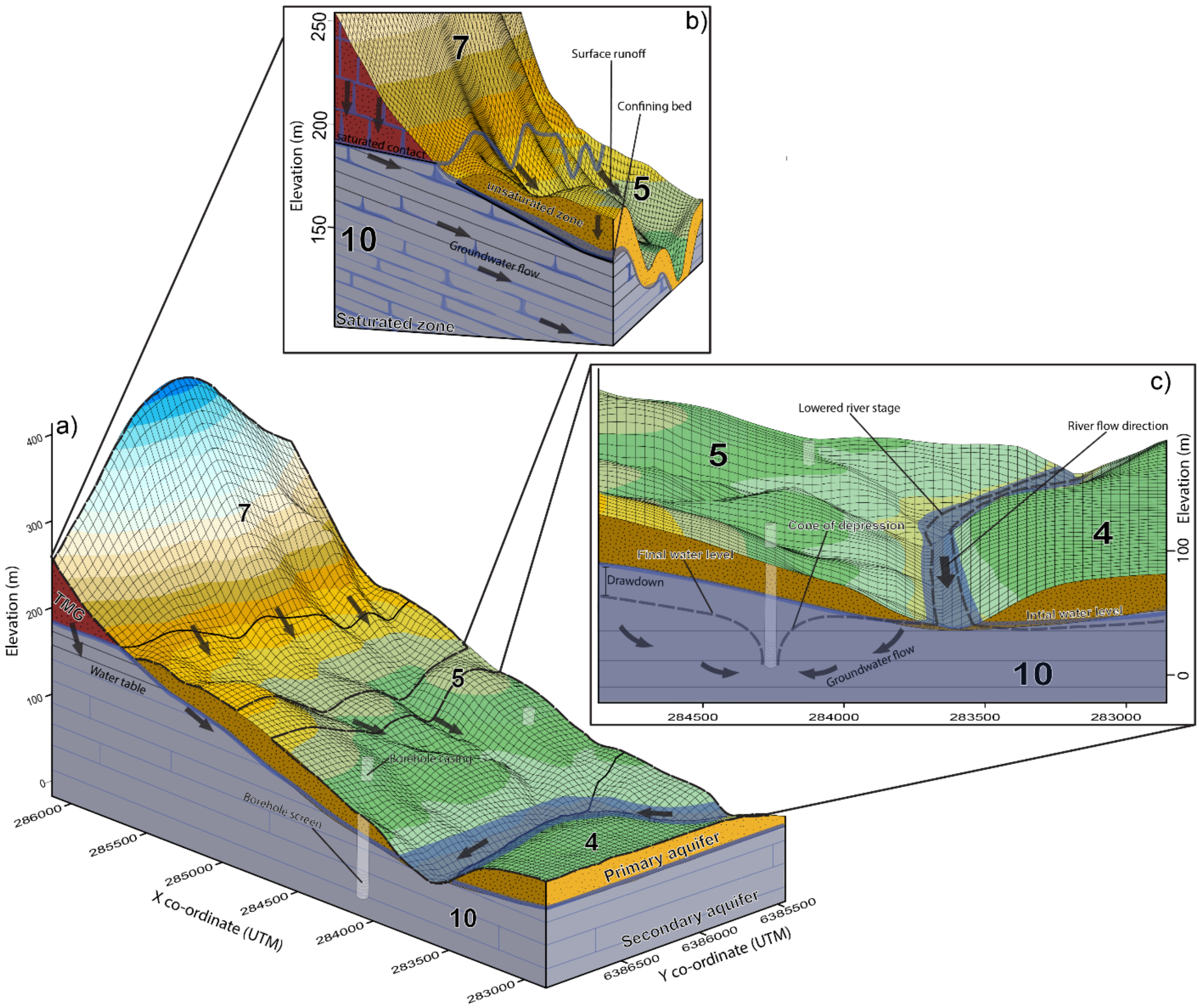

Fig. 2. a) Conceptual model of the Krom Antonies catchment with main aquifer types, b) Mechanism of transfer of groundwater from the TMG into the secondary and primary aquifer, c) Gaining stream of the main tributary illustrating the influence that abstraction could potentially have on the catchment baseflow.

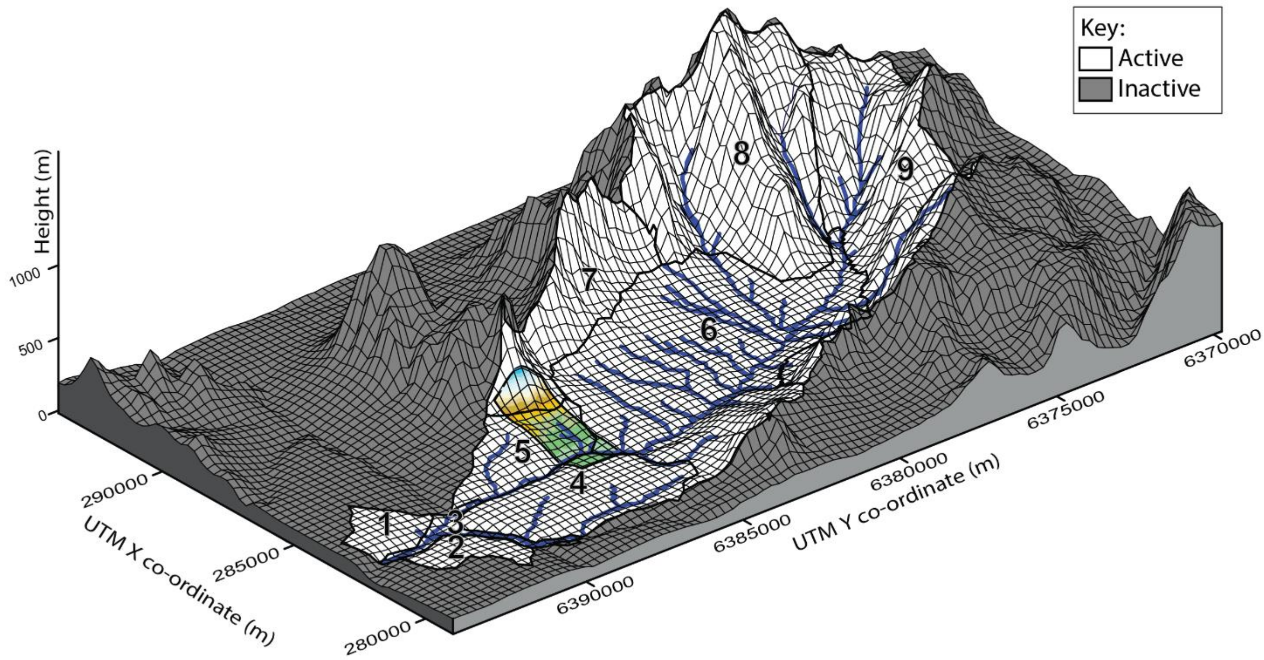

Fig. 3. The MODFLOW sub-catchment hydraulic zones that form the net recharge and hydraulic conductivity boundaries with the primary aquifer (hydraulic zones 1-6), the TMG aquifer (hydraulic zones 7-9) and the secondary aquifer (hydraulic zone 10) (not shown).

4. MODFLOW Methodology

4.1. *Input data*

4.1.1. Lithology units and faults

The extent of the primary and TMG aquifers, the presence of faults, dip direction and strike orientations were obtained from geological and land-type classification maps (De Beer, 2003; Pike and Schulze, 1995) as well as limited borehole logs. These datasets indicate that the primary aquifer has six different hydraulic zones (zones 1-6) on the basis of soil type variations, while the TMG aquifer is divided into three hydraulic zones (zones 7-9) because of the presence of large faults (Fig. 1e). The secondary aquifer underlying both the TMG aquifer and primary aquifer is incorporated within one zone (zone 10) because there is not enough geological information and very limited outcrop, to warrant dividing into additional hydraulic zones. Hydraulic zones 1-9 are modelled as an unconfined aquifer layer within the conceptual model whilst hydraulic zone 10 is modelled as a confined aquifer layer. Transferring these hydraulic zones into a numerical model requires estimating the average thickness of these two layers. The primary aquifer was assumed to be a uniform 15 m thick which was determined by drill chip analysis at location KKB04 (Fig. 1). The thickness of the TMG aquifer was determined using published geological data (De Beer, 2003) and interpolation techniques. The thickness of the secondary aquifer was determined using interpolation of borehole casing depths as well as borehole drill logs from the nearby town of Piketberg (SRK, 2009). An average thickness of 70 m was used as an approximation, and derived from 22 drill records, hydrogeological consulting reports (SRK, 2009) and National Groundwater Archive (NGA) casing depths. The NGA shows thickness of between 20-50 m in the study area but the Malmesbury Group shale itself can be significantly thicker in other regions with some farmers reporting drilling through over 300 m of shale. Therefore, the 70 m used in this study is likely to be a minimum thickness.

4.1.2. Hydraulic Properties

The hydraulic zones 1-10 are differentiated in the numerical model by hydraulic conductivity, storage coefficient and specific yield of the individual zones. A range of material hydraulic properties were obtained from theoretical and previous studies (Table 1), which allowed for the model to vary these parameters during calibration to reduce residuals between modelled and observed water levels. The TMG aquifer has a relatively high hydraulic conductivity and low storage capacity due to its fractured nature. Previous estimates of hydraulic conductivity of TMG formations varied greatly due to different calculation techniques and locations. Estimates of hydraulic conductivity for the Peninsula Formation are between 1.99 to 1.99×10^{-3} m.d^{-1} (UMVOTO-SRK, 2000), although other studies have much lower estimates (e.g 0.0007 m.d^{-1}; Xu et al., 2009). Storage coefficient values range from 1.3×10^{-3} m to 7×10^{-4} for the TMG aquifer (UMVOTO-SRK, 2000). The secondary aquifer is dominated by phyllite shale/schist, which has a low hydraulic conductivity with values of 0.1-0.001 m.d^{-1} (SRK, 2009) in the direction of the bedding plane. Other relatively low storage shale formations in the Karoo (South Africa) are between 0.01-0.0001 m.d^{-1}

(Sami, 1996), with estimates from shale elsewhere in the world between 0.0001-0.00005 (Williams and Paillet, 2002).

4.1.3. *Potential recharge*

Potential recharge for the Krom Antonies sub-catchment was modelled between 2010 and 2016 by Watson et al. (2018) using the J2000 rainfall/runoff model (Supplementary: Table 1). Potential recharge was modelled per HRU but generally divides between high potential recharge in the mountains and low potential recharge in the valleys. During the wet years (2010, 2013, 2014) potential recharge for the sub-catchment was between 3-255 $mm.yr^{-1}$, with the Piketberg Mountains receiving 162-255 $mm.yr^{-1}$ and the valley 3-47 $mm.yr^{-1}$ (Supplementary: Table 1) . During the dry years (2011, 2015, 2016) potential recharge for the sub-catchment was between 5 and 150 $mm.yr^{-1}$, with the Piketberg Mountains receiving 114-150 $mm.yr^{-1}$ and the valley receiving 5-45 $mm.yr^{-1}$(Supplementary: Table 1). Based on these results, average potential recharge is between 22.6 and 25.1 % of rainfall for the Piketberg Mountains, and between 0.8 and 5.1 % of rainfall in the valley (Supplementary: Table 1).

4.1.4. *Water levels*

Groundwater levels in boreholes and piezometers in the primary and secondary aquifers were used for model as the National Groundwater Archive (NGA) were used for steady state calibration. Forty static water levels were available for calibration with 11 in the primary aquifer and 29 in the secondary aquifer (Fig. 4). The static water levels were weighted equally (weight 1) to reduce bias in the calibration, although primary aquifer water levels were censored to represent the maximum water level, as it is likely that streamflow impacted these measurements. For transient state calibration, dynamic water levels for both the primary and secondary aquifers were used. Actual water level data used for both steady state and transient state calibration were taken from Watson et al. (2018).

4.2.*Model Setup*

4.2.1. *Model grid*

The MODFLOW-NWT (Newton solver) numerical model was used to model groundwater flow within the Krom Antonies sub-catchment. The model domain consisted of 97172 cells, of which 316 columns by 204 rows covered an area of 140 km2. The pixel size of the cells in the catchment was 120 m, where refinement to 60 m was made in the middle of the catchment near observation targets and pumped boreholes. The input digital elevation model (DEM) was the aggregated Stellenbosch University Digital Elevation Model (SUDEM), which has a 5 m resolution (van Niekerk and Joubert, 2011).

4.2.2. *Catchment boundaries*

The Piketberg Mountains form the eastern, southern and western portions of the model boundary with the northern boundary (where the tributary exits the sub-catchment) defined by topographic slope along the valley ridges. In the model, this boundary is represented by a no-flow boundary, which restricts groundwater flow to active cells, and forces all water to exit the sub-catchment via the tributary which is represented as a river (Fig. 3). Dip direction of the bedding plane could theoretically influence groundwater flow into or out of the catchment. However, a methodology has not yet been developed to incorporate this complexity within the groundwater model and therefore the catchment boundary was created based solely on drainage of surface water flow into the sub-catchment (Watson et al., 2017). Ongoing refinement of the model will attempt to address this limitation in the sub-catchment model boundary conditions.

Table 1
Results from the model calibration showing literature (a: Domenico and Schwartz (1990), b: Driscoll, (1986), c: UMVOTO-SRK (2000):, d: Lin (2008), e: SRK (2009), f: Sami (1996), g: Williams and Paillet (2002)) and calibrated hydraulic conductivity (K), storage coefficient/specific storage (Ss) and specific yield (Sy) for the primary aquifer (Quaternary sediments), TMG and secondary aquifer (Malmesbury shale). Where * represents unoptimized units.

			Hydraulic conductivity (m/day)			Storage coefficent/specific storage			Specific yield		
Layer	Aquifer Host rock	Hydraulic Zones	Literature range	Initial	Calibrated	Literature range	Initial	Calibrated	Literature range	Initial	Calibrated
1	Quaternary sediments	1-6	0.0002-50[a]	0.52-1.97	0.01-1				0.15-0.25[b]	0.2	0.25*
1	TMG	7-9	0.00199-1.99[c]	1.00	0.7-1.2				0.0001-0.001[d]	0.001	0.00011-0.0036*
2	Malmesbury Shale	10	0.1-0.001[e]	0.0505	0.0021	0.1-0.001[e] 0.01-0.0001[f], 0.0001-0.00005[g]	0.0000025	0.000003			

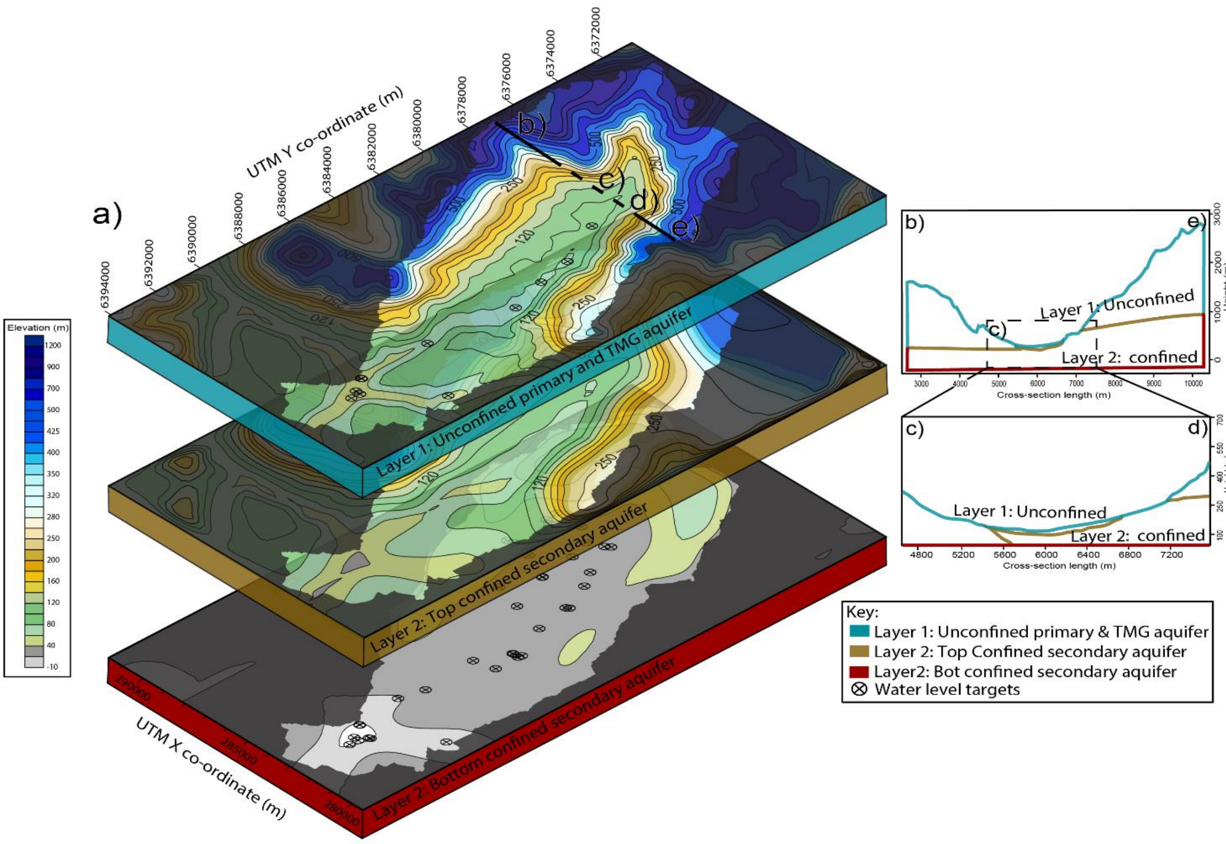

Fig. 4. Simplified model layers, thicknesses and aquifer type showing 11 primary and 29 secondary aquifer targets (static water levels).

4.2.3. Aquifer baseflow and drawdown

The elevation of the aquifer water level at any given location was determined by taking 2016 abstraction records for thirty actively pumped boreholes within the sub-catchment and assuming that the same abstraction volumes per month per borehole applied for the period 2010-2015. This assumption is regarded as valid as farmers in the area have a consistent irrigation programme from year to year and typically do not do crop rotation. These abstraction scenarios were calculated using multi-node wells (MNW) (Halford and Hanson, 2002) so that a single well (ie borehole) could penetrate more than one model unit. Using this approach, the MNW package was able to remove water from both the primary and secondary aquifer. This assumed that abstraction was from both the primary and secondary aquifer based on yield and water quality.

When the aquifer water level elevation is beneath the topographic elevation of the river, water in the river will move into the aquifer, resulting in a losing stream being simulated. For water in the river to flow correctly and exit the catchment, the topographic elevation of the upstream river cell needs to be higher than the cell adjacent to it. Improvement of the river elevation was achieved through manual removal of depressions and high points from the SUDEM to ensure correct river functioning. Using the river boundary condition, MODFLOW calculates baseflow (B) in $m^3.s^{-1}$ as:

$$B = C \times \left(\frac{Rh}{dh}\right) \quad (1)$$

where, C is the aquifer conductance, Rh is the river stage elevation set by the user, and dh the change in hydraulic head of the aquifer (in m). For baseflow to be generated, $\left(\frac{Rh}{dh}\right) > 1$. The aquifer conductance is calculated as:

$$C = \frac{KLW}{D} \quad (2)$$

where, K is the hydraulic conductivity of the river bed material in m.d-1, L is the length of the river in the cell (60 m in the middle of sub-catchment and 120 m on the outskirts), W is the width of the river in the cell and D is the thickness of the river bed material (15 m). The length of the river and thickness of the stream bed were derived from grid size and layer thickness with saturated soil hydraulic conductivity being determined from the soil-plant-air-water (SPAW) pedotransfer function model (Saxton and Willey, 2005). The river widths were determined using a multi-routing algorithm (Pfennig et al., 2009) based on the SUDEM to determine flow accumulation and sub-catchment basin areas. Stream width (W) was calculated according to:

$$W = a \times area^b \quad (3)$$

where $area$ is the sub-basin extent at each reach outlet (km^2) and a and b are calibration factors with default values of 0.824 and 0.674 based on previous studies (Pfennig et al., 2009). The area of the Krom Antonies sub-basin was determined using:

$$area = \frac{fa \times grid\ size}{1000000} \quad (4)$$

where, fa is the flow accumulation (number of grid cells that contribute to flow in a certain cell), and $grid\ size$ is in m^2 (Pfennig et al., 2009).

4.2.4. *Open water bodies and dams*

Open water bodies or reservoirs have an impact on baseflow generation within the study area sub-catchment, with the majority of dams having inadequate spillways and being positioned in the main drainage networks of the Krom Antonies tributary (Fig. 1). Although a time varying constant head would allow the head to be set during different stages of the simulation making it theoretically more representative of reservoir water levels, actual water levels are not recorded in this sub-catchment. Hence, a constant head approach was used to simulate the baseflow impact in MODFLOW. For each reservoir the constant head was the average elevation across the area of the reservoir taken from the SUDEM plus the height of the dam wall. The Krom Antonies sub-catchment has 13 reservoirs (Fig. 1) that are used to supplement irrigation during summer, when rainfall is limited. The surface area of the reservoirs was digitized using aerial photographs. The dams vary in size from 2424 m^2 to 123 819 m^2.

4.2.5. *Translation of HRUs to Hydraulic Zones*

Watson et al. (2018) modelled potential recharge in 596 HRUs in the Krom Antonies sub-catchment where each HRU was delineated using the SRTM DEM and assigned soil type, land type, geology and climate properties. Climate properties were based on measured data from nearby weather stations with reference evaporation calculated using the Penman Monteith equation and HRU rainfall corrected for elevation. To translate the 596 HRUs into the identified 9 hydraulic zones, yearly potential recharge and rainfall from 2010 to 2016 were clipped for each hydraulic zone and sorted into upper, median and lower quartile ranges (Supplementary: Fig. 1). The upper and lower quartiles of potential recharge for each hydraulic zone were the ranges used during net recharge calibration for 2016 (Table 1).

