Distributive rainfall/runoff modelling to understand runoff to baseflow proportioning and its impact on the determination of reserve requirements of the Verlorenvlei estuarine lake, west coast, South Africa

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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

25   has to be estimated or calculated through the incorporation of hydrogeological variables. In

26   this study, the groundwater components within the J2000 rainfall/runoff model were distributed

27   to provide daily baseflow and streamflow estimates needed for reserve determination. The

28   modelling approach was applied to the RAMSAR-listed Verlorenvlei estuarine lake system on

29   the west coast of South Africa which is under threat due to agricultural expansion and climatic

30   fluctuations. The sub-catchment consists of four main tributaries, Krom Antonies, Hol,

31   Bergvallei and Kruismans. Of these, Krom Antonies was initially presumed the largest

32   baseflow contributor, but was shown to have significant streamflow variability, attributed to

33   the highly conductive nature of the Table Mountain Group sandstones and quaternary

34   sediments. Instead, Bergvallei was identified as the major contributor of baseflow. Hol was the

35   least susceptible to streamflow fluctuations due to the higher baseflow proportion (56 %), as

36   well as the dominance of less conductive Malmesbury shales that underlie it. The estimated

37   flow exceedance probabilities indicated that during the 2008-2017 wet cycle average lake

38   inflows exceeded the average evaporation demand, although yearly rainfall is twice as variable

39   in comparison to the first wet cycle between 1987-1996. During the 1997-2007 dry cycle,

40   average lake inflows are exceeded 85 % of the time by the evaporation demand. The

41   exceedance probabilities estimated here suggest that inflows from the four main tributaries are

42   not enough to support Verlorenvlei, with the evaporation demand of the entire lake being met

43   only 35 % of the time. This highlights the importance of low occurrence events for filling up

44   Verlorenvlei, allowing for regeneration of lake-supported ecosystems. As climate change

45   drives increased temperatures and rainfall variability, the length of dry cycles is likely to

46   increase into the future and result in the lake drying up more frequently. For this reason, it is

47   important to ensure that water resources are not overallocated during wet cycles, hindering

48   ecosystem regeneration and prolonging the length of these dry cycle conditions.

1. Introduction

Functioning river systems offer numerous economic and social benefits to society including water supply, nutrient cycling and disturbance regulation amongst others (Costanza et al., 1997; Nelson et al., 2009; Postel and Carpenter, 1997). As a result, many countries worldwide have endeavoured to protect river ecosystems, although only after provision has been made for basic human needs (Gleick, 2003; Richter et al., 2012; Ridoutt and Pfister, 2010). However, the implementation of river protection has been problematic, because many river courses and flow regimes have been severely altered due to socio-economic development (Gleeson and Richter, 2018; O'Keeffe, 2009; Richter, 2010). River health problems were thought to only result from low-flow conditions and if minimum flows were kept above a critical level, the river's ecosystem would be protected (Poff et al., 1997; Tennant, 1976). 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 (Arthington et al., 2018; Bunn and Arthington, 2002; Olden and Naiman, 2010; 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 hydrological components of what is called the ecological reserve, the minimum environmental conditions needed to maintain the ecological health of a river system (Hughes, 2001; King and Louw, 1998; Richter et al., 2003). A variety of different methods have been developed to incorporate various river health factors into ecological reserve determination (Acreman and Dunbar, 2004; Bragg et al., 2005). One of the simplest and most widely applied, is where compensation flows are set below reservoirs and weirs, using flow duration curves to

74 derive mean flow or flow exceedance probabilities (e.g. Harman and Stewardson, 2005). This

75 approach focusses purely on hydrological indices, which are rarely ecologically valid (e.g.

76 Barker and Kirmond, 1998; Lancaster and Downes, 2010).

77 More comprehensive ecological reserve estimates such as functional analysis are focused on

78 the whole ecosystem, including both hydraulic and ecological data (e.g. ELOHA: Poff et al.,

79 2010; Building Block Methodology: King and Louw, 1998). While these methods consider that

80 a variety of low, medium and high flow events are important for maintaining ecosystem

81 diversity, they require specific data regarding the hydrology and ecology of a river system,

82 which in many cases does not exist, has not been recorded continuously or for sufficient

83 duration (Acreman and Dunbar, 2004; Richter et al., 2012). To speed up ecological reserve

84 determination, river flow records have been used to analyse natural seasonality and variability

85 of flows (e.g. Hughes and Hannart, 2003). However, this approach requires long-term

86 streamflow and baseflow timeseries. Whilst streamflow is a measurable quantity subject to a

87 gauging station being in place, baseflow has to be modelled based on hydrological and

88 hydrogeological variables.

89 Rainfall/runoff models can be used to calculate hydrological variables using distributive

90 surface water components (e.g. J2000: Krause, 2001; SWAT: Arnold et al., 1998) but the

91 groundwater components are generally lumped within conventional modelling frameworks. In

92 contrast, groundwater models, which distribute groundwater variables (e.g. MODFLOW:

93 Harbaugh et al., 2000; FEFLOW: Diersch, 2002) are frequently set up to lump climate

94 components. In order to accurately model daily baseflow, which is needed for reserve

95 determination, modelling systems need to be setup such that both groundwater and climate

96 variables are treated in a distributed manner (e.g Bauer et al., 2006; Kim et al., 2008).

97 Rainfall/runoff models, which use Hydrological Response Units (HRUs) as an entity of

98 homogenous climate, rainfall, soil and landuse properties (Flügel, 1995; Leavesley and

99   Stannard, 1990), are able to reproduce hydrographs through model calibration (Wagener and

100 Wheater, 2006; Young, 2006). However, they are rarely able to correctly proportion runoff and

101 baseflow components (e.g. Willems, 2009; Hughes, 2004). To correctly determine

102 groundwater baseflow using rainfall/runoff models such as the J2000, aquifer components need

103 to be distributed. This can be achieved using net recharge and hydraulic conductivity collected

104 through aquifer testing or groundwater modelling.

105 To better understand river flow variability, a rainfall/runoff model was distributed to

106 incorporate aquifer hydraulic conductivity within model HRUs using calibrated values from a

107 MODFLOW groundwater model (Watson, 2018). The model was setup for the RAMSAR

108 listed Verlorenvlei estuarine lake on the west coast of South Africa, which is under threat from

109 climate change, agricultural expansion and mining exploration. The rainfall/runoff model used

110 was J2000 as this model had previously been set up in the region and model variables were

111 well established (e.g Bugan, 2014; Schulz et al., 2013). While the estuarine lake's importance

112 is well documented (Martens et al., 1996; Wishart, 2000), the lake's reserve is not well

113 understood, due to the lack of streamflow and baseflow estimates for the main feeding

114 tributaries of the system. The modelling framework developed in this study aimed to

115 understand the flow variability of the lake's feeding tributaries, to provide the hydrological

116 components (baseflow and runoff proportioning) of the tributaries needed to understand the

117 lake reserve. The surface water and groundwater components of the model were calibrated for

118 two different tributaries which were believed to be the main source of runoff and baseflow for

119 the sub-catchment. The baseflow and runoff rates calculated from the model indicate not only

120 that the lake system cannot be sustained by baseflow during low flow periods but also that the

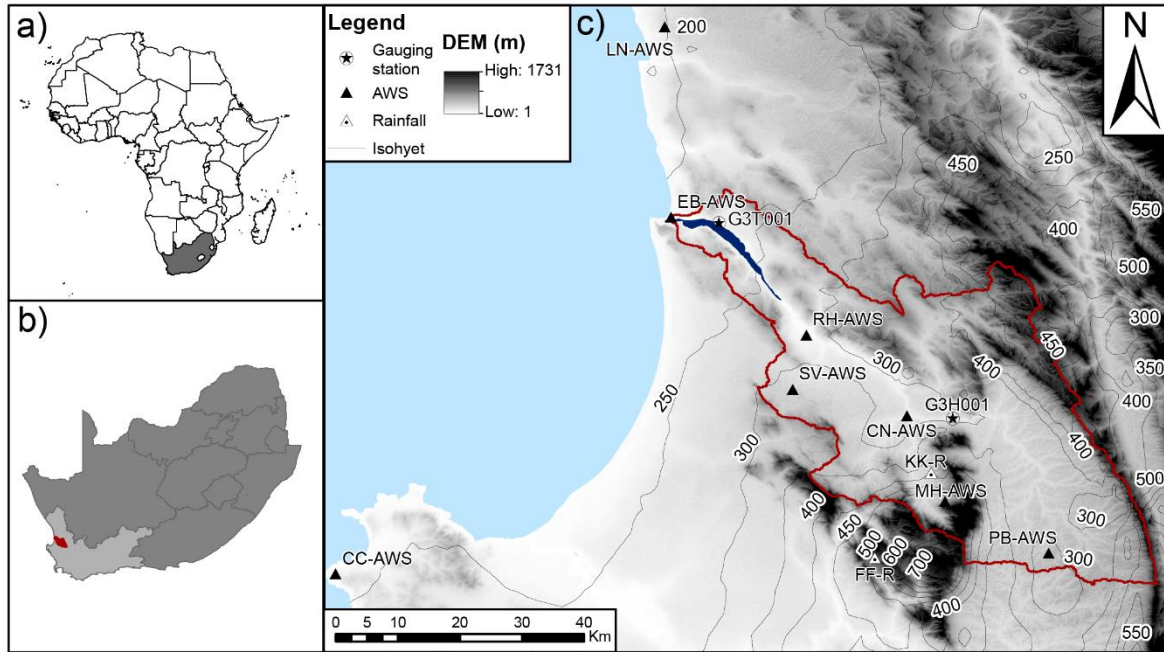
121 initial understanding of which tributaries are key to the sustainability of the lake system was

122 not correct. The results have important implications for how we understand water dynamics in

123 water stressed catchments and the sustainability of ecological systems in these types of
124 environments generally.

