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**Water balance estimation under the challenge of data scarcity in a hyper-arid to Mediterranean region**

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**Abstract:**

Water budget analyses are important for the evaluation of the water resources in semi-arid and arid regions. The lack of observed data is the major obstacle for hydrological modelling in arid regions. The aim of this study is the analysis and calculation of the natural water resources of the Western Dead Sea subsurface catchment, one which is highly sensitive to rainfall resulting in highly variable temporal and spatial groundwater recharge. We focus on the subsurface catchment and subsequently apply the findings to a large-scale groundwater flow model to estimate the groundwater discharge to the Dead Sea.

We apply a semi-distributed hydrological model (J2000g), originally developed for the Mediterranean, to the hyper-arid region of the Western Dead Sea catchment, where runoff data and meteorological records are sparsely available. The challenge is to simulate the water budget, where the localized nature of extreme rainstorms together with sparse runoff data results in few observed runoff and recharge events. To overcome the scarcity of climate input data we enhance the database with mean monthly rainfall data. The rainfall data of two satellites are shown to be unsuitable to fill the missing rainfall data due to underrepresentation of the steep hydrological gradient and temporal resolution. Hydrological models need to be calibrated against measured values, hence the absence of adequate data can be problematic. Therefore, our calibration approach is based on a nested strategy of diverse observations. We calculate a direct surface runoff of the Western Dead Sea surface area (1801 km$^2$) of 3.4 mm/a and an average recharge (36.7 mm/a) for the 3816 km$^2$ subsurface drainage basin of the Cretaceous aquifer system.

**Keywords:** data scarcity, hydrological model, J2000g, water balance, Dead Sea

**Introduction**

Water resources in the Dead Sea Basin are strongly limited due to the largely arid to hyper-arid climate. Therefore, reliable estimates are necessary for sustainable evaluation of available water resources, especially for groundwater recharge as the main source of replenishment of water resources.

In the study area the strong seasonal climate of highly variable rainfall in autumn and spring and the specific landscape characteristics of steep slopes and considerable variation in soil moisture require consideration in the simulation of a water balance. Hydrological models for water balance simulations establish a link between hydrological and climatic variables. Thereby the models have to be sufficiently complex to simulate the dominant hydrological processes, but not too complex to be over parameterized (Blöschl and Grayson, 2002). For our study area different simple and partly uncalibrated hydrological models have been applied (e.g. Al-Jabari et al., 2009; Gunkel et al., 2012; Lange et al., 1999; Rozalis et al., 2010; Sheffer et al., 2010). They focused on the description of a single hydrological process variable (groundwater recharge, surface runoff in the form of flash floods or rainfall-runoff relations) whereas other terms of the water budget equation were neglected.

The optimal simulation of a hydrological process variable is highly dependent on the temporal resolution of the input parameters (Cherkauer, 2009). For example, daily time steps are common for the modelling of rainfall extremes, realistic stream flow variability or soil erosion. In contrast, monthly time steps are sufficient to model drought and broad scale hydrology (Hutjes et al., 1998; Kabat et al., 2004). In the Dead Sea catchment the spatio-temporal dynamics of surface runoff are largely determined by frontal synoptics in winter but by local heavy rain events, which are often a rain event per month or even per season (Dayan and Morin, 2006). Frequency analysis of rain events in the study area show that the depth of most rain events is under 20 mm/day, the daily rain depth is >20 mm for about 12% of the monthly rainy days and the probability of a >50 mm/day rainfall event is 1.5 days per year (Ben-Zvi, 2009; Goldreich, 2003; Morin et al., 2001 and Morin and Yakir, 2012). However, in the study area the limited data availability of rainfall data for the 1970 - 2010 time period required a coarsening of the temporal resolution to a monthly scale. Self-evidently, increasing the time step reduces the accuracy and the dynamics of the sub-daily processes (e.g., Cherkauer, 2009; Hughes, 1995). These and other similar studies showed that the simulated water discharge peaks are most sensitive to model time step, coarser time steps yielding fewer peaks but the mean water volume is nearly constant. The latter and the fact that surface runoff to a large extent depends on single months during which strong rainfall events could be exploited for calibration of the time of occurrence and the volume of surface runoff events. Admittedly, monthly rain time steps cannot yield realistic outputs of streamflow variability. Our objective is to apply a model of intermediate complexity, such as J2000g (Kralisch and Krause, 2006), for a combined simulation of the general water budget for the Western Dead Sea catchment. We thereby make use of the existing experience with the J2000g model from a similar, nearby area in Jordan (Rödiger et al., 2014; Schulz et al., 2013).

A major problem for hydrological modelling in most regions is the lack of observed data and the quality of the measurements. For example, standard error analysis in different studies show that individual discharge measurements have a range between 2% under ideal conditions to about 20% when conditions are poor and shortcut methods are used. Furthermore, most measurements have a standard error in the range 3-6% (Sauer and Meyer, 1992). Additionally, the calibration process is affected by the location of the measurements. Because limited runoff data are based on gauging stations located relatively far from the source of runoff generation. The absence of data leads to soft model statements and partially uncalibrated models.

To overcome this common situation, a multi-objective calibration was carried out by fitting model state variables (e.g., groundwater recharge) to suitable observations in addition to classical calibration with regard to runoff (Klemeš, 1986). This approach was successfully applied in previous studies (Rödiger et al., 2014; Sharda et al., 2006; Sophocleous, 1991). The objective of this study is to estimate the water balance of a catchment that is highly sensitive to the spatial and temporal distribution of rainfall and under the challenge of data scarcity. Among others, an appropriate method to overcome data scarcity is the refill of rainfall data series based on data analysis. A calibration parameter for maximum infiltration was considered to illustrate the effect of increasing soil surface dryness depending on aridity. Final we present a numerical hydrological model with calibrated data series and estimate the water balance of the catchment. This method is applied to the Western Dead Sea catchment.

**Study area**

The Dead Sea is an endorheic lake located in the Jordan Rift Valley. The study area (Figure 1) represents 3816 km² of a western subsurface drainage basin, directly connected to the Dead Sea (Gräbe et al., 2013). The surface catchment is affected by a topographic gradient from about 1,000 m.a.s.l. in the N-S-oriented mountainous regions to currently -431.07 m.b.s.l. (eingedi.co.il, 2017) at the Dead Sea over a distance of about 30 km. The climate has respective average 25-39°C summer and 10-23°C winter temperatures. The most important climatological feature is a strong rainfall gradient, stretching from Mediterranean mountainous regions (Hebron: 595 mm/a, Al Arroub: 632 mm/a (PMD, 2015)) to hyper-arid in the Jordan Valley (< 50 mm/a: e.g. Dragot: 14 mm/a). Rainfall usually occurs between October and April. The convective rainstorms, which are the principal synoptic rain in the western Dead Sea area, are either caused by an Eastern Mediterranean Low - Cyprus and Gaza depressions (Alpert et al., 1990) or by Red Sea Trough systems (Kidron, 2000). The aridity of the Dead Sea region is due to the rain shadow affecting the water availability in the Judean Mountains.

Observations in the field demonstrate that rain events lead to stronger and faster surface runoff processes in arid and hyper-arid areas. Reasons are to be found in the dry soil surface, in the presence of physical and biological crusting and lack of a vegetation cover (Kidron et al., 1999; Kidron et al., 2003; Yair, 1990) favouring increased surface runoff.

The geological settings represent the fractured and karstified lime- and dolomite aquifers of Upper Cretaceous age. Both aquifers belong to the Eastern Mountain Aquifer system, which described comparable double porosity aquifer systems. Inside the Jordan Rift Valley, Quaternary fluviatile sediments around Wadi mouths interfinger with the lacustrine to hypersaline lake sediments that form the heterogeneous coastal Dead Sea Group aquifer. Nine west-east oriented Wadis drain much of the surface catchment (Figure 1, Table 1), where ephemeral flash floods are generated after significant rainfall events. The runoff behaviour is characterized by infrequent flash floods with short duration hydrographs, some of which have high peak discharges and flood volumes. Due to the southerly increase in aridity, the situation is sharpened in the southern Wadis, which may be dry several years (Cohen and Laronne, 2005). The channels in all of these Wadis are characterized by narrow valley floors, an upper and middle steep reach, and a short lower channel reach with thick gravelly-sandy alluvium (Laronne and Shlomi, 2007). Five Wadi catchments were selected for this study: Qilt, Darga, Tkoa, Arugot and Rahaf. The main soil types from the mountain area down to the Jordan Valley are: (a) Terra Rossa, brown Rendzina and lithogenic Rendzina; (b) calcareous desert steppe lithosols and (c) bare rock and calcareous desert lithosols (Dan and Koyumdjisky, 1963; Dan, 1983; Shapiro, 2006; Singer, 2007). The depth of soils varies considerably and strongly depends on the dominating vegetation cover. Predominant soil types features are A- and C-horizons with a missing B-horizon. In areas with poor vegetation and steep slopes the A-horizon is also missing. During rain events the dry and immature soils possess only small capability to store an adequate amount of water.

Thus, in combination with the steep morphology, rainfall excess accumulates quickly and induces flash floods. The study area belongs to the Mediterranean and the Irano-Turanian vegetation zones (Zohary 1973). The large majority of the Wadi catchment areas are located in the Judean Desert in an Irano-Turanian zone mainly consisting of sparsely located shrubs and bushes.

