Evaluation of floodwater spreading for groundwater recharge in Gareh Bygone Plain, southern Iran

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Dedicated to the messenger of

faith, peace and justice for all human kind

Imam Aba Saleh Al-Mahdi (PUH)

Remembering my late Father Aziz

my first and ever teacher
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Evaluatie van de spreiding van de afstroming na hevige regenbuien op de aanvulling van het grondwater in de Gareh Bygone Plain in het zuiden van Iran

Cover illustration:
The top photo illustrates the ponded infiltration basin of the floodwater spreading system at Gareh Bygone, Iran after the flooding event of 08/04/2013. The bottom photo shows the same location before implementing the system in 1981.

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In the name of Allah, most benevolent, ever-merciful

He (Allah) sends down from the sky, rain, and valleys flow according to their capacity, and the torrent carries rising foam. And from that [ore] which they heat in the fire, desiring adornments and utensils, is foam like it. Thus Allah presents [the example of] truth and falsehood. As for the foam, it vanishes, [being] cast off; but as for that which benefits the people, it remains on the earth. Thus does Allah present parables.

Holly Quran - 13: 17
Summary

Water scarcity due to climate change and a growing water demand in different consumption sectors is a major environmental crisis that drives arable lands to the state of degradation, especially in dry regions. Artificial recharge of groundwater (ARG) through floodwater spreading (FWS) which is a potential measure for reversing this emerging trend is investigated in this thesis.

The overall objective of this dissertation is to evaluate a floodwater spreading system that was installed in 1981 at the Gareh Bygone Plain, southern Iran for recharging the groundwater table. The main research objectives are i) to evaluate spatial and temporal changes in both evapotranspiration ($ET$) and crop coefficients ($K_c$) based on a calibrated surface energy balance system (SEBS) model, ii) to assess the recharge of the groundwater by the floodwater spreading system using a combination of two types of saturated zone methods, water table fluctuation (WTF) and water budget, and iii) to assess the recharge by the same system using a vadose modelling approach based on soil water content measured over the vadose zone using probes that were particularly calibrated for the stony layers of interest.

The surface energy balance system (SEBS) model was used to estimate actual $ET$ using non-cloudy images of Landsat 5 TM from May 2009 to October 2010 for 32 dates. Calibration through improved parameterization of SEBS for selecting the appropriate sources and methods of acquiring the parameters was performed by maximizing the fitness between the estimated $ET_a$ and the reference $ET_r$ for the water reservoir pixels. A cross validation of seasonal $ET_a$ showed minor differences between modelled and water budget $ET_a$ for wheat and forage corn crops, but major discrepancies for the pastures outside the FWS (PO) and bare soils were observed. Due to a lack of reliable data on water consumption for the pastures inside the FWS systems (PI) and tree plantations (TP), the estimated $ET_a$ could not be compared with the water budget $ET_a$ and evaluated by means of $K_c$. The mid-season $K_c$ obtained for main crops, TP compared well to the published $K_c$ under similar conditions; however, PI and PO $K_c$ showed higher values than expected. These findings were used in further parts of the dissertation to evaluate the floodwater spreading for recharging the groundwater table in which $ET$ data are needed to define the upper boundary conditions of the system.
In a first ‘saturated zone’ approach, the effect of flooding events on groundwater recharge in the Gareh Bygone Plain by using water table fluctuation (WTF) and water budget methods. Fluctuations in groundwater table depth were examined monthly during 1993-2012 in six observation wells installed inside (N=3) and around (N=3) the FWS systems. The depth of rainfall and the volume of diverted floodwater into the FWS systems were also measured. Three experimental wells were hand drilled and the typical layers were sampled for physical characterization needed to determine specific yield, i.e. the ratio of the volume of water that can be drained after saturation by gravity to its own volume. The overall hydrograph of the whole Gareh Bygone Plain showed a long term descending trend with an overall water level drop of six meters since the start of the recordings in 1993. However, the observation wells located inside the FWS systems revealed a greater resistance to dry periods and extractions than the other wells in the area. The hydrograph displayed a substantial disparity of water level rise with the other wells, particularly in two major floods in 2004 and 2005 (water level rise of 0.205 m inside FWS systems vs. 0.50 m outside FWS systems). A water budget calculated for the hydrological year 2010-2011 (October 2011 to September 2012), the only period that the needed full data set was available, showed a depletion of 4.13 million m$^3$ (Mm$^3$) from the aquifer storage during this period and the return flow as 3.2 Mm$^3$. The recharge was calculated at 7.94 Mm$^3$, which was a consequence of both artificial recharge and natural replenishment. The artificial recharge data in the same period during the flooding events from 28 January to 2 February 2011 show a total volume of 6.92 Mm$^3$ of retained flood water in the FWS systems. Artificial recharge from these events, which ponded an average depth of 0.34 m on the system surface, was calculated as 0.24 cm. The artificial recharge was calculated by the two methods (flow data and water budget, respectively) as 4.84 and 4.46 therefore, 56 to 61% of the recharge could be assigned to the impact of the FWS systems for this hydrological year.