2. Study site

126 Verlorenvlei is an estuarine lake situated on the west coast of South Africa, approximately 150
127 km north of the metropolitan city of Cape Town (Fig. 1). The west coast, which is situated in
128 the Western Cape Province of South Africa, is subject to a Mediterranean climate where the
129 majority of rainfall is received between May to September. The Verlorenvlei lake, which is
130 approximately 15 km^2 in size draining a watershed of 1832 km^2, forms the southern sub-
131 catchment of the Olifants/Doorn water management area (WMA). The lake hosts both Karroid
132 and Fynbos biomes, with a variety of vegetation types (e.g Arid Estuarine Saltmarsh, Cape
133 Inland Salt pans) sensitive to reduced inflows of freshwater (Helme, 2007). A sandbar created
134 around a sandstone outcrop (Table Mountain Group; TMG) allows for an intermittent
135 connection between salt and fresh water. During storms or extremely high tides, water scours
136 the sand bar allowing for a tidal exchange, with a constant inflow of salt water continuing until
137 the inflow velocity decreases enough for a new sand bar to form (Sinclair et al., 1986).

138 The lake is supplied by four main tributaries which are Krom Antonies, Bergvallei, Hol and
139 Kruismans (Fig. 2). The main freshwater sources are presumed to be Krom Antonies and
140 Bergvallei, which drain the mountainous regions to the south (Piketberg) and north of the sub-
141 catchment respectfully. Hol and Kruismans tributaries are variably saline (Sigidi, 2018), due
142 to high evaporation rates in the valley. Average daily temperatures during summer within the
143 sub-catchment are between 20-30 °C, with estimated potential evaporation rates of 4 to 6 mm.d-
144 ^{1} (Muche et al., 2018). In comparison, winter daily average temperatures are between 12-20 °C,
145 with estimated potential evaporation rates of 1 to 3 mm.d^{-1} (Muche et al., 2018).

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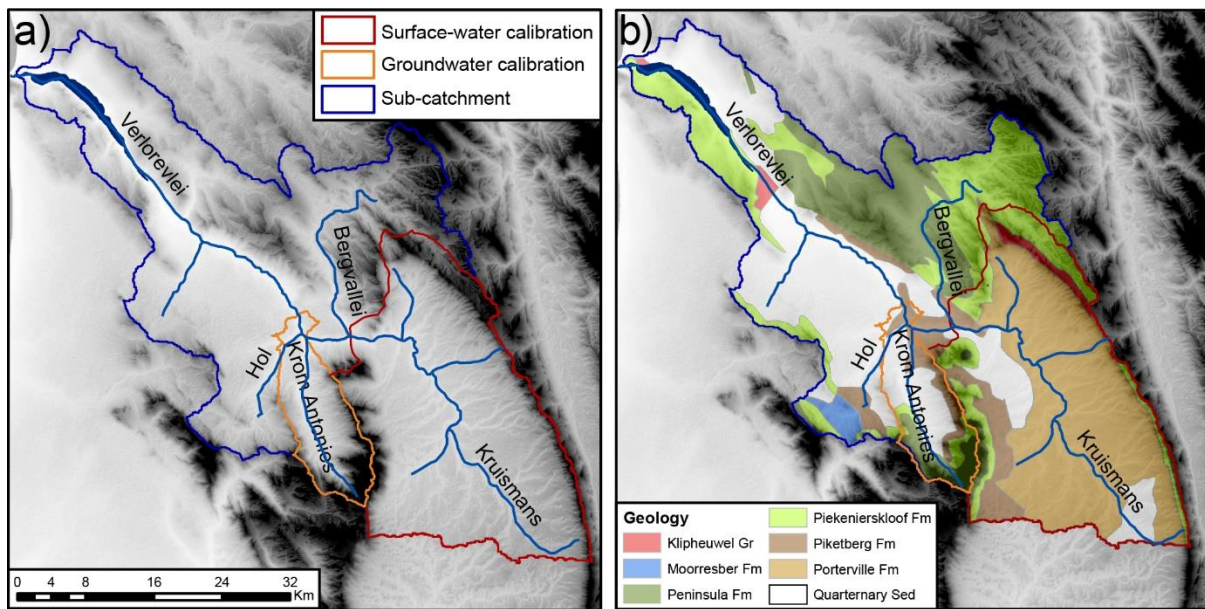
147   Figure 1: a) Location of South Africa, b) the location of the study catchment within the Western

148   Cape and c) the extend of the Verlorenvlei sub-catchment with the climate stations, gauging

149   station (G3H001), measured lake water level (G3T001) and rainfall isohyets

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151   Figure 2: a) The Verlorenvlei sub-catchment with the surface water calibration tributary

152   (Kruismans) and groundwater calibration tributary (Krom Antonies) and b) the hydrogeology

153   of the sub-catchment with Malmesbury shale formations (MG; Klipheuwel, Mooresberg,

154   Porterville, Piketberg), Table Mountain Group formations (Peninsula, Piekenierskloof) and

155   quaternary sediments

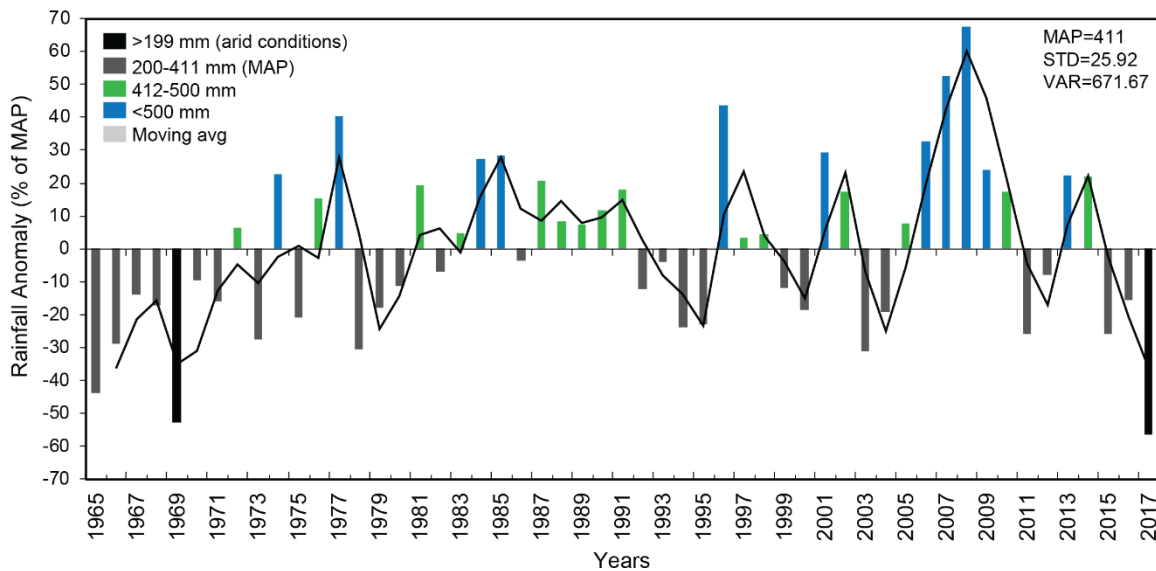
156   Rainfall for the sub-catchment, recorded over the past 52 years by local farmers at KK-R (Fig.

157   1) shows large yearly variability (26 %) between the Mean Annual Precipitation (MAP) (411

158   mm) and measured rainfall (Fig. 3). Where rainfall was greater than 500 mm.yr^{-1} (2006-2010),

159   it is presumed that the lake is supported by a constant influx of streamflow from the feeding

160   tributaries. Recently, where rainfall was less than 50 % of the MAP (2015-2017), concerns

161   over the amount of streamflow required to support the lake have been raised.

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163   Figure 3: The difference between MAP and measured rainfall (plotted as rainfall anomaly) for

164   52 years (1965-2017) at location KK-R in the valley of Krom Antonies (after Watson *et al.,*

165   2018).

166   While rainfall varies greatly between years in the sub-catchment, it is also spatially impacted

167   by elevational differences. The catchment valley which receives the least MAP 100-350 mm.yr^{-}

168   ^{1} (Lynch, 2004), is between 0-350 masl and is comprised of quaternary sediments that vary in

169   texture, although the majority of the sediments in the sub-catchment are sandy in nature. The

170 higher relief mountainous regions of the sub-catchment between 400-1300 masl receive the

171 highest MAP 400-800 $mm.yr^{-1}$ (Lynch, 2004), are mainly comprised of fractured TMG

172 sandstones, (youngest to oldest): Peninsula, Graafwater (not shown), and Piekernerskloof

173 formations (Fig. 2) (Johnson et al., 2006). Underlying the sandstones and quaternary sediments

174 are the MG shales, which are comprised of the Mooresberg, Piketberg and Klipheuwel

175 formations (Fig. 2) (Rozendaal and Gresse, 1994). Agriculture is the dominant water user in

176 the sub-catchment with an estimated usage of 20 % of the total recharge (Conrad et al., 2004;

177 DWAF, 2003), with the main food crop being potatoes. The MG shales and quaternary

178 sediments, which host the secondary and primary aquifer respectfully, are frequently used to

179 supplement irrigation during the summer months of the year. During winter, the majority of

180 the irrigation water needed for crop growth is supplied by the sub-catchment tributaries or the

181 lake itself. The impact of irrigation on the lake is still regarded as minimal (Meinhardt et al.,

182 2018) but further investigation is still required. For additional information regarding the study

183 site refer to Watson *et al.*, (2018) and Conrad *et al.*, (2004).