## 1. Methods

The present study provides a description and quantification of water balance parameters for a Mediterranean to hyper-arid catchment. These are summarized in the following simplified water balance equation (Equation 1):

$$P = Q + ET + R \qquad (1)$$

where $P$ = precipitation, in our case it is solely rainfall, $Q$ = runoff, $ET$ = evapotranspiration and $R$ = groundwater recharge. Based on long-term data series of rainfall, the remaining water budget components are estimated by the conceptional hydrological model J2000g.

3.1. Estimation of groundwater recharge

Published data and empirical relation data (Equation 2-4) are used to derive plausible parameter ranges for groundwater recharge in J2000g. Guttman (2000) summarised previous studies in this area based on spring discharge and suggested an empirical equation relating $R$ to transient $P$:

$$R = 0.8 \times (P - 360) \qquad if\ P > 650\ mm/a \qquad (2)$$

$$R = 0.534 \times (P - 216) \qquad if\ 650\ mm/a < P > 300\ mm/a \qquad (3)$$

$$R = 0.15 \times P \qquad if\ P < 300\ mm/a \qquad (4)$$

This approach allows a rough estimation of groundwater recharge based on the distribution of rainfall. In contrast, the process-based hydrological model considers further parameters of the soil zone in humid areas as well as the surface characterization in more arid areas.

3.2. Estimation of aridity index

The aridity index (AI) (UNEP 1992, 1997) is calculated by the ratio of rainfall to potential evapotranspiration ($ET_{pot}$) (Equation 5):

$$AI = \frac{P}{ET_{pot}} \qquad (5)$$

In this method the calculation of $ET_{pot}$ is based on the simple Thornthwaite approach (Thornthwaite, 1948) due to an appropriate application in areas where only small-equipped meteorological stations exist.

3.3. Water-balance model J2000g

J2000g is implemented in the Jena Adaptable Modelling System (JAMS) (Kralisch et al., 2007). JAMS is an open-source software framework for component-based development and application of environmental models. It is a process-based, distributive model for water-balance estimation (Krause and Hanisch, 2009) simulating the important hydrological processes at the meso-scale in a daily or monthly time step. The model is a simplified version of J2000 (Krause, 2001), which was developed for long-term simulation of hydrological processes in large catchments. J2000g is characterized by a small number of calibration parameters and robust prediction capability in terms of water-balance (Krause and Hanisch, 2009). It calculates the hydrological processes for an arbitrary number of spatial modelling entities, the so-called Hydrological Response Units (HRUs) (e.g. Flügel, 1995; Krause, 2001). The model aggregates the simulated values to the desired spatial and temporal resolutions. Graphs and maps are generated to present the results of the water-balance simulation for further analysis. In the last step goodness of fit measures (e.g. coefficient of determination $r^2$) are determined.

**Step 1 (Input Data):** The J2000g model requires spatially distributed information on topography, land-use, soil type and hydrogeology to estimate specific attribute values for each HRU. Therefore, a mesh of irregular triangular HRU elements with a mesh base length between 200 and 2000 m was generated. The derivation of the HRUs is based on representation of the most important hydrological structures (wadi stream course) and in considering a later coupling with a groundwater flow model (Gräbe et al., 2013). The model also requires meteorological input data of rainfall, minimum, average and maximum temperature, sunshine duration, wind speed and relative humidity.

**Step 2 (Correction and spatial interpolation):** For each time step (daily or monthly) data from rain gauges are corrected according to Richter (1995). The measurements of all climatological variables are then interpolated by using inverse distance weighting and optional elevation correction (Shepard, 1968).

**Step 3:** For each time step and for each HRU the water balance components are calculated according to the climatological input data in the following order:

**Step 3a (Infiltration and Excess Water):** The central element of the J2000g model is the Soil Water Storage module. J2000g simulates single soil water storage for each modelling entity, representing the Middle Pore Storage (*MPS*). The capacity of this storage is defined by soil thickness and field capacity and can be adapted by the calibration parameter (*FCA*). In the model the amount of infiltration is limited by i) the soil water storage and ii) the maximum infiltration rate per time step (*maxInf*). After saturation, excess water (*EW1*) is distributed to evaporation, and the unevaporated portion of *EW1* is used as input (*EW2*) for the generation of direct runoff and groundwater recharge.

**Step 3b (Radiation and Potential Evapotranspiration):** J2000g estimates incoming solar radiation for each HRU based on observed average sunshine duration and topographical HRU properties (slope, aspect). Potential evapotranspiration is calculated using the Penman–Monteith formula based on observed data of wind-speed, temperature and relative humidity data (Allen et al., 1998).

The potential evapotranspiration rates are first taken from the *EW1* and *EW2* excess water storages. If potential evapotranspiration exceeds *EW1* and *EW2*, actual evapotranspiration (*actET*) is calculated depending on soil water saturation and a reduction parameter $\beta$ *(Equation 6)* actMPS*:*

$$actET = \frac{\Delta actET \cdot actMPS}{\beta \cdot maxMPS} [mm/timestep] \quad (6)$$

where *actMPS* = actual Middle Pore Storage and *maxMPS* = maximum Middle Pore Storage.

**Step 3c (Overland flow):** If infiltration excess water (*EW1*) is not completely evaporated, overland flow is generated (*dirQ*) and directly routed to the catchment outlet.

**Step 3d (Runoff component calculation):** The remaining amount of saturation excess water (EW2) is distributed to direct runoff (*dirQ*) (Equation 7) and groundwater recharge (*GWR*) (Equation 8):

$$dirQ = \tan a \cdot LVD \cdot EW2 \; [mm/timestep] \quad (7)$$

$$GWR = (1 - \tan \alpha) \cdot LVD \cdot EW2\ [mm/timestep] \quad (8)$$

The distribution depends on the slope of the HRU ($\alpha$) and a calibration parameter (*LVD),* a factor controlling the lateral and vertical distribution of infiltration. The direct runoff (*dirQ*) is routed together with the overland flow (from step 3c) to the catchment outlet within the same time step.

**Step 3e (Groundwater recharge):** Potential groundwater recharge calculated in step 3d is limited by a maximum percolation rate (*maxPerc*), defined by landscape characteristics and adapted by the parameter "*percAdaptation*". Inside the groundwater module the generated *GWR* is partitioned into two groundwater reservoirs (*GWS1*, *GWS2*), realised by the calibration coefficient $\alpha$. The two-groundwater reservoirs allow the simulation of a fast and slow groundwater component - double continuum aquifer (Rödiger et al., 2014; Sauter, 1993). The outflow (base flow: *basQ*) from *GWS1* and *GWS2* is computed with single linear storage cascades (Nash, 1958) parameterised by *n* linear reservoirs and the retention coefficient *k*.

## 2. Data

***Spatial data***: i) Topographical parameters (slope, aspect and elevation) were obtained from a digital elevation model (DEM, 25 x 25 m). The slopes and their related aggregated surface ratios (A) were classified into four groups (as per Tilch et al., 2002), in which the study area is characterized as moderate steep catchment, with steep slopes along the main faults. The aspects of each spatial unit and the respective surface area ratio were classified into eight 45° segments (Table 2).

ii) Land cover was classified by using an ASTER image (30 m grid) dated March 2008 (Table 3). The parameterization of classes is based on previous studies (e.g., Krause and Hanisch, 2009; Rödiger et al., 2014; Schulz et al., 2013).

iii) Soil properties (Table 4) are based on the soil map classification of Israel (Dan et al., 1976; Ravikovitch, 1969) and outcomes of the GLOWA Jordan River project - Global Change and the Hydrological Cycle (Gunkel et al., 2012). Fifteen main soil types, which can be separated in 29 subclasses with regard to slope, were identified in the study area. The information about average soil depth and field capacity of each soil class was estimated by field measurements and published results (Dan et al., 1980; Dan, 1983; Mimi and Assi, 2009; Singer, 2007 and Smith et al., 1985).

***Meteorological data***: Time series (mostly monthly records) of rainfall, relative humidity, sunshine duration, temperature and wind speed were available for the period January 1977 - December 2010. Data from 33 stations (Israel Met. Service, Palestine Met. Service and Internet databases) were used for the meteorological description of the study area (Table 5). A complicating factor is the spatial and temporal lack of meteorological data, a major problem for hydrological modelling in the region (Cudennec et al., 2007). Most of the ground-based meteorological data are concentrated in urban areas (Jerusalem, Jericho, Hebron and Arad), hence large areas are characterized by data scarcity. The spatial data scarcity is compounded by sparse time series for most of the station data. Therefore, satellite data products from the Tropical Rainfall Measuring Mission (TRMM) and Climate Prediction Centre Morphing Technique (CMORPH) were used to overcome the spatial and temporal data scarcity (Joyce et al., 2004). The CMORPH data set is characterized by a spatial resolution of 0.0727° latitude and longitude grid (8 km at the equator) and temporal data are available at half-hour intervals up to monthly data (Joyce et al., 2004). Rainfall data of the TRMM satellite mission are characterized by a larger pixel grid size of 0.25° by 0.25°and temporal resolution of 3 hours up to monthly data (Huffmann et al., 2007).