4.3. *Model sensitivity*

A composite sensitivity analysis (SENSAN: Doherty, 2016) was conducted on recharge (R), vertical (Kz) and horizontal (Kx,y) hydraulic conductivity, the storage coefficient (Ss) and specific yield (Sy) (Table 2). In general, highly sensitive parameters are greater than 10% of the maximum composite sensitivity, while parameters with less than 1% are considered insensitive (Necpálová et al., 2015; Hill and Tiedeman, 2006). This relies on having an equal distribution of observation points in all zones which was not the case here. Parameters identified as sensitive, that impact calibration, were optimised for both steady and transient state to reduce the residual between observed and modelled values. Poor data coverage had a significant impact on sensitivity results for transient state calibration of Ss and Sy. The primary and secondary aquifers were therefore optimised for both horizontal (Kx,y) and vertical hydraulic conductivity (Kz) in all hydraulic zones (1-10). Net recharge was optimised for hydraulic zones 3, 5, 7, 8 and 9. During transient state calibration, only zone 10 had enough data to calibrate Ss. No other zones had sufficient observation points to allow for calibration of Ss and assessment of sensitivity to Ss. Moreover, none of the primary aquifer zones (1-9) could be calibrated for Sy. The impact of this is assessed in the discussion.

Table 2
Local sensitivity analysis (SENSAN) produced during PEST (Watermark computing) autocalibration of horizontal hydraulic conductivity (Kx), vertical hydraulic conductivity (Kz), net recharge (R), Storage coefficient (Ss) and Specific yield (Sy) for model zones 1-10.

	Steady state			Transient state	
Zone	R	Kx	Kz	Ss	Sy
1	0.711	0.254	1.043		0.000
2	0.711	0.468	0.780		0.002
3	0.349	7.066	0.271		0.000
4	0.185	3.779	3.002		0.001
5	0.654	1.139	1.139		0.000
6	0.200	1.155	19.967		0.006
7	0.088	17.647	4.669		0.000
8	0.272	0.673	32.783		0.000
9	0.392	1.372	1.748		0.000
10		1.442	3.543	0.500	

4.4. *PEST autocalibration*

4.4.1. *Net recharge and hydraulic conductivity calibration*

Net recharge and hydraulic conductivity were calculated in steady state calibration (Fig. 5). Potential recharge in the J2000 model is defined as:

$$R_{pot} = P - [I + Q + ET] - IT_f \quad (6)$$

where R_{pot} is potential recharge, P is precipitation, I is vegetation interception, Q is surface runoff, ET is evapotranspiration and IT_f is interflow. In the J2000 model P, I, Q and ET are defined per HRU and are hence variables, but IT_f is a constant for the entire sub-catchment. To calculate net recharge, the above equation must be modified so that

hydraulic conductivity controls interflow proportioning. In this basis, potential recharge can be regarded as:

$$R_{net} = R_{pot} - IT_f[f(k_{x,y}, k_z)] \quad (7)$$

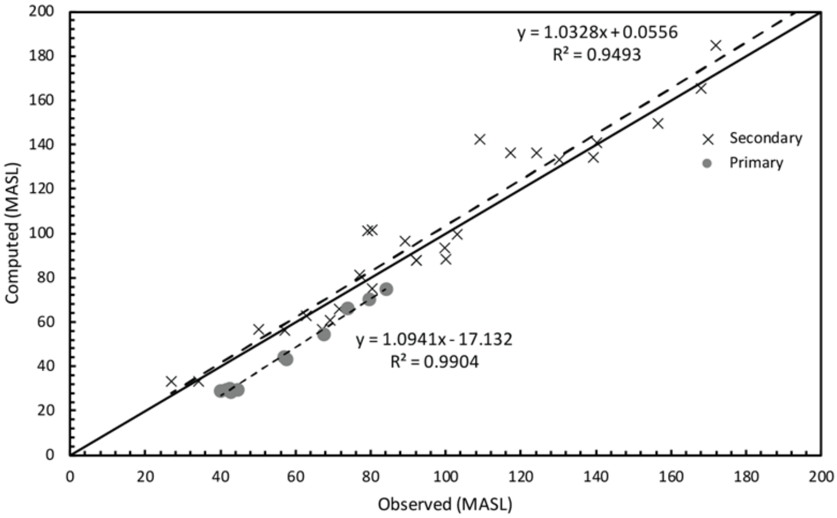

Fig. 5. Steady state calibration results showing the residual between simulated and measured water levels (Meters Above Sea Level) in the primary and secondary aquifer.

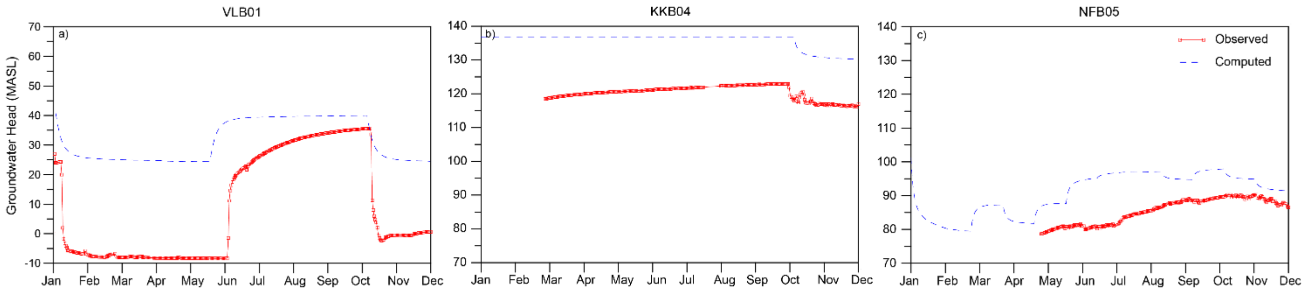

Fig. 6. Transient state calibration showing the residual between simulated and measured water levels in pumped boreholes: a) VLB01, b) KKB04 and c) NFB05 during 2016.

where R_{net} is the net recharge, and $IT_f[f(k_{x,y}, k_z)]$ is interflow as a function of horizontal ($k_{x,y}$) and vertical (k_z) hydraulic conductivity. In this situation, R_{net} drives a change in the aquifer water level dh that controls baseflow according to equation 1. The J2000 model was used to constrain the upper and lower limits of potential recharge rates (Supplementary: Fig. 1) using literature estimates of $k_{x,y}$ (Table 1). MODFLOW calculates the aquifer water level using Richards equation based on various iterations of R_{net} and IT_f per hydraulic zone to reduce the difference between the observed and modelled water levels.

4.4.2. Aquifer storage calibration

Ss was calibrated for the secondary aquifer (zone 10) (Table 1) using transient state calibration and continuous water levels that were monitored between January and December 2016 (Fig. 6). Although continuous water levels were included in the model, only three boreholes (VLB01, KKB04 and NFB05) influenced Ss as these were actively pumped and therefore the calibration effectiveness was hindered by data limitations (Fig. 6). Ss is calculated iteratively in MODFLOW to reduce the residual between the modelled and observed hydraulic head. This process was only performed on hydraulic zone 10 since, primary aquifer zones 1-6 and TMG aquifer zones 7-9 are unconfined (Table 1).

5. Results

5.1. *Net Recharge*

MODFLOW results for net recharge for the primary and TMG aquifer were between 0.85 to 107.05 mm which corresponds to 0.3 to 11.4 % of yearly rainfall for 2010-2016 (Table 3 and Supplementary: Fig. 1). During the wet years (2010, 2013, 2014) net recharge across the primary aquifer was between 1.4-25.3 mm, while for the TMG aquifer net recharge was between 57.5-107.1 mm (Table 3). During the dry years (2011, 2015, 2016) net recharge for the primary aquifer was between 0.9-17.7 mm, while for the TMG aquifer net recharge was between 33.7-73.6 mm (Table 3). The primary aquifer receives 26.0 % of the total net recharge, with zones 4, 5 and 6 receiving 10.0, 5.0 and 8.0 % (Fig. 7a). Net recharge for zones 1, 2 and 3 corresponded to less than 1.0 % of the total net recharge for the primary aquifer (Fig.7a). The TMG aquifer receives 74.0 % of total net recharge with zones 7, 8 and 9 receiving 20, 21 and 33.0 % (Fig. 7a).

Table 3

Net recharge for primary aquifer zones (1-6) and TMG aquifer zones (7-9) with percentage net recharge to HRU median simulated rainfall.

	Valley (Primary aquifer)						Piketberg Mountains (TMG aquifer)		
Year	Zone 1	Zone 2	Zone 3	Zone 4	Zone 5	Zone 6	Zone 7	Zone 8	Zone 9
2016	1.05	2.40	2.40	24.75	8.70	17.68	45.03	47.63	73.57
2015	0.85	1.88	1.96	21.62	7.78	15.76	33.73	33.31	53.79
2014	1.44	3.20	3.26	34.54	12.37	25.19	59.77	63.99	97.94
2013	1.52	3.39	3.43	35.52	12.65	25.33	62.98	69.93	107.05
2012	1.13	2.53	2.53	27.62	9.78	19.52	47.94	55.93	89.83
2011	0.94	2.10	2.12	22.66	8.02	15.93	39.01	45.87	73.31
2010	1.43	3.18	3.24	34.05	12.11	24.20	57.48	62.18	96.78
Net recharge (%)	0.31	0.71	0.72	6.93	2.39	4.75	8.36	6.30	11.43

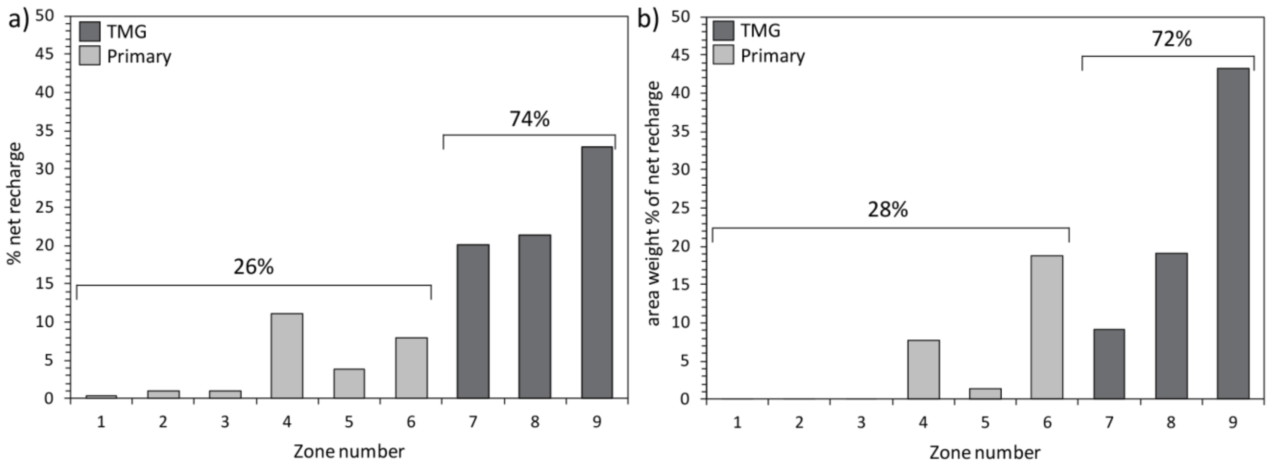

Fig. 7. Percentage of recharge received between the TMG and primary aquifer based on a) net percentage recharge and b) area weighted net recharge.

5.2.*Drawdown*

The variations in monthly abstraction volumes across the primary and secondary aquifer of the Krom Antonies influenced the degree of drawdown modelled by MODFLOW. In borehole VLB01 (Fig. 1), a maximum of 15 m drawdown between January and March 2016 was modelled, with a 15 m recovery between May and July 2016 when abstraction ceased (Fig. 6). In borehole KKB04, no abstraction took place during January to March 2016, although a 19 m drawdown was modelled between November and December 2016 (Fig. 5). In borehole NFB05, a 10 m drawdown was modelled between January and March 2016, recovering in May 2016 when abstraction ceased (Fig. 5). In the headwaters of the Krom Antonies, near other major abstraction points, a 16 m drawdown was modelled in the secondary aquifer between January and March 2016, recovering in May when abstraction ceased (not shown). In abstraction boreholes in the middle of the catchment near the Krom Antonies tributary, a drawdown of 8 m was observed in the secondary aquifer, while in the primary aquifer a 6 m drawdown was modelled (not shown).

5.3.*Baseflow*

For the Krom Antonies tributary, the average daily baseflow modelled between 2010 and 2016 varied from 13,800 – 19,000 $m^3.d^{-1}$ (0.16-0.22 $m^3.s^{-1}$) with wet years (2010, 2013, 2014) having a baseflow of between 18,000 – 19,000 $m^3.d^{-1}$ (0.21-0.22 $m^3.s^{-1}$) and dry years (2011, 2015, 2016) between 13,800 – 16,000 $m^3.d^{-1}$ (0.16-0.19 $m^3.s^{-1}$) (Table 4). The average daily baseflow was compared with baseflow for 2016, which was the calibrated period. Wet years (2010, 2013, 2014) had an increased baseflow of between 18.0 - 21.0 % compared to 2016 estimates (Supplementary: Fig. 2). 2012 was an intermediate year with respect to rainfall and had an increased baseflow of 8.0 % compared with 2016 estimates. Dry years (2011, 2015) had a baseflow decrease of between 5.0 - 11.0 % in comparison with 2016 estimates.

Table 4
Calculated baseflow for the Krom Antonies tributary between 2010 to 2016.

	Year						
Baseflow	2016	2015	2014	2013	2012	2011	2010
m3/d-1	15086	13447	19121	19413	16322	14330	18436
m3/s-1	0.17	0.16	0.22	0.22	0.19	0.17	0.21

6. Discussion

The conversion of potential recharge from a J2000 rainfall/runoff model (Watson *et al.*, 2018) to net recharge was performed in this study using the MODFLOW groundwater model. This was done to account for the variable nature of the hydraulic conductivities in the different aquifer zones. The resultant net recharge rates were compared to abstraction volumes to generate average yearly baseflow for the Krom Antonies tributary. This tributary was considered the main source of freshwater to the Verlorenvlei lake and hence the determination of baseflow could help in quantifying the ecological reserve for the estuarine system. Before the significance of the baseflow volumes can be assessed, the method of calculating net recharge rates as well as the effectiveness of the model calibration given the data limitations, needs to be evaluated.

6.1. *Model performance*

As model results are a function of the input data and model assumptions, these need to be critically assessed. Model parameters are adjusted during calibration but the degree of adjustment is constrained by the number of field observations available. The limited number of field observations from this sub-catchment, meant that different model parameter combinations could yield similar results in terms of water level residuals and therefore it is necessary to evaluate the impact of the model framework on the net recharge and baseflow results.

6.1.1. Impact of model assumptions

During the construction of the groundwater model, assumptions were made regarding the model boundary conditions. The most important of these can be summarised as follows: (1) groundwater flow only occurs within the boundaries of the sub-catchment; (2) the primary aquifer has a uniform thickness of 15 m and the secondary aquifer has a thickness of between 70 and 90 m; (3) reservoirs within the sub-catchment had fixed water levels that did not fluctuate during the modelling period; and (4) 2016 groundwater abstraction volumes are representative of the annual periods between 2010 and 2016. Whilst these assumptions are all to do with boundary conditions, assumption (2) impacts on steady state calibration of primary and secondary aquifer water levels and will be discussed in the following section.

If the sub-catchment is restricted within a boundary defined using surface water flow accumulation directions, this does not allow for possible inflows and outflows within the sub-catchment caused by geological dip orientation. However, hydrochemical results from Eilers (2018) suggest that there is a groundwater component flowing into the Krom Antonies from the Hol sub-catchment to the west. This inflow is only partially balanced by outflows along the south-east boundary of the sub-catchment towards Piketberg (Fig. 1). It is difficult to evaluate the impact of this assumption on the modelling results here because of data limitations. The two boundaries in question are defined by mountain ranges and there is no monitoring data along these mountains to determine where the groundwater flow boundary is located. Hence, there is scope for developing a means to test the impact of geological dip orientation in this region on the delineation of the groundwater flow boundary. However, the most likely impact of this assumption is that the hydraulic conductivity estimated by the model is too high for the TMG aquifer units within the sub-catchment and this probably results in baseflow being over-estimated.

Baseflow might also be overestimated by assuming that water levels in reservoirs along the Krom Antonies tributary remain fixed throughout the modelling period. This assumption was built into the model by simulating a constant head boundary condition because of limited reservoir water level records. While this could impact baseflow from the sub-catchment, if reservoirs were set as dry at the start of the simulation (ie by setting the reservoir head below topographic elevation), the resultant baseflow reduction is only 8%. Given that the reservoirs are not dry throughout the entire year, this reduction in baseflow is likely to be substantially less than 8% and hence does not significantly impact baseflow volumes.

To calibrate aquifer storage capacity both abstraction volumes and continuous aquifer water levels are required. The most significant model assumption relates to abstraction volumes whereby the 2016 groundwater abstraction scenario is considered representative of the yearly average abstraction between 2010 and 2016. This assumption was based on the fact that the crops planted in this region are relatively fixed with little crop rotation or cover crops being implemented. Even so, the rainfall record for the period 2010 to 2016 is variable and 2016 was a relatively dry year where annual rainfall was 10% less than mean annual precipitation (MAP) (Watson et al., 2018). During the years 2011 and 2015, when rainfall was even lower than 2016, the pumping volumes would have been higher, resulting in a larger baseflow reduction. In 2010, 2012, 2013 and 2014 when MAP was higher than 2016, baseflow reduction would be lower. Regardless of the yearly rainfall record, the daily abstraction rate from the secondary aquifer of 7000 $m^3.d^{-1}$ (based on farmer's records and estimates) is likely to be an underestimate and would result in baseflow being over-estimated. How far over-estimated it is not possible to evaluate given the paucity of abstraction records.

6.1.2. Steady state limitations

Steady state calibration was used to calculate net recharge and aquifer hydraulic conductivity, using potential recharge and aquifer conductance estimates. While the R^2 of the steady state calibration produced a correlation of 0.95 between observed and modelled water levels, the modelled primary aquifer water levels were on average 17 m lower than observed (Fig. 5). The primary aquifer water level deviation is linked to model setup limitations because a uniform primary aquifer thickness of 15 m was used. It is more likely, that near (> 5m) tributaries, the primary aquifer is between 1-2 m thick. Primary aquifer water levels were only measured on the banks of the tributary, and with the secondary aquifer being confined and therefore fully saturated, a reduced primary aquifer thickness would bring the model residual down to 2 m. The difference between a uniform 15 m thick primary aquifer and a reduced primary aquifer of 1-2 m near the tributaries resulted in a 5% difference in the average aquifer thickness. It therefore seems unlikely that primary aquifer thickness limitations would have a significant influence on the modelled baseflow.

The thickness of the secondary aquifer was determined from previous consulting reports and borehole logs. In some parts of the sub-catchment, the secondary aquifer is probably considerably thicker than modelled, resulting in increased aquifer storage capacity. The average difference between observed and modelled water levels for the secondary aquifer were relatively small, with an average residual of 0.06 m, although there are a number of data outliers. The boreholes used for this part of the study were existing boreholes with static water levels and borehole information (construction and borehole logs) derived from the National Groundwater Archive. However, in some cases the records are incomplete, and it was difficult to ascertain the groundwater host rock. While most of the water levels were presumed to be from the secondary aquifer, the influence of the TMG aquifer on the calibration cannot be ignored without more detailed borehole logs.

Hydraulic conductivity values for the aquifer units in the sub-catchment are not well constrained. The only published hydraulic conductivities from the sub-catchment are from the secondary aquifer (SRK, 2009) and the calibrated model hydraulic conductivity falls on the lower end of these published values. The calibrated hydraulic conductivity for the primary aquifer is on the lower end of theoretical primary aquifer hydraulic conductivities (Domenico and Schwartz 1990), whereas the calibrated hydraulic conductivity for the TMG aquifer is on the upper half of previously published TMG estimates (UMVOTO-SRK, 2000) (Table 1). While the calibrated aquifer properties were within known ranges, this does not necessarily imply that the model correctly represents the sub-catchment. It is possible that various combinations of hydraulic conductivity and net recharge would produce similar observation residuals within the model, a problem known as regularisation.

To avoid regularisation, fixed and calibrated variables are needed to minimise residuals and to ensure that the model is representative of the water dynamics in the sub-catchment. Sensitivity analysis can be used to determine which variables are fixed and which are calibrated in each hydraulic zone. The problem arises when the hydraulic zone is not sensitive to a variable, but the value of this fixed variable is not well constrained in the sub-catchment. The implication is that for zones that were not sensitive to any variable, in this case the primary aquifer zones, the model could not calibrate net recharge or hydraulic conductivity. However, given that most recharge is received in the mountains into the TMG aquifer (Watson et al., 2018), this is not surprising. All the TMG aquifer zones calibrated for both net recharge and aquifer hydraulic conductivity and thus the average net recharge and baseflow is considered representative of the sub-catchment.

6.1.3. Transient state limitations

During transient state calibration, storage parameters were calculated by fitting observed to modelled water levels in pumped boreholes, but could not be fully calibrated due to data limitations. Unoptimised storage coefficients were a single order of magnitude higher than similar rock types from the Karoo Basin (Sami, 1996), and consistent with world-wide compilations of shale aquifer estimates (Williams and Paillet, 2002). The uncalibrated storage properties resulted in poor correlation between observed and modelled water levels in boreholes VLB01 and KKB04. While borehole NFB05 showed some agreement between the observed and modelled water levels, this did not allow for confidence in storage calibration and therefore baseflow reductions could not be calculated.