184 **3. Methodology**

185 In this study, the J2000 coding was adapted to incorporate distributed groundwater components

186 for the model HRU's (Fig. 4). This was done by aligning the MODFLOW recharge estimates

187 and previous studies (Conrad et al., 2004; Miller et al., 2017; Vetger, 1995; Weaver and Talma,

188 2005; Wu, 2005) with those of the J2000, through adjustment of aquifer hydraulic conductivity

189 from the MODFLOW groundwater model of Krom Antonies (Watson, 2018) (Fig. 5). The

190 assigned hydraulic conductivity for each geological formation was thereafter transferred across

191 the entire J2000 model of the sub-catchment. The adaption applied to the groundwater

192 components influenced the proportioning of water routed to runoff and baseflow within the

193 J2000 model. To validate the outputs of the model, an empirical mode decomposition (EMD)

194   (Huang et al., 1998) was applied to compute the proportion of variation in discharge timeseries

195   that attributed to a high and low water level change at the sub-catchment outlet. The streamflow

196   estimates were thereafter compared with the lake evaporation demand, to understand the sub-

197   catchment water balance.

198   The J2000 model incorporated distributed climate, soil, landuse and hydrogeological

199   information, with aquifer hydraulic conductivity transferred from MODFLOW as described

200   above (Fig. 4). The measured streamflow was used to both calibrate and validate the model,

201   with the landuse dataset being selected according to the period of measured streamflow.

202   Changes in the recorded lake level were used alongside remote sensing to estimate the lake

203   evaporation rate. The impact of irrigation was not included in the model, as there is not enough

204   information available regarding agricultural water use. This is currently one of the major

205   limitations with the study approach presented here and will be the focus of future work. The

206   HRU delineation, model regionalisation, water balance calculations, lateral and reach routing

207   as well as the lake evaporation procedure are presented. Thereafter the input data for the model,

208   the calibration and validation procedures as well as the EMD protocol used, is described.

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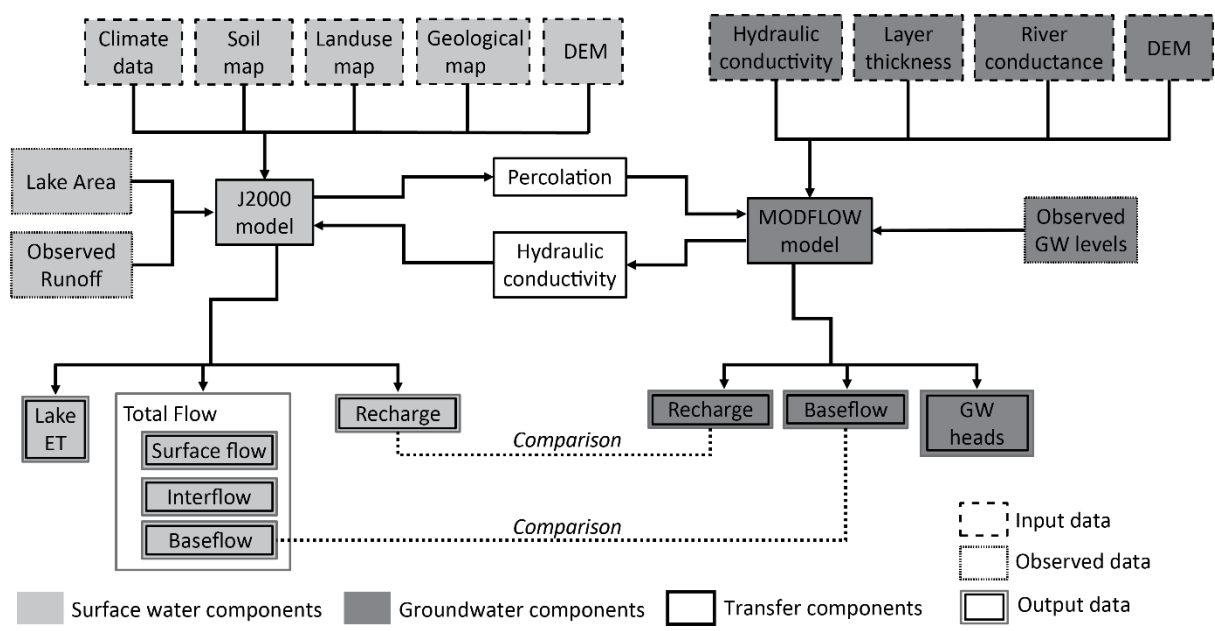


Figure 4: Schematic of the model structure, showing the processors simulated by the J2000 and MODFLOW and the components that were transferred from the MODFLOW model

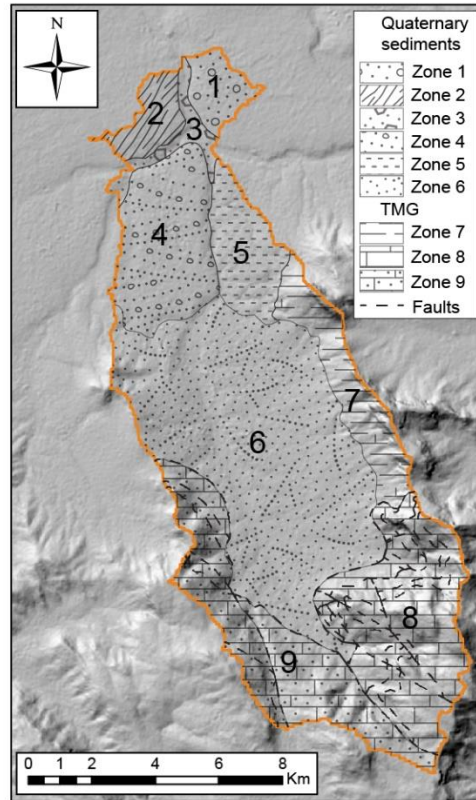



Figure 5: The aquifer hydraulic zones used for the groundwater calibration of the J2000 (after Watson, 2018)

3.1 Hydrological Response Unit Delineation

HRUs and stream segments (reaches) are used within the J2000 model for distributed 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/km^2 being defined

223   (Pfannschmidt, 2008). After the delineation of HRUs, dominant soil, land use and geology

224   properties were assigned to each. The hydrological topology was defined for each HRU by

225   identifying the adjacent HRUs or stream segments that received water fluxes.

3.2 Model regionalisation

227   Rainfall and relative humidity are the two main parameters that are regionalised within the

228   J2000 model. While a direct regionalisation using an inverse-distance method (IDW) and the

229   elevation of each HRU can be applied to rainfall data, the regionalisation of relative humidity

230   requires the calculation of absolute humidity. The regionalisation of rainfall records was

231   applied by defining the number of weather station records available and estimating the

232   influence on the rainfall amount for each HRU. A weighting for each station using the distance

233   of each station to the area of interest was applied to each rainfall record, using an elevation

234   correction factor (Krause, 2001). The relative humidity and air temperature measured at set

235   weather stations was used to calculate the absolute humidity. Absolute humidity was thereafter

236   regionalised using the IDW method, station and HRU elevation. After the regionalisation had

237   been applied, the absolute humidity was converted back to relative humidity through

238   calculation of saturated vapor pressure and the maximum humidity.

3.3 Water balance calculations

240   The J2000 model is divided into calculations that impact surface water and groundwater

241   processors. The J2000 model distributes the regionalised precipitation (P) calculated for each

242   HRU using a water balance defined as:

$$P = R + Int_{max} + ETR + \Delta Soil_{sat} \tag{1}$$

243   where R is runoff (mm) (RD1 - surface runoff; RD2 - interflow), Int_{max} is vegetation canopy

244   interception (mm), ETR is 'real' evapotranspiration and $\Delta Soil_{sat}$ is change in soil saturation.

245   The surface water processes have an impact on the amount of modelled runoff and interflow,

246 while the groundwater processors influence the upper and lower groundwater flow

247 components.

248 *3.3.1 Surface water components*

249 Potential evaporation (ETP) within the J2000 model is calculated using the Penman Monteith

250 equation. Before evaporation was calculated for each HRU, interception was subtracted from

251 precipitation using the leaf area index and leaf storage capacity for vegetation (a_rain)

252 (Supplementary: Table 1). Evaporation within the model considers several variables that

253 influence the overall modelled evaporation. Firstly, evaporation is influenced by a slope factor,

254 which was used to reduce ETP based on a linear function. Secondly, the model assumed that

255 vegetation transpires until a particular soil moisture content where ETP is reached, after which

256 modelled evaporation was reduced proportionally to the ETP, until it becomes zero at the

257 permanent wilting point.

258 The soil module in the J2000 model is divided up into processing and storage units. Processing

259 units in the soil module include soil-water infiltration and evapotranspiration, while storage

260 units include middle pore storage (MPS), large pore storage (LPS) and depression storage. The

261 infiltrated precipitation was calculated using the relative saturation of the soil, and its maximum

262 infiltration rate (SoilMaxInfSummer and SoilMaxInfWinter) (Supplementary: Table 1).

263 Surface runoff was generated when the maximum infiltration threshold was exceeded. The

264 amount of water leaving LPS, which can contribute to recharge, was dependant on soil

265 saturation and the filling of LPS via infiltrated precipitation. Net recharge (R_{net}) was estimated

266 using the hydraulic conductivity ($SoilMaxPerc$), the outflow from LPS (LPS_{out}) and the slope

267 ($slope$) of the HRU according to:

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

268   The hydraulic conductivity, $SoilMaxPerc$ and the adjusted LPS_{out}.were thereafter used to

269   calculate interflow (IT_f) according to:

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

270   with the interflow calculated representing the sub-surface runoff component RD2 and is routed

271   as runoff within the model.

272   **_3.3.2 Groundwater components_**

273   The J2000 model for the Verlorenvlei sub-catchment was set up with two different geological

274   reservoirs: (1) the primary aquifer (upper groundwater reservoir - RG1), which consists of

275   quaternary sediments with a high permeability; and (2) the secondary aquifer (lower

276   groundwater reservoir- RG2), made up of MG shales and TMG sandstones (Table 1).