***Runoff data***: Due to the low frequency of direct runoff events in the Dead Sea region, five hydrometeorological stations (Qilt, Darga, Tkoa, Arugot and Rahaf) provide hydrograph records of a small number of events. The Israel Hydrological Service (IHS) provided the observed runoff events for the period 1977 – 2010. None of the registered runoff events occurred contemporaneously in all Wadis during the 33 years of record. Most of the discharge events are recorded in one catchment. It is apparent that most of the runoff events take place during winter (around 54%) and at approximately equal proportions (around 22%) in spring and autumn. The quality of most of hydrological data is poor because of the flashy nature of the hydrographs and the weakly armoured river beds, which allow scour and fill during floods (Hassan, 1990; Laronne et al., 1994). Due to the event-based characteristic of flash floods, water volume could be used as a calibration value for the monthly time steps, since the specific runoff volume of the partly singular rain event per month is coupled. For the modelling purposes, the event-based runoff data were accumulated to monthly runoff volumes.

## 3. Model setup

In a first step, published parameter values (e.g. Table 3 and 4) are used to build up a priori model with best guess parameter estimates (Table 8). First assumptions regarding soil data (*FCA*) were taken from field studies and published data (Sachse, 2015). The default settings of *FCA, β, LVD* and *maxInf* were set to 1 for use as the initial values to run the model without calibration. The coefficients for the baseflow dynamic (*n* and *k*) are set as described for a comparable double porosity (karstified) aquifer system in Jordan (Rödiger et al., 2014).

An accurate representation of spatial and temporal rainfall data is required for water balance modelling as uncertainties in rainfall estimates have a considerable effect on model simulations and predictions in semi-arid and arid areas. Due to the lack and the uncertainties of measured rainfall data in terms of spatial and temporal availability, reliability tests were undertaken before those of the calibration process (Figure 2).

**Initial test**: J2000g uses an inverse distance weighting correction with elevation for interpolation of rainfall observations. This adds a challenge additional to the existing difficulties: inhomogeneous distribution of stations, sparse and sometimes incomplete time series and the strong W-E rainfall gradient from the mountains to the Jordan Rift Valley. The largest data set was achieved only by additional consideration of rain gauges with monthly data. Consequently, model simulations were done on a monthly basis. The first simulations between simulated and observed runoff do not match well. An overestimation of surface runoff was simulated in the northern as well as central part and an underestimation of surface runoff in the southern part of the study area (Figure 7).

**Reliability test A:** The satellite products are only available for the 2000-2010 year time period. The comparison with ground-based data shows that the CMORPH grid data estimate too low rainfall event data. In contrast, TRMM shows a west-east orientation-dependence overestimation or underestimation of rain events. This may be due to an inappropriate spatial resolution of pixel data, which may not reflect the strong rain gradient within short distances along the graben flank. To conclude: both satellite products underestimate annual rainfall data (89 mm CMOPRPH, 189 mm TRMM) in contrast to the known annual rainfall data (313 mm) from the rain gauges of the study area (Figure 3) and are not useful for an appropriate modelling approach.

R**eliability test B:** Missing rainfall information was systematically filled with average monthly rainfall data belonging to the local rain gauge. This kind of rainfall data reconstruction is a widely accepted method in hydrological modelling (Stooksbury et al., 1999). Due to the processing method, the refill leads to harmonize time series without extreme events in the rainfall distribution. However, the enhanced time series represent the actual characteristic of this region more closely than the satellite products. Notwithstanding, rain gauges for which the calculation of mean values is based on one hydrological year only must be viewed critically due to the large variability in annual rainfall distribution (rain stations Mizpe Jericho, Asify Bar, Bet Govrin, Shufat and Doma). In the subsequent calibration and validation process, we show that the systematically filling by average monthly rainfall data compensates the lack of information and ensures a reliable analysis of Wadi runoff volumes due to the provision of continuous time series for simulation.

### 4.1. Calibration

A classical runoff calibration and validation approach (e.g., Klemeš, 1986) is not possible due to the short runoff time series. Instead, a nested multi-response calibration strategy was applied (Figure 2) based on prior experience (Rödiger et al., 2014; Sharda et al., 2006; Sophocleous, 1991). This approach uses diverse observations to calibrate sub-parts of the nested model, whereby available data used to derive realistic ranges for single model parameters or subsets of model parameters are identified.

1) The calibration of surface runoff volume was done mainly by modification of two calibration parameters: $\beta$, which modifies evaporation and hence soil water saturation, and the parameter *FCA*, which controls soil water capacity. In Figure 4 the observed and simulated discharges for runoff events of the five Wadis are generalized for different combinations of $\beta$ and *FCA*. An increase of $\beta$ causes higher rates of evapotranspiration resulting in a decrease of runoff generation. Increasing *FCA* leads to decrease in both surface runoff and groundwater recharge (Figure 4). The calibration process was stopped because a better adaptation of the model was not possible without an unnatural modification of both calibration parameters (Figure 7). Although runoff was overestimated, both calibration

parameters generate a significant improvement of the modelling results. The combination *FCA*=0.9 and *β*=0.8 results in a moderate correlation between simulated and observed direct runoff volume although runoff volumes at Wadi outlets are still overestimated (Figure 7, dark grey bars; Table 8).

2) The calibration of mean groundwater recharge was mainly driven by adapting two calibration parameters: the parameter *LVD* and the parameter *maxInf*. Published groundwater recharge information is used to estimate the remaining calibration parameters (*LVD, maxInf*) listed together with the used method (Table 6). The modelling domain is extended to the subsurface groundwater catchment to compare modelled groundwater recharge to published values. The comparison of the outcome of groundwater recharge by the first calibration process (31 mm/a) versus the published values (Table 6) shows that the model estimation was too low. In the subsequent step, the ratio of vertical to horizontal discharge by the *LVD* parameters was changed. The modifications were accompanied by field infiltration tests in the Judean Desert. Indeed, considerable vertical water movement takes place due to the presence of fissures and the karst system as confirmed elsewhere (Ries et al., 2015; Schmidt, 2014). Increasing the vertical drainage component gave the required effect, for which groundwater recharge attained comparable values with published data (Figure 5). This effect was accompanied by a decline in surface runoff, such that in the dry Mediterranean western part of the study area very low runoff was present and in the arid to hyper-arid eastern area runoff hardly existed. Due to an increase in soil surface dryness with increase in aridity, the calibration parameter *maxInf* was developed and implemented in J2000g. The parameter *maxInf* was set in the arid and hyper-arid zones and parameterized with 10 mm/month limited infiltration rates (Figure 6). In these areas the application led to a decline in groundwater recharge, and to a significant increase of surface runoff. The implementation of the new calibration parameter led to a breakthrough in the water balance simulation for the study area.

The nested multi-response calibration approach resulted in the following parameter set: *FCA* = 0.9, $\beta$ = 0.8, *LVD* = 0.1, *maxInf* = 10 mm (Table 8). The final calibration data set was achieved on sub-catchments (Qilt, Darga, Tkoa, Arugot and Rahaf) by obtaining, which a robust model result in terms of the water balance parameters of runoff (Figure 7, final model) and groundwater recharge (Figure 5).

**4.2. Validation**

The validation process was launched on the Wadi scale and ended on the scale of the groundwater catchment of the entire model area. The following steps were undertaken: (a) simulated surface runoff versus observed runoff of Wadi Darga , (b) simulated versus measured actual evapotranspiration and (c) distribution and amount of the simulated versus empirical groundwater recharge data (Guttman, 2000).

a) Figure 8 presents the good fit between simulated and measured runoff data for Wadi Darga. The majority of the runoff events were simulated well with respect to volume ($r^2$=0.6). b) Simulation of the evapotranspiration yielded between 180 (southern part) and 380 mm/a (northern part). These values of evapotranspiration are comparable to 360 mm annual evapotranspiration for this area (Al-Jabari et al., 2009; Goldschmidt, 1955; Schmidt et al., 2014) and show that the model is able to represent the evaporation.

c) The resulting groundwater recharge of 139.9 MCM and the simulated groundwater recharge distribution for the Western Dead Sea Escarpment are comparable with groundwater recharge results based on an empirical rainfall runoff relation (Guttman and Zuckerman, 1995). The simulated groundwater recharge volumes, both for Wadis and the entire study area, match well previously published groundwater recharge data (Table 6, Figure 5). The recharge values of the northern Wadis are in a good agreement with data from Eastern Basin analyses and the southernmost Wadi Rahaf corresponds with its considerably lower recharge in comparison to published data for the Northern Negev.

## 4. Modelling Results

The final results show that the hydrological model reproduces the hydrological character of the heterogeneous study area. Using the nested multi-response calibration approach, the model tends to better match the water budget components than a model with a priori values. For the total Western Dead Sea catchment the final model estimates 3.4 mm/a of direct runoff, 270 mm/a of actual evapotranspiration and 36.7 mm/a of groundwater recharge for the period 1977–2010. Table 7 contains the estimated water balance components of each Wadi.