6.2. *Comparison of recharge estimates*

Previous recharge estimates for the Verlorenvlei sub-catchment by Conrad et al. (2004) were between 0.2-10.0 % of MAP for the primary aquifer and 6 – 22% for the TMG aquifer. These recharge estimates were derived from a variety of different methods and interpolated to derive spatially resolved recharge across the sub-catchment. The daily estimates of potential recharge derived from the rainfall/runoff model from Watson et al. (2018) are lower for the primary aquifer (0.8 - 5.1%) but higher for the TMG aquifer (22.7 - 25.2%) (Supplementary: Table 2). The calibrated net recharge modelled in this study was slightly lower than potential recharge estimates for the primary aquifer, with a reduction to 0.3 – 4.8% of rainfall. However, for the TMG aquifer calibrated net recharge was 50% less than potential recharge with values of 6.3 - 11.4% of rainfall. The contribution of interflow plus baseflow from the primary and TMG aquifers, which is determined as the residual between potential and net recharge, was estimated to be 0.3 % of potential recharge for the primary aquifer and 14.0 % for the TMG aquifer. The lumped hydraulic properties of the primary aquifer in the J2000 model were similar to those estimated by MODFLOW, and this accounts for the small interflow plus baseflow residual. For the TMG aquifer, the interflow plus baseflow components were significantly higher than the primary aquifer, as the hydraulic conductivity of TMG sandstones is considerably more than that of the unconsolidated sediments in the primary aquifer. In addition, the topographic slope of TMG formations generated a higher interflow component which reduced the amount of net recharge possible in the TMG aquifer.

6.3. *Recharge utilisation*

Many semi-arid environments are likely to be impacted by rainfall variability, with Verlorenvlei likely to be influenced into the future. Between 2010 and 2017, the Verlorenvlei sub-catchment experienced several consecutive years of below average rainfall, with agriculture and the local municipality becoming strongly dependent on groundwater. Considering this situation, it becomes important to evaluate whether groundwater abstraction rates are “safe”, as sustainability cannot be assessed without the inclusion of provisions for the ecological reserve. One way in which safe abstraction rates can be determined is by calculating the amount of recharge utilised by groundwater abstraction. Agricultural activities in the Verlorenvlei sub-catchment typically utilise groundwater from either the primary or the secondary aquifer or both. However, as shown by the conceptual model, the secondary aquifer is recharged via the TMG aquifer. Therefore, to evaluate the proportion of recharge utilised, net recharge to abstraction rates were compared for both the TMG aquifer (as a proxy for recharge utilisation in the secondary aquifer) and the primary aquifer.

The TMG aquifer receives the bulk of rainfall and net recharge (72.0 %), while the primary aquifer receives considerably less (28.0 %). Between 2010 and 2016, recharge utilised during summer was between 8.0 and 41.0 % (Fig. 8), with the highest recharge utilisation occurring in January and February (dry season) 2015. Over the same period, recharge utilised during winter was between 1.5 to 8.0 % (Fig. 8), with the lowest recharge utilisation ocurring in July and August (wet season) 2013. The modelled results estimate recharge utilised to be around 40 % during the peak of abstraction, using 20.0 % of recharge from the TMG aquifer based on the 2016 abstraction records. These estimates of recharge utilisation must be regarded as a minimum because, as already pointed out, the abstraction scenarios used are likely to underestimate the amount of abstraction taking place in the sub-catchment to support agricultural activities. Given this, baseflow rates modelled here are likely to be over-estimated. Hence with continual decreases in MAP into the future as a result of climate change (Abiodun et al., 2017), groundwater abstracted from the sub-catchment will over-exploit recharge needed for sustaining the ecological reserve of the estuarine lake.

6.4. *Baseflow vs ecological reserve*

For the estimation of comprehensive ecological reserves, daily baseflow estimates are required. Daily baseflow estimates require the inclusion of climate variables in modelling framework. Therefore groundwater models which lump climate variables can only provide average daily estimates of baseflow. The average daily baseflow modelled between 2010 and 2016 for the Krom Antonies, showed significant variation (~14,000-19,000 $m^3.d^{-1}$), which was mainly attributed to changes in net recharge. The average daily baseflow for the Krom Antonies modelled in this study could be used as a proxy for the ecological reserve.

While, the water level in the Verlorenvlei lake is supplied by runoff from the four main tributaries, baseflow is also an additional source of freshwater. The water level in the lake fluctuates substantially over the course of a year (Watson et al., 2018), with the primary cause of water level decline being evaporation. The steep slope where the lake level increases reflects a large runoff component that supplies the lake. However, rainfall in the region is received over a four-month period meaning that for the rest of the year, baseflow must be compensating for evaporative demands. With the approximate size of the lake being 15 km^2, an average daily evaporation of 0.06 $m.d^{-1}$ would equate to a volume loss of 90,000 $m^3.d^{-1}$ if the lake was fill. Given that the maximum baseflow estimated in this study is 19,000 $m^3.d^{-1}$, the baseflow contribution from the Krom Antonies alone does not seem sufficient to sustain the lake system. Consequently, there must be significant contributions of baseflow from tributaries other than the Krom Antonies. Given that all the other tributaries in the region have lower water quality than the Krom Antonies (Sigidi, 2018), this implies that the saline contribution from the other tributaries is larger than the freshwater inputs from the Krom Antonies. This has long-term implications for the health of the estuarine lake and future work will be needed to constrain the baseflow from each of the input sources.

7. Conclusion

Estimations of baseflow reduction and aquifer depletion are important for determining sustainable groundwater abstraction regimes. In many catchments in semi-arid southern Africa, the poor availability of data required for comprehensive groundwater modelling is the most serious limitation to determining recharge and baseflow rates. In this study potential recharge from a rainfall/runoff model was transferred into a groundwater model and calibrated for net recharge after which recharge utilisation could be quantified.

Baseflow was modelled using monitoring and abstraction data collected during 2016, to provide estimates of recharge utilised and baseflow between 2010 and 2016. Of the total net recharge 72.0 % was modelled to be in the mountainous regions of the catchment, with groundwater abstraction using as much as 40.0 % of this amount. Average daily estimates of groundwater baseflow were between 14,000-19,000 $m^3.d^{-1}$, with 2015 having the lowest estimated baseflow. The sensitivity analysis suggested that hydraulic properties in and below the mountains, controlled much of the steady state calibration and directly impacted on recharge amounts. Due to data limitations average aquifer thickness was used in this study, which accounted for much of the primary aquifer deviation between observed and modelled water levels, resulting in overestimation of baseflow.

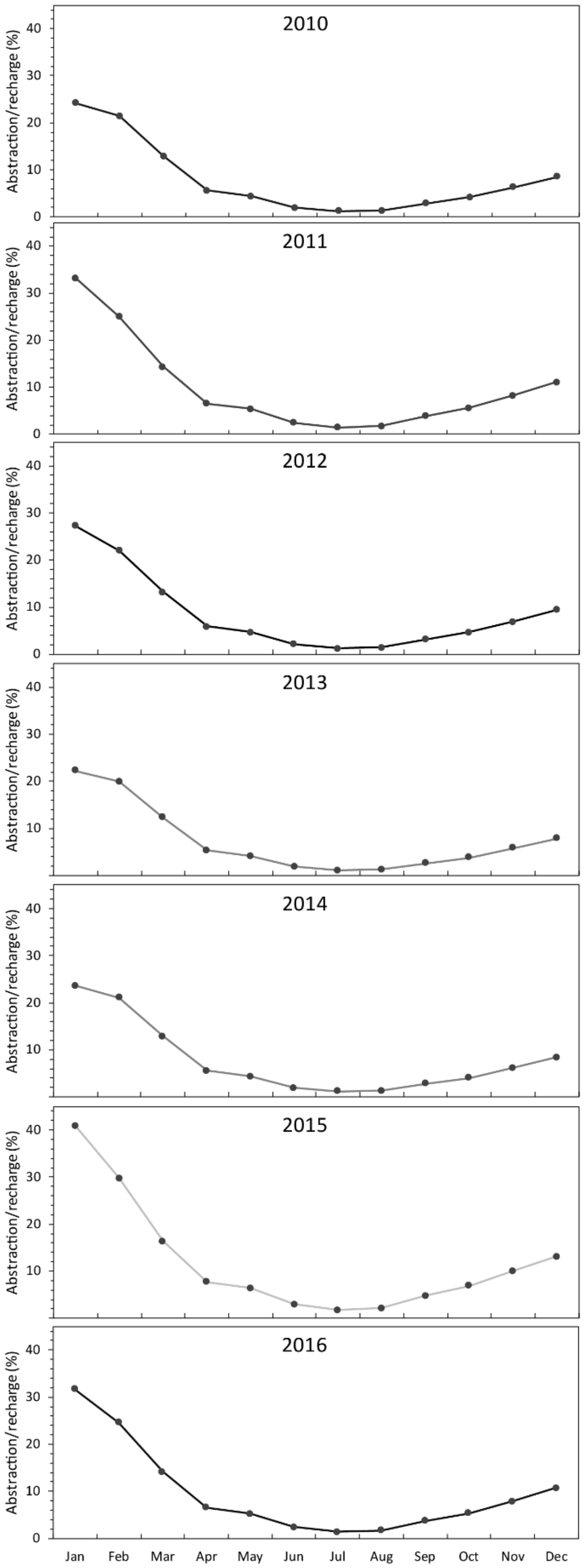

Fig. 8. Average baseflow difference between 2016 and years 2015-2010 showing the influence that changes in recharge and abstraction have on aquifer baseflow.

A comparison between the MODFLOW net recharge to J2000 potential recharge highlighted the impact that lumping groundwater components could have on regional recharge estimates. A further adjustment to rainfall/runoff models could be made by incorporating distributive aquifer hydraulic conductivity, which would improve the percolation to interflow proportioning. In this study the largest variation in baseflow was due to changes in yearly recharge, where baseflow variability is likely to increase due to climate change. The most likely impact of climate variability will be a reduction in annual rainfall volumes which will reduce baseflow rates further. Given that baseflow rates modelled in this study seem to be on the low side for sustaining the Verlorenvlei lake, additional sources of baseflow must be contributing to streamflow. By distributing groundwater components within rainfall/runoff models, daily baseflow estimates can be modelled. Daily baseflow rates underpinned by sufficient monitoring data are central to providing the hydrological components needed for accurate ecological reserve determination, which are required to ensure sustainable groundwater usage in this water scarce region.

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9. References

Arnold, J.., Muttiah, R.., Srinivasan, R., Allen, P.., 2000. Regional estimation of base flow and groundwater recharge in the Upper Mississippi river basin. J. Hydrol. 227, 21–40.

Arnold, J.G., Allen, P.M., 1999. Automated methods for estimating baseflow and ground water recharge from streamflow records. J. Am. Water Resour. Assoc. 35, 411–424. doi:10.1111/j.1752-1688.1999.tb03599.x

Bulygina, N., Mcintyre, N., Wheater, H., 2009. Conditioning rainfall-runoff model parameters for ungauged catchments and land management impacts analysis. Hydrol. Earth Syst. Sci 13, 893–904. doi:10.5194/hessd-6-1907-2009

Conrad, J., Nel, J., Wentzel, J., 2004. The challenges and implications of assessing groundwater recharge: A case study-northern Sandveld , Western Cape, South Africa. Water SA 30, 75–81.

De Beer, C., 2003. The geology of the Sandveld area between Lambert's Bay and Piketberg (Project 5510). CGS report.

Doherty, J., 2016. PEST: Model-Independent Parameter Estimation, User Manual Part I:PEST, SEANSAN and Global Optimisers. Watermark Numer. Comput. 360.

Domenico, P.A., Schwartz, F.W., 1990. Physical and Chemical Hydrogeology. John Wiley and Sons, Inc., New York.

DWAF, 2003. Sandveld Preliminary (Rapid) Reserve Determinations. Langvlei, Jakkals and Verlorenvlei Rivers. Olifants-Doorn WMA G30. Surface Volume 1: Final Report Reserve Specifications. DWAF Project Number: 2002-227.

Eilers, A., 2018. Deep groundwater characterisation and recharge estimation in the Verlorenvlei catchment.

Flügel, W., 1995. Delineating hydrological response units by geographical information system analyses for regional hydrological modelling using PRMS/MMS in the drainage basin of the River Bröl, Germany. Hydrol. Process. 9, 423–436.

Haarsma, R.J., Selten, F.M., Weber, S.L., Kliphuis, M., 2005. Sahel rainfall variability and response to greenhouse warming. Geophys. Res. Lett. 32, 1–4. doi:10.1029/2005GL023232

Halford, K.J., Hanson, R.T., 2002. User guide for the drawdown-limited, multi-node well (MNW) package for the US Geological Survey's modular three-dimensional finite-difference ground-water flow model, versions MODFLOW-96 and MODFLOW-2000. US Department of the Interior, US Geological Survey.

Harbaugh, B.A.W., Banta, E.R., Hill, M.C., Mcdonald, M.G., 2000. MODFLOW-2000 , The U .S . Geological Survey modular graound-water model — User guide to modularization concepts and the ground-water flow process. U.S. Geol. Surv. 130.

Hill, M.C., Tiedeman, C.R., 2006. Effective groundwater model calibration: with analysis of data, sensitivities, predictions, and uncertainty. John Wiley & Sons.

Kim, N.W., Chung, I.M., Won, Y.S., Arnold, J.G., 2008. Development and application of the integrated SWAT-MODFLOW model. J. Hydrol. 356, 1–16. doi:10.1016/j.jhydrol.2008.02.024

Konikow, L.F., Kendy, E., 2005. Groundwater depletion: A global problem. Hydrogeol. J. 13, 317–320.

Lin, L., 2007. HYDRAULIC PROPERTIES OF THE TABLE MOUNTAIN GROUP (TMG) AQUIFERS Dissertation submitted to the University of the Western Cape in the Department of Earth Sciences Faculty of Natural Sciences , University of the Western Cape Supervisor Co-supervisor.

McMichael, A.J., Woodruff, R.E., Hales, S., 2006. Climate change and human health: Present and future risks. Lancet 367, 859–869. doi:10.1016/S0140-6736(06)68079-3

Muche, G., Kruger, S., Hillman, T., Josenhans, K., Ribeiro, C., Bazibi, M., Seely, M., Nkonde, E., de Clercq, W.P., Strohbach, B., Kenabatho, K.., Vogt, R., Kaspar, F., Helmschrot, J., Jürgens, N., 2018. Climate change and adaptive land management in southern Africa – assessments, changes, challenges, and solutions, in: Revermann, R., Krewenka, K.M., Schmiedel, U., Olwoch, J.., Helmschrot, J., Jürgens, N. (Eds.), Biodiversity & Ecology. Klaus Hess Publishers, Göttingen & Windhoek, pp. 34–43.

Ndomba, P., Mtalo, F., Killingtveit, A., 2008. SWAT model application in a data scarce tropical complex catchment in Tanzania. Phys. Chem. Earth 33, 626–632. doi:10.1016/j.pce.2008.06.013

Necpálová, M., Anex, R.P., Fienen, M.N., Del Grosso, S.J., Castellano, M.J., Sawyer, J.E., Iqbal, J., Pantoja, J.L., Barker, D.W., 2015. Understanding the DayCent model: Calibration, sensitivity, and identifiability through inverse modeling. Environ. Model. Softw. 66, 110–130. doi:10.1016/j.envsoft.2014.12.011

Perrin, C., Oudin, L., ANDREASSIAN, V., ROJAS-SERNA, C., MICHEL, C., MATHEVET, T., 2007. Impact of limited streamflow data on the efficiency and the parameters of rainfall—runoff models. Hydrol. Sci. J. 52, 131–151. doi:10.1623/hysj.52.1.131

Pfennig, B., Kipka, H., Wolf, M., Fink, M., Krause, P., Flügel, W. a., 2009. Development of an extended routing scheme in reference to consideration of multi-dimensional flow relations between hydrological model entities. 18th World IMACS Congr. MODSIM09 Int. Congr. Model. Simulation, Cairns, Aust. 1972–1978.

Pike, A., Schulze, R.E., 1995. AUTOSOILS: A program to convert ISCW soils attributes to variables usable in hydrological models. Pietermaritzburg, South Africa Univ. KwaZulu-Natal, Sch. Bioresour. Eng. Environ. Hydrol.

Rotstayn, L.D., Collier, M.A., Dix, M.R., Feng, Y., Gordon, H.B., O'Farrell, S.P., Smith, I.N., Syktus, J., 2010. Improved simulation of Australian climate and ENSO-related rainfall variability in a global climate model with an interactive aerosol treatment. Int. J. Climatol. 30, 1067–1088. doi:10.1002/joc.1952

Rozendaal, A., Gresse, P.G., 1994. Structural setting of the Riviera W-Mo deposit, western Cape, South Africa. South African J. Geol. 97, 184.

Sami, K., 1996. Evaluation of the variations in borehole yield from a fractured Karoo aquifer, South Africa. Ground Water 34, 114–120. doi:10.1111/j.1745-6584.1996.tb01871.x

Saxton, K.E., Willey, P.H., 2005. The SPAW model for agricultural field and pond hydrologic simulation. Watershed Model. CRC Press. Boca Raton, FL. Doi 10, 401–422.

Scanlon, B.R., Healy, R.W., Cook, P.G., 2001. Choosing appropriate technique for quantifying groundwater recharge. Hydrogeol. J. 10, 18–39. doi:10.1007/s10040-0010176-2

SRK, 2009. Preliminary Assessment of Impact of the Proposed Riviera Tungsten Mine on Groundwater Resources Preliminary Assessment of Impact of the Proposed Riviera Tungsten Mine on Groundwater Resources.

UMVOTO-SRK, 2000. Reconnaissance investigation into the development and utilization of the Table Mountain Group Artesian Groundwater, using the E10 catchment as a pilot study area.

van Niekerk, A., Joubert, S.J., 2011. Input variable selection for interpolating high-resolution climate surfaces for the western cape. Water SA 37, 271–280. doi:10.4314/wsa.v37i3.68475

Wada, Y., Van Beek, L.P.H., Van Kempen, C.M., Reckman, J.W.T.M., Vasak, S., Bierkens, M.F.P., 2010. Global depletion of groundwater resources. Geophys. Res. Lett. 37, 1–5. doi:10.1029/2010GL044571

Watson, A., Miller, J., de Clercq, W.P., 2017. Investigating Potential Additional Sources of Groundwater Flow into a Defined Watershed. Procedia Earth Planet. Sci. 17, 546–549. doi:10.1016/j.proeps.2016.12.138

Watson, A.P., Miller, J.A., Fleischer, M., de Clercq, W.P., 2018. Estimation of groundwater recharge via percolation outputs from a rainfall/runoff model for the Verlorenvlei estuarine system, west coast, South Africa. J. Hydrol. 558, 238–254. doi:10.1016/S1532-0464(03)00032-7

Weber, K.A., Perry, R.G., 2006. Groundwater abstraction impacts on spring flow and base flow in the Hillsborough River Basin, Florida, USA. Hydrogeol. J. 14, 1252–1264. doi:10.1007/s10040-006-0040-5

Williams, J.H., Paillet, F.L., 2002. Using flowmeter pulse tests to define hydraulic connections in the subsurface: A fractured shale example. J. Hydrol. 265, 100–117. doi:10.1016/S0022-1694(02)00092-6

Xu, Y., Lin, L., Jia, H., 2009. Groundwater Flow Conceptualization and Storage Determination of the Table Mountain Group (TMG) Aquifers.

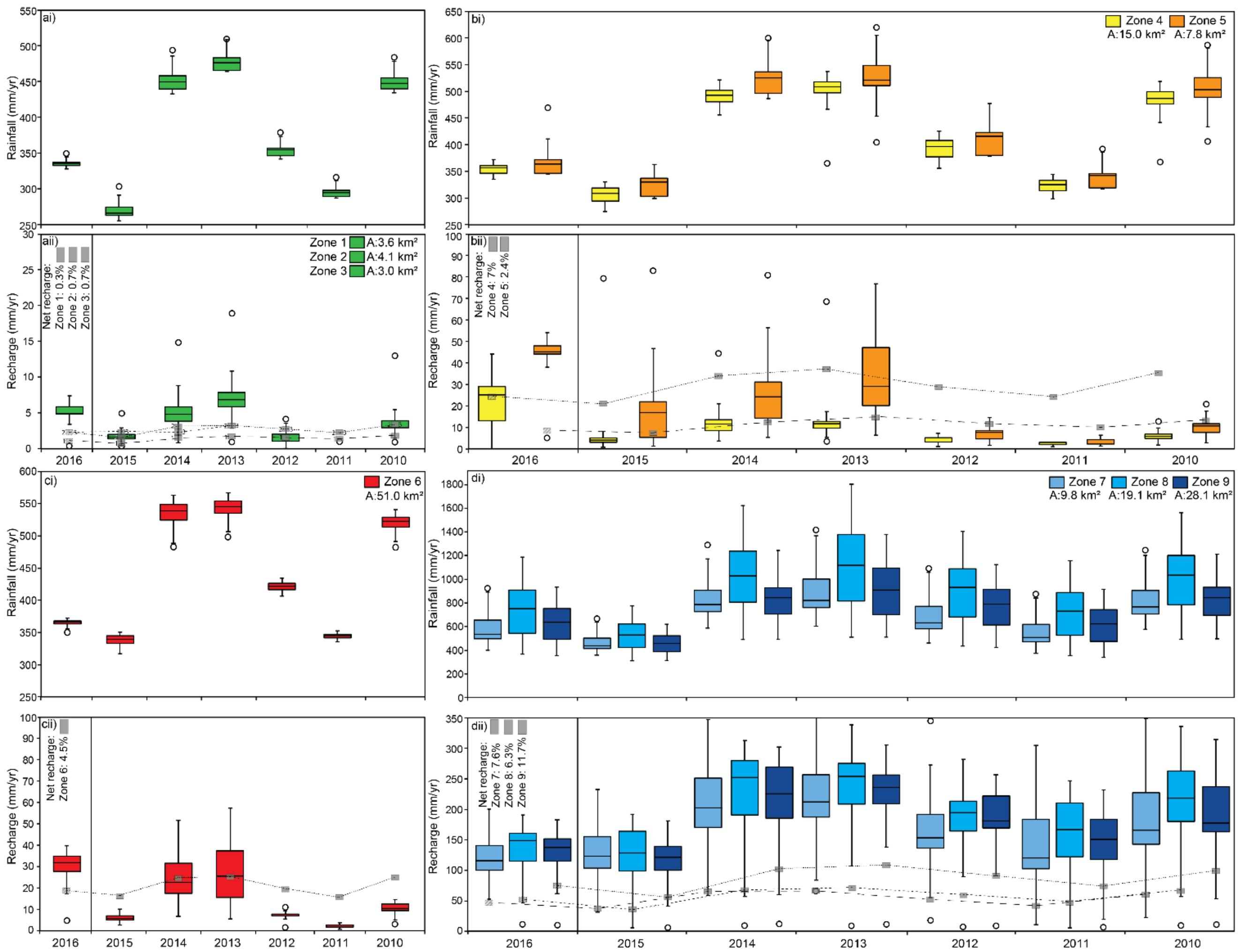
Zhou, Y., 2009. A critical review of groundwater budget myth, safe yield and sustainability. J. Hydrol. 370, 207–213.