In a second ‘unsaturated zone’ approach, recharge of the groundwater table was evaluated by measuring and modelling water content within the vadose zone inside the FWS systems. Three wells were dug in a 32 year old recharge basin of the system, seven layers were identified above the water table (at 28.6 to 31.5 m depth) and their hydraulic properties were determined. In one of the wells, which was insulated with concrete rings, TDR probes were placed at
0.3, 1.0 and 2.0 m depth intervals for depths of 0-3, 3-10, and 10-28 m, respectively and maximum care was taken not to disturb the soil while insulating. New calibration curves for measuring water content from bulk dielectric properties with Time Domain Reflectometry (TDR) probes were established and tested for the stony materials of interest prior to insertion to the well. Since probes were installed very deep into the profile (~30 m), the effect of cable length was tested as well. The equations provided in this study for converting the measured dielectric permittivity \(K_a\) to volumetric water content \(\theta_v\) and for compensating the effect of cable length resulted in improving the reliability of measured \(\theta_v\) by the TDR. The RMSE values were improved from 0.04, 0.03 and 0.02 to 0.0024, 0.0019 and 0.0015 \((m^3 \ m^{-3})\) respectively for the three soil types due to applying the new equations as compared to traditional TDR equation.

With the calibrated probes, \(\theta_v\) was measured weekly from August 2010 to December 2013 before flood occurrence, twice a day for 30 days after recharge events, and daily thereafter for 60 days. Rainfall depth, ponding height and duration of ponding were also recorded. The soil-water budget (SWB) method and the Hydrus-1D (H1D) model that simulates water transport in variously saturated porous media were employed to evaluate recharge. The results show reliable and consistent readings of \(\theta_v\) for each soil layer before the recharge events. A stepwise increase in \(\theta_v\) in the various layers was observed as influenced by flooding event to a depth of 4.0 m. Calibration of the H1D model by inverse solution resulted in \(r^2\) values of simulated vs. observed \(\theta_v\) of 0.94 to 0.96 and RMSE of 0.02 to 0.05 \((m^3 \ m^{-3})\) for different subsurface layers. Calculations using the SWB and H1D methods, indicated that out the 51.8 cm of rainfall and ponded floodwater added to the site during the 16 January to 23 August 2011 period, 29.6 cm of cumulative flux (recharge) occurred, showing an efficiency of 57%. The calibration results of the H1D model, which includes the optimized hydraulic parameters of the representative layers in aquifer profile, can be applied in future studies at this research site when attempting to up-scale our findings.

To recapitulate briefly, it might be concluded that, notwithstanding the decline in groundwater table observed in recent decades in the Gareh Bygone Plain, the FWS systems that was installed there in 1981 seems to be effective in recharging the groundwater table. Two independent approaches suggest that 57 to 61% of rainfall effectively flows to the groundwater table.
چکیده:

با آن که مجموعه برنامه‌های آبخوان‌داری از طریق یخبندان سیلاب در گربه‌گرانگیچی فسا موجب تبدیل یک محيط بی‌بندانی به یک سردانی آب‌دانگی گسترش و پوشش سی و جنگلی و مرتعی شده است، اما اثر یخبندان در منابع آب زیرزمینی دشت بناب به شواهدی مورد تردید بخری قرار دارد. سپس رو به کاهش سطح آب زیرزمینی کاهش تراویه سطح خاک شهری آب سیلاب، نیاز آب درختان ترمیم در داخل سامانه‌ی سیلاب سیلاب از جمله این شواهد است. تحقیق حاضر با هدف یافتن پاسخ‌هایی کمی به اهمیت باد شده.