In Figure 9 the spatial patterns of mean annual rainfall, direct surface runoff (Figure 9A) and groundwater recharge (Figure 9B) are shown. The spatial patterns of annual rainfall (Figure 9A) represent the strong rainfall gradient from 750 mm in the mountain range (north-west) to < 40 mm at the Dead Sea (south-east). It is visible that the water balance of the study area is strongly influenced by topographic and rainfall gradient. The spatial distribution of groundwater recharge shows a maximum along the mountain ridge representing the areas with the highest annual rainfall and a moderate slope in the western mountain range. In contrast, the highest potential evapotranspiration rates (ca. 1180 mm/a) are registered along the Dead Sea and accompanied with very low amounts of groundwater recharge. In Figure 10 time series of rainfall and groundwater recharge are shown for the model period 1977–2010. A large variability in groundwater recharge is visible with a 20-100 mm/a range of for 33 hydrological years. The simulation of groundwater recharge indicates that only large rainfall events result in groundwater infiltration. This is typical of arid areas where rainfall distribution is highly heterogeneous. Thus, average values of groundwater recharge are only conditionally suitable for water management in arid regions, since several consecutive years with very low or very high recharge can occur.

During the calibration process it was found that each sub-catchment needs a specific catchment rainfall to generate direct runoff as evidenced by the hydrographs of the Wadi outlets. Rainfall below the critical magnitude can also lead to runoff events, but may not be registered in the hydrograph, since either transmission losses occurred or the discharge was too low to be monitored.

A general relationship is derived between catchment runoff and rainfall based on the analysis of a long time series of simulated and observed runoff data vs. contemporaneous rainfall (Figure 11). Indeed, in the northern catchments Qilt and Darga variable rainfall amounts lead to relative low but constant runoff volumes (< 0.1 $m^3/s$). In contrast, rainfall amounts of >100 mm per month generated higher direct runoff volumes in Tkoa and Arugot. From this threshold onwards the model underestimates runoff volumes. Possible reasons are i) too high soil storage, where water is not provided as direct runoff and ii) significant underestimation of assumed rainfall. However, very small rainfall amounts are sufficient to generate large outflows in the southernmost Wadi Rahaf.

The hydrological model is only in part capable to reflect the runoff - rainfall relationship. For some events, especially low runoff due to mean rainfall, the model is well suited to predict correct runoff volumes. The hydrological model always underestimates observed very high outflow discharges.

## 5. Conclusions

The proposed conceptual hydrological model applied to the Western Dead Sea catchment enables assessment of water balance parameters and groundwater recharge in the arid to hyper-arid catchment, where hydrological and meteorological records are insufficient for direct estimation of infiltration. The estimation of groundwater recharge is a typical challenge in semiarid to arid catchments, where groundwater recharge is the main source of replenishment of water resources (e.g. Mohammadi et al., 2014; Sharda et al., 2006). We show that different data sets were used in a nested multi-response calibration approach to confine model parameters and to reduce parameter uncertainty in an area characterized by data scarcity as done previously (Sophocleous, 1991; Rödiger et al., 2014).

The study show that rainfall is the most sensitive parameter in the hydrological modelling of arid to hyper-arid areas; rainfall uncertainties dominate uncertainties in runoff prediction and related parameters of the water balance. To overcome the scarcity of rainfall data we successfully enhanced the database with mean monthly rainfall data. Tests to fill the missing rainfall data from satellite products show these are unsuitable for the study area.

During the calibration approach we show that a model with a priori parameter values tends to overestimate direct runoff by as much as 10% and to underestimate recharge rates by up to 10%. An increase in the vertical drainage component by the *LVD* parameters gave the required effect. At the same time, the surface discharge declined, especially in the arid and semiarid areas. An obvious limitation of the a priori model for arid and hyper-arid zones has been detected. Therefore, the model calibration parameter *maxInf* was developed for arid and hyper-arid zones. The application limited infiltration rates and led to a decline in modelled groundwater recharge and to a significant increase in surface runoff. This led to a breakthrough in the water balance simulation.

Wadi catchment runoff has been shown to depend on catchment rainfall. In the northern catchments variable rainfall amounts lead to low but constant runoff volumes. In the centre of the Western Dead Sea catchment a rainfall threshold >100 mm per month generates higher direct runoff volumes whereas in the southern Wadis small rainfall amounts (<50 mm per month) are sufficient to generate large outflows.

The ongoing research aims to implement the calculated groundwater recharge as a boundary condition for the transient groundwater flow model (Gräbe et al., 2013). As such, the temporal and spatial distribution of groundwater recharge to the Cretaceous Judea Group aquifer system allows estimating more precisely actual and future water management of this stressed limestone aquifer. The application of the model (calibration parameter data set) to other semi-arid to hyper-arid Wadis results in good agreement between the groundwater recharge for the Darga catchment and those of Tkoa, Qilt and Rahaf.

## 6. Acknowledgments

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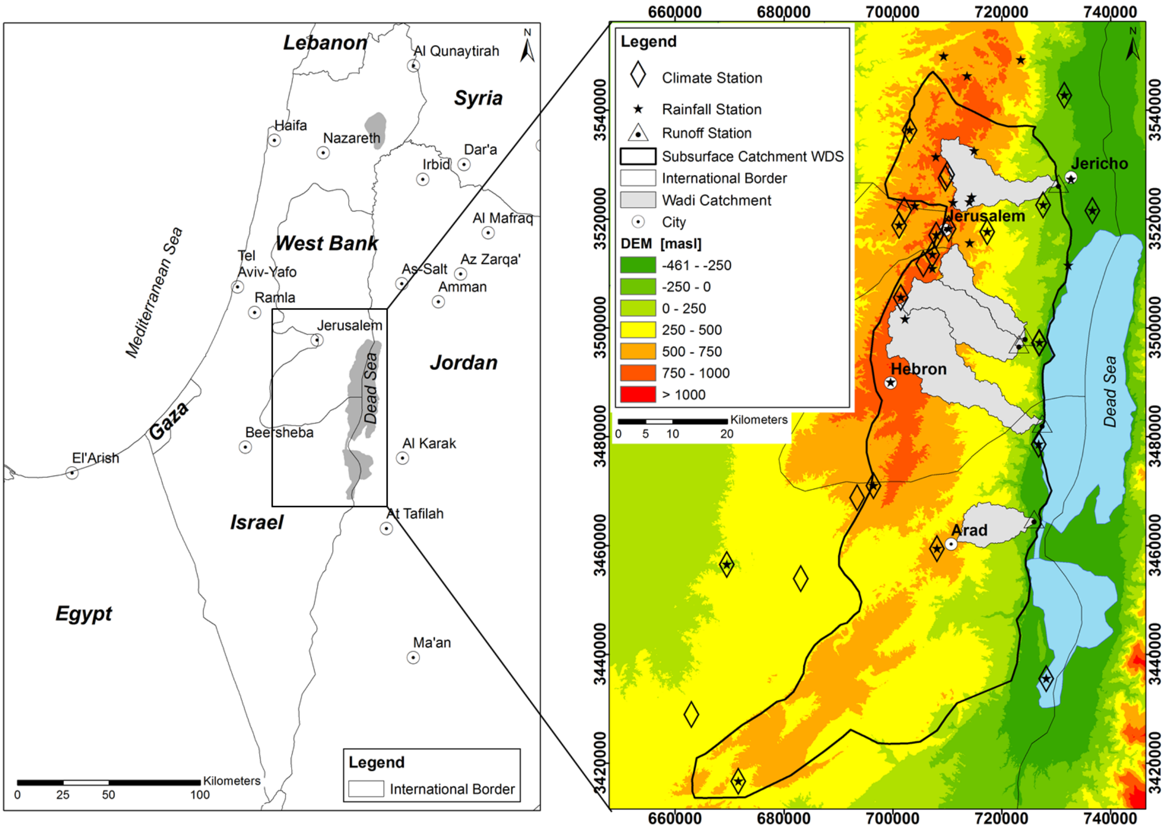

Lebanon
Syria
West Bank
Jordan
Israel
Egypt
Gaza
Mediterranean Sea
Dead Sea
Al Qunaytirah
Haifa
Nazareth
Dar'a
Irbid
Al Mafraq
Tel Aviv-Yafo
Ramla
As-Salt
Az Zarqa'
Amman
Jerusalem
El'Arish
Beersheba
Al Karak
At Tafilah
Ma'an
Kilometers
0 25 50 100
Legend
International Border
Legend
Climate Station
Rainfall Station
Runoff Station
Subsurface Catchment WDS
International Border
Wadi Catchment
City
DEM [masl]
-461 - -250
-250 - 0
0 - 250
250 - 500
500 - 750
750 - 1000
> 1000
Kilometers
0 5 10 20
Jericho
Jerusalem
Hebron
Arad
Dead Sea
660000 680000 700000 720000 740000
3540000 3520000 3500000 3480000 3460000 3440000 3420000