Aquifer	Formation	Type	RG1_max (mm)	RG2_max (mm)	RG1_k (d)	RG2_k (d)	RG1_active (n/a)	Kf_geo (mm/d)	depthRG1 (cm)
Primary	Quarternary Sediments	Sediments	50	700	100	431	1	500	1750
Secondary/MG	Moorresberg 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

277

278   Table 1: The J2000 hydrogeological parameters RG1_max, RG2_max, RG1_k, RG2_Kf_geo

279   and depthRG1 assigned to the primary and secondary aquifer formations for the Verlorenvlei

280   sub-catchment

281   The model therefore considered two baseflow components, a fast one from RG1 and a slower

282   one from RG2. The filling of the groundwater reservoirs was done by net recharge, with

283   emptying of the reservoirs possible by lateral subterranean runoff as well as capillary action in

284   the unsaturated zone. Each groundwater reservoir was parameterised separately using the

285   maximum storage capacity (maxRG1 and maxRG2) and the retention coefficients for each

286   reservoir ($recRG1$ and $recRG2$). The outflow from the reservoirs was determined as a function

287   of the actual filling ($actRG1$ and $actRG2$) of the reservoirs and a linear drain function.

288 Calibration parameters $recRG1$ and $recRG2$ are storage residence time parameters. The

289 outflow from each reservoir was defined as:

$$OutRG1 = \frac{1}{gwRG1Fact \times recRG1} \times actRG1 \tag{4}$$

$$OutRG2 = \frac{1}{gwRG2Fact \times recRG2} \times actRG2 \tag{5}$$

290 where $OutRG1$ is the outflow from the upper reservoir, $OutRG2$ is the outflow from the lower

291 reservoir and $gwRG1Fact$/ $gwRG2Fact$ are calibration parameters for the upper and lower

292 reservoir used to determine the outflow from each reservoir. To allocate the quantity of net

293 recharge between the upper (RG1) and lower (RG2) groundwater reservoirs, a calibration

294 coefficient $gwRG1RG2sdist$ was used to distribute the net recharge for each HRU using the

295 HRU slope. The influx of groundwater into the shallow reservoir ($inRG1$) was defined as:

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

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

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

297 with the combination of $OutRG1$ and $OutRG2$ representing the baseflow component that is

298 routed as an outflow from the model.

3.4 Lateral and reach routing

300 Lateral routing was responsible for water transfer within the model and included HRU influxes

301 and discharge through routing of cascading HRUs from the upper catchment to the exit stream.

302 HRUs were either able to drain into multiple receiving HRUs or into reach segments, where

303 the topographic ID within the HRU dataset determined the drain order. The reach routing

304 module was used to determine the flow within the channels of the river using the kinematic

305 wave equation and calculations of flow according to Manning and Strickler. The river

306   discharge was determined using the roughness coefficient of the stream (Manning roughness),

307   the slope and width of the river channel and calculations of flow velocity and hydraulic radius

308   calculated during model simulations.

309   **3.5 Calculations of lake evaporation rate**

310   The lake evaporation rate was based on the ETP calculated by the J2000 and an estimated lake

311   surface area. The lake was modelled as a unique HRU (water as the land-cover type), with a

312   variable area which was estimated using remote sensing data from Landsat 8 and Sentinel-2

313   and the measured lake water level at G3T001 (Fig. 1). To infill lake surface area when remote

314   sensing data was not available, a relationship was created between the estimated lake's surface

315   area and the measured water level between 2015-2017. Where lake water level data was not

316   available (before 1999), an average long-term monthly value was used for the lake evaporation

317   calculations.

318   **3.6 J2000 Input data**

319   *3.6.1 Surface water parameters*

320   Climate and rainfall: Rainfall, windspeed, relative humidity, solar radiation and air temperature

321   were monitored by Automated Weather Stations (AWS) within and outside of the study

322   catchment (Fig. 1). Of the climate and rainfall data used during the surface water modelling

323   (Watson et al., 2018), data was sourced from seven AWS's of which four stations were owned

324   by the South African Weather Service (SAWS) and three by the Agricultural Research Council

325   (ARC). Two stations that were installed for the surface water modelling, namely Moutonshoek

326   (M-AWS) and Confluence (CN-AWS) were used for climate and rainfall validation due to their

327   short record length. Additional rainfall data collected by farmers at high elevation at location

328   FF-R and within the middle of the catchment at KK-R were used to improve the climate and

329   rainfall network density.

330 Landuse classification: The vegetation and landuse dataset that was used for the sub-catchment

331 (CSIR, 2009) included five different landuse classes: 1) wetlands and waterbodies, 2)

332 cultivated (temporary, commercial, dryland), 3) shrubland and low fynbos, 4) thicket,

333 bushveld, bush clumps and high fynbos and 5) cultivated (permanent, commercial, irrigated).

334 Each different landuse class was assigned an albedo, root depth and seal grade value based on

335 previous studies (Steudel et al., 2015)(Supplementary: Table 2). The Leaf Area Index (LAI)

336 and vegetation height varies by growing season with different values of each for the particular

337 growing season. While surface resistance of the landuse varied monthly within the model, the

338 values only vary significantly between growing seasons.

339 Soil dataset: The Harmonized World Soil Database (HWSD) v1.2 (Batjes et al., 2012) was the

340 input soil dataset, with nine different soil forms within the sub-catchment (Supplementary:

341 Table 3). Within the HWSD, soil depth, soil texture and granulometry were used to calculate

342 and assign soil parameters within the J2000 model. MPS and LPS which differ in terms of the

343 soil structure and pore size were determined in Watson et al. (2018), using pedotransfer

344 functions within the HYDRUS model (Supplementary: Table 3).

345 Streamflow and water levels: Streamflow, measured at the Department of Water Affairs

346 (DWA) gauging station G3H001 between 1970-2009, at the outlet of Kruismans tributary (Het

347 Kruis) (Fig 1 and 3), was used for surface water calibration. The G3H001 two-stage weir could

348 record a maximum flow rate of 3.68 $m^2.s^{-1}$ due to the capacity limitations of the structure. After

349 2009, the G3H001 structure was decommissioned due to structural damage, although repairs

350 are expected in the near future due to increasing concerns regarding the influx of freshwater

351 into the lake. Water levels measured at the sub-catchment outlet at DWA station G3T001 (Fig

352 1) between 1994 to 2018 were used for EMD filtering.

3.6.2 Groundwater parameters

Net recharge and hydraulic conductivity: The hydraulic conductivity values used for the groundwater component adaptation were collected from detailed MODFLOW modelling of Krom Antonies tributary (Fig. 5) (Watson, 2018). The net recharge and aquifer hydraulic conductivity for 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 total number of voids and cavities, 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 Krom Antonies sub-catchment (Watson, 2018). 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).

3.7 J2000 model calibration

3.7.1 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

377   conducted by Watson *et al*., (2018) showed that for high flow conditions (E2) (Nash-Sutcliffe

378   efficiency in its standard squared), model outputs are most sensitive to the slope factor, while

379   for low flow conditions (E1) (modified Nash-Sutcliffe efficiency in a linear form) the model

380   outputs were most sensitive to the maximum infiltration rate of the soil (ie. the parameter

381   maxInfiltrationWet) (Supplementary: Figure 1). The max percolation was moderately sensitive

382   during wet and dry conditions, and together with the slope factor, controlled the interflow to

383   percolation proportioning that was calibrated in this study.

384   ***3.7.2 Surface water calibration***

385   The surface water parameters of the model were calibrated for Kruismans tributary (688 km^2)

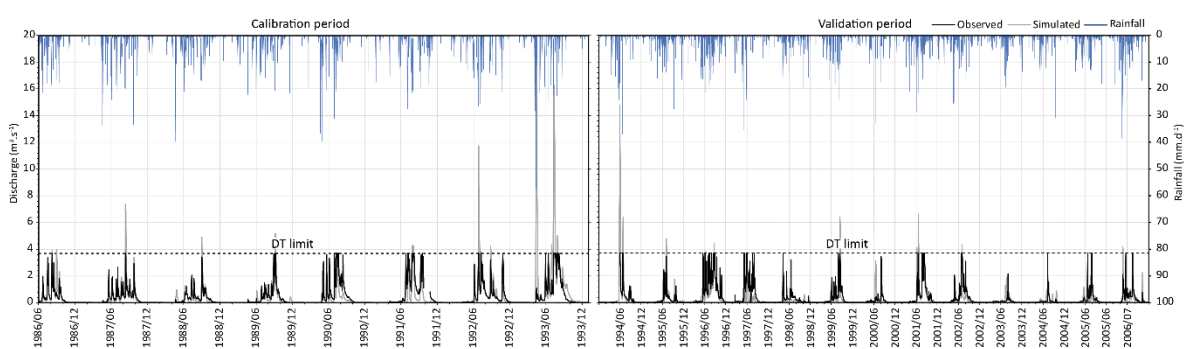
386   (Fig. 3) using the gauging data from G3H001 (Fig. 6 and Table 1). The streamflow data used

387   for the calibration was between 1986-1993, with model validation between 1994 to 2007 (Fig.

388   6). This specific calibration period was selected due to the wide range of different runoff

389   conditions experienced at the station, with both low and high flow events being recorded. For

390   the calibration, the modelled discharge was manipulated in the same fashion, with a DT limit

391   (discharge table) of 3.68 m³/s, so that the tributary streamflow behaved as measured discharge.