در گام نخست برای دست‌یابی به میزان تبخیر-تعرق واقعی سالانه، یکی از شیوه‌های مدل‌های تاز گرمایه‌ای ابتدا مورد استفاده و تحقیق قرار گرفته و سپس با کمک داده های سنجش از دور، نقشه‌های تبخیر-تعرق مربوط به زمان‌های پاییز در طول یک سال آب تهیه شد. به این ترتیب پراکندگی زمین و مکانی کشیده آب کاربری‌های مختلف موجود در دشت گربه‌گرانگی به دست آمد. شبهی‌اندازی تغییرات SEBS و در روند تحقیق اصلاحات در معرفی فراستج (باران‌متر) ها نیز یکی از معادلات آن اعمال شد تا نتایج آن به حداکثر تطبیق با واقعیت برسد. نتایج نشان دادن در سال‌های 1389-1391 میلادی 10.9 میلیون متر مکعب (متر مکعب) آب توسط گیاهان مهی‌زراعی و 1 میلیون متر مکعب گیاهان گنجگی از آب زیرزمینی دشت گربه‌گرانگی به مصرف رسیده است. در گام بعدی برای یافتن مهم تغییر می‌باشد که اثر یخبندان سیلاب کلیه اجزای تراز آب شناسی کل دشت در یک سال آب تعبین شد. این کار بر اساس تلفیق روش‌های تاز تاز آب (water budget) با روش تغییرات سطح آب زیرزمینی (water table fluctuation) داده‌های آب ورودی شامل پاشش، بارش خشک‌شده و نیز داده‌های خروجی شامل گزارش و تغییرات سطح آب زیرزمینی همگی انداده‌گری هستند. با اعمال ضوابطی مدل‌سازی که معادله بیان که مقدار تغییره است تعبین شود. داده‌های بارش از ایستگاه هواشناسی، داده‌های سیل با انداده‌گری مستقیم مقدار ورودی به شکه‌ها، داده‌های رویای با انداده‌گری مقدار خروجی از شکه‌ها، داده‌های آبیاری با انداده‌گری به مقدار آب ورودی به مزرعه نمونه، داده‌های سطح آب زیرزمینی با استفاده از آمار جاهای مشاهده‌ی 6 گاهی موجود در دشت به دست آمدند. نتایج نشان دادن با آن که روند تغییرات سطح آب زیرزمینی در سال‌های اخیر تقلید به داشته است، با این حال پس از هر واقعه سیل افزایش قابل توجهی در ذخایر آب سفره زیرزمینی رخ داده، و بیشترین تأثیر در سفره‌های مجاور سامانه یخبندان سیلاب بوده است.

در یک سال آبی 1391-1390 از 14.83 میلیون متر مکعب آب که از سفره گردده شده تنهای سه برابر مقدار آب بگونشی کشوارژی به مقدار 3.2 میلیون متر مکعب، حدود کاهش حجم آب در سفره گردده است. با توجه به مقدار آب بگونشی کشوارژی به مقدار 4.13 میلیون متر مکعب، حدود
پخش سیلاب و بنابراین نسبت تغذیه مصنوعی به تغذیه طبیعی حدود 57% تا 61% بوده است.

در گام پایانی برای اندازه‌گیری دقیق تر میزان نفوذ آب بر اثر پخش سیلاب تحقیقاتی بر اساس روش بیلان طراحی شد. به این منظور 3 حلقه چاه در یکی از شهرهای پخش سیلاب با عمق 30 متر حفر و ضمن برداشت نمونه‌های آب از تمام لایه‌ها و یزدگی‌های هیدرولیکی خاکها در محل و در آزمایشگاه اندازه‌گیری شد. سپس دستگاه TDR واسنجی شد تا ضریب صحیح آن برای اندازه‌گیری رطوبت خاک‌های مورد بررسی تعیین گردد. یکی از جوانه‌های تجربیات اندازه‌گیری رطوبت لایه‌ها با روش TDR مجزه شد. به این منظور ابتدا بدن‌های چاه با جداره سیمانی عابق بنده و سپس حسگرهای TDR دستگاه در لایه‌های با فواصل 30 سانتی‌متری تا عمق 30 متری کار گذشتند. از مرداد ماه 1389 تا کنون مقادیر رطوبت لایه‌ها بطور دایر اندازه‌گیری شده است. داده‌های وقوع سیل و بارش به همراه داده‌های رطوبت لایه‌ها و تبخیر - تعرق با روش بیلان آب خاک مورد تحلیل قرار گرفت. همچنین با کاربرد نرم‌افزار های‌دورس یک بعدی (Hydrus1D) و پس از واسنجی آن، حرکت آب در خاک شیبی و سایر نتایج نشان داد که در واقعیت سیل بهمین 1389 برحسب 51.8 سانتی‌متر آب بارش و سیل مقدار 29.6 سانتی‌متر به اعماق لایه‌های زیرین نفوذ کرده و این امر در طی 7 ماه اتفاق افتاده است که 57% تغذیه پر اثر نفوذ را گواهی می‌دهد. بنابراین با جمع بندی دو روش بررسی یاد شده، گزارش سامانه پخش سیلاب در تغذیه

سفره آب زیرزمینی 57 تا 61 درصد می‌باشد.