10. Appendix

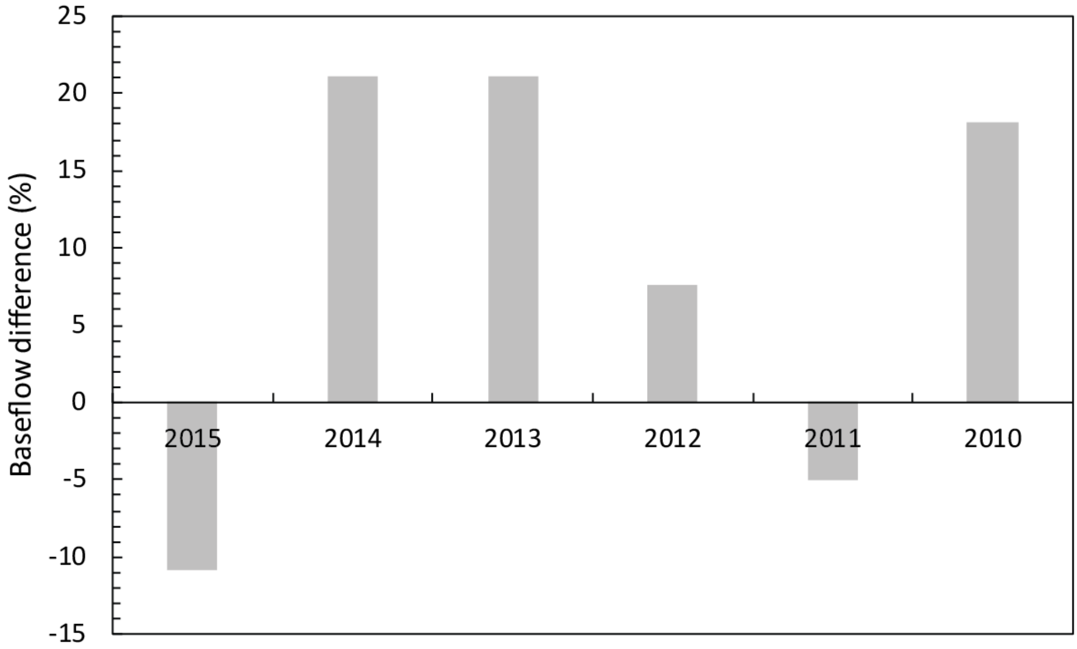
Supplementary Table 1

Simulated rainfall from the J2000 rainfall/runoff model for primary aquifer (hydraulic zones 1-6) and TMG aquifer (hydraulic zones 7-9) zones between 2010 to 2016 (after Watson *et al.*, (2018) * refer to section 4.1.4 and Fig 3 hydraulic zones description.

	Valley (Primary aquifer)						Piketberg Mountains (TMG aquifer)		
Year	Zone 1	Zone 2	Zone 3	Zone 4	Zone 5	Zone 6	Zone 7	Zone 8	Zone 9
2016	337	337	333	357	365	372	539	755	644
2015	273	265	274	309	331	347	442	533	461
2014	461	451	456	494	526	555	783	1024	839
2013	485	479	479	508	538	558	825	1119	917
2012	361	357	354	395	416	430	628	895	770
2011	300	297	297	324	341	351	511	734	628
2010	457	449	452	487	515	533	753	995	829



Supplementary Fig. 1. Box and whisker plots of ai) simulated rainfall for primary aquifer zones 1-3 between 2010-2016, aii) Net and HRU recharge for primary aquifer zones 1-3 between 2010-2016, bi) simulated rainfall for primary aquifer zones 4-5 between 2010-2016, bii) Net and HRU recharge for primary aquifer zones 4-5 between 2010-2016, ci) simulated rainfall for primary aquifer zone 5 between 2010-2016, cii) Net and HRU recharge for primary aquifer zone 6 between 2010-2016, di) simulated rainfall for TMG aquifer zones 7-9 between 2010-2016, dii) Net and HRU recharge for TMG aquifer zones 7-9 between 2010-2016.



Supplementary Fig. 2. Average baseflow difference between 2016 and years 2015-2010 showing the influence that changes in recharge and abstraction have on aquifer baseflow

Supplementary Table 2

Simulated potential recharge from the J2000 rainfall/runoff model for primary aquifer (hydraulic zones 1-6) and TMG aquifer (hydraulic zones 7-9) between 2010 to 2016 with average percentage HRU recharge to HRU simulated rainfall * refer to section 4.1.4 and Fig 3 hydraulic zones description.

	Valley (Primary aquifer)						Piketberg Mountains (TMG aquifer)		
Year	Zone 1	Zone 2	Zone 3	Zone 4	Zone 5	Zone 6	Zone 7	Zone 8	Zone 9
2016	5.00	5.00	11.00	25.00	45.00	44.00	114.00	150.00	144.00
%	1.48	1.48	3.30	7.00	12.35	11.83	21.17	19.86	22.37
2015	1.50	1.65	2.12	3.89	16.60	9.27	118.32	126.67	119.34
%	0.55	0.62	0.77	1.26	5.02	2.67	26.78	23.76	25.90
2014	5.00	5.00	7.00	11.00	24.00	38.00	201.00	255.00	245.50
%	1.08	1.11	1.54	2.23	4.56	6.85	25.67	24.90	29.26
2013	7.00	7.00	9.00	11.00	28.00	47.00	212.00	253.00	236.00
%	1.44	1.46	1.88	2.17	5.20	8.42	25.70	22.61	25.74
2012	1.00	1.00	1.00	4.00	7.00	9.00	148.00	198.00	191.00
%	0.28	0.28	0.28	1.01	1.68	2.09	23.57	22.12	24.82
2011	0.00	0.00	0.00	1.00	3.00	3.00	115.00	163.00	147.00
%	0.00	0.00	0.00	0.31	0.88	0.85	22.50	22.21	23.41
2010	3.00	3.00	4.00	6.00	11.00	14.00	162.00	230.00	202.00
%	0.66	0.67	0.88	1.23	2.14	2.63	21.51	23.12	24.37
Average	0.78	0.80	1.24	2.17	4.55	5.05	23.84	22.65	25.12

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Distributive rainfall/runoff modelling to determine runoff to baseflow proportioning and its impact on the determination of the ecological reserve.

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ABSTRACT

River systems that support high biodiversity profiles are conservation priorities world-wide. Understanding river eco-system thresholds to low flow conditions is important for the conservation of these systems. While climatic variations are likely to impact the streamflow variability of many river courses into the future, understanding specific river flow dynamics with regard to streamflow variability and aquifer baseflow contributions are central to the implementation of protection strategies. While streamflow is a measurable quantity, baseflow has to be estimated or calculated through the incorporation of hydrogeological variables. In this study, the groundwater components within the J2000 rainfall/runoff model were distributed to provide daily baseflow and streamflow estimates needed for ecological reserve determination. The modelling approach was applied to the RAMSAR-listed Verlorenvlei estuarine lake system on the west coast of South Africa which is under threat due to agricultural expansion and climatic fluctuations. The sub-catchment consists of four main tributaries, the Krom Antonies, Hol, Bergvallei and Kruismans. Of these, the Krom Antonies tributary was initially presumed the largest baseflow contributor, but was shown to have significant streamflow variability, attributed to the highly conductive nature of the Table Mountain Group sandstones and quaternary sediments. The Bergvallei tributary was instead identified as the major contributor of baseflow. The Hol tributary was the least susceptible to streamflow fluctuations due to the higher baseflow proportion (56 %), as well as the dominance of less conductive Malmesbury shales which underlie this tributary. The estimated flow exceedance probabilities indicated that during the wet cycle (2007-2017) the average inflow supported the evaporative demands if the lake was below 40 % capacity, while during the dry cycle (1997-2008), inflow would only support 15 % of the lake's surface area. The exceedance probabilities estimated in this study suggest that inflows from the four main tributaries are not enough to support the lake during dry cycles, with the evaporation demand of the entire lake being met only 38 % of the time. This study highlighted the importance of low occurrence events for filling up the lake, allowing for regeneration of lake supported ecosystems. While the increased length of dry cycles is predicted to occur more often, resulting in the lake drying up more frequently, it is important to ensure that water resources are not overallocated during wet cycles, hindering ecosystem regeneration and prolonging the length of these dry cycle condition.

1. Introduction

Functioning river systems offer numerous economic and social benefits to society including water supply, nutrient cycling and disturbance regulation amongst others (e.g Costanza et al., 1997; Postel and Carpenter, 1997). As a result, many countries worldwide have endeavoured to protect river ecosystems after provision has been made for basic human needs (eg Rowlston and Palmer, 2002). However, the implementation of river protection has been problematic (Richter, 2010), because many river courses and flow regimes have been severely altered due to socio-economic development. Previously, river health problems were thought to be only a result of low-flow conditions and therefore, if minimum flows were kept above a critical level, the river's ecosystem would be protected (Poff et al., 1997). It is now recognised that a more natural flow regime, which includes floods as well as low and medium flow conditions is required for sufficient ecosystem functioning (Postel and Richter, 2012). For these reasons, before protection strategies can be developed or implemented for a river system, a comprehensive understanding of the river flow regime dynamics is necessary.

River flow regime dynamics include consideration of not just the surface water in the river but also other water contributions including runoff, interflow and baseflow which are all essential for the maintenance of the discharge requirements. Taken together these factors all contribute to the determination of what is called the ecological reserve, the minimum environmental conditions needed to maintain the ecological health of a river system (Hughes, 2001). A variety of different methods have been developed to incorporate various river health factors into ecological reserve determination (Acreman and Dunbar, 2004). One of the simplest and most widely applied, is where compensation flows are set below reservoirs and weirs, using flow duration curves to derive mean flow or flow exceedance probabilities (e.g. Souchon and Keith, 2001). This approach focusses purely on hydrological indices, which are rarely ecologically valid (e.g. Barker and Kirmond, 1998).

More comprehensive methods such as functional analysis are focused on the whole ecosystem, including both hydraulic and ecological data (e.g. Building Block Methodology: King and Louw, 1998). While these methods consider that a variety of low, medium and high flow events are important for maintaining ecosystem diversity, they require specific data regarding the hydrology and ecology of a river system, which in many cases does not exist, or has not been recorded continuously or for sufficient duration (Acreman and Dunbar, 2004). To speed up ecological reserve determination, river flow records have been used to analyse natural seasonality and variability of flows (e.g. Hughes and Hannart, 2003). However, this approach requires long-term streamflow and baseflow timeseries. Whilst streamflow is a measurable quantity subject to a gauging station being in place, baseflow has to be modelled based on hydrological and hydrogeological variables.

While rainfall/runoff models can be used to calculate hydrological variables using distribute surface water components (e.g. J2000: Krause, 2001), groundwater variables are lumped within the modelling framework. In contrast, groundwater models which distribute groundwater variables (e.g. MODFLOW: Harbaugh, Arlen, 2005), are frequently setup to lump climate components. In order to accurately model daily baseflow, which is needed for ecological reserve determination, modelling systems need to be setup such that both groundwater, climate and surface water variables are treated in a distributive manner (e.g Kim et al., 2008). Rainfall/runoff models which use Hydrological Response Units (HRUs) as an entity of homogenous climate, rainfall, soil and landuse properties (Flügel, 1995) are able to reproduce hydrographs through model calibration (Wagener and Wheater, 2006). However, they are rarely able to correctly proportion runoff and baseflow components (e.g. Robson et al., 1992). To correctly determine groundwater baseflow using rainfall/runoff models such as the J2000, aquifer components need to be distributed. This can be achieved using net recharge and hydraulic conductivity collected through aquifer testing or inverse numerical modelling

To better understand the nature of river flow regime dynamics, a J2000 rainfall/runoff model previously setup to simulate surface water processes (Watson et al., 2018) was distributed to incorporate aquifer hydraulic conductivity within model HRUs using calibrated values from a groundwater model (Watson, submitted). The model was setup for the RAMSAR listed Verlorenvlei estuarine system on the west coast of South Africa, which is under threat from climate change and agricultural expansion. While the estuarine lake's importance is well documented (Martens et al., 1996; Wishart, 2000), the ecological reserves of the main feeding tributaries have not yet been set, partially due to a lack of streamflow and baseflow estimates within the sub-catchment. The modelling framework developed in this study aimed to provide the hydrological components, (baseflow and runoff proportioning) of the tributaries needed to set the ecological reserve. The surface water and groundwater components of the model were calibrated for two different tributaries which were believed to be the main source of runoff and baseflow for the sub-catchment. The baseflow and runoff rates calculated from the model indicate not only that the lake system cannot be sustained by baseflow during low flow periods but also that the initial understanding of which tributaries are key to the sustainability of the lake system was not correct. The results have important implications for how we understand water dynamics in water stressed catchments and the sustainability of ecological systems in these environments.

2. Study site

Verlorenvlei is an estuarine lake situated on the west coast of South Africa, approximately 150 km north of the metropolitan city of Cape Town (Fig. 1). The west coast, which is situated in the Western Cape Province of South Africa, is subject to a Mediterranean climate where the majority of rainfall is received between May to September. The Verlorenvlei lake, which is approximately 15 km2 in size draining a watershed of 1832 km2, forms the southern sub-catchment of the Olifants/Doorn quaternary catchment. The estuarine lake supports both Karroid and Fynbos biomes, due to the intermittent connection between salt and fresh water. A sandbar created around a sandstone outcrop (Table Mountain Group) allows for freshwater to exit the lake to the sea, as well as reducing sea water flow within the lake. The lake is supplied by four main tributaries which are the Krom Antonies, Bergvallei, Hol and Kruismans. The main

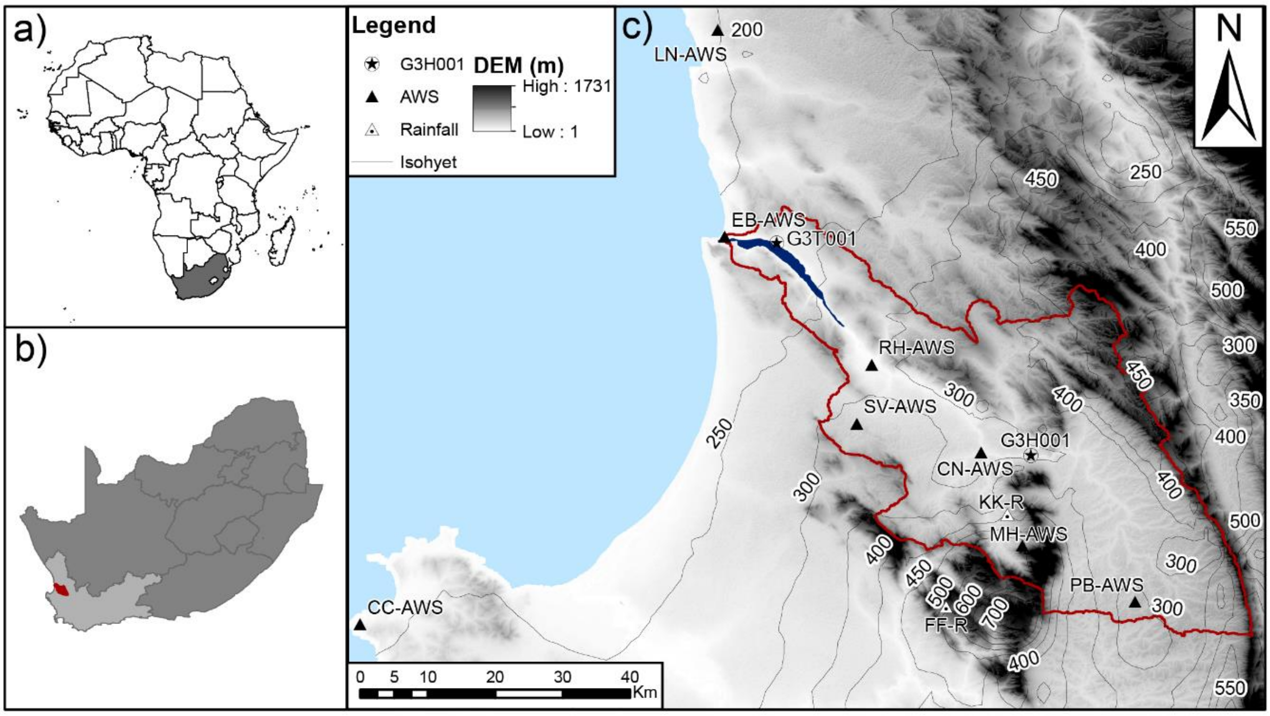

Fig. 1. a) Location of South Africa, b) the location of the study catchment within the Western Cape and c) the extend of the Verlorenvlei sub-catchment with the climate stations, gauging station (G3H001), measured lake water level (G3T001) and rainfall isohets

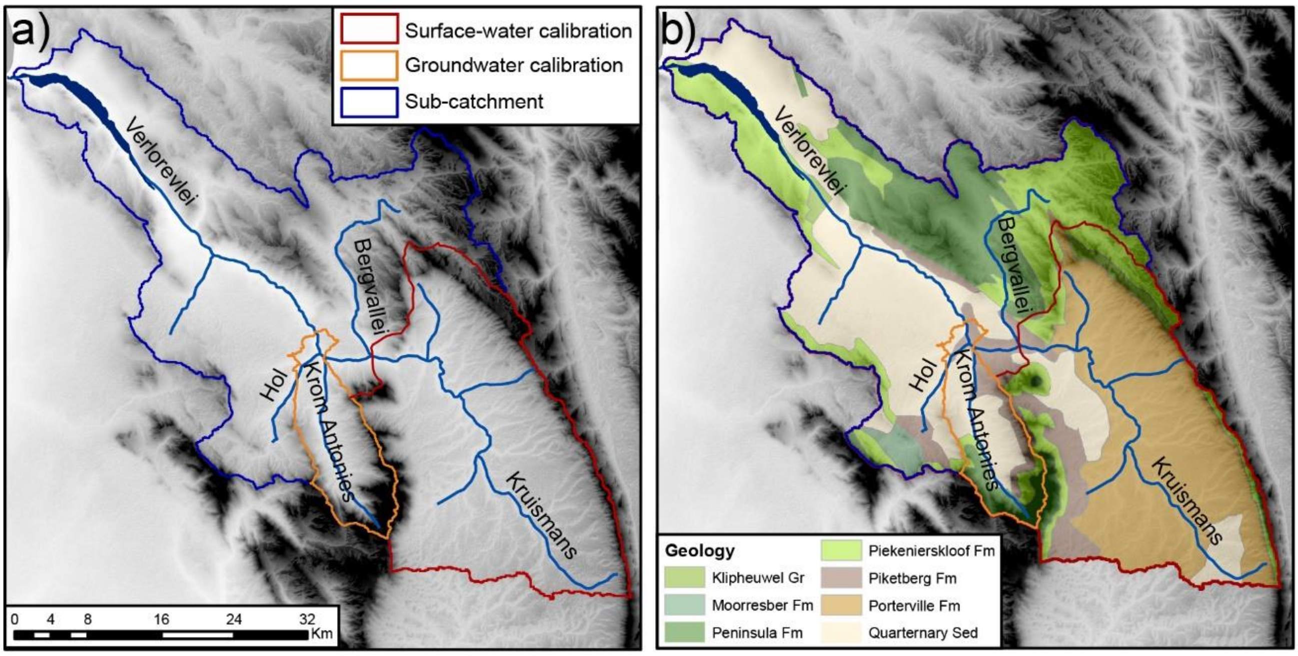

Fig. 2. a) The Verlorenvlei sub-catchment with the surface water calibration tributary (Kruismans) and groundwater calibration tributary (Krom Antonies) and b) the hydrogeology of the sub-catchment with Malmesbury shale formations (Klipheuwel, Mooresberg, Porterville, Piketberg), Table Mountain Group formations (Peninsula, Piekenierskloof) and quaternary sediments

freshwater sources are suggested to be the Krom Antonies (Sigidi, 2018) and the Bergvallei, which drain the mountainous regions to the south (Piketberg) and north of the sub-catchment respectively. The Hol and Kruismans tributaries are variably saline (Sigidi, 2018), due to high evaporation rates in the valley. Average daily temperatures during summer within the sub-catchment are between 20-30 °C, with estimated potential evaporation rates of 4 to 6 mm.d-1 (Muche et al., 2018). In comparison, winter daily average temperatures are between 12- 20°C, with estimated potential evaporation rates of 1 to 3 mm.d-1 (Muche et al., 2018).

The sub-catchment is comprised of three main lithological units, namely: 1) quaternary sediments 2) Table Mountain Group (TMG) sandstones and 3) Malmesbury (MG) shales (Fig 2). The quaternary sediments make up the primary aquifer within the sub-catchment, while the TMG sandstones and MG shales make up the secondary aquifer. The catchment valley, which receives the least mean annual precipitation (MAP) (150-500 mm.yr-1: Lynch, 2004), is comprised of quaternary sediments that vary in texture, although the majority of the sediments in the sub-catchment are sandy in nature. The higher relief mountainous regions of the sub-catchment, which receive the highest MAP (550-1000 mm.yr-1: Lynch, 2004), are mainly comprised of fractured TMG sandstones, (youngest to oldest): Peninsula, Graafwater (not shown), and Piekernerskloof formations (Fig. 2). Underlying the sandstones and quaternary sediments are the MG shales, which are comprised of the Mooresberg, Piketberg and Klipheuwel formations (Fig. 2). The MG shales and quaternary sediments which host the secondary and primary aquifer respectfully, are frequently used to supplement irrigation during the summer months of the year. During winter, the majority of the irrigation water needed for crop growth is supplied by the sub-catchment tributaries or lake itself. Agriculture is the dominant water user in the sub-catchment with an estimated usage of 20 % of the total recharge (DWAF, 2003; Watson et al, submitted), with the main food crop being potatoes. For further information regarding the study site refer to Watson et al., (2018).