392

393   Figure 6: The surface water calibration (1986-1993) and validation (1994-2006) of the J2000

394   model using gauging data from the G3H001

395 An automated model calibration was performed using the "Nondominating Sorting Genetic

396 Algorithm II" (NSGA-II) multi-objective optimisation method (Deb et al., 2002) with 1024

397 model runs being performed. Narrow ranges of calibration parameters (FC_Adaptation,

398 AC_Adaptation, soilMAXDPS, gwRG1Fact and gwRG2Fact) were chosen to (1) ensure that

399 the modelled recharge from J2000 was within an order of magnitude of recharge from the

400 MODFLOW model and previous studies; (2) achieve a representative sub-catchment

401 hydrograph. As objective functions, Nash-Sutcliffe-Efficiency based on absolute differences

402 (E1) and squared differences 2 (E2) as well as the average bias in % (Pbias) were utilized for

403 the calibration (Krause et al., 2005) (Table 2). The choice of the optimized parameter set was

404 made to ensure that E2 was better than 0.57 (best value was 0.57) and the Pbias better than 5

405 % (Table 1). From the automated calibration, 308 parameter sets were determined with the best

406 E1 being chosen to ensure that the model is representative of low flow conditions (Table 1).

407 *3.7.3 Model validation*

408 Observed vs modelled streamflow: For the surface water model validation, the streamflow

409 records between 1994-2007 were used, where Nash-Sutcliffe-Efficiency (E1 and E2) were

410 reported. The Pbias was also used as an objective function to report the model performance by

411 comparison between measured and modelled streamflow (Table 2). Although gauging station

412 limitations resulted in good objective functions from the model, the performance of objective

413 functions E1, E2, Pbias reduced between the validation and calibration period (Table 2). During

414 the calibration period there was a good fit between modelled and measured streamflow (Pbias=-

415 1.82), with a significant difference between modelled and measured streamflow during the

416 validation period (Pbias=-19.2). The calibration was performed over a wet cycle (1986-1997),

417 which resulted in a more common occurrence of streamflow events that exceeded 3.68 $m^3.s^{-1}$,

418 thereby reducing the number of calibration points. In contrast the validation was performed

419 over a dry cycle (1997-2007), which resulted in more data points as few streamflow events

420 exceeded 3.68 $m^3.s^{-1}$.

	Calibration 1987-1993	Validation 1994-2007
E1	0.55	0.53
E2	0.57	0.56
LogE1	0.28	0.10
LogE2	0.46	0.19
AVE	-19.24	-269.20
R^2	0.62	0.58
Pbias	-1.82	-19.23758
KGE	0.79	0.67417

421

422 Table 2: Value of the objective functions E1, E2, logarithmic versions of E1 and E2, absolute

423 volume error (AVE), coefficient of determination (R^2), Pbias and Kling Gupta Efficiency

424 (KGE) (Gupta et al., 2009) for surface water calibration (1987-1993) and validation (1994-

425 2007)

426 The J2000 and MODFLOW recharge estimates: With adjustment of hydraulic conductivities

427 from MODFLOW to J2000 it was possible to converge the net recharge estimates between 1.3

428 % with a range of recharge of 0.65-10.03 % for the J2000 and 0.3-11.40 % for MODFLOW.

429 Recharge estimates from previous studies of the primary aquifer indicate recharge rates of 0.2-

430 3.4 % (Conrad et al., 2004), and 8 % (Vetger, 1995), while for the TMG aquifer 13 % (Wu,

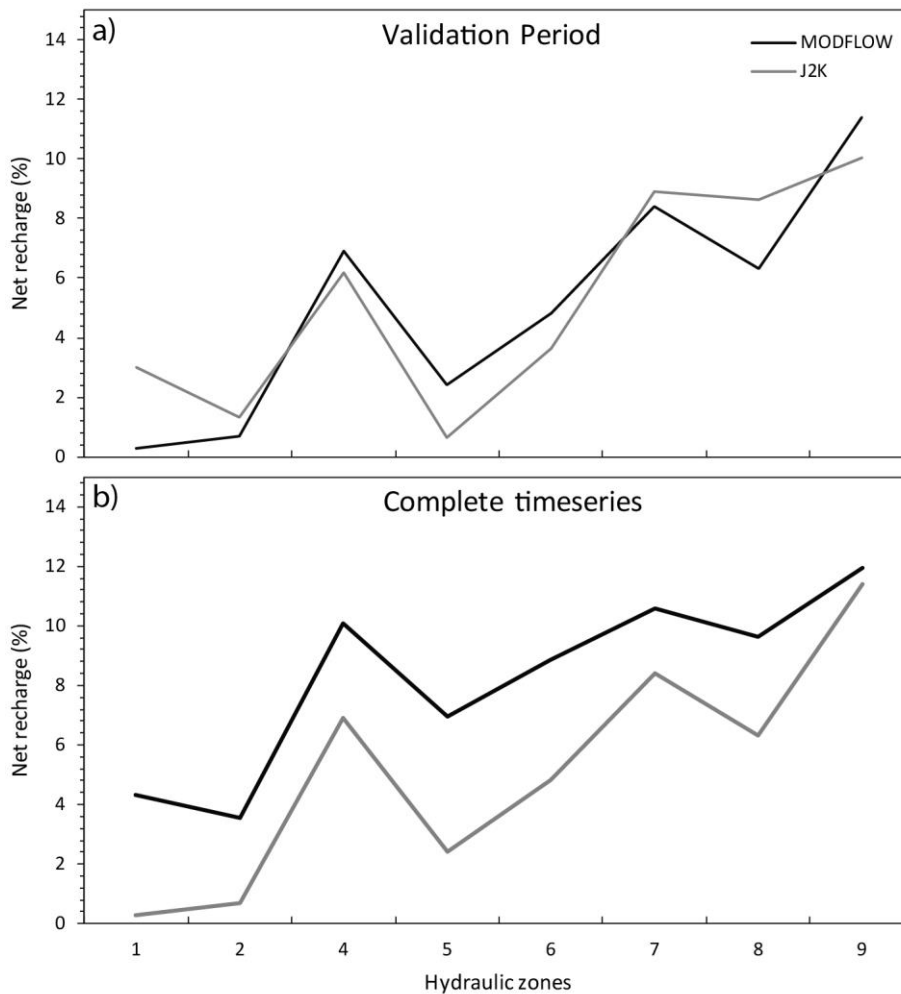
431 2005), 27 % (Miller et al., 2017) and 17.4 % (Weaver and Talma, 2005) of MAP. J2000

432 estimates had an average value of 5.30 % while MODFLOW was 5.20% for the eight hydraulic

433 zones of Krom Antonies. The coefficient of determination (R^2) between net recharge from the

434 J2000 and MODFLOW was 0.81. Across the entire dataset J2000 overestimated groundwater

435 recharge by 2.75 % relative to MODFLOW, although the coefficient of determination produced

436 an R^2 of 0.92 which is better than during the validation period.

438 Figure 7: The groundwater calibration for each hydraulic zone with a) net recharge for the

439 J2000 and MODFLOW during the model calibration (2016) and b) the net recharge deviation

440 between MODFLOW and J2000 across the entire modelling timestep (1986-2017)

441 **3.8 EMD filtering**

442 To account for missing streamflow data between 2007-2017, an Empirical Mode

443 Decomposition (EMD) (Huang et al., 1998) was applied to the measured water level data at

444 the sub-catchment outlet (G3T001)(Fig. 1) between 1994 to 2018 (Fig 8a). EMD is a method

445 for the decomposition of nonlinear and nonstationary signals into sub-signals of varying

446 frequency, so-called intrinsic mode functions (IMF), and a residuum signal. By removing one

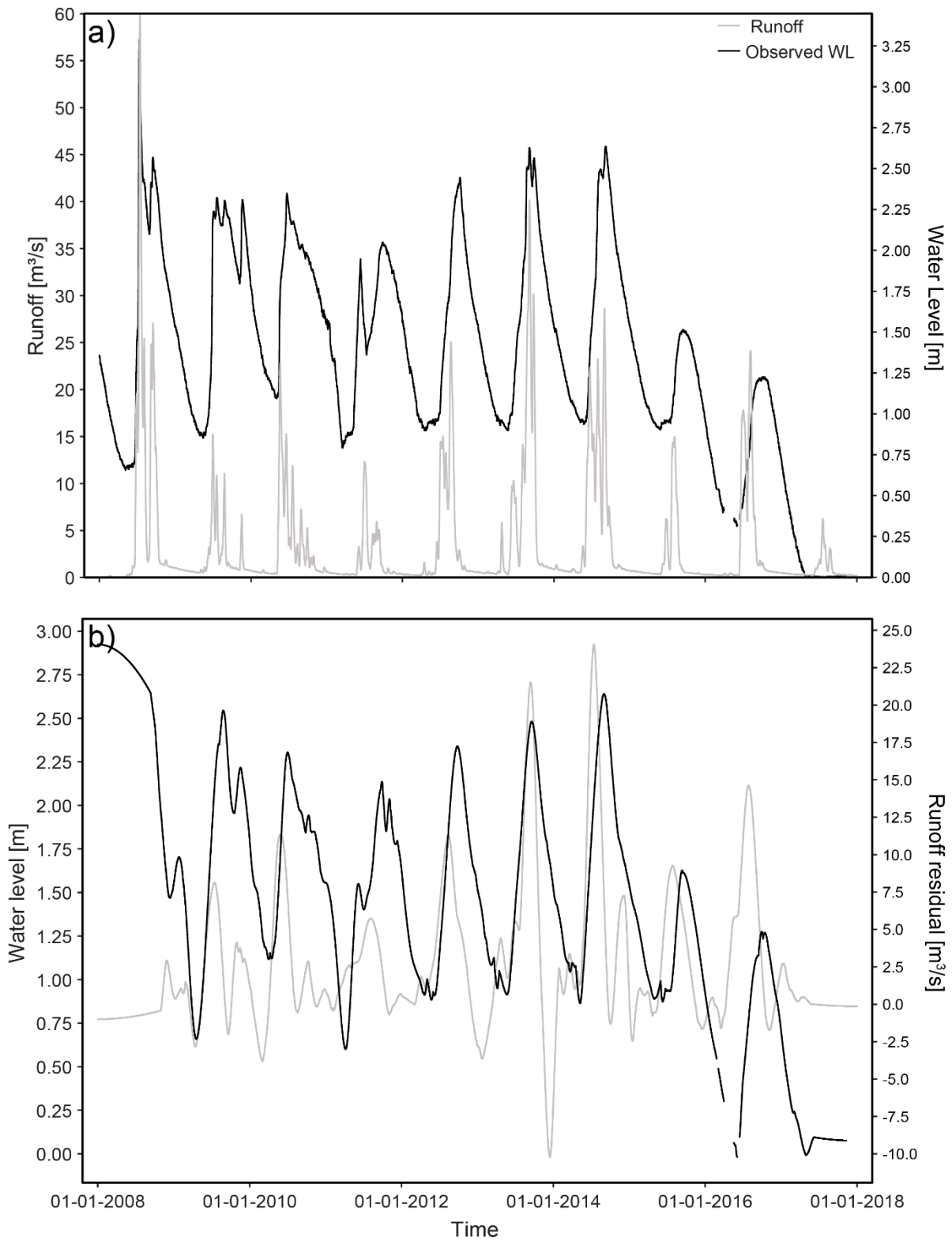
447 or more IMF or the residuum signal, certain frequencies (e.g. noise) or an underlying trend can