Figure 1

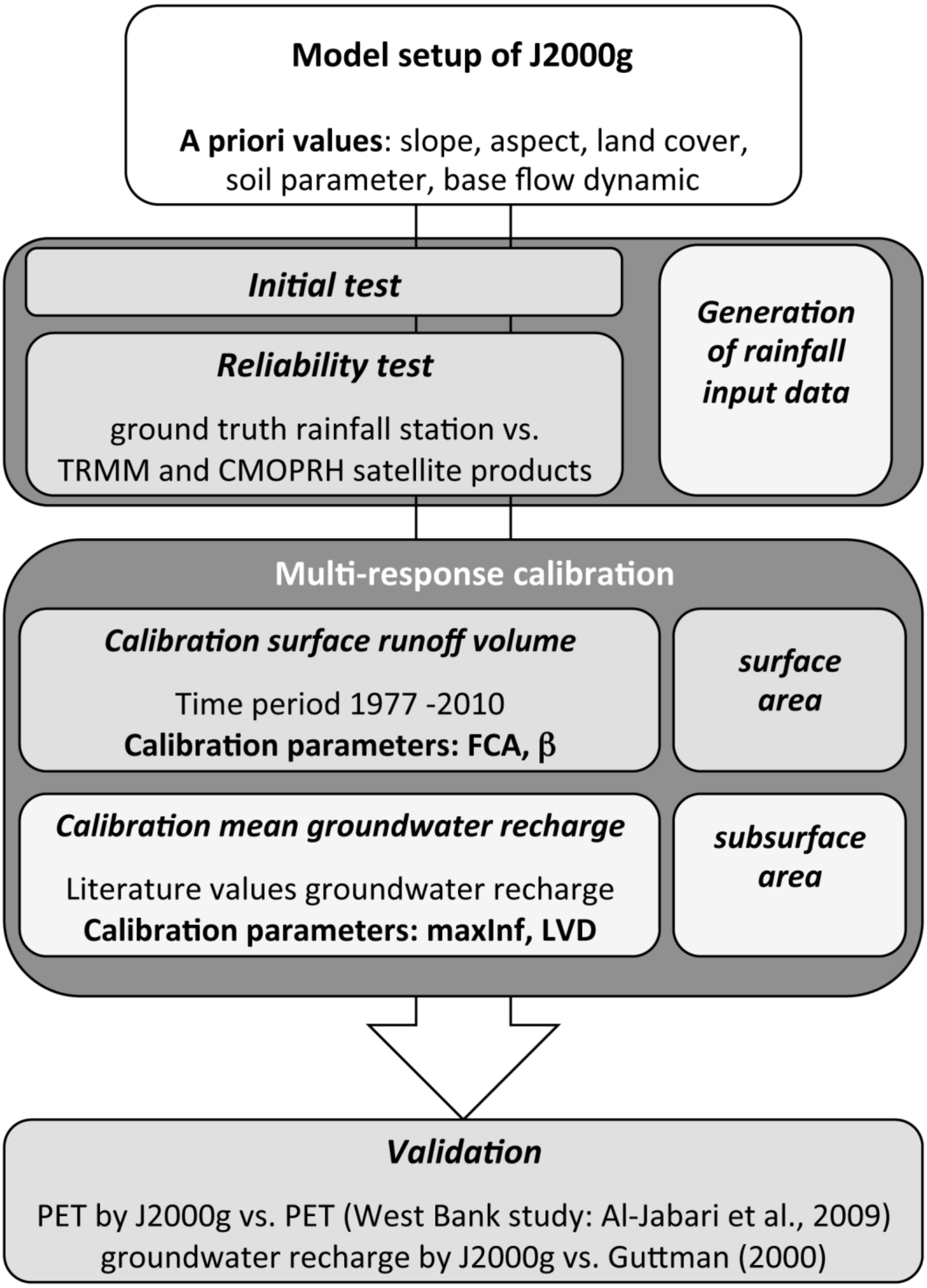

Model setup of J2000g
A priori values: slope, aspect, land cover, soil parameter, base flow dynamic
Initial test
Reliability test
ground truth rainfall station vs. TRMM and CMOPRH satellite products
Generation of rainfall input data
Multi-response calibration
Calibration surface runoff volume
Time period 1977 -2010
Calibration parameters: FCA, β
surface area
Calibration mean groundwater recharge
Literature values groundwater recharge
Calibration parameters: maxInf, LVD
subsurface area
Validation
PET by J2000g vs. PET (West Bank study: Al-Jabari et al., 2009)
groundwater recharge by J2000g vs. Guttman (2000)

Figure 2

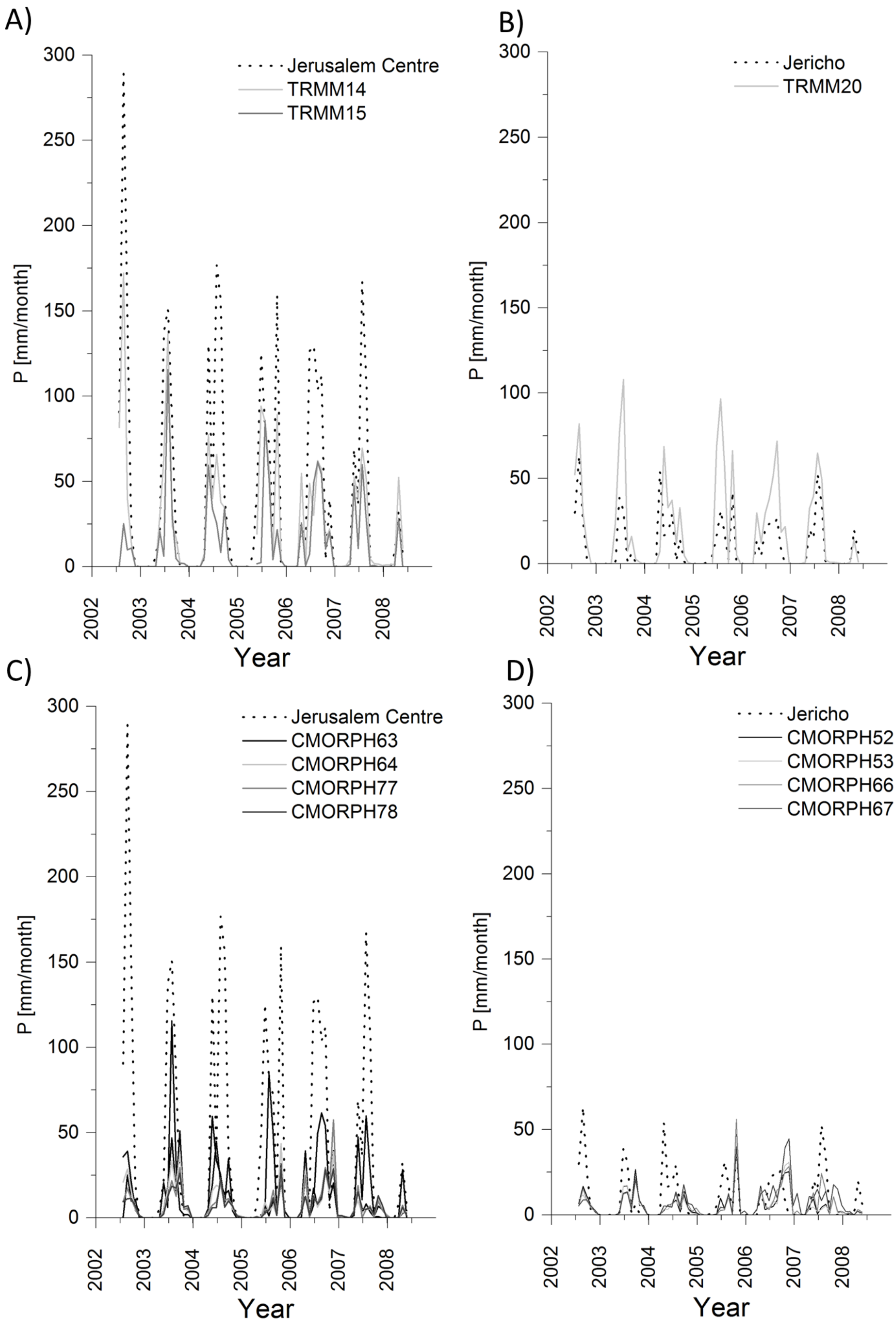

A)
Jerusalem Centre
TRMM14
TRMM15
P [mm/month]
Year
B)
Jericho
TRMM20
P [mm/month]
Year
C)
Jerusalem Centre
CMORPH63
CMORPH64
CMORPH77
CMORPH78
P [mm/month]
Year
D)
Jericho
CMORPH52
CMORPH53
CMORPH66
CMORPH67
P [mm/month]
Year