3. Methodology

For this study, the J2000 coding was adapted to incorporate distributive groundwater components. To do this, the hydraulic conductivity or maximum percolation value for specific geological formations was assigned to the model HRU's, using net recharge and calibrated aquifer hydraulic conductivity from a groundwater model (Watson, submitted) determined for eight hydraulic zones (Fig. 3).

The adaption was applied to the groundwater components of the J2000 coding which influenced the portioning of water routed to runoff and baseflow. To validate the outputs of the model, an empirical mode decomposition (EMD) (Huang et al., 1998) was applied to compute the proportion of variation in discharge timeseries that attributed to a high and low water level change at the sub-catchment outlet.

3.1. Hydrological Response Unit Delineation

HRUs and stream segments (reaches) are used within the J2000 model for distributive topographic and physiological modelling. In this study, the HRU delineation made use of a digital elevation model, with slope, aspect, solar radiation index, mass balance index and topographic wetness being derived. Before the delineation process, gaps within the digital elevation model were filled using a standard fill algorithm from ArcInfo (Jenson and Domingue, 1988). The AML (ArcMarkupLanguage) automated tool (Pfennig et al., 2009) was used for the HRU delineation, with between 13 and 14 HRUs/km² being defined (Pfannschmidt, 2008). After the delineation of HRUs, dominant soil, land use and geology properties were assigned to each. The hydrological topology was defined for each HRU by identifying the adjacent HRUs or stream segments that received water fluxes.

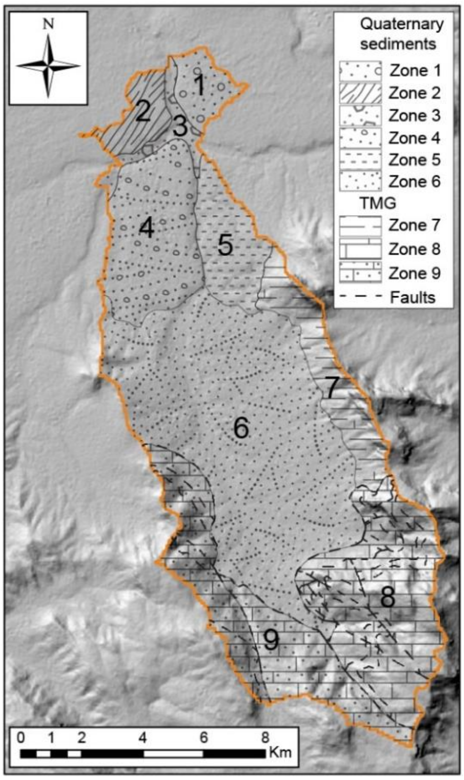

Fig. 3. The aquifer hydraulic zones used for the groundwater calibration of the J2000 (after Watson, submitted).

3.2. Model regionalisation

Rainfall and relative humidity are the two main parameters that are regionalised within the J2000 rainfall/runoff model. While a direct regionalisation using an inverse-distance method (IDW) and the elevation of each HRU can be applied to rainfall data, the regionalisation of relative humidity requires the calculation of absolute humidity. The regionalisation of rainfall records was applied by defining the number of weather station records available and estimating the influence on the rainfall amount for each HRU. A weighting for each station using the distance of each station to the area of interest was applied to each rainfall record, using an elevation correction factor (Watson et al., 2018). The relative humidity and air temperature measured at set weather stations was used to calculate the absolute humidity. Absolute humidity was thereafter regionalised using the IDW method, station and HRU elevation. After the regionalisation had been applied, the absolute humidity was converted back to relative humidity through calculation of saturated vapor pressure and the maximum humidity.

3.3. Water balance calculations

The J2000 model is divided into calculations that impact surface water and groundwater processes. The J2000 model distributes the regionalised precipitation (P) calculated for each HRU using a water balance defined as:

$$P = R + Int_{max} + ETR + \Delta Soil_{sat} \qquad (1)$$

where R is runoff (mm) (RD1 - surface runoff; RD2 - interflow), Int_{max} is vegetation canopy interception (mm), ETR is 'real' evapotranspiration and $\Delta Soil_{sat}$ is change in soil saturation. The surface water processes have an impact on the amount of modelled runoff and interflow, while the groundwater processors influence the upper and lower groundwater flow components.

3.3.1. Surface water components

Potential evaporation (ETP) within the J2000 model is calculated using the Penman Monteith equation. Before evaporation was calculated for each HRU, interception was subtracted from precipitation using the leaf area index and leaf storage capacity for vegetation (a_rain) (Supplementary: Table 1). Evaporation within the model considers several variables that influence the overall modelled evaporation. Firstly, evaporation is influenced by a slope factor, which was used to reduce ETP based on a linear function. Secondly, the model assumed that vegetation transpires until a particular soil moisture content where ETP is reached, after which modelled evaporation was reduced proportionally to the ETP, until it became zero at the permanent wilting point.

The soil module in the J2000 model is divided up into processing and storage units. Processing units in the soil module include soil-water infiltration and evapotranspiration, while storage units include middle pore storage (MPS), large pore storage (LPS) and depression storage. The infiltrated precipitation was calculated using the relative saturation of the soil, and its maximum infiltration rate (SoilMaxInfSummer and SoilMaxInfWinter) (Supplementary: Table 1). Surface runoff was generated when the maximum infiltration threshold was exceeded. The amount of water leaving LPS, which can contribute to recharge, was dependant on soil saturation and the filling of LPS via infiltrated precipitation. Net recharge (R_{net}) was estimated using the hydraulic conductivity ($SoilMaxPerc$), the outflow from LPS (LPS_{out}) and the slope ($slope$) of the HRU according to:

$$R_{net} = LPS_{out} \times (1 - \tan\ (slope\ )\ SoilMaxPerc) \qquad (2)$$

The hydraulic conductivity, $SoilMaxPerc$ and the adjusted LPS_{out}.were thereafter used to calculate interflow (IT_f) according to:

$$IT_f = LPS_{out} \times (\tan(slope\ )\ SoilMaxPerc) \qquad (3)$$

with the interflow calculated representing the sub-surface runoff component RD2 and is routed as runoff within the model.

3.3.2. Groundwater components

The J2000 model for the Verlorenvlei sub-catchment was set up with two different geological reservoirs: (1) the primary aquifer (upper groundwater reservoir - RG1), which consists of quaternary sediments with a high permeability; and (2) the secondary aquifer (lower groundwater reservoir- RG2), made up of MG shales and TMG sandstones (Table 1).

The model therefore considered two baseflow components, a fast one from the RG1 and a slower one from RG2. The filling of the groundwater reservoirs was done by net recharge, with emptying of the reservoirs possible by lateral subterranean runoff as well as capillary action in the unsaturated zone. Each groundwater reservoir was parameterised separately using the maximum storage capacity (maxRG1 and maxRG2) and the retention coefficients for each reservoir ($recRG1$ and $recRG2$). The outflow from the reservoirs was determined as a function of the actual filling ($actRG1$ and $actRG2$) of the reservoirs and a linear drain function. Calibration parameters $recRG1$ and $recRG2$ are storage residence time parameters.

The outflow from each reservoir was defined as:

$$OutRG1 = \frac{1}{gwRG1Fact \times recRG1} \times actRG1 \qquad (4)$$

$$OutRG2 = \frac{1}{gwRG2Fact \times recRG2} \times actRG2 \qquad (5)$$

where $OutRG1$ is the outflow from the upper reservoir, $OutRG2$ is the outflow from the lower reservoir and $gwRG1Fact/\ gwRG2Fact$ are calibration parameters for the upper and lower reservoir used to determine the outflow from each reservoir. To allocate the quantity of net recharge between the upper (RG1) and lower (RG2) groundwater reservoirs, a calibration coefficient $gwRG1RG2sdist$ was used to distribute the net recharge for each HRU using the HRU slope. The influx of groundwater into the shallow reservoir ($inRG1$) was defined as:

$$inRG1 = R_{net} \times (1 - (1\ \tan(slope)) \times gwRG1RG2sdist \qquad (6)$$

Table 1
The J2000 hydrogeological parameters RG1_max, RG2_max, RG1_k, RG2_Kf_geo and depthRG1 assigned to the primary and secondary aquifer formations for the Verlorenvlei sub-catchment

			RG1_max	RG2_max	RG1_k	RG2_k	RG1_active	Kf_geo	depthRG1
Aquifer	Formation	Type	(mm)	(mm)	(d)	(d)	(n/a)	(mm.d^{-1})	(cm)
Primary	Quarternary Sediments	Sediment	0	700	100	431	1	500	1750
Secondary/MG	Moorreesberg Formation	Shale Greywacke	0	580	0	350	0	950	1750
Secondary/MG	Porterville Formation	Shale Greywacke	0	560	0	335	0	2	1750
Secondary/MG	Piketberg Formation	Shale Greywacke	0	1000	0	600	0	950	1750
Secondary/MG	Klipheuwel Group	Shale Greywacke	0	500	0	300	0	950	1750
Secondary/TMG	Peninsula Formation	Sandstone	0	1000	0	600	0	950	1750
Secondary/TMG	Piekenierskloof Formation	Sandstone	0	600	0	400	0	1	1750

The influx of net recharge into the lower groundwater reservoir ($inRG2$) was defined as:

$$inRG2 = R_{net} \times (1 - \tan(slope)) \times gwRG1RG2sdist \tag{7}$$

with the combination of $OutRG1$ and $OutRG2$ representing the baseflow component that is routed as an outflow from the model.

3.4. Lateral and reach routing

Lateral routing was responsible for water transfer within the model and included HRU influxes and discharge through routing of cascading HRUs from the upper catchment to the exit stream. HRUs were either able to drain into multiple receiving HRUs or into reach segments, where the topographic ID within the HRU dataset determined the drain order. The reach routing module was used to determine the flow within the channels of the river using the kinematic wave equation and calculations of flow according to Manning and Strickler. The river discharge was determined using the roughness coefficient of the stream (Manning roughness), the slope and width of the river channel and calculations of flow velocity and hydraulic radius calculated during model simulations.

3.5. J2000 Input data

After the above adaptations of the J2000 model coding, input data representing both the surface water and groundwater components were required.

3.5.1. Surface water components

Climate and rainfall: Rainfall, windspeed, relative humidity, solar radiation and air temperature were monitored by Automated Weather Stations (AWS) within and outside of the study catchment (Fig. 1). Of the climate and rainfall data used during the surface water modelling (Watson et al., 2018), data was sourced from six AWS's of which four stations were owned by the South African Weather Service (SAWS) and three by the Agricultural Research Council (ARC). Two stations that were installed for the surface water modelling, namely Moutonshoek (M-AWS) and Confluence (CN-AWS) were used for climate and rainfall validation due to their short record length. Additional rainfall data collected by farmers at high elevation at location FF-R and within the middle of the catchment at KK-R were used to improve the climate and rainfall network density.

Landuse classification: The vegetation and landuse dataset that was used for the sub-catchment (CSIR, 2009) included five different landuse classes: 1) wetlands and waterbodies, 2) cultivated (temporary, commercial, dryland), 3) shrubland and low fynbos, 4) thicket, bushveld, bush clumps and high fynbos and 5) cultivated (permanent, commercial, irrigated). Each different landuse class was assigned an albedo, root depth and seal grade value of soil surface based on previous studies (Steudel et al., 2015)(Supplementary: Table 2). The Leaf Area Index (LAI) and vegetation height varies by growing season with different values of each for the particular growing season. While surface resistance of the landuse varied monthly within the model, the values only vary significantly between growing seasons (Steudel et al., 2015).

Soil dataset: The Harmonized World Soil Database (HWSD) v1.2 (Batjes et al., 2012) was the input soil dataset, with nine different soil forms within the sub-catchment (Supplementary: Table 3). Within the HWSD, soil depth, soil texture and granulometry were used to calculate and assign soil parameters within the J2000 model. MPS and LPS which differ in terms of the soil structure and pore size were determined in Watson et al. (2018), using pedotransfer functions within the HYDRUS model (Supplementary: Table 3).

Streamflow and water levels: Streamflow, measured at the Department of Water Affairs (DWA) gauging station G3H001 between 1970-2009, at the outlet of the Kruismans tributary (Het Kruis) (Fig 1 and 2), was used for surface water calibration. The G3H001 two-stage weir could record a maximum flow rate of 3.675 m2.s-1 due to the DT limitations of the structure. After 2009, the G3H001 structure was decommissioned due to structural damage, although repairs are expected in the near future due to increasing concerns regarding the influx of freshwater into the lake. Water levels measured at the sub-catchment outlet at DWA station G3T001 (Fig 1) between 1994 to 2018 were used for EMD filtering.

3.5.2. Groundwater components

Net recharge and hydraulic conductivity: The net recharge and hydraulic conductivity values used for the groundwater model calibration were collected from detailed MODFLOW modelling conducted for the Krom Antonies tributary and applied to the entire sub-catchment (Fig. 3) (Watson, submitted). The net recharge and aquifer hydraulic conductivity for the Krom Antonies tributary, was estimated through PEST autocalibration using hydraulic conductivities from previous studies (SRK, 2009; UMVOTO-SRK, 2000) and potential recharge estimates (Watson et al., 2018).

Hydrogeology: Within the hydrogeological dataset, parameters assigned include maximum storage capacity (RG1 and RG2), storage coefficients (RG1 and RG2), the minimum permeability/maximum percolation (Kf_geo of RG1 and RG2) and depth of the upper groundwater reservoir (depthRG1). The maximum storage capacity was determined using an average thickness of each aquifer and the porosity, where the primary aquifer thickness was assumed to be between 15-20 m (Conrad et al., 2004), and the secondary aquifer between 80-200 m (SRK, 2009). The maximum percolation of the different geological formations was assigned hydraulic conductivities using the groundwater model for the Krom Antonies sub-catchment (Watson et al., submitted). The J2000 geological formations were assigned conductivities to modify the maximum percolation value to ensure internal consistency with recharge values calculated using MODFLOW (Table 1).

3.6. Model calibration

3.6.1. Model sensitivity

The J2000 sensitivity analysis for Verlorenvlei sub-catchment was presented in Watson *et al.*, (2018) and therefore only a short summary is presented here. In this study, parameters that were used to control the ratio of interflow to percolation were adjusted, which in the J2000 model include a slope (SoilLatVertDist) and max percolation value. The sensitivity analysis conducted by Watson *et al.*, (2018) showed that for high flow conditions (E2) (Nash-Sutcliffe efficiency in its standard squared), model outputs are most sensitive to the slope factor, while for low flow conditions (E1) (modified Nash-Sutcliffe efficiency in a linear form) the model outputs were most sensitive to the maximum infiltration rate of the soil (ie. the parameter maxInfiltrationWet) (Supplementary: Figure 1). The max percolation was moderately sensitive during wet and dry conditions, and together with the slope factor, controlled the interflow to percolation portioning that was calibrated in this study.

3.6.2. Surface water calibration

The surface water parameters of the model were calibrated for the Kruismans tributary (688 km^2) (Fig. 2) using the gauging data from G3H001 (Fig. 4 and Table 1). The streamflow data used for the calibration was between 1986-1993, with model validation between 1994 to 2007 (Fig. 4). This specific calibration period was selected due to the wide range of different runoff conditions experienced at the station, with both low and high flow events being recorded. For calibration, the modelled discharge was manipulated in the same fashion, with a maximum value of 3.675 m^3/s, so that the tributary streamflow behaved as measured discharge. An automated model calibration was performed using the "Nondominating Sorting Genetic Algorithm II" (NSGA-II) multi-objective optimisation method (Deb et al., 2002) with 1023 model runs being performed. Narrow ranges of calibration parameters (FC_Adaptation, AC_Adaptation, soilMaxDPS, gwRG1Fact and gwRG2Fact) were used to (1) achieve a representative sub-catchment hydrograph and (2) ensure J2000 net recharge estimates agreed with modelled values from the MODFLOW model for the Krom Antonies.

As objective functions, the E2, E1 and the average bias in % (Pbias) were utilized for the calibration (Krause et al., 2005) (Table 2). The choice of the optimized parameter set was made to ensure that E2 was better than 0.57 (best value was 0.574302) and the Pbias better than 5 % (Table 1). From the automated calibration, 308 parameter sets were determined with the best E1 being chosen to ensure that the model is representative of low flow conditions (Table 1).

3.6.3. Model validation

For the surface water model validation, the streamflow records between 1994-2007 were used, where absolute values (E1) and squared differences (E2) of the Nash Sutcliffe efficiency were reported. The Pbias was also used as an objective function to report the model performance by comparison between measured and modelled streamflow (Table 2). Although gauging station limitations resulted in good objective functions from the model, the performance of objective functions E1, E2 and Pbais reduced between the validation and calibration period (Table 2). During the calibration period there was a good fit between modelled and measured streamflow (Pbias=-1.82, AVG=-19.23, KGE=0.78), with a significant difference between modelled and measured streamflow during the validation period (Pbias=-19.2, AVG=-269.20, KGE=0.67). The calibration was performed over a wet cycle (1986-1997), resulting in a few occurrences where streamflow events exceeded the DT limit of 3.675 m^3.s^{-1}, thereby reducing the number of calibration points. In contrast

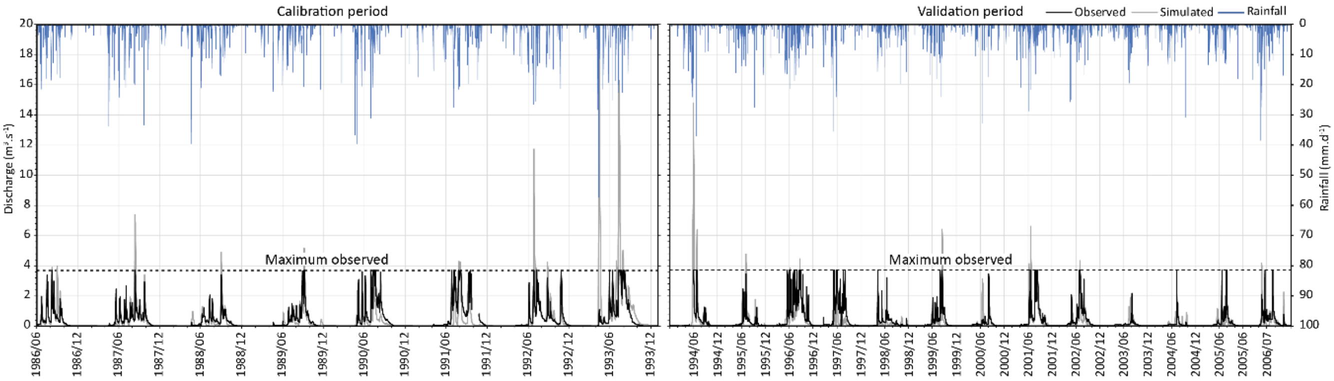

Fig. 4. The surface water calibration (1986-1993) and validation (1986-2006) of the J2000 model using gauging data from the G3H001

the validation was performed over a dry cycle (1997-2007), which resulted in more data points as few streamflow events exceeded 3.675 $m^3.s^{-1}$.

The groundwater recharge values from MODFLOW were validated with J2000 recharge estimates (Fig. 5). During the calibration period the groundwater recharge proportion for the eight calibrated hydraulic zones (Fig. 3) achieved a good fit, with an average value from J2000 of 4.71 % and from MODFLOW of 4.58 %. The coefficient of determination (R2) between the J2000 and MODFLOW was 0.81. Across the entire dataset J2000 overestimates groundwater recharge by 2.75 %, although the coefficient of determination produced an R^2 of 0.92 which is higher than the calibration period.

Table 2
The objective functions E1, E2, logarithmic versions of E1 and E2, average error (AVG) coefficient of determination R2, Pbias and Kling Gupta efficiency (KGE)(Gupta et al., 2009) used for the surface water calibration (1987-1993) and validation (1994-2007)

	Calibration 1987-1993	Validation 1994-2007
E1	0.55085	0.53312
E2	0.57156	0.55736
LogE1	0.28007	0.10097
LogE2	0.46028	0.19007
AVE	-19.23814	-269.20398
R^2	0.61788	0.58067
Pbias	-1.82301	-19.23758
KGE	0.78527	0.67417

3.7. EMD filtering

To account for missing streamflow data between 2007-2017, an Empirical Mode Decomposition (EMD) (Huang et al., 1998) as applied to the measured water level data at the sub-catchment outlet (G3T001)(Fig. 1) between 2008 to 2018 (Fig 6a). EMD is a method for the decomposition of nonlinear and nonstationary signals into sub-signals of varying frequency, so-called intrinsic mode functions (IMF), and a residuum signal. By removing one or more IMF or the residuum signal, certain frequencies (e.g. noise) or an underlying trend can be removed from the original time series data. This approach was successfully applied to the analysis of river runoff data (Huang et al., 2009) and forecasting of hydrological time series (Kisi et al., 2014). In this study, EMD filtering was used to remove high frequency sub-signals from simulated runoff and measured water level data to compare the more general seasonal variations of both signals (Fig. 6b).