448   be removed from the original time series data. This approach was successfully applied to the

449   analysis of river runoff data (Huang et al., 2009) and forecasting of hydrological time series

450   (Kisi et al., 2014). In this study, EMD filtering was used to remove high frequency sub-signals

451   from simulated runoff and measured water level data to compare the more general seasonal

452   variations of both signals (Fig. 8b).

453

454 Figure 8: a) The water level fluctuations at station G3T001 with modelled runoff and b) the

455 EMD filtering showing the variation in discharge timeseries attributed a water level change at

456 the station

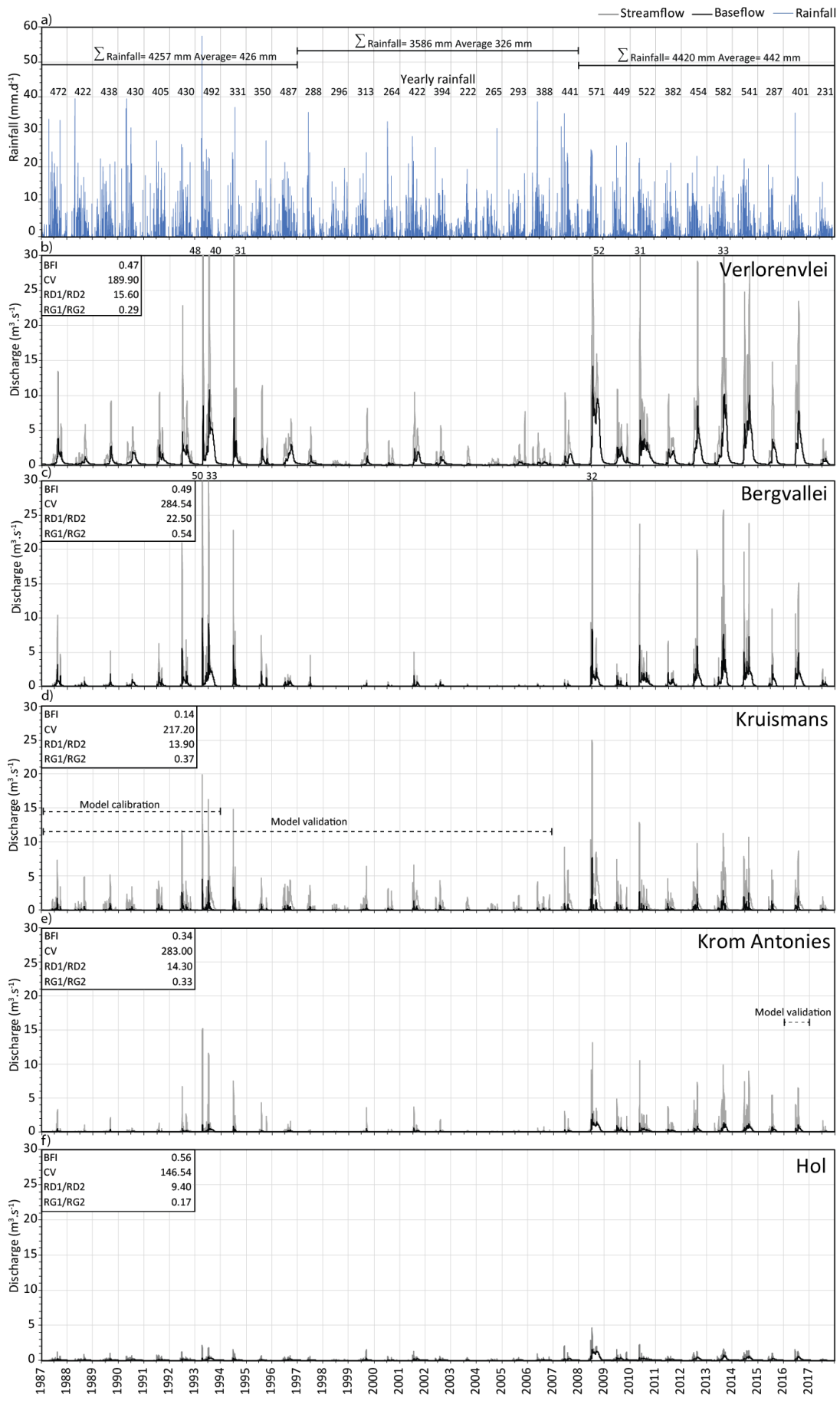
4. Results

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 (1987-1996 and 2008-2017), separated by a dry period (1997-2007) (Fig. 9). Yearly sub-catchment rainfall volumes between 1987-1996 were between 288 and 492 mm/yr^{-1}, with an average of 426 mm.yr^{-1} and standard deviation (STD) of 51 mm.yr^{-1}. For this period, average yearly streamflow 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, where 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 326 mm.yr^{-1} and STD of 69 mm.yr^{-1} (Fig. 9). For this same period, average yearly streamflow 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. During the second wet period between 2008-2017 sub-catchment yearly rainfall was between 231 and 582 mm.yr^{-1} with an average of 442 mm.yr^{-1} and STD of 112 mm.yr^{-1} (Fig. 9). Over this same period, average yearly streamflow was 2.5 m^3.s^{-1} with an average baseflow contribution of 1.3

480    $m^3.s^{-1}$. The modelled streamflow reached a maximum of 52 $m^3.s^{-1}$ in 2008, with 13 $m^3.s^{-1}$ of

481   baseflow generated after two consecutive rainfall events each of 25 mm.

483   Figure 9: a) Average sub-catchment rainfall between 1987-2017 showing wet cycles (1987-

484   1997 and 2008-2017) and dry cycle (1997-2007), Modelled streamflow and baseflow inflows

485   for the b) Verlorenvlei, c) Bergvallei, d) Kruismans, e) Krom Antonies and f) Hol tributaries

486   with estimated BFI, CV, RD1/RD2, RG1/RG2

487   **4.2 Tributary contributions**

488   The four main feeding tributaries (Bergvallei, Kruismans, Hol and Krom Antonies) together

489   contribute 81 % of streamflow for the Verlorenvlei, with the additional 19 % from small

490   tributaries near Redelinghuys (Fig. 10). Kruismans contributes most of the total streamflow at

491   32 %, but only 15 % of the area-weighted contribution as its sub-catchment is the largest of the

492   four tributaries at 688 km^2 (Fig. 10). Bergvallei with a sub-catchment of 320 km^2, contributes

493   29 % of the total flow with an area weighted contribution of 28 %. Krom Antonies has the

494   largest area weighted contribution of 30 % due to its small size (140 km^2) in comparison to the

495   other tributaries, although Krom Antonies contributes only 13 % of the total flow (Fig. 10).

496   Hol sub-catchment at 126 km^2 makes up the smallest contribution to the total flow of only 7

497   %, but has a weighted contribution of 17 % (Fig. 10).