Figure 3

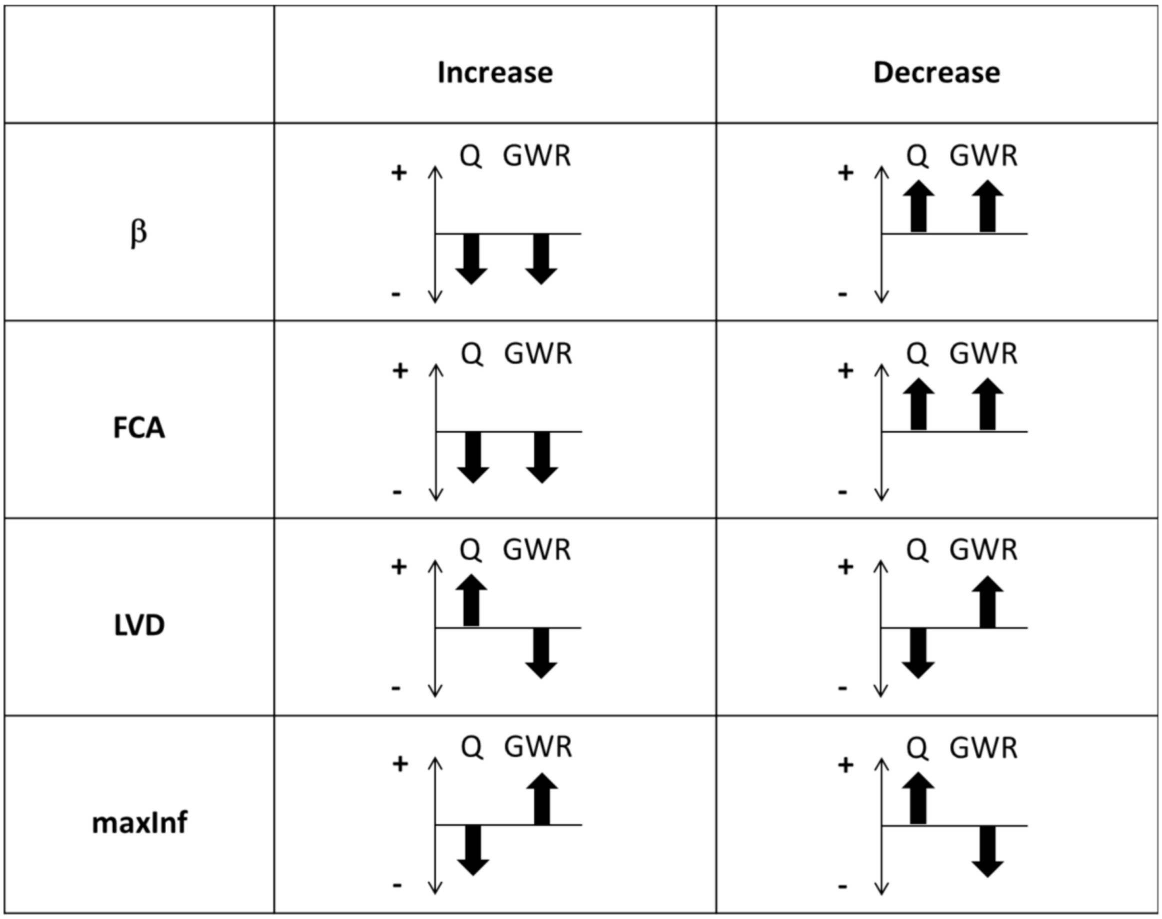
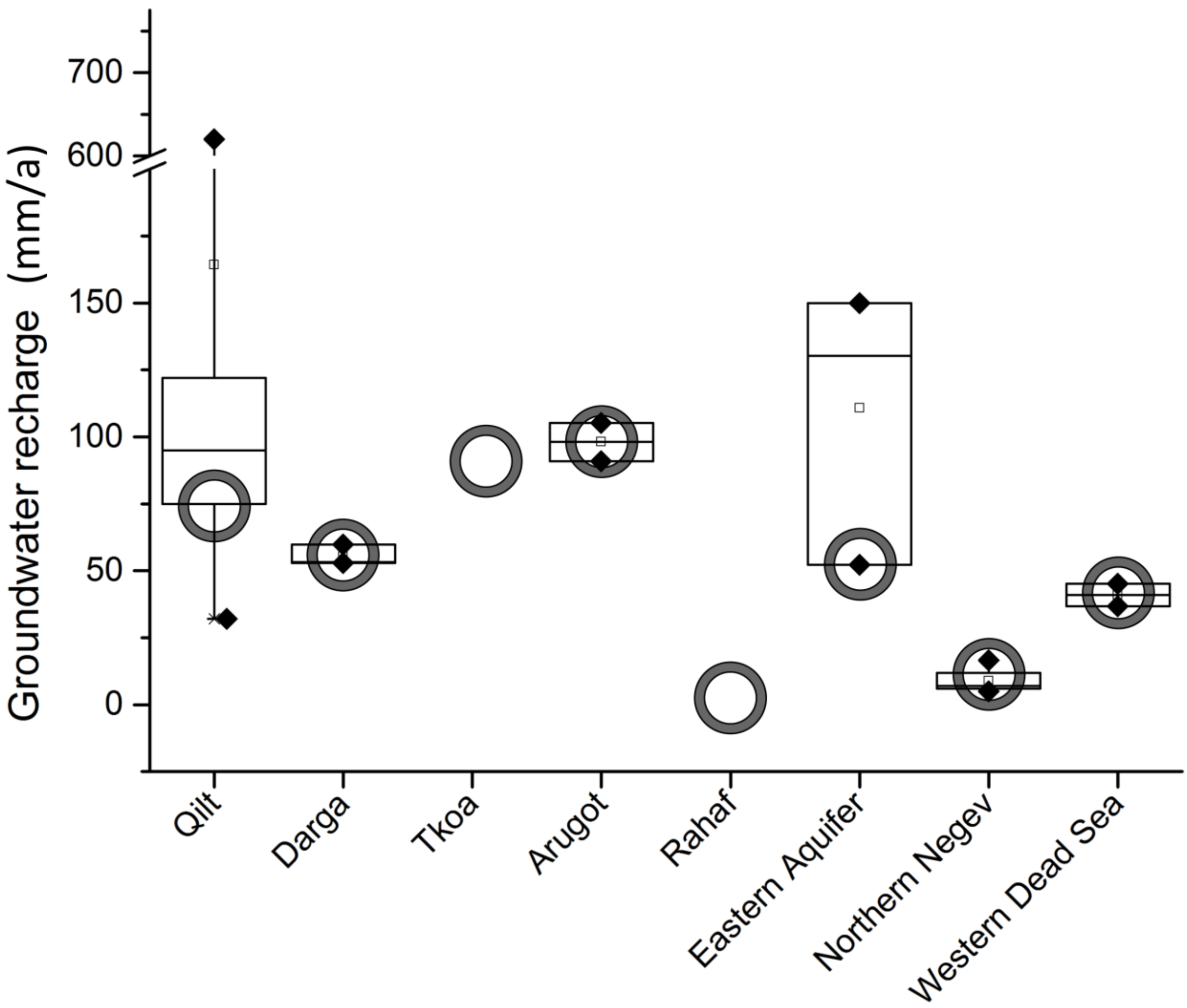



Figure 4



Groundwater recharge (mm/a)
700
600
150
100
50
0
Qilt
Darga
Tkoa
Arugot
Rahaf
Eastern Aquifer
Northern Negev
Western Dead Sea