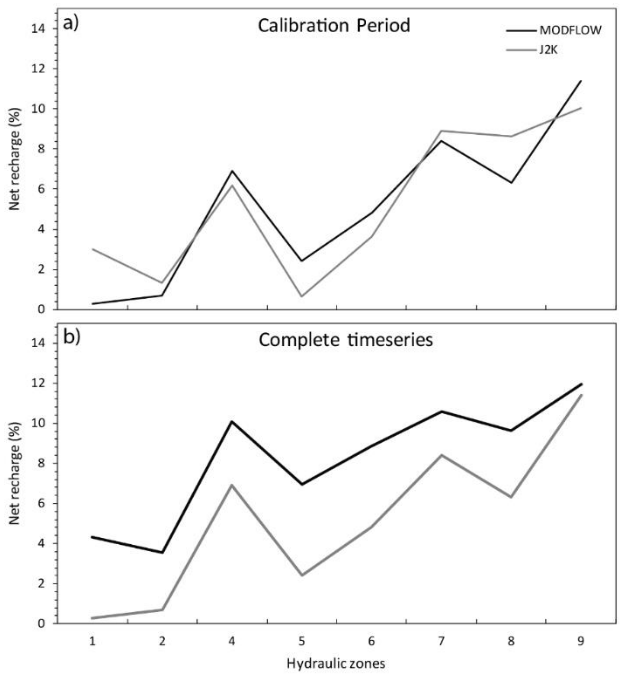

Fig. 5. The groundwater calibration for each hydraulic zone with a) net recharge for the J2000 and MODFLOW during the model calibration (2016) and b) the net recharge deviation between MODFLOW and J2000 across the entire modelling timestep (1986-2017)

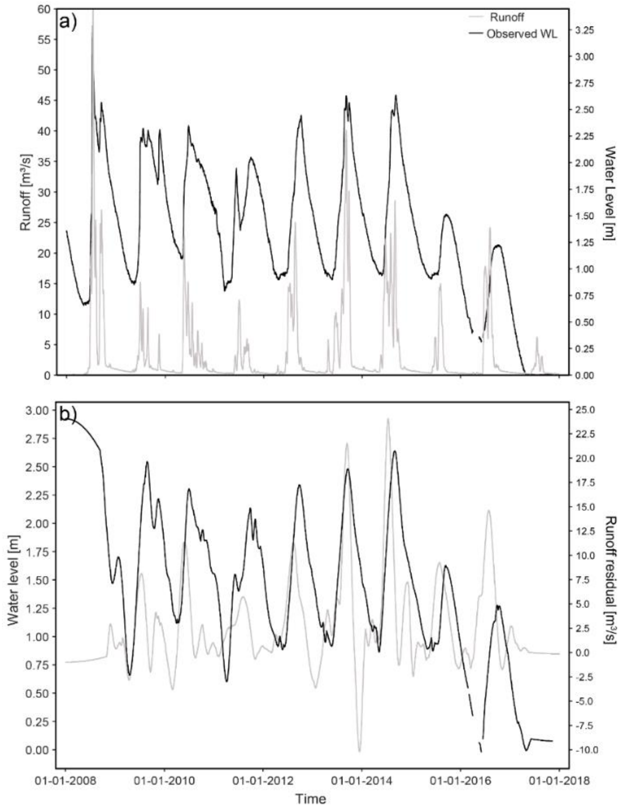

Fig. 6. a) The water level fluctuations at station G3T001 with modelled runoff and b) the EMD filtering showing the variation in discharge timeseries attributed a water level change at the station.

4. Results

The J2000 model was used to simulate both runoff and baseflow, with runoff being comprised of direct surface runoff (RD1) and interflow (RD2) and baseflow simulated from the primary (RG1) and secondary aquifer (RG2). Below, the results of the modelled streamflow and baseflow are presented, along with the total flow contribution of each tributary, the runoff to baseflow proportioning and stream exceedance probabilities. The coefficient of variation (CV) was used to determine the streamflow variability of each tributary, while the baseflow index (BFI) was used to determine the baseflow and runoff proportion.

4.1. Streamflow and baseflow

Streamflow for the sub-catchment shows two distinctively wet periods (1987-1997 and 2007-2017), separated by a dry period (1997-2007) (Fig. 7). Yearly sub-catchment rainfall volumes between 1987-1997 were between 288 and 492 $mm.yr^{-1}$, with an average of 404 $mm.yr^{-1}$. For this period, average yearly streamflow between 1987-1997 was 1.4 $m^3.s^{-1}$, with an average baseflow contribution of 0.63 $m^3.s^{-1}$. The modelled streamflow reached a maximum of 48 $m^3.s^{-1}$ in 1993, when 5 $m^3.s^{-1}$ of baseflow was generated after 58 mm of rainfall was received. Between 1997-2007 (dry period) sub-catchment yearly rainfall was between 222 and 394 $mm.yr^{-1}$ with an average of 330 $mm.yr^{-1}$ (Fig. 7). For this period, average yearly streamflow between 1997-2007 was 0.44 $m^3.s^{-1}$, with an average baseflow contribution of 0.18 $m^3.s^{-1}$. The modelled streamflow reached a maximum of 11 $m^3.s^{-1}$ in 2002, with a baseflow contribution of 2.5 $m^3.s^{-1}$ after 28 mm of rainfall was received. Between 2007-2017 (wet period) sub-catchment yearly rainfall was between 231 and 582 $mm.yr^{-1}$ with an average of 427 $mm.yr^{-1}$ (Fig. 7). Over this period, average yearly streamflow between 2007-2017 was 2.5 $m^3.s^{-1}$ with an average baseflow contribution of 1.3 $m^3.s^{-1}$. The modelled streamflow reached a maximum of 52 $m^3.s^{-1}$ in 2008, with 13 $m^3.s^{-1}$ of baseflow generated after two consecutive rainfall events each of 25 mm.

4.2. Tributary contributions

The four main feeding tributaries (Bergvallei, Kruismans, Hol and Krom Antonies) together contribute 81% of streamflow for the Verlorenvlei, with the additional 19% from small tributaries near Redelinghuys (Fig. 7). The Kruismans contributes most of the total streamflow with 32.4 %, although due to the sub-catchment being the largest of the tributaries (688 km²), the area weighted contribution is 16.4 % (Fig. 7). The Bergvallei (320 km²), which is smaller than the Kruismans, contributes 29 % of the total flow with an area weighted contribution of 32 %. The Krom Antonies has the largest area weighted contribution of 33 % due to its small size (140 km²) in comparison to the other tributaries, although the Krom Antonies contributes only 13 % of the total flow (Fig. 7). The Hol (126 km²) contributes the least total flow with 6.79 %, with a weighted contribution of 18.69 % (Fig. 7).

4.3. Flow variability

Streamflow that enters Verlorenvlei has a large daily variability with a coefficient of variation of 189.90 (Fig. 7). Verlorenvlei's streamflow is mainly comprised of surface runoff (RD1/RD2=15.6) as opposed to interflow. The total groundwater flow contribution for the Verlorenvlei is 47 % (BFI=0.47) with the majority of groundwater baseflow from the secondary aquifer (RG1/RG2=0.29). The Kruismans has large daily streamflow variability with a CV of 217.20 (Fig. 7). The Kruismans tributary is mainly comprised of surface runoff (RD1/RD2=13.9) with a small interflow contribution. The total groundwater flow contribution for the Kruismans tributary is relatively low, with groundwater making up 14 % of streamflow (BFI=0.14), where the majority of baseflow is from the secondary aquifer (RG1/RG2=0.37). The Bergvallei has the highest streamflow variability, with a CV of 284.54, and the highest surface runoff to interflow proportion (RD1/RD2=22.50), with a total groundwater contribution of 49 % (BFI=0.49) (Fig. 7). The secondary aquifer contributes the majority of baseflow for the Bergvallei, with the secondary

Table 3
The exceedance probabilities for sub-catchment rainfall and Verlorenvlei, Kruismans, Bergvallei, Krom Antonies and Hol streamflow in $m^3.s^{-1}$ and $m^3.d^{-1}$

Exceedance percentile	Rainfall	Verlorenvlei		Kruismans		Bergvallei		Krom Antonies		Hol	
	mm/yr^{-1}	$m^3.s^{-1}$	$m^3.d^{-1}$	$m^3.s^{-1}$	$m^3.d^{-1}$	$m^3.s^{-1}$	$m^3.d^{-1}$	$m^3.s^{-1}$	$m^3.d^{-1}$	$m^3.s^{-1}$	$m^3.d^{-1}$
95	227	0.054	4702	0.004	346	0.001	69	0.001	109	0.002	176
90	264	0.074	6356	0.007	604	0.002	191	0.003	232	0.003	269
85	282	0.088	7628	0.010	830	0.004	366	0.004	319	0.004	353
80	290	0.104	8979	0.012	1072	0.007	596	0.005	392	0.005	434
75	296	0.119	10303	0.015	1291	0.010	839	0.005	459	0.006	508
70	324	0.136	11759	0.018	1517	0.013	1104	0.006	534	0.007	587
65	357	0.155	13373	0.021	1791	0.016	1381	0.007	602	0.008	676
60	387	0.054	4702	0.024	2104	0.019	1657	0.008	685	0.009	786
55	397	0.203	17575	0.029	2506	0.023	1965	0.009	772	0.011	913
50	405	0.237	20498	0.035	3032	0.027	2309	0.010	882	0.012	1058
45	422	0.286	24669	0.043	3755	0.032	2807	0.012	1024	0.014	1222
40	430	0.371	32023	0.058	5022	0.041	3511	0.015	1258	0.017	1439
35	437	0.516	44598	0.089	7699	0.053	4613	0.020	1745	0.021	1790
30	444	0.710	61310	0.156	13511	0.076	6599	0.033	2824	0.029	2481
25	454	1.067	92204	0.291	25182	0.123	10619	0.062	5387	0.044	3814
20	481	1.571	135726	0.489	42242	0.223	19295	0.110	9511	0.065	5655
15	498	2.399	207275	0.780	67408	0.421	36354	0.192	16594	0.096	8262
10	537	3.759	324746	1.324	114432	0.885	76477	0.359	31045	0.141	12191
5	575	6.939	599535	2.490	215152	1.884	162795	0.929	80305	0.224	19312

aquifer contribution being more than double the primary aquifer (RG1/RG2=0.54). The Krom Antonies has significant variability in daily streamflow, with a CV of 283.00 (Fig. 7). The runoff from the Krom Antonies is mainly comprised of surface runoff with interflow being a minor contribution (RD1/RD2=14.30). The Krom Antonies has a relatively high groundwater component (BFI=0.34), with the secondary aquifer contributing the most baseflow (RG1/RG2=0.33). The Hol tributary has the lowest variability in daily streamflow with a CV of 146.54 (Fig. 7). The Hol tributary is mainly comprised of surface runoff (RD1/RD2=9.40), although the interflow is the highest proportion within the sub-catchment. The Hol tributary is mainly comprised of groundwater (BFI=0.56), with the majority of baseflow being derived from the secondary aquifer (RG1/RG2=0.17).

4.4. Flow exceedance probabilities

The results for the flow exceedance probabilities includes flow volumes for each tributary and the lake's inflow which are exceeded 95 %, 75 %, 50 %, 25 % and 5 % of the time. The 95 percentile corresponds to a lake inflow of 0.054 $m^3.s^{-1}$ or 4,702 $m^3.d^{-1}$, with between 0.001-0.004 $m^3.s^{-1}$ from the feeding tributaries (Table 3). The 75-percentile flow, which is exceeded 3/4 of the time corresponds to an inflow of 0.119 $m^3.s^{-1}$ or 10,303 $m^3.d^{-1}$, with between 0.005-0.015 $m^3.s^{-1}$ from the feeding tributaries. Average (50 percentile) streamflow flowing into the Verlorenvlei is 0.237 $m^3.s^{-1}$ or 20,498 $m^3.d^{-1}$, with between 0.012-0.012 $m^3.s^{-1}$ from the feeding tributaries. The 25-percentile flow, which is exceeded ¼ of the time corresponds to a lake inflow of 1,067 $m^3.s^{-1}$ or 92,204 $m^3.d^{-1}$ with between 0.044-0.291 $m^3.s^{-1}$ from the feeding tributaries. The lake inflows that are exceeded 5 % of the time correspond to 6.939 $m^3.s^{-1}$ or 599,535 $m^3.d^{-1}$ with between 0.224-2.49 $m^3.s^{-1}$ from the feeding tributaries.

5. Discussion

5.1. Modelling in sub-Saharan Africa

A major limitation facing the development and construction of comprehensive modelling systems in sub-Saharan Africa is the availability of appropriate climate and streamflow data. For this study, while there was access to over 20 years of streamflow records, the station was only able to measure a maximum of 3.675 $m^3.s^{-1}$, which hindered calibration of the model for high flow events. As such, the confidence in the model's ability to simulate high streamflow events using climate records is limited. While the availability of measured data is a limitation that could affect the modelled streamflow, discontinuous climate records also hindered the estimations of long time series streamflow. Over the course of the 30-year modelling period, a number of climate stations used for regionalisation were decommissioned and were replaced by stations in different areas. This required climate regionalisation adaption for simulations over the entire 30-year period to incorporate the measured streamflow from the gauging station. To account for missing streamflow records since 2007, an EMD filtering protocol was applied to the runoff data (Fig. 6). The results from the EMD filtering showed that after removing the first nine IMFs, the local maxima of both signals match the seasonal water level maxima during most of the years. While considerable improvement can be made to the EMD filtering, the results show some agreement which suggested that the simulated runoff was representative of inflows into the lake. In data scares catchments it is important to make use of all available data in an effort to improve the understanding of the catchment dynamics. To account for historical gauging data a number of adaptions were made to the climate regionalisation, as well as an EMD filtering protocol to use water level data at the sub-catchment outlet. Consequently, the model performed particularly well considering the streamflow and climate station limitations, although the model is yet to be tested regarding its ability to simulate high flow events.

5.2. Catchment dynamics

Factors that impact on streamflow variability are important for understanding river flow regime dynamics. Previously, factors that affect streamflow variability such as CV and BFI values have been used to determine how susceptible particular river systems are to drought (e.g Hughes and Hannart, 2003). While CV values have been used to account for climatic impacts such as dry and wet cycles, BFI values are associated with runoff generation processes that impact the catchment. For South African river systems, BFI values are generally below 1 implying that runoff exceeds baseflow. In comparison CV values can be in excess of 10 implying high variability in streamflow volumes (Hughes and Hannart, 2003). Generally, CV and BFI measures have been applied to quaternary river systems in southern Africa. For this study, these two measurements have been used to understand river flow dynamics in much smaller tributaries.

The highest proportion of streamflow needed to sustain the Verlorenvlei lake water level is received from the Bergvallei tributary, although the area weighted contribution from the Krom Antonies is more significant (Fig. 7). However, CV values for the Bergvallei indicate high streamflow variability. This is partially due to the high surface runoff component in modelled streamflow within the Bergvallei in comparison to the minor interflow contribution, suggesting little sub-surface runoff. While streamflow from the Bergvallei tributary is 47% groundwater, which would suggest a more sustained streamflow, due to the TMG dominance as well as a high primary aquifer contribution, baseflow from the Bergvallei is driven by highly conductive rock and sediment materials. Similarly, CV values for the Krom Antonies indicate high streamflow variability due to the presence of a high baseflow contribution from the conductive TMG and primary aquifers. Although the Krom Antonies has a larger interflow component, which would reduce streamflow variability, the dominant TMG presence within this tributary partially compensates for the subsurface flow contributions.

In contrast, the Hol has a much smaller daily streamflow variability in comparison to both the Bergvallei and the Krom Antonies (Fig. 7). While streamflow from the Hol tributary is mainly comprised of baseflow (56 %), the dominance of low conductive shale rock formations as well as a large interflow component result in reduced streamflow variability. While the larger shale dominance in this tributary not only results in a more sustained baseflow from the secondary aquifer, it also results in large interflow due to the limited conductivity of the shale formations. Compounding the more sustained baseflow from the Hol tributary, the reduced presence of the primary aquifer results in a dominance in slow groundwater flow from this tributary. Similarly, the Kruismans is dominated by shale formations which result in a larger interflow contribution, although due to the limited baseflow contribution (14 %) the streamflow from this tributary is highly variable, which impacts on its susceptibility to drought.

The results from this study have shown that while the Krom Antonies was initially believed to be the major flow contributor, the Bergvallei is in fact the most significant or most likely the largest contributor, with streamflow from the four tributaries being highly variable and baseflow from the Hol tributary being the only constant input source. The presence of conductive TMG sandstones and quaternary sediments in both the Krom Antonies and Bergvallei result in quick baseflow responses with little flow attenuation. The potential implication of a constant source of groundwater being provided from the Hol tributary, is that if the groundwater is of poor quality this would result in a constant input of saline groundwater, with the Krom Antonies and Bergvallei providing freshwater after sufficient rainfall.

5.3. Baseflow comparison

The groundwater components of the J2000 model were adjusted using calibrated net recharge and aquifer hydraulic conductivity from a MODFLOW model of one of the main feeding tributaries of the Verlorenvlei. The Krom Antonies was selected for this calibration as it was previously believed to be the largest input of groundwater to Verlorenvlei (Fig. 2). Baseflow for the Krom Antonies tributary was previously calculated using a MODFLOW model (Watson et al., 2018), by considering aquifer hydraulic conductivity and average groundwater recharge. Due to the fact that average recharge was used, baseflow estimates from MODFLOW are likely to fall on the upper end of daily baseflow values. For the Krom Antonies sub-catchment Watson et al., (2018) estimated baseflow between 14,000 to 19,000 $m^3.d^{-1}$ for 2010-2016. In this study, similar daily estimates were only exceeded 10 % of

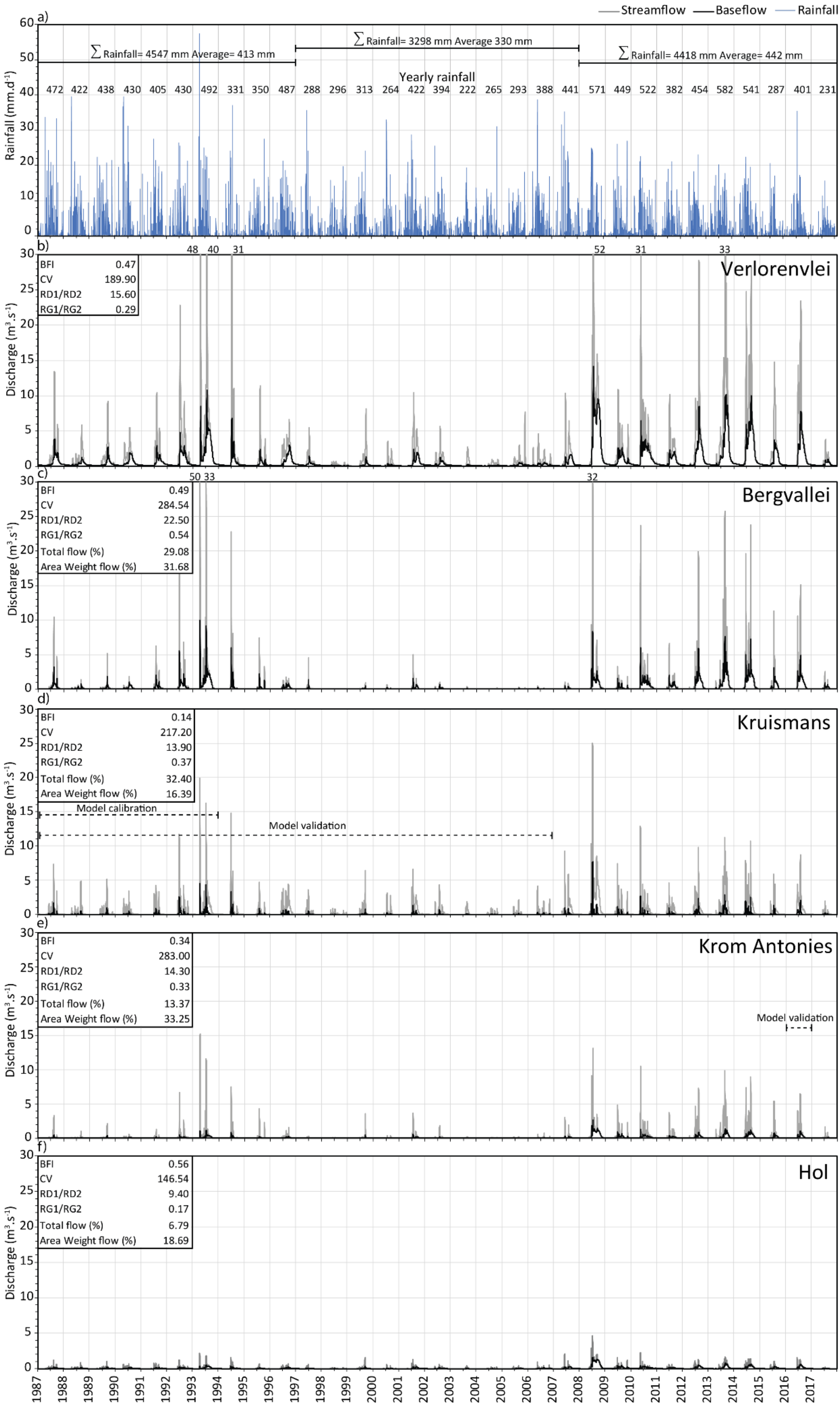
Streamflow
Baseflow
Rainfall
a)
∑Rainfall= 4547 mm Average= 413 mm
∑Rainfall= 3298 mm Average 330 mm
∑Rainfall= 4418 mm Average= 442 mm
Yearly rainfall
472 422 438 430 405 430 492 331 350 487 288 296 313 264 422 394 222 265 293 388 441 571 449 522 382 454 582 541 287 401 231
Rainfall (mm.d⁻¹)
b)
Verlorenvlei
BFI 0.47
CV 189.90
RD1/RD2 15.60
RG1/RG2 0.29
48 40 31 52 31 33
Discharge (m³.s⁻¹)
c)
Bergvallei
BFI 0.49
CV 284.54
RD1/RD2 22.50
RG1/RG2 0.54
Total flow (%) 29.08
Area Weight flow (%) 31.68
50 33 32
d)
Kruismans
BFI 0.14
CV 217.20
RD1/RD2 13.90
RG1/RG2 0.37
Total flow (%) 32.40
Area Weight flow (%) 16.39
Model calibration
Model validation
e)
Krom Antonies
BFI 0.34
CV 283.00
RD1/RD2 14.30
RG1/RG2 0.33
Total flow (%) 13.37
Area Weight flow (%) 33.25
Model validation
f)
Hol
BFI 0.56
CV 146.54
RD1/RD2 9.40
RG1/RG2 0.17
Total flow (%) 6.79
Area Weight flow (%) 18.69
1987 1988 1989 1990 1991 1992 1993 1994 1995 1996 1997 1998 1999 2000 2001 2002 2003 2004 2005 2006 2007 2008 2009 2010 2011 2012 2013 2014 2015 2016 2017

Fig. 7. a) The average sub-catchment rainfall between 1987-2017 showing wet cycles (1987-1997 and 2008-2017), the modelled streamflow and baseflow inflows for the b) Verlorenvlei, c) Bergvallei, d) Kruismans, e) Krom Antonies and f) Hol with estimated BFI, CV, RD1/RD2, RG1/RG

the time, with average estimates (50 %) of 1,036 $m^3.d^{-1}$ over course of the modelling period (Fig. 7).