498

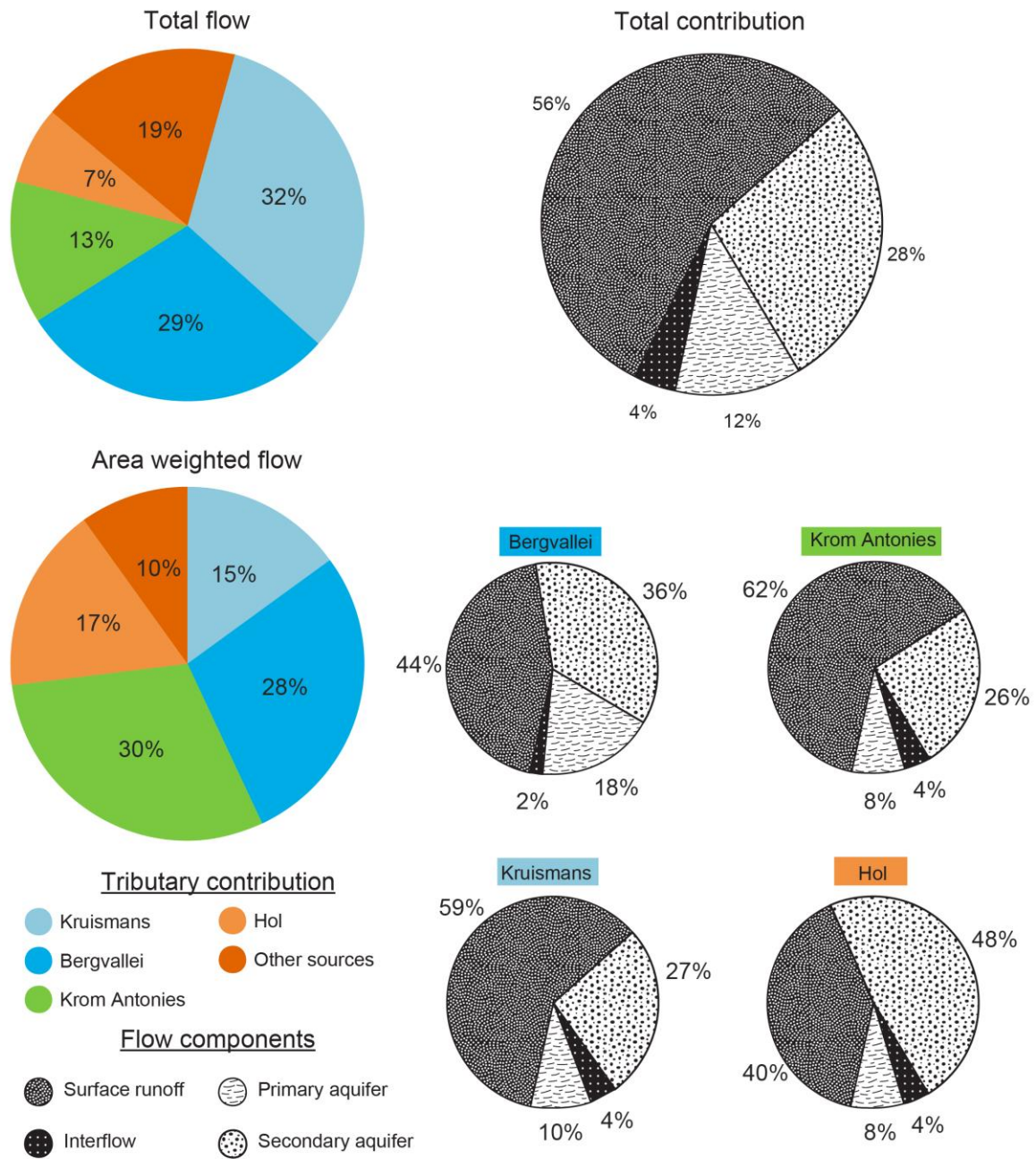



Figure 10: The Verlorenvlei reserve flow contributions (total flow and area weighted flow) of

Kruismans, Bergvallei, Krom Antonies and Hol as well as flow component separation into

surface runoff (RD1), interflow (RD2), primary aquifer flow (RG1) and secondary aquifer

flow (RG2).

4.3 Flow variability

Streamflow that enters Verlorenvlei has a large daily variability with a coefficient of variation (CV) of 189.90 (Fig. 9). This is mainly due to high streamflow variability from Kruismans (32 %) with a CV of 217.20, which is the major total flow contributor (Fig 10). Bergvallei and Krom Antonies, which both have high streamflow variability with CV values of 284.54 and 283.00 respectfully (Fig. 9), further contribute to the high variability of streamflow that enters the lake. While Hol reduces the overall streamflow variability with a CV of 146.54, it is a minor total flow contributor (7 %) and therefore does not reduce the overall streamflow variability significantly (Fig. 10).

Streamflow that enters Verlorenvlei is dominated by surface runoff which makes up 56 % of total flow, with groundwater and interflow contributing 40 % and 4 % respectfully (Fig. 10). The large surface runoff dominance in streamflow entering the lake, is due to a high surface runoff contribution from Kruismans and Krom Antonies, which contribute 26 % of total flow from surface runoff. However, for Bergvallei and Hol, surface runoff contributions are less dominant with 16 % of the total, while the total groundwater contribution is 20 % from these tributaries. Across all four tributaries, the secondary aquifer is the dominant baseflow component with 28 % of total flow, with the primary aquifer contributing 12 %. Bergvallei and Kruismans contribute the majority of primary aquifer baseflow with 8 % of the total. The secondary aquifer baseflow is mainly contributed by Kruismans and Bergvallei, where together 18 % of the total is received. Interflow across the four tributaries is uniformly distributed with 0.3 – 1 % of the total flow being contributed from each tributary.

4.4 Flow exceedance probabilities

The flow exceedance probability, which is a measure of how often a given flow is equalled or exceeded was calculated for each of the tributaries as well as the lake water body. The results

527   for the flow exceedance probabilities includes flow volumes which are exceeded 95 %, 75 %,

528   50 %, 25 % and 5 % of the time. The 95 percentile corresponds to a lake inflow of 0.054 $m^3.s^{-1}$

529   or 4,702 $m^3.d^{-1}$, with between 0.001-0.004 $m^3.s^{-1}$ from the feeding tributaries (Fig. 11 and

530   Table 3). The 75-percentile flow, which is exceeded 3/4 of the time corresponds to an inflow

531   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.

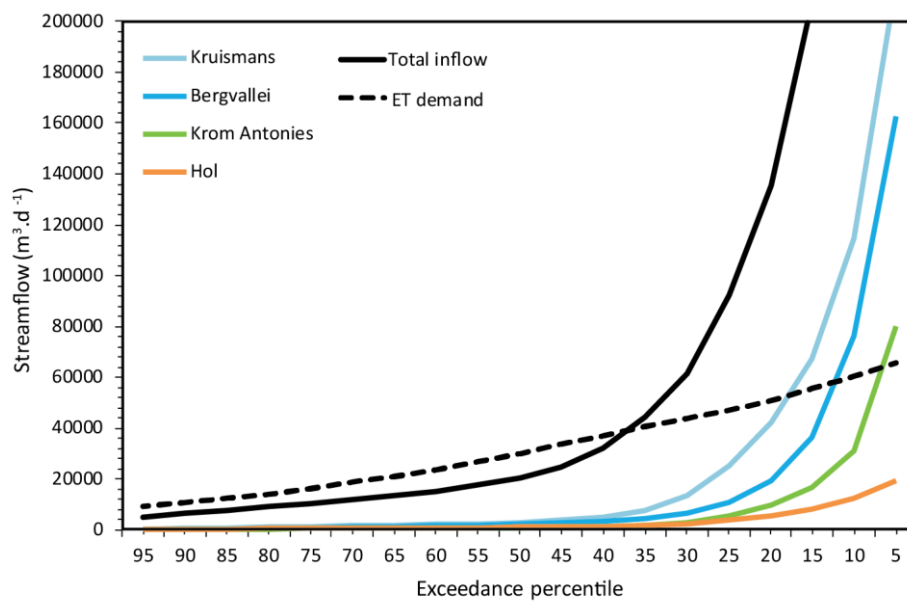
532   Average (50 percentile) streamflow flowing into the Verlorenvlei is 0.237 $m^3.s^{-1}$ or 20,498

533    $m^3.d^{-1}$, with between 0.010-0.035 $m^3.s^{-1}$ from the feeding tributaries. The 25-percentile flow,

534   which is exceeded ¼ of the time corresponds to a lake inflow of 1,067 $m^3.s^{-1}$ or 92,204 $m^3.d^{-1}$

535   with between 0.044-0.291 $m^3.s^{-1}$ from the feeding tributaries. The lake inflows that are

536   exceeded 5 % of the time correspond to 6.939 $m^3.s^{-1}$ or 599,535 $m^3.d^{-1}$ with between 0.224-

537   2.49 $m^3.s^{-1}$ from the feeding tributaries.

538

539   Figure 11: The streamflow exceedance percentiles and evaporation demand of the Verlorenvlei

540   reserve, with the contributions from each feeding tributary

Exceedance percentile	Lake ET m³.d⁻¹	Verlorenvlei m³.s⁻¹	m³.d⁻¹	Kruismans m³.s⁻¹	m³.d⁻¹	Bergvallei m³.s⁻¹	m³.d⁻¹	Krom Antonies m³.s⁻¹	m³.d⁻¹	Hol m³.s⁻¹	m³.d⁻¹
95	9158	0.054	4702	0.004	346	0.001	69	0.001	109	0.002	176
90	10956	0.074	6356	0.007	604	0.002	191	0.003	232	0.003	269
85	12559	0.088	7628	0.010	830	0.004	366	0.004	319	0.004	353
80	14249	0.104	8979	0.012	1072	0.007	596	0.005	392	0.005	434
75	16330	0.119	10303	0.015	1291	0.010	839	0.005	459	0.006	508
70	18653	0.136	11759	0.018	1517	0.013	1104	0.006	534	0.007	587
65	21152	0.155	13373	0.021	1791	0.016	1381	0.007	602	0.008	676
60	23791	0.176	15180	0.024	2104	0.019	1657	0.008	685	0.009	786
55	26979	0.203	17575	0.029	2506	0.023	1965	0.009	772	0.011	913
50	30057	0.237	20498	0.035	3032	0.027	2309	0.010	882	0.012	1058
45	33467	0.286	24669	0.043	3755	0.032	2807	0.012	1024	0.014	1222
40	36760	0.371	32023	0.058	5022	0.041	3511	0.015	1258	0.017	1439
35	40391	0.516	**44598**	0.089	7699	0.053	4613	0.020	1745	0.021	1790
30	43814	0.710	61310	0.156	13511	0.076	6599	0.033	2824	0.029	2481
25	47062	1.067	92204	0.291	25182	0.123	10619	0.062	5387	0.044	3814
20	50997	1.571	135726	0.489	42242	0.223	19295	0.110	9511	0.065	5655
15	55797	2.399	207275	0.780	**67408**	0.421	36354	0.192	16594	0.096	8262
10	60162	3.759	324746	1.324	114432	0.885	**76477**	0.359	31045	0.141	12191
5	65418	6.939	599535	2.490	215152	1.884	162795	0.929	**80305**	0.224	19312

541

542   Table 3: The streamflow exceedance percentiles and lake evaporation demand for the

543   Verlorenvlei reserve, with the contributions from Kruismans, Bergvallei, Krom Antonies and

544   Hol (m³.s⁻¹ and m³.d⁻¹)

5. Discussion

546   The adaptation of the J2000 rainfall/runoff model was used to understand the flow

547   contributions of the main feeding tributaries, the proportioning of baseflow to surface runoff

548   as well as how often the inflows exceed the lake evaporation demand. Before a comparison

549   with previous baseflow estimates can be made and the impact of evaporation on the lake

550   reserve, the model limitations and catchment flow dynamics must also be assessed.