نتایج نشان از این دارد که سامانه پخش سیلاب در تغذیه سفره آب زیرزمینی پس از 32 سال کارکرد و رسوب گیری از کارایی مطلوب برخوردار است. همچنین میزان مصرف پوشش درختی، که عمداً کالیپسوس کامالیوستیس می‌باشد در مقایسه با مصرف آب گیاهی نقش پیشروی نقش پیشروی گیاهی کچک‌تری در برداشت آب در دشت گربه‌باغ دارد.
List of symbols

Chapter 2

Roman symbols

\( at\text{-surf} \) \quad at surface (earth skin) reflectance
\( A_{cm} \) \quad adjustment coefficient for the, \( K_c \)
\( C_p \) \quad specific heat of the air, kJ kg\(^{-1}\) K\(^{-1}\)
\( d \) \quad the earth–sun distance in astronomical units
\( d_o \) \quad zero plane displacement height, m
\( d_c \) \quad geometrical distribution of the surface and the internal reflections
\( e_a \) \quad actual vapour pressure, hPa
\( e(i) \) \quad residual term (measured-predicted) at each observation number \( i \)
\( e_s \) \quad saturated vapour pressure, hPa
\( ESUN \) \quad mean solar exo-atmospheric irradiances, W/(m\(^2\) sr \( \mu \)m)
\( ET \) \quad evapotranspiration, mm day\(^{-1}\)
\( EF_r \) \quad relative evaporation
\( EF \) \quad evaporative fraction
\( ET_{a} \) \quad actual evapotranspiration, mm day\(^{-1}\)
\( ET_{o} \) \quad grass reference evapotranspiration, mm day\(^{-1}\)
\( f_v \) \quad fraction of ground covered by vegetation
\( G \) \quad soil heat flux, W m\(^{-2}\)
\( \bar{G} \) \quad soil heat flux for 24 hours, W m\(^{-2}\)
\( h \) \quad vegetation height, m
\( H \) \quad sensible heat flux, W m\(^{-2}\)
\( H_0 \) \quad reference value of sensible heat flux, W m\(^{-2}\)
\( h_{max} \) \quad maximum vegetation height, L
\( h_{min} \) \quad minimum vegetation height, L
$H_{dry}$  dry limit of sensible heat, W m\(^{-2}\)

$H_{wet}$  wet limit of sensible heat, W m\(^{-2}\)

$K_i$  incoming shortwave radiation, W m\(^{-2}\)

$K_c$  crop coefficient in $ET$ calculations

$K_c\ adj$  adjusted $K_c$

$K_{c\ mid}$  mid season $K_c$

$K_{c\ end}$  end season $K_c$

$K_{c\ ini}$  initial $K_c$

$K_s$  stress coefficient in $ET$ calculations

$kb^{-1}$  parameter that relates the roughness lengths of heat and momentum transfer

$LAI$  leaf area index, m\(^2\) m\(^{-2}\)

$LAI_{stress}$  LAI of the vegetation under stress, m\(^2\) m\(^{-2}\)

$LAI_{dense}$  LAI of dense vegetation, m\(^2\) m\(^{-2}\)

$LST$  Land surface temperature

$L_A$  spectral radiance at the sensor’s aperture, W/(m\(^2\) sr \(\mu m\))

$NDVI$  normalized deference vegetation index

$NDVI_{soil}$  NDVI of bare soil surface

$NDVI_{veg}$  NDVI of vegetated surface

$O_s$  solar zenith angle in degrees (90° minus Sun elevation)

$P$  atmospheric pressure in SEBS model, hPa

$P_v$  vegetation proportion

$q$  specific humidity, kg kg\(^{-1}\)

$r_e$  external or aerodynamic resistance

$r_i$  internal or surface resistance

$r^2$  coefficient of determination

$r.sun$  module inside GRASS software for global radiation mapping
**List of symbols - v**