Watson et al., (2018) estimates were applied over the course of a wet cycle (2016), and in comparison average baseflow from J2000 for 2016 was 8, 214 $m^3.d^{-1}$. The daily timestep nature of the J2000 is likely to result in far lower baseflow estimates, as recharge is only received over a 6-month period as opposed to a yearly average estimate. The possible implication of this is that while common groundwater abstraction scenarios have been based on yearly recharge, abstraction is likely to exceed sustainable volumes during dry months or dry cycles which could hinder the ability of the aquifer to supply baseflow. While the groundwater components of the J2000 have been distributed to allow for improved baseflow estimates, the groundwater calibration was applied to the Krom Antonies. However, this study showed that Bergvallei has been identified as the largest water contributor. In hind sight, the use of geochemistry to identify dominant tributaries could have aided the groundwater calibration, as well as the geochemistry data being used within the model calibration. While it would have been beneficial to calibrate the groundwater components of the J2000 using the Bergvallei, incorporating one tributary that is dominated by TMG outcrops and one by shale would have improved the representativeness of the baseflow estimates from the model. While the distribution of aquifer components improved modelled baseflow, including groundwater abstraction scenarios in baseflow modelling in the sub-catchment is important for future water management for this ecologically significant area.

5.4. Ecological reserve and evaporative demand

Exceedance probabilities have been used as approximate estimates of minimum river flow requirements. The exceedance percentiles used for ecological reserve determination are streamflow values that are exceeded 95 % of the time (Barker and Kirmond, 1998). For this study, exceedance probabilities were estimated through rainfall/runoff modelling for the previous 30 years within the Verlorenvlei sub-catchment. The exceedance probabilities were determined for each tributary, as well as the total inflows into the lake. These exceedance probabilities were compared with the evaporative demand of the lake, to understand whether inflows are in surplus or whether evaporation demands exceed inflows. As an approximation of the evaporative demand of the Verlorenvlei, an average evaporation loss of 5 mm.d-1 (Schulze et al., 2007) was assumed across the lake's surface area (15 km^2).

The 95th percentile streamflow contribution, which is the ecological reserve percentile, corresponds to a lake inflow of 4,702 $m^3.d^{-1}$, meeting the evaporation demand if the lake was at 7 % capacity. From this it does not seem that the 95th percentile is enough to balance the evaporation demand of the lake. Furthermore, an average streamflow (50th percentile) would only meet the evaporation demand of 1/4 of the lake's surface area. Considering the exceedance probability of the wet cycle period (2007-2017), the 95th percentile corresponded to 7,093 $m^3.d^{-1}$, meeting the evaporation demand if the lake was at 10 % capacity, while an average inflow (50 percentile) would meet demands if the lake was at 40 % capacity. In contrast, for the dry cycle (1997-2007), the 95 percentile would correspond to a streamflow of 3,438 $m^3.d^{-1}$, meeting the demand if the lake was at 5 % capacity, while on average (50 percentile) the demands of 15 % of the lake's capacity would be met.

From the exceedance probabilities generated in this study, the lake is predominately fed by less frequent large discharge events, where on average the daily inflows to the lake do not sustain the water level above 40 % capacity. This is particularly evident in the measured water level data from station G3T001, where measured water levels have a large daily standard deviation (0.62) (Watson et al., 2018). With climate change likely to impact the length and severity of dry cycles, it is likely that the lake will dry up more frequently into the future, which could have severe implications on the biodiversity that relies on the lake's habitat for survival. Of importance to the lake's survival is the protection of river inflows during wet cycles, where the lake requires these inflows for regeneration and the overallocation of resources could result in prolonged dry cycle conditions.

6. Conclusions

Understanding river flow regime dynamics is important for the management of ecosystems that are sensitive to streamflow fluctuations. While climatic factors impact rainfall volumes during wet and dry cycles, factors that control catchment runoff and baseflow are key to the implementation of river protection strategies. In this study, groundwater components within the J2000 model were distributed to improve baseflow and runoff proportioning for the Verlorenvlei sub-catchment. The J2000 was distributed using groundwater model values for the dominant baseflow tributary, while calibration was applied to the dominant streamflow tributary. The model calibration was hindered by the maximum gauging station resolution, which reduced the confidence in modelling high flow events, although an EMD filtering protocol was applied to account for the resolution limitations and missing streamflow records. The modelling approach would likely be transferable to other partially gauged semi-arid catchments, provided that groundwater recharge is well constrained. The daily timestep nature of the J2000 model allowed for an in-depth understanding of tributary flow regime dynamics, showing that while streamflow variability is influenced by the runoff to baseflow proportion, the host rock or sediment in which groundwater is held is also a factor that must be considered. The modelling results showed that on average the streamflow influxes were not able to meet the evaporation demand of the lake. High-flow events, although they occur infrequently, are responsible for regeneration of the lake's water level and ecology, which illustrates the importance of wet cycles in maintaining biodiversity levels in semi-arid environments. With climate change likely to impact the length and occurrence of dry cycle conditions, wet cycles are important for ecosystem regeneration.

7. Acknowledgements

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8. References

Acreman, M.C., Dunbar, M.J., 2004. Defining environmental river flow requirements – a review. Hydrol. Earth Syst. Sci. 8, 861–876. doi:10.5194/hess-8-861-2004

Barker, I., Kirmond, A., 1998. Managing surface water abstraction in Wheater, H. and Kirby, C.(eds) Hydrology in a changing environment, vol1. Br. Hydrol. Soc. pp249-258.

Batjes, N., Dijkshoorn, K., Van Engelen, V., Fischer, G., Jones, A., Montanarella, L., Petri, M., Prieler, S., Teixeira, E., Wiberg, D., 2012. Harmonized World Soil Database (version 1.2). Tech. rep., FAO and IIASA, Rome, Italy and Laxenburg, Austria.

Conrad, J., Nel, J., Wentzel, J., 2004. The challenges and implications of assessing groundwater recharge: A case study- northern Sandveld , Western Cape, South Africa. Water SA 30, 75–81.

Costanza, R., Arge, R., Groot, R. De, Farberk, S., Grasso, M., Hannon, B., Limburg, K., Naeem, S., O'Neill, R. V, Paruelo, J., Raskin, R.G., Suttonkk, P., van den Belt, M., 1997. The value of the world ' s ecosystem services and natural capital. Nature 387, 253–260. doi:10.1038/387253a0

CSIR, 2009. Development of the Verlorenvlei estuarine management plan: Situation assessment. Report prepared for the C.A.P.E. Estuaries Programme 142.

Domenico, P.A., Schwartz, F.W., 1990. Physical and Chemical Hydrogeology. John Wiley and Sons, Inc., New York.

Flügel, W., 1995. Delineating hydrological response units by geographical information system analyses for regional hydrological modelling using PRMS/MMS in the drainage basin of the River Bröl, Germany. Hydrol. Process. 9, 423–436.

Gupta, H. V., Kling, H., Yilmaz, K.K., Martinez, G.F., 2009. Decomposition of the mean squared error and NSE performance criteria: Implications for improving hydrological modelling. J. Hydrol. 377, 80–91. doi:10.1016/j.jhydrol.2009.08.003

Harbaugh, Arlen, W., 2005. MODFLOW-2005 , The U . S . Geological Survey Modular Ground-Water Model — the Ground-Water Flow Process. U.S. Geol. Surv. Tech. Methods 253.

Huang, N.E., Shen, Z., Long, S.R., Wu, M.C., Shih, H.H., Zheng, Q., Yen, N.-C., Tung, C.C., Liu, H.H., 1998. The empirical mode decomposition and the Hilbert spectrum for nonlinear and non-stationary time series analysis, in: Proceedings of the Royal Society of London A: Mathematical, Physical and Engineering Sciences. The Royal Society, pp. 903–995.

Huang, Y., Schmitt, F.G., Lu, Z., Liu, Y., 2009. Analysis of daily river flow fluctuations using empirical mode decomposition and arbitrary order Hilbert spectral analysis. J. Hydrol. 373, 103–111.

Hughes, D.A., 2001. Providing hydrological information and data analysis tools for the determination of ecological instream flow requirements for South African rivers. J. Hydrol. 241, 140–151.

Hughes, D.A., Hannart, P., 2003. A desktop model used to provide an initial estimate of the ecological instream flow requirements of rivers in South Africa. J. Hydrol. 270, 167–181. doi:10.1016/S0022-1694(02)00290-1

Jenson, S.K., Domingue, J.O., 1988. Extracting topographic structure from digital elevation data for geographic information system analysis. Photogramm. Eng. Remote Sensing 54, 1593–1600.

Kim, N.W., Chung, I.M., Won, Y.S., Arnold, J.G., 2008. Development and application of the integrated SWAT-MODFLOW model. J. Hydrol. 356, 1–16. doi:10.1016/j.jhydrol.2008.02.024

King, J., Louw, D., 1998. Instream flow assessments for regulated rivers in South Africa using the Building Block Methodology. Aquat. Ecosyst. Health Manag. 1, 109–124.

Kisi, O., Latifoğlu, L., Latifoğlu, F., 2014. Investigation of empirical mode decomposition in forecasting of hydrological time series. Water Resour. Manag. 28, 4045–4057.

Krause, P., 2001. Das hydrologische Modellsystem J2000. Beschreibung und Anwendung in großen Flussgebieten, in: Umwelt/Environment, Vol. 29. Jülich: Research Centre.

Martens, K., Davies, B.R., Baxter, A.J., Meadows, M.E., 1996. A contribution to the taxonomy and ecology of the Ostracoda (Crustacea) from Verlorenvlei (Western Cape, South Africa). African Zool. 31, 22–36.

Muche, G., Kruger, S., Hillman, T., Josenhans, K., Ribeiro, C., Bazibi, M., Seely, M., Nkonde, E., de Clercq, W.P., Strohbach, B., Kenabatho, K.., Vogt, R., Kaspar, F., Helmschrot, J., Jürgens, N., 2018. Climate change and adaptive land management in southern Africa – assessments, changes, challenges, and solutions, in: Revermann, R., Krewenka, K.M., Schmiedel, U., Olwoch, J.., Helmschrot, J., Jürgens, N. (Eds.), Biodiversity & Ecology. Klaus Hess Publishers, Göttingen & Windhoek, pp. 34–43.

Pfannschmidt, K., 2008. Optimierungsmethoden zur HRU-basierten N/A-Modellierung für eine operationelle Hochwasservorhersage auf Basis prognostischer Klimadaten des Deutschen Wetterdienstes: Untersuchungen in einem mesoskaligen Einzugsgebiet im Thüringer Wald.

Pfennig, B., Kipka, H., Fink, M., Wolf, M., Krause, P., Flügel, W.-A., 2009. Development of an extended routing scheme in reference to consideration of multi-dimensional flow relations between hydrological model entities,. 18th World IMACS / MODSIM Congr. Cairns, Aust. 13-17 July 2009.

Poff, N.L., Allan, J.D., Bain, M.B., Karr, J.R., Prestegaard, K.L., Richter, B.D., Sparks, R.E., Stromberg, J.C., 1997. A paradigm for river conservation and restoration. Bioscience 47, 769–784. doi:10.2307/1313099

Postel, S., Carpenter, S., 1997. Freshwater ecosystem services. Nature's Serv. Soc. Depend. Nat. Ecosyst. 195.

Postel, S., Richter, B., 2012. Rivers for life: managing water for people and nature. Island Press.

Richter, B.D., 2010. Re-thinking environmental flows: from allocations and reserves to sustainability boundaries. River Res. Appl. 26, 1052–1063.

Robson, A., Beven, K., Neal, C., 1992. Towards identifying sources of subsurface flow: A comparison of components identified by a physically based runoff model and those determined by chemical mixing techniques. Hydrol. Process. 6, 199–214. doi:10.1002/hyp.3360060208

Rowlston, W.S., Palmer, C.G., 2002. Processes in the development of resource protection provisions on South African Water Law, in: Proceedings of the International Conference on Environmental Flows for River Systems, Cape Town March.

Schulze, R.E., Maharaj, M., Warburton, M.L., Gers, C.J., Horan, M.J.C., Kunz, R.P., Clark, D.J., 2007. South African atlas of climatology and agrohydrology. Water Res. Comm. Pretoria, RSA, WRC Rep. 1489, 6.

Souchon, Y., Keith, P., 2001. Freshwater fish habitat: science, management, and conservation in France. Aquat. Ecosyst. Health Manag. 4, 401–412.

SRK, 2009. Preliminary Assessment of Impact of the Proposed Riviera Tungsten Mine on Groundwater Resources Preliminary Assessment of Impact of the Proposed Riviera Tungsten Mine on Groundwater Resources.

Steudel, T., Bugan, R., Kipka, H., Pfennig, B., Fink, M., de Clercq, W., Flügel, W.-A., Helmschrot, J., 2015. Implementing contour bank farming practices into the J2000 model to improve hydrological and erosion modelling in semi-arid Western Cape Province of South Africa. Hydrol. Res. 46, 192.

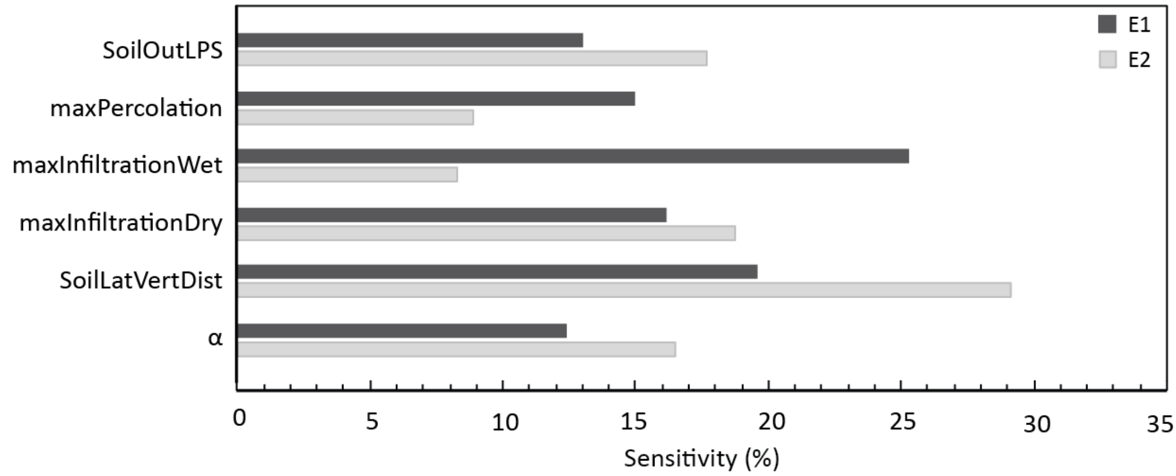
UMVOTO-SRK, 2000. Reconnaissance investigation into the development and utilization of the Table Mountain Group Artesian Groundwater, using the E10 catchment as a pilot study area.

Wagener, T., Wheater, H.S., 2006. Parameter estimation and regionalization for continuous rainfall-runoff models including uncertainty. J. Hydrol. 320, 132–154. doi:10.1016/j.jhydrol.2005.07.015

Watson, A.P., Miller, J.A., Fleischer, M., de Clercq, W.P., 2018. Estimation of groundwater recharge via percolation outputs from a rainfall/runoff model for the Verlorenvlei estuarine system, west coast, South Africa. J. Hydrol. 558, 238–254. doi:10.1016/S1532-0464(03)00032-7

Wishart, M.J., 2000. The terrestrial invertebrate fauna of a temporary stream in southern Africa. African Zool. 35, 193–200.

9. Appendix



Supplementary Fig. 1. The sensitivity analysis conducted for the Verlorenvlei sub-catchment calibration parameters using Nash Sutcliffe Efficiency with squared difference (e2) and absolute values (e1) (Watson et al., 2018)

Supplementary Table 1:
The model calibration parameters for the Verlorenvlei sub-catchment with minimum, maximum and optimized values

Name of parameter	Description of parameter	min	max	optimized value
FC_Adaptation	multiplies the volume of the middle pore storage of soil	0.8	1.5	1.472503
AC_Adaptation	multiplies the volume of the large pore storage of soil	0.8	1.5	0.804489
gw_CapRise	parameter to govern the amount of capillary rise to replenish the soil storage out of the groundwater	1	10	4.412625
gwRG1_RG2_dist	distribution parameter for the slow and fast groundwater runoff (influences overall hydrograph and especially the baseflow)	0	10	3.023532
gwRG1Fact	fast groundwater (slow interflow) delay f (influences overall hydrograph)	0	0.2	0.065366
gwRG2Fact	Base flow delay (influences especially the baseflow)	0	0.2	0.197137
a_rain	Plant canopy interception parameter (mm/LAI) (Influences basically ET)	1	10	6.570599
soilMaxPerc	Conductivity adaption parameter for leaching water to the groundwater storage. (influences baseflow and overall hydrograph, distribution of runoff components)	0	20	5.161027
soilConcRD2	Interflow delay parameter (influences the quick flow components)	2	5	4.94513
soilConcRD1	Surface runoff delay parameter (influences the quick flow components)	1	2	1.33439
soilOutLPS	Outflow parameter of the large pore storage. (Influencing ET, hydrograph and distribution of runoff components)	0.1	1	0.34583
soilMAXDPS	Parameter describing the maximum storage on the soil surface. (influences ET and surface runoff)	3	10	6.681899
soilMaxInfSummer	Decribing the maximum infiltration capacity of soil in the winter period (influences ET and surface runoff)	10	200	118.514
soilMaxInfWinter	Decribing the maximum infiltration capacity of soil in the winter period (influences ET and surface runoff)	10	200	29.30617
soilLinRed	Actual ET parameter, governing the reduction of potential ET according to the soil moisture (influences the ET and therefore the water balance)	0	1	0.333001
flowRouteTA	Stream routing parameter (overall damping of the hydrograph)	1	100	10.17903

Supplementary Table 2

The landuse dataset used for the Verlorenvlei sub-catchment model with the albedo, root depth, sealed grade, LAI, height and surface resistance for each landuse type across 4 different growing seasons

				Growing season											
				1			2			3			4		
		Root depth	Sealed grade	LAI	Height	Surface resistance	LAI	Height	Surface resistance	LAI	Height	Surface resistance	LAI	Height	Surface resistance
Landuse	Albedo	(dm)			(m)	(s/m)		(m)	(s/m)		(m)	(s/m)		(m)	(s/m)
Wetlands and waterbodies	150	0	0	1	0.05	150	1	0.05	150	1	0.05	150	0.05	1	1
Cultivated (temporary, commercial and dryland)	0.25	5	0.3	7	2	50	3.5	1.5	90	0.2	0.2	120	3	0.4	60
Shrubland and low fynbos	0.2	15	0	5	2.5	55	4	2.5	80	1	2	80	2	2	50
Thicket, bushveld, bush clumps, high fynbos	0.15	20	0	7	3	50	10	15	75	2	15	75	7	3	50
Cultivated (permanent, commercial, irrigation)	75.125	2.5	0.15	4	1.025	100	2.25	0.775	120	0.6	0.125	135	1.525	0.7	30.5

Supplementary Table 3

The soil parameter dataset used for the Verlorenvlei sub-catchment with the depth and texture used to estimate MPS, LPS as well as aircap and field capacity for each soil type horizon.

Soil type	Horizon	Depth (mm)	Sand , Silt, Clay (%)	Aircap (mm)	FC Sum (mm)	Waterholding MPS	LPS
Arenosol	A	300	89,6,5	214.43	125.51	33.03	65.37
	B	700	90,5,5			73.36	154.42
					Total	106.39	219.79
Leptosol	A	100	43,29,28	5.87	28.02	28.02	5.87
					Total	28.02	5.87
Solonetz	A	300	35,37,28	137.45	197.63	91.47	13.62
	B	700	27,37,36			221.27	33.95
					Total	312.74	47.57
Fluvisol	A	300	44,33,23	192.52	142.5	85.47	16.38
	B	700	45,31,24			183.47	40.53
					Total	268.94	56.91
Planosol	A	300	56,25,19	138.73	187.45	77.46	22.26
	B	700	44,23,33			191.87	42.98
					Total	269.33	65.24
Regosol	A	300	69,19,12	204.57	116.67	69.63	31.59
	B	700	70,17,13			160.93	72.87
					Total	230.56	104.46
Lixisols	A	300	63,15,22	234.89	125.42	73.41	22.92
	B	700	53,13,34			181.65	43.26
					Total	255.06	66.18
Cambisol	A	300	42,26,32	214.13	120.44	83.79	17.94
	B	700	41,25,34			196.49	41.72
					Total	280.28	59.66
Luvisol	A	300	51,22,27	210.14	122.09	78.69	19.95
	B	700	45,21,34			190.19	43.4
					Total	268.88	63.35

6. Conclusions and Recommendations

Rainfall/runoff modelling has been used to understand water balances in catchments where gauging data is limited or where there is a need to understand the impact that changes in land use and climate could pose to the availability of water resources (e.g Bronstert et al., 2002). In areas that have existing gauging data, models can be used for future forecasting (e.g Warburton et al., 2010). However, in ungauged or partially gauged regions, it is necessary to transfer model calibration parameters from gauged catchments to understand streamflow fluctuations (e.g Patil and Stieglitz, 2014). However, if rainfall/runoff models can be successfully constructed, they have the ability to calculate recharge and baseflow on a daily timestep. In semi-arid environments daily variability is particularly important for groundwater recharge estimates, especially when dealing with highly variable seasonal climate conditions. In contrast to rainfall/runoff models, groundwater models have been used to understand the impact of long-term stresses, such as changes in recharge and abstraction rates, on groundwater sustainability. Although groundwater models lump climate components, they are distributed in terms of aquifer properties, improving the proportioning of interflow to recharge, and thus making estimates of groundwater recharge more spatially and temporally representative. The modelling of recharge and baseflow in semi-arid environments therefore requires the coupling of both rainfall/runoff and groundwater approaches to consider both climatic and aquifer variability. This is the approach implemented in this study to understand the hydrological system in the Verlorenvlei sub-catchment on the west coast of South Africa. The results of the study have important conclusions on three different levels. The first are the implications for the management of the Verlorenvlei system itself. Verlorenvlei is an important small-scale analogue of the greater Cape Town water management situation. Recent droughts have highlighted the vulnerability of the City of Cape Town municipal water supply system, which is dominated by surface water reservoirs, to climate fluctuations. Abstraction of groundwater is an important mechanism for drought relief but most abstraction will occur in the agricultural hinterland to Cape Town. The results of the modelling in Verlorenvlei can help us to understand the potential impacts to the greater Cape region. Secondly, the lessons learned in this study can be applied to other semi-arid environments and particularly those experiencing severe drought such as Melbourne and California. Lastly, the study resulted in significant improvement in the J2000 modelling code by driving the incorporation of distributed groundwater components. This is an important advancement for this model and has direct benefit to all users of the J2000 code.