Figure 5

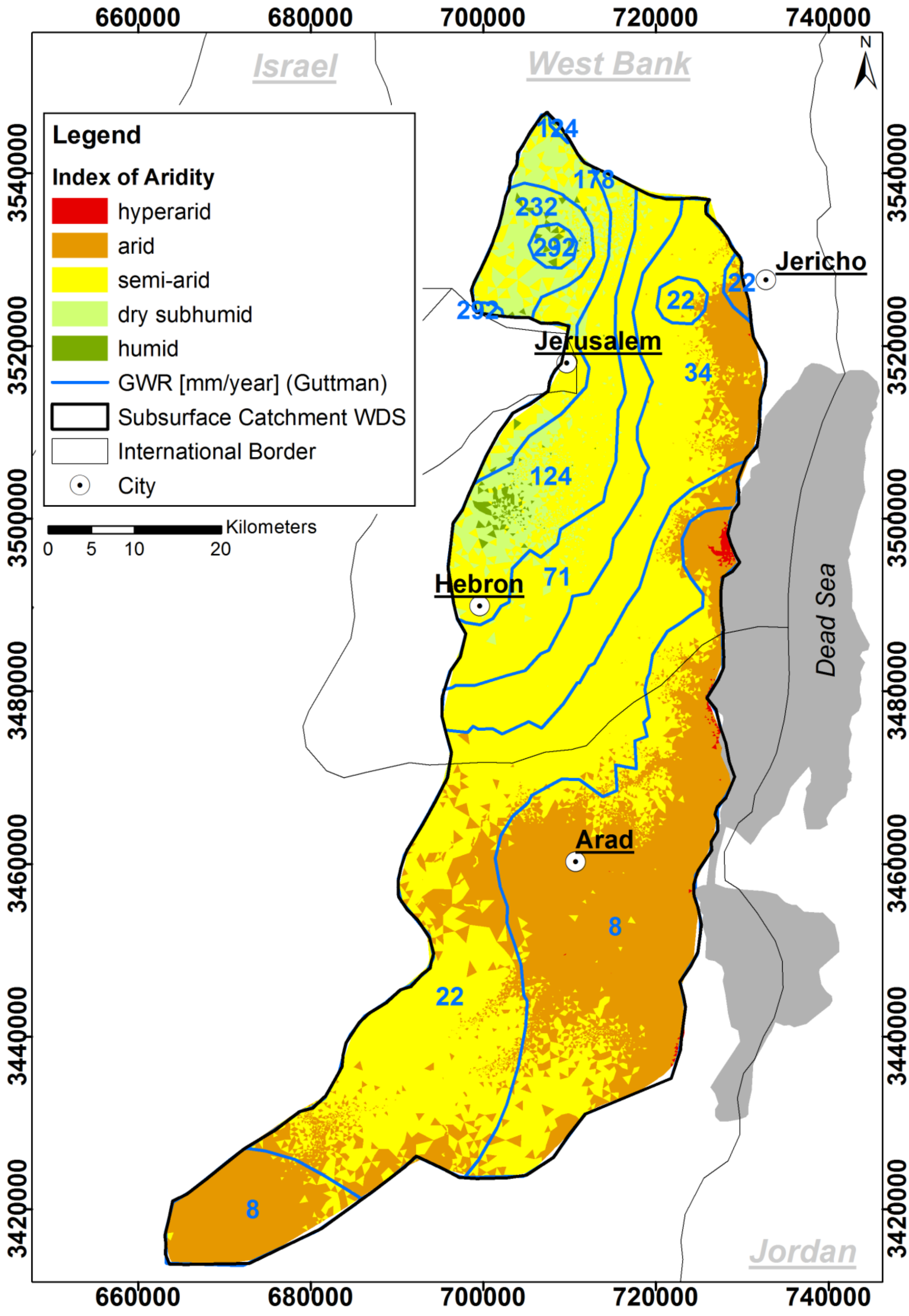

660000
680000
700000
720000
740000
Israel
West Bank
N
Legend
Index of Aridity
hyperarid
arid
semi-arid
dry subhumid
humid
GWR [mm/year] (Guttman)
Subsurface Catchment WDS
International Border
City
Kilometers
0 5 10 20
124
178
232
292
292
22
22
34
124
71
22
8
8
Jericho
Jerusalem
Hebron
Arad
Dead Sea
Jordan
3540000
3520000
3500000
3480000
3460000
3440000
3420000

Figure 6

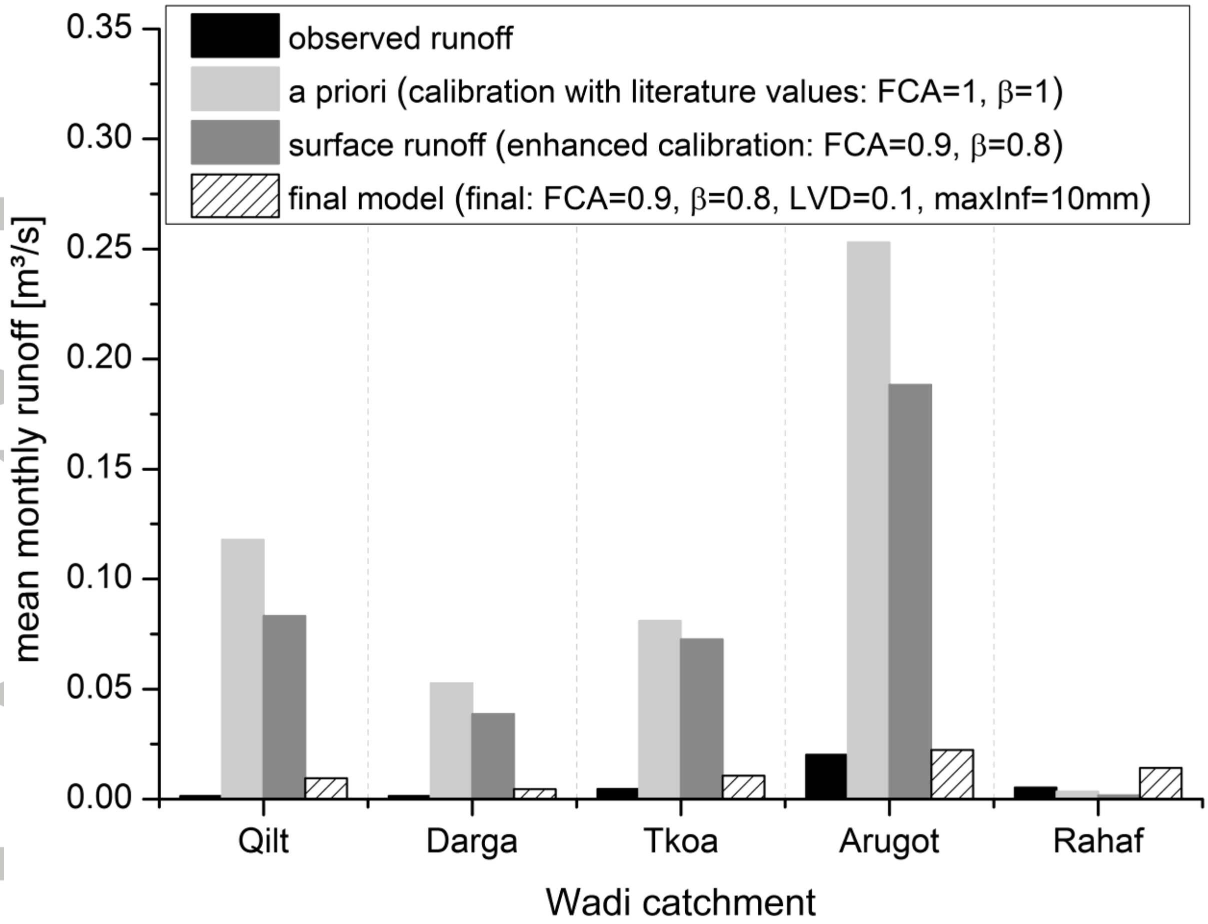
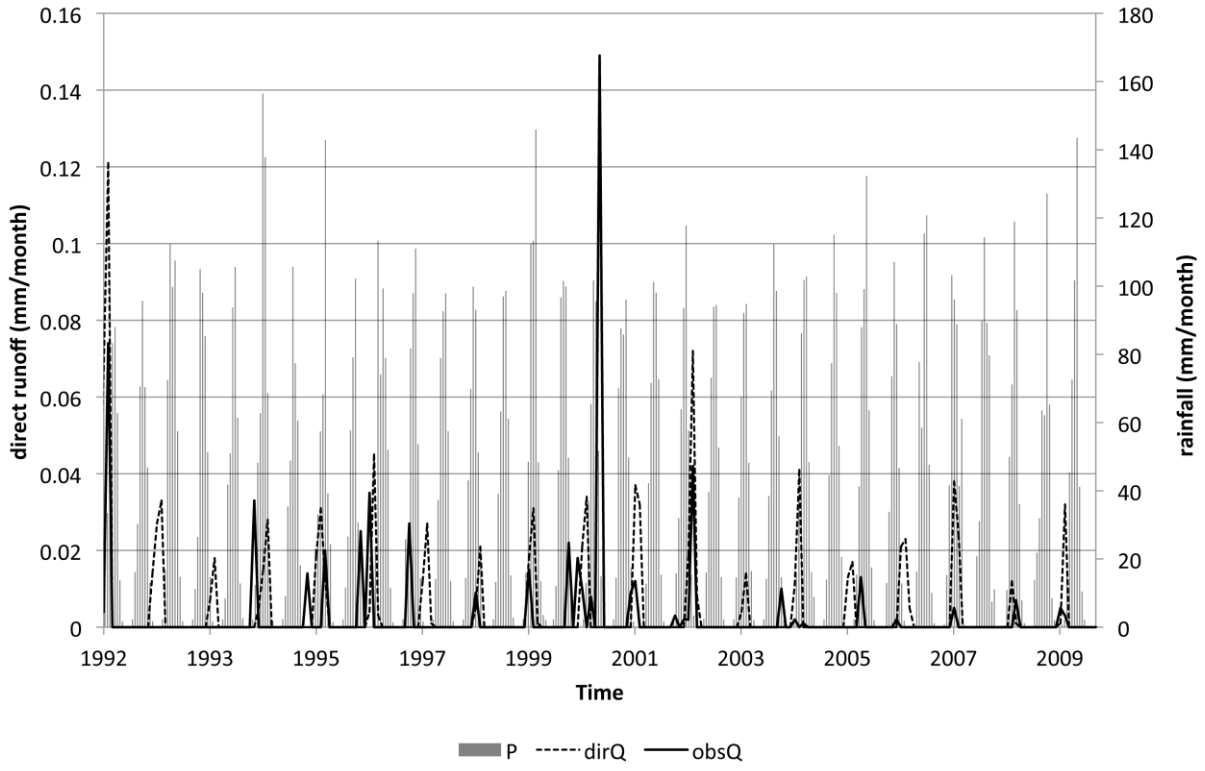

observed runoff
a priori (calibration with literature values: FCA=1, β=1)
surface runoff (enhanced calibration: FCA=0.9, β=0.8)
final model (final: FCA=0.9, β=0.8, LVD=0.1, maxInf=10mm)
mean monthly runoff [m³/s]
0.35
0.30
0.25
0.20
0.15
0.10
0.05
0.00
Qilt
Darga
Tkoa
Arugot
Rahaf
Wadi catchment

Figure 7



0.16
0.14
0.12
0.1
0.08
0.06
0.04
0.02
0
direct runoff (mm/month)
180
160
140
120
100
80
60
40
20
0
rainfall (mm/month)
1992
1993
1995
1997
1999
2001
2003
2005
2007
2009
Time
P
dirQ
obsQ