5.1 Model limitations and performance

552   A major limitation facing the development and construction of comprehensive modelling

553   systems in sub-Saharan Africa is the availability of appropriate climate and streamflow data.

554   For this study, while there was access to over 20 years of streamflow records, the station was

555   only able to measure a maximum of 3.68 m³.s⁻¹, which hindered calibration of the model for

556   high flow events. As such, the confidence in the model's ability to simulate high streamflow

557   events using climate records is limited. While the availability of measured data is a limitation

558   that could affect the modelled streamflow, discontinuous climate records also hindered the

559   estimations of long time series streamflow.

560   Over the course of the 31-year modelling period, a number of climate stations used for

561   regionalisation were decommissioned and were replaced by stations in different areas. This

562   required adaption of climate regionalisation for simulations over the entire 31-year period to

563   incorporate the measured streamflow from the gauging station. To account for missing

564   streamflow records since 2007, an EMD filtering protocol was applied to the runoff data (Fig.

565   6). The results from the EMD filtering showed that after removing the first nine IMFs, the local

566   maxima of both signals match the seasonal water level maxima during most of the years. While

567   considerable improvement can be made to the EMD filtering, the results show some agreement

568   which suggested that the simulated runoff was representative of inflows into the lake.

569   **5.2 Catchment dynamics**

570   Factors that impact streamflow variability are important for understanding river flow regime

571   dynamics. Previously, factors that affected streamflow variability such as CV and BFI values

572   were used to determine how susceptible particular river systems were to drought (e.g Hughes

573   and Hannart, 2003). While CV values have been used to account for climatic impacts such as

574   dry and wet cycles, BFI values are associated with runoff generation processes that impact the

575   catchment. For most river systems, BFI values are generally below 1 implying that runoff

576   exceeds baseflow. In comparison CV values can be in excess of 10 implying high variability

577   in streamflow volumes (Hughes and Hannart, 2003). In this study, these two measurements

578   have been applied to tributaries as opposed to quaternary river systems, to understand the

579   streamflow input variability into the Verlorenvlei.

580   The highest proportion of streamflow needed to sustain the Verlorenvlei lake water level is

581   received from the Bergvallei tributary, although the area weighted contribution from Krom

582   Antonies is more significant (Fig. 10). However, CV values for the Bergvallei indicate high

583   streamflow variability. This is partially due to the high surface runoff component in modelled

584   streamflow within the Bergvallei in comparison to the minor interflow contribution, suggesting

585   little sub-surface runoff. While streamflow from the Bergvallei tributary is 54 % groundwater,

586   which would suggest a more sustained streamflow, due to the TMG dominance as well as a

587   high primary aquifer contribution, baseflow from the Bergvallei is driven by highly conductive

588   rock and sediment materials. Similarly, CV values for Krom Antonies indicate high streamflow

589   variability due to the presence of a high baseflow contribution from the conductive TMG and

590   primary aquifers. Although Krom Antonies has a larger interflow component, which would

591   reduce streamflow variability, the dominant TMG presence within this tributary partially

592   compensates for the subsurface flow contributions.

593   In contrast, Hol has a much smaller daily streamflow variability in comparison to both

594   Bergvallei and Krom Antonies (Fig. 9). While streamflow from Hol tributary is mainly

595   comprised of baseflow (56 %), the dominance of low conductive shale rock formations as well

596   as a large interflow component results in reduced streamflow variability. While the larger shale

597   dominance in this tributary not only results in a more sustained baseflow from the secondary

598   aquifer, it also results in a large interflow component due to the limited conductivity of the

599   shale formations. Compounding the more sustained baseflow from Hol tributary, the reduced

600   extent of the primary aquifer results in a dominance in slow groundwater flow from this

601   tributary. Similarly, Kruismans is dominated by shale formations which result in a larger

602 interflow contribution, although due to the limited baseflow contribution (37 %) the streamflow

603 from this tributary is highly variable, which impacts on its susceptibility to drought.

604 The results from this study have shown that while Krom Antonies was initially believed to be

605 the major flow contributor, Bergvallei is in fact the most significant, although streamflow from

606 the four tributaries is highly variable, with baseflow from Hol tributary the only constant input

607 source. The presence of conductive TMG sandstones and quaternary sediments in both Krom

608 Antonies and Bergvallei, results in quick baseflow responses with little flow attenuation. The

609 potential implication of a constant source of groundwater being provided from Hol tributary,

610 is that if the groundwater is of poor quality this would result in a constant input of saline

611 groundwater, with Krom Antonies and Bergvallei providing freshwater only after sufficient

612 rainfall has been received.

613 **5.3 Baseflow comparison**

614 The groundwater components of the J2000 model were adjusted using aquifer hydraulic

615 conductivity from a MODFLOW model of one of the main feeding tributaries of the

616 Verlorenvlei. Krom Antonies was selected as it was previously believed to be the largest input

617 of groundwater to Verlorenvlei (Fig. 2). Baseflow for Krom Antonies tributary was previously

618 calculated using a MODFLOW model (Watson, 2018), by considering aquifer hydraulic

619 conductivity and average groundwater recharge. As average recharge was used, baseflow

620 estimates from MODFLOW are likely to fall on the upper end of daily baseflow values

621 estimated by the J2000 model. For Krom Antonies sub-catchment, Watson, (2018) estimated

622 baseflow between 14,000 to 19,000 $m^3.d^{-1}$ for 2010-2016 using MODFLOW. Similar daily

623 baseflow estimates from the J2000 were only exceeded 10 % of the time, with average

624 estimates (50 %) of 1,036 $m^3.d^{-1}$ over the course of the modelling period (Fig. 9).

625   The MODFLOW estimates were applied over the course of a wet cycle (2016). In comparison

626   to the MODFLOW estimates (14,000 to 19,000 $m^3.d^{-1}$) average baseflow from J2000 for 2016

627   was 8, 214 $m^3.d^{-1}$. The daily timestep nature of the J2000 is likely to result in far lower baseflow

628   estimates, as recharge is only received over a 6-month period as opposed to a yearly average

629   estimate. One possible implication of this is that while common groundwater abstraction

630   scenarios have been based on yearly recharge, abstraction is likely to exceed sustainable

631   volumes during dry months or dry cycles and this could hinder the ability of the aquifer to

632   supply baseflow. While the groundwater components of the J2000 have been distributed to

633   allow for improved baseflow estimates, the groundwater calibration was applied to Krom

634   Antonies. However, this study showed that Bergvallei has been identified as the largest water

635   contributor. In hindsight, the use of geochemistry to identify dominant tributaries could have

636   aided the groundwater model adaption. While it would have been beneficial to adapt the

637   groundwater components of the J2000 using the dominant baseflow contributor, considering

638   the geological heterogeneity between tributaries is more important for identifying how to adapt

639   the groundwater components of the J2000. While the distribution of aquifer components

640   improved modelled baseflow, including groundwater abstraction scenarios in baseflow

641   modelling in the sub-catchment is important for future water management for this ecologically

642   significant area.

643   **5.4 The Verlorenvlei reserve and the evaporative demand**

644   For this study, exceedance probabilities were estimated through rainfall/runoff modelling for

645   the previous 31 years within the Verlorenvlei sub-catchment. The exceedance probabilities

646   were determined for each tributary, as well as the total inflows into the lake. These exceedance

647   probabilities were compared with the evaporative demand of the lake, to understand whether

648   inflows are in surplus or whether the evaporation demand exceeds inflow.

649 From the exceedance probabilities generated in this study, the lake is predominately fed by less

650 frequent large discharge events, where on average the daily inflows to the lake do not sustain

651 the lake water level. This is particularly evident in the measured water level data from station

652 G3T001, where measured water levels have a large daily standard deviation (0.62) (Watson *et*

653 *al*., 2018). The daily inflows of water into the Verlorenvlei has also been subject to significant

654 rainfall variability, with yearly rainfall between the second wet cycle (2007-2017) being twice

655 as variable in comparison to the first wet cycle (1987-1996). The change in rainfall variability

656 has had a significant impact on soil moisture conditions, resulting in not only larger peak

657 discharges but also lengthened low flow conditions. With climate change likely to impact the

658 length and severity of dry cycles, it is likely that the lake will dry up more frequently into the

659 future, which could have severe implications on the biodiversity that relies on the lake's habitat

660 for survival. Of importance to the lake's survival is the protection of river inflows during wet

661 cycles, where the lake requires these inflows for regeneration.

662 While the impact of irrigation could not be incorporated, over allocation of water resources

663 may potentially have a significant impact on the catchment water balance, especially during

664 wet cycles when ecosystems are recovering from dry conditions. The increased irrigation

665 during wet cycles as a result of agricultural development, could be a further impact on the

666 recovery of sensitive ecosystems. This type of issue is not limited to Verlorenvlei but applies

667 to many wetlands or estuarine lakes around the world, while they have been classified as

668 protected areas, water resources within the catchments are required for food security. As

669 climate change drives increased temperatures and variability in rainfall, the ± 10-year cycles

670 of dry and wet conditions may no longer be valid anymore, where these conditions may shorten

671 or lengthen. With the routine breaking of weather records across the world (Bruce, 2018; Davis,

672 2018), it is becoming increasingly evident that conditions are changing and becoming more

673   variable, which could impact sensitive ecosystems around the world, highlighting the need for

674   effective water management protocols during times of limited rainfall.

675

6. Conclusion

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 DT limit, 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, with yearly rainfall becoming more variable. 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 become particularly important for ecosystem regeneration, especially for semi-arid regions such as the Verlorenvlei.

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