6.1. The Verlorenvlei Hydrological System

The Verlorenvlei is a water-stressed sub-catchment with numerous users competing for the same water resource. As such the modelling approach used in this study to quantify recharge rates, baseflow volumes and understand streamflow exceedance probabilities can be used to manage this competition, as well as making provisions for the long-term survival of the Verlorenvlei. The modelling approach developed highlighted the impact that geology has on the aquifers ability to receive recharge, as well as the flow attenuation and lag of baseflow from the primary, secondary and TMG aquifer. The modelling conducted for the Verlorenvlei would have benefited from a more extensive monitoring network, although the results can be used to aid in understanding the resource base available within this water stressed region.

6.1.1. *Groundwater recharge*

The potential recharge estimates varied from 1-5 % for the primary aquifer and between 22-25 % for the TMG aquifer. The MODFLOW model further reduced the TMG aquifer recharge to between 6-11 % (reduction of 14-16%), while for the primary aquifer recharge values did not vary between the two models. The baseflow estimates from MODFLOW were between 0.16 to 0.22 m3.s-1 for the Krom Antonies. In comparison, average baseflow values from the J2000 were 0.01 m3.s-1, with similar MODFLOW values only being achieved 15 % of the time (0.192 m3.s-1). The resultant output from the rainfall/runoff modelling for the sub-catchment, is that the Hol tributary has the largest proportion of baseflow to runoff (50%), while the Bergvallei contributes the largest amount of baseflow due to the higher rainfall and larger area. The conductive nature of primary and TMG aquifers as well as the low conductance of the secondary aquifer were major factors in baseflow/runoff proportioning. While the Hol with a larger secondary aquifer contribution, albeit more saline, contributes a constant supply of baseflow that is required to maintain biodiversity levels in the lake.

6.1.2. *Catchment dynamics*

The coefficient of variability (CV) and baseflow index (BFI) are values that can be used to understand the flow variability of river systems. While CV values have been used to understand streamflow variability, BFI values are used to relate the proportion of baseflow to runoff (e.g. Eckhardt, 2008). In addition to CV and BFI values being derived for each tributary, the rainfall/runoff model inherently calculates interflow to runoff and primary to secondary aquifer baseflow. In the context of the Verlorenvlei, higher interflow to runoff ratios result in reduced CV values due to a larger flow attenuation from the tributary. This is evident by comparison of CV values between the Kruismans (217.20) and the Bergvallei (284.54). While the Kruismans is mainly comprised of runoff (0.37), interflow ratios are higher (13.90) than the Bergvallei (22.50), which accounted for the reduction in CV values. Similarly, higher secondary aquifer to primary aquifer flow ratios result in lower CV values, if the tributary is mainly comprised of shale as opposed to TMG sandstones. This is evident in comparison between CV values of the Krom Antonies (283.00) and the Hol (146.54), where the Hol is mainly driven by shale dominated secondary aquifer flow (0.17) and the Krom Antonies by the TMG secondary aquifer flow (0.33). While tributaries that are dominated by TMG formations have larger net recharge values, baseflow responses have less lag attenuation and therefore are more variable. The indexes and proportions used in this study provided valuable insight into the flow dynamics of the tributaries that feed the Verlorenvlei. This type of approach can be applied to other semi-arid environments provided that an in-depth understanding is available of the recharge dynamics and variation across the catchment.

6.1.3. *Improvement of hydrological modelling results for the sub-catchment*

In light of both the domestic and international significance of the Verlorenvlei, more effective water management protocols will be required in this part of the Olifants/doorn WMA in the future. It is likely that to accommodate and develop improved water management protocols, new gauging stations, improved climate and aquifer water level monitoring will need to be implemented. At the station (Het Kruis) that was used for model calibration in this study, the two-stage weir's DT limit (Discharge Table) was exceeded on 105 occasions between 1986-2006, with an average of five times a year. The DT limit reduced the confidence in the rainfall/runoff model's ability to reproduce high flow events using climate data. As the DT limit was only exceeded seldomly, there is good confidence in the model to reproduce streamflow for the Kruismans tributary. However, as the model was calibrated for the largest streamflow contributor (total contribution), the calibrated model parameters might not be representative of other tributary streamflow behavior. The Hol was a position of a previous station, G3H005 (1973-1981), but no data was available due to unreliable records or unrealistic DT limits. It would be beneficial for water management in the Verlorenvlei to have a gauging structure on the Hol to assess if the model estimates align to the actual behavior of this tributary, as the model found that this is the main contributor during low flow conditions.

In this study, an EMD filtering protocol was applied to the runoff data, which resulted in a water level change at the sub-catchment outlet, to identify whether calibrated parameters resemble the behavior of combined streamflow from all the tributaries. The EMD filtering protocol showed agreement between the measured and simulated water levels, although further improvements could still be made to the runoff lag and attenuation. Future gauging structures in this area should be sized using the flow exceedance probabilities determined using the modelling approach developed in this study. For future modelling endeavors, more climate measurements at high elevations, such as on the peaks of the Piketberg Mountain would improve regionalisation of precipitation. In addition, a more extensive groundwater monitoring network across multiple aquifer units would also improve the representativeness of aquifer properties, as well as allowing for determination of the impact of groundwater abstraction on aquifer baseflow.

6.1.4. *Water use competition*

The Sandveld is made up of multiple competing users that require water for either domestic or agricultural purposes. At present the competition within the region is contentious due to the limited availability of both surface and groundwater, as well as the need for economic development and poverty alleviation. While job creation is of interest to the 2020 sustainability development goals (Assembly, 2014), in water-stressed regions like the Sandveld there is limited opportunity for further agricultural expansion. This study highlighted the limited water availability within the sub-catchment, even though the Sandveld is still regarded as an important potato production and agricultural area. While the sub-catchment is important for agriculture, this study showed that the tributaries are particularly sensitive to climate due to high streamflow variability, as well as the limited ability of the aquifer to buffer climatic variations as groundwater stores do not seem to be particularly extensive. Climate change is likely to have a significant influence on rainfall into the future in the Sandveld, which will further reduce the water availability for agriculture. It is expected that the length and occurrence of dry cycles is likely to become more frequent, resulting in the Verlorenvlei drying up more often, which will raise concerns of the ability of the area to support agricultural requirements.

Continued expansion of agriculture could have severe impacts on water resources into the future, although due to data restrictions agricultural impacts where not incorporated into models used in this study. Instead the initial frameworks that had the potential to assess these impacts were developed, which can be used into the future when more data becomes available. The assessment of the impact that land use change has on water resources is particularly important in the Sandveld, as over the course of this study there has been significant changes in land use, with conversion of virgin land to cultivated land. As such the cultivation of virgin land should be restricted until these agricultural impacts can be fully understood. A similar approach has been applied in northern Kwa-Zulu Natal were certain areas have been delegated closed catchments due to their water producing ability and the overwhelming impact of land use on streamflow reduction (Dye and Jarmain, 2004; Smith and Scott, 1992; Van Wyk, 1987). There is need for further collaboration between scientists and farmers to provide more realistic estimates of the availability of water in the Sandveld, to ensure that water is not over allocated to agricultural requirements at the expense of the future survival of the Verlorenvlei.

6.2. Hydrological modelling in semi-arid environments

6.2.1. *Catchment boundaries*

The delineation of catchment or sub-catchment boundaries is an important component of hydrological modelling, where the catchment boundary is used to restrict flow within the model domain. While in some cases modelers have used influx/efflux variables to include/exclude either surface or groundwater flow from outside of the catchment, this approach requires a comprehensive understanding of the physical system to be able to quantify the inflows/outflows both temporally and spatially. While surface water variables have been used to constrain boundaries for many hydrological models, the incorporation of geological dip for groundwater flow remains an uncommon modelling practice. In this study, a test tributary (the Krom Antonies), was investigated to understand the possible impacts that different types of catchment boundaries could have on the water balance of a model constructed for the watershed. The tributary selected was presumed to be the most significant groundwater contributor in the sub-catchment, and therefore the boundary could have the largest bearing on the model results. Two different boundaries were generated for the tributary which included a boundary that was solely reliant on surface water flow directions and one that incorporated the geological dip direction of bedding planes. The study concluded that while the tributary had a number of different geological formations which could impact the catchment water balance, the dip was relatively shallow and therefore presumed negligible for this region. As geological dip remains a factor that could influence the catchment water balance, there is still a need to investigate aquifer hydraulic conductivity to assess this influence in other regions.

6.2.2. *Recharge in semi-arid environments*

Groundwater recharge is one of the most difficult processes to quantify and measure correctly, especially in semi-arid environments (Scanlon et al., 2006) like the Verlorenvlei sub-catchment. Recharge methods such as chloride mass balance, which has been applied across different climatic conditions, have proven ineffective in semi-arid environments. Similarly, yearly recharge estimates of groundwater recharge do not consider seasonal fluctuations. In semi-arid environments there are significant differences in climatic conditions between summer and winter (Sivapalan et al., 2005). In catchments with a Mediterranean climate, little to no recharge occurs during summer as evaporative demand generally exceeds rainfall. Rainfall also occurs less frequently resulting in low antecedent moisture conditions and rainfall volumes are generally not enough to satisfy soil-moisture demand. In contrast, during winter when rainfall is received and with evaporation demand lower, recharge potential is considerably higher and soil antecedent conditions are more favourable for recharge. Although, daily estimates of groundwater recharge are important for semi-arid environments, modelling approaches that do not also consider aquifer spatial variability, have the potential to over or under estimate recharge as shown in this study. Ultimately, when dealing with daily climatic variations, modelling approaches can over estimate recharge during the winter when rainfall is received. The recharge estimates in this study bridged a knowledge gap between previous low groundwater recharge estimates in semi-arid environments and their inherent high biodiversity levels. Although potential recharge methods that do not consider the spatial variation in aquifer hydraulic conductivity may over estimate recharge, this daily timestep is important in understanding groundwater recharge in semi-arid environments.

6.2.3. *Aquifer Baseflow Contribution*

Daily aquifer baseflow estimates, which respond to climatic fluctuations and seasonality are important for water management, as well as understanding catchment responses. These estimates can be used to restrict the over allocation of groundwater resources as well as help understand the hydrological components that allow the ecological reserve to be constrained. In this study, net recharge is used to determine average baseflow from the Krom Antonies sub-catchment. The average baseflow estimates suggest that there must be an additional larger source of baseflow from another tributary to meet the evaporation demand of the Verlorenvlei. This was confirmed by the rainfall/runoff model of the entire sub-catchment that identified the Bergvallei as the largest groundwater source and the Hol they most consistent. While the baseflow contribution was on average not enough to meet the evaporation demand of the Lake, the modelling suggested that the lake was mainly supplied by less frequent high flow events. Inherently, the streamflow from these tributaries is more important to protect and preserve as these tributaries are responsible for supplying the bulk of low flow volumes.

6.3. Improvements in Hydrological Modelling for Water Stressed Catchments

The lessons learned from the Verlorenvlei sub-catchment have direct parallels to other water stressed catchments, particularly ones in semi-arid environments. Worldwide municipal institutions are tasked with meeting the supply demands of urban centres, as well as accommodating the need for increased food production. While these needs are the main priority for water managers, the maintenance of natural environments is also an important consideration that needs to be included. To facilitate the needs of the human population, the environment and continued agricultural development, a multi-faceted and multi-disciplinary approach is required.

6.3.1. *The Importance of Monitoring Data*

The performance and ability of hydrological models to replicate catchment water balances is limited by the amount of data available, as well as the temporal and spatial resolution of the data. Climate conditions which vary both spatially and temporally and have a large bearing on results, are important to incorporate for effective rainfall/runoff modelling. As such modelers have applied regionalization to try to account for spatial variability between measuring points. While some climate data does not require regionalisation (e.g. air temperature, windspeed) as it is often uniform within a sub-catchment, rainfall which has both high spatial and temporal resolution should be regionalized within hydrological modelling systems. Rainfall regionalisation in semi-arid sub-Saharan Africa is hindered by station availability, where the availability of reliable climate/rainfall data is usually dependent on the proximity to metropolitan cities or adequate infrastructure. As many climate stations make use of cell phone reception for telemetry, station location is limited by line of sight to nearby cell phone towers. In the context of the Verlorenvlei sub-catchment, rainfall regionalisation was controlled by climate stations of nearby towns, with almost no records at high elevation. To improve the representativeness of rainfall spatial variability, farmers rainfall records were used to assist with regionalisation with one particular record being important due to its high elevation. Farmers records are thus a valuable source of information for hydrological models in regions that suffer from low climate station density. However, while records from farmers improve hydrological modelling in data scarce catchments, these records require more data scrutiny than measurements from well-maintained climate stations. Groundwater modelling in sub-Saharan Africa is also usually limited by the spatial resolution of adequate hydrogeological information, which include site specific aquifer hydraulic conductivity and storage. For many semi-arid catchments in sub-Saharan Africa the adequate documentation of geological information or access to geological information limits the implementation of groundwater modelling solutions. As such modelers have resorted to using bulk literature values of aquifer parameters, which usually encompass a large range of values and might not be representative of catchment conditions. The documentation of abstraction volumes not only influences the model's ability to simulate baseflow reductions, but also influences model calibration. While in this study, access to groundwater abstraction information was provided from farmers records, the bulk of farmers provided rough estimates that could be very different to actual groundwater abstraction volumes. In the context of recent drought conditions in the Western Cape, the monitoring and documentation of groundwater levels and abstraction volumes are essential for future groundwater management, especially as the competition for groundwater use increases. It is most likely that more expanded groundwater models will be required in other semi-arid catchments, where over abstraction is a concern and models will likely be faced with similar data limitations.

6.3.2. *Climate Change Impacts*

With climate change likely to have an impact on rainfall volumes across many semi-arid environments (Lema and Majule, 2009; Warburton et al., 2010), the length and occurrence of dry cycle conditions is likely to increase. In this study, the impact of changing rainfall on both streamflow and baseflow was assessed over a 30-year period for the Verlorenvlei sub-catchment. Over the 30 years, the region was subject to 10-year cycles of dry and wet conditions, marked by particularly low and high streamflow volumes. Of concern to the area is that the past 3 years (2015-2018) which resemble drought conditions, fall within the wet cycle between 2008-2018, highlighting that cycle length may have already altered and could further change in the future. It is expected that dry cycle conditions will occur more frequently and last for longer periods of time. The prolonged occurrence of low flow conditions will likely result in increased catchment discharge, due to low antecedent conditions resulting in more lag variability.

The hydrological models developed in this study were capable of reproducing streamflow using climate data for low flow conditions, although confidence in reproducing high flow events is less satisfactory. With historical climate records producing a unique calibration solution, future conditions might result in models not being able to replicate streamflow fluctuations. Consequently, many models might not be able to replicate streamflow fluctuations using climate records, with the potential that models will need to be regularly calibrated when data becomes available. In contrast to the concerns of increased periods of low rainfall, it is expected that climate change will also influence the occurrence of more extreme events such as floods and droughts, putting stress on city and town infrastructure.

6.3.3. *Integrating hydrological and hydrogeological modelling*

The coupling or integration of hydrological and hydrogeological models for semi-arid catchments has the potential to improve and make models more representative of catchment dynamics. The method applied in this study can be effectively transferred to other water-stressed catchments, where an understanding of recharge and baseflow as well as river regime dynamics is required. While a number of coupling approaches have been developed such as SWAT-MODFLOW (Kim et al., 2008), it is important to have a good understanding of model parameter ranges, especially in partial or ungauged catchments for effective implementation. In this study a J2000-MODFLOW coupling was applied as the Verlorenvlei is currently an ungauged catchment (although there is historical data), with parameter ranges been previously defined for an adjacent sub-catchment (Bugan, 2014). Unlike the SWAT-MODFLOW coupling, which is a hard couple (coded together) the J2000-MODFLOW coupling developed in this study relied on a transfer of components from one model to the other. A benefit to this is that the two modelling systems do not interfere with one another, as well as reducing the computational requirements. While the transfer of components means that modellers have to modify different data sets for both temporal and spatial resolutions (which could be automated), it allows for the inclusion of model advantages across both systems.

A further drawback to this is that further modification is required to the J2000 to allow for simulating aquifer water levels as well as groundwater abstraction, that will be essential for future modelling development.

6.4. Recommendations

In summary a few recommendations have been made regarding the future management of water resources in the sub-catchment, the contribution of this study as well as recommendations for future research in the area.

(A) Contribution of this study

- For high spatial and temporal resolutions of recharge, rainfall/runoff models can be used, although these models require distribution using aquifer properties
- To account for seasonal variability in estimates of baseflow, groundwater and surface water models can be coupled together, using the transfer of model parameters from one model to the other
- To understand river flow regime dynamics required for more effective management of water courses, indexes such as BFI and CV's can be used
- To understand baseflow attenuation and lag, the spatial distribution of the host geology of the aquifer needs to be included in hydrological modelling systems

(B) Future management of water in the Verlorenvlei

- Promote the reduced cultivation of virgin land until modelling systems are equipped to incorporate the impact that agricultural irrigation as well as land use change has on both groundwater and surface water availability in the sub-catchment
- Encourage farmers to record and monitor groundwater abstraction volumes as well as aquifer water levels, so that more information is available to construct better groundwater modelling systems for the region into the future
- Reduce the construction of dams as well as diversion mechanisms that could impact streamflow and baseflow required to maintain the water level in the Verlorenvlei
- Encourage the use of more efficient irrigation systems such as drip irrigation, as well as promoting the use of monitoring systems that help to determine optimum irrigation cycles and amounts

(C) Future research in Verlorenvlei

- Encouraged the construction of new groundwater models for each of the tributaries to understand the groundwater resource base that is available to supplement the growing need for food production
- Improve the rainfall and climate station density through the installation of new weather stations, particularly at higher elevations to improve the regionalisation of rainfall in hydrological modelling
- Construct new and repair existing gauging structures to improve water management as well as provide data that can be used to calibrate future models, which are able to assess the impact of land use and climate change on water resources in the Verlorenvlei

6.5. References

Assembly, U.N.G., 2014. Report of the open working group of the General Assembly on Sustainable Development Goals. Gen. Assem.

Doc. A/69/970, New York 12.

Bronstert, A., Niehoff, D., Brger, G., 2002. Effects of climate and land-use change on storm runoff generation: Present knowledge and modelling capabilities. Hydrol. Process. 16, 509–529.

Bugan, R.D.H., 2014. Modeling and Regulating Hydrosalinity Dynamics in the Sandspruit River Catchment (Western Cape). Stellenbosch University. Unpublished PhD thesis, 1-216

Dye, P., Jarmain, C., 2004. Water use by black wattle (Acacia mearnsii): implications for the link between removal of invading trees and catchment streamflow response: working for water. S. Afr. J. Sci. 100, 40–44.

Kim, N.W., Chung, I.M., Won, Y.S., Arnold, J.G., 2008. Development and application of the integrated SWAT-MODFLOW model. J. Hydrol. 356, 1–16.

Lema, M.A., Majule, A.E., 2009. Impacts of climate change, variability and adaptation strategies on agriculture in semi arid areas of Tanzania: The case of Manyoni District in Singida Region, Tanzania. African J. Environ. Sci. Technol. 3, 206–218.

Patil, S., Stieglitz, M., 2014. Modelling daily streamflow at ungauged catchments: What information is necessary? Hydrol. Process. 28, 1159–1169.

Scanlon, B.R., Keese, K.E., Flint, A.L., Flint, L.E., Gaye, C.B., Edmunds, W.M., Simmers, I., 2006. Global synthesis of groundwater recharge in semiarid and arid regions. Hydrol. Process. 20, 3335–3370.

Sivapalan, M., Blöschl, G., Merz, R., Gutknecht, D., 2005. Linking flood frequency to long-term water balance: Incorporating effects of seasonality. Water Resour. Res. 41.

Smith, R.E., Scott, D.F., 1992. The effects of afforestation on low flows in various regions of South Africa. Water S. A. 18, 185–194.

Van Wyk, D.B., 1987. Some effects of afforestation on streamflow in the Western Cape Province, South Africa. Water Sa 13, 31–36.

Warburton, M.L., Schulze, R.E., Jewitt, G.P.W., 2010. Confirmation of ACRU model results for applications in land use and climate change studies. Hydrol. Earth Syst. Sci. 14, 2399–2414.

Wessels, P., Rooseboom, A., 2009. Flow-gauging structures in South African rivers Part 2: Calibration. Water SA 35, 11–20.