Figure 8

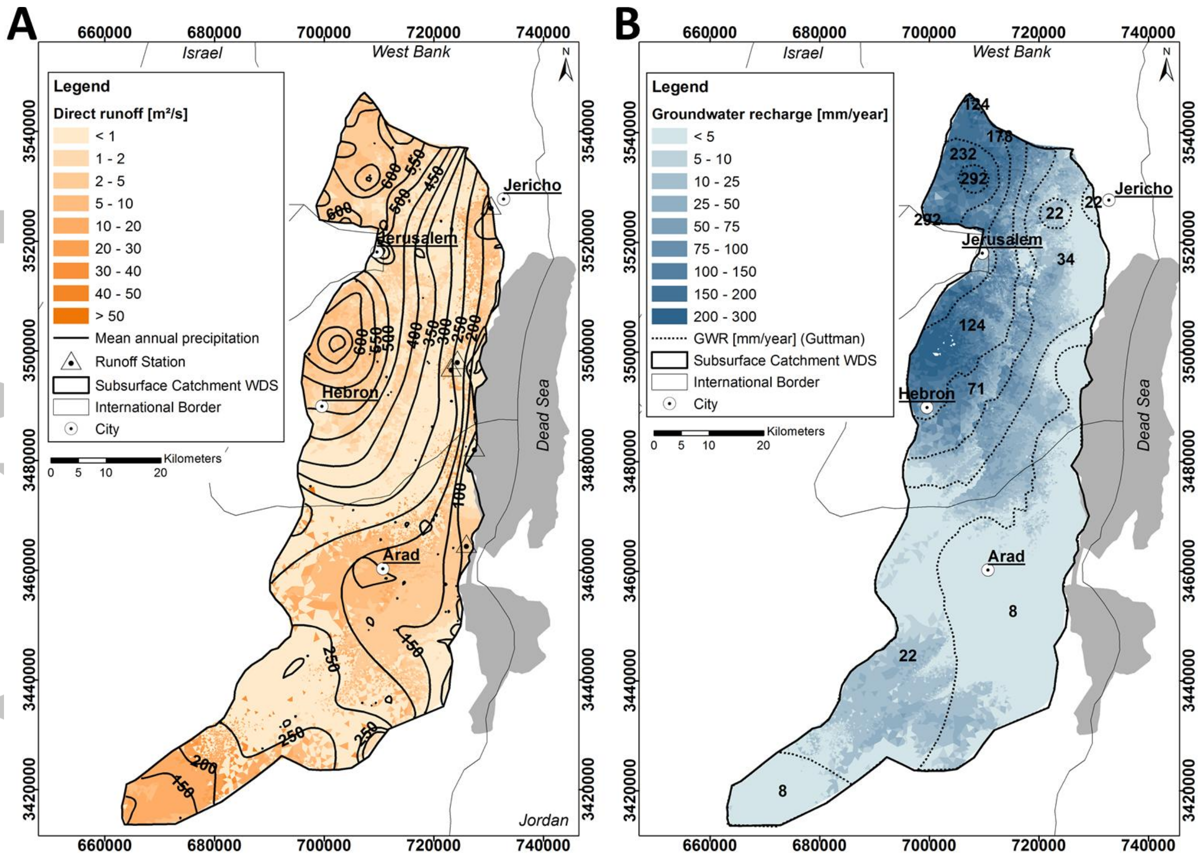
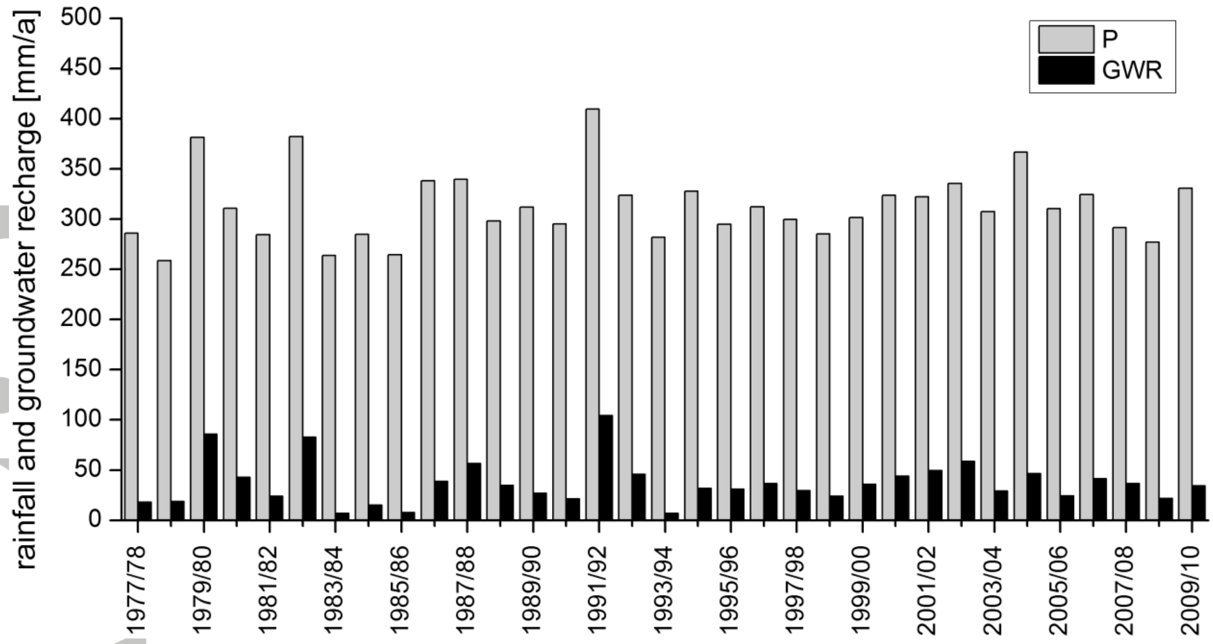

A
Legend
Direct runoff [m³/s]
< 1
1 - 2
2 - 5
5 - 10
10 - 20
20 - 30
30 - 40
40 - 50
> 50
Mean annual precipitation
Runoff Station
Subsurface Catchment WDS
International Border
City
Kilometers
Israel
West Bank
Jericho
Jerusalem
Hebron
Arad
Dead Sea
Jordan
B
Legend
Groundwater recharge [mm/year]
< 5
5 - 10
10 - 25
25 - 50
50 - 75
75 - 100
100 - 150
150 - 200
200 - 300
GWR [mm/year] (Guttman)
Subsurface Catchment WDS
International Border
City
Kilometers

Figure 9



rainfall and groundwater recharge [mm/a]
500
450
400
350
300
250
200
150
100
50
0
P
GWR
1977/78
1979/80
1981/82
1983/84
1985/86
1987/88
1989/90
1991/92
1993/94
1995/96
1997/98
1999/00
2001/02
2003/04
2005/06
2007/08
2009/10


Figure 10

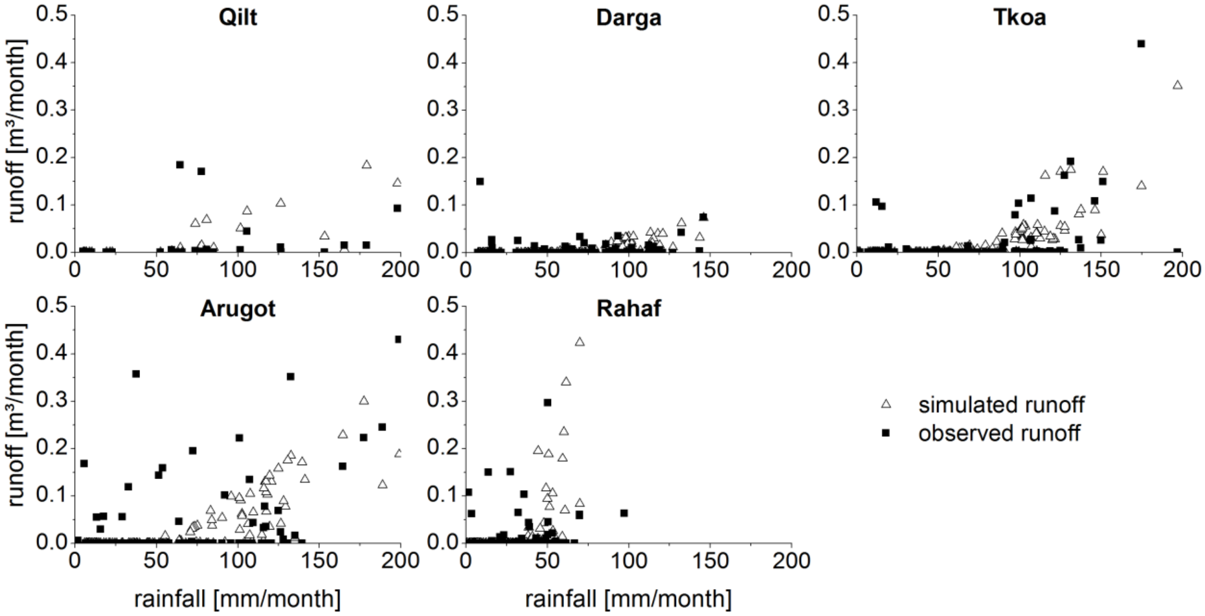

Qilt
Darga
Tkoa
Arugot
Rahaf
runoff [m³/month]
rainfall [mm/month]
simulated runoff
observed runoff

Figure 11