\( RF \)  
return flow, L

\( RH \)  
relative humidity, %

\( Ri \)  
unit less planetary reflectance

\( Rt \)  
net radiation, W m\(^{-2}\)

\( \overline{Rn} \)  
net radiation for 24 hours for daily \( ET_a \) calculation, W m\(^{-2}\)

\( S_i \)  
sensitivity value in sensitivity analysis of the parameter \( i \)

\( S_{i} \)  
simulated value at time \( i \) in \( r^2 \) calculation

\( t \)  
observation number in SEE and SEP calculation

\( T \)  
mean daily air temperature, °C

\( T_a \)  
air temperature, °C

\( ToA \)  
top of atmosphere reflectance

\( T_{rad} \)  
thermal band ToA product

\( T_s \)  
surface temperature, °C

\( u \)  
wind speed, m s\(^{-1}\)

\( u_{adj} \)  
Adjusted wind speed, m s\(^{-1}\)

\( u_{inst} \)  
instantaneous wind speed at the time of satellite overpass, m s\(^{-1}\)

\( u_{max} \)  
maximum daily wind speed, m s\(^{-1}\)

\( z_{om} \)  
roughness height for momentum, m

**Greek symbols**

\( \alpha \)  
broadband albedo in \( ET \) calculation by remote sensing

\( \Delta \)  
slope of vapor pressure curve, kPa °C\(^{-1}\)

\( \varepsilon \)  
broadband emissivity

\( \varepsilon_s \)  
soil emissivity

\( \varepsilon_v \)  
emissivity of the vegetation

\( \gamma \)  
psychrometric constant, kPa °C\(^{-1}\)

\( \lambda \)  
latent heat of vaporization, MJ kg\(^{-1}\)

\( \lambda E \)  
latent heat flux, W m\(^{-2}\)
\( \lambda E_{dry} \)  
latent heat flux at dry limit, W m\(^{-2}\)

\( \lambda E_{wet} \)  
latent heat flux at wet limit, W m\(^{-2}\)

\( \rho \)  
mean air density at constant pressure, kg m\(^{-3}\)

\( \rho_r \)  
reflectivity values in the red band (0.63-0.68 \( \mu \)m)

\( \rho_w \)  
density of water, \( \sim \)1000 kg m\(^{-3}\)

**Chapter 3**

*Roman symbols*

\( A \)  
area, L\(^2\)

\( AR \)  
artificial recharge in water budget, L

\( C_s \)  
concentration of suspended material when infiltrating, g L\(^{-1}\)

\( C_w \)  
discharge coefficient in flow calculation, m\(^{0.5}\) s\(^{-1}\)

\( D \)  
deepth of water applied for recharge, m

\( DF \)  
degree of freedom

\( e \)  
open water evaporation during the period of infiltration, mm day\(^{-1}\)

\( E \)  
extration from the aquifer, L

\( ET^{gw} \)  
ET from groundwater, L

\( F \)  
flooding height, L

\( FC \)  
field capacity

\( h_w \)  
hydraulic head in flow calculation, m

\( I \)  
irrigation, L

\( L \)  
weir length in flow calculation, m

\( MP \)  
measuring point in observation wells

\( MS \)  
mean of squares

\( n \)  
Gauckler–Manning coefficient in Manning’s formula, (s m\(^{-1/3}\))

\( NR \)  
natural recharge in water budget, L

\( OW \)  
observation well
List of symbols

\(P\) precipitation, L
\(P_{\text{eff}}\) effective precipitation, L
\(Q\) flow rate, m\(^3\) s\(^{-1}\)
\(Q_{\text{off}}\) subsurface flow from the groundwater, L T\(^{-1}\)
\(Q_{\text{on}}\) subsurface flow to the groundwater, L T\(^{-1}\)
\(Ro\) runoff or runon, L
\(R\) recharge, L
\(RF\) return flow, L
\(R_h\) hydraulic radius in Manning’s formula, m
\(RL\) representative layers
\(S_h\) slop of the hydraulic grade line in Manning’s formula, m m\(^{-1}\)
\(SS\) sum of squares
\(S_y\) specific yield
\(V\) volume, L\(^3\)
\(V\) velocity in flow calculation, m s\(^{-1}\)
\(V_r\) volume fraction of stones in stony soils in Bauwer equation
\(WB\) water budget
\(WL\) water level
\(WT\) water table
\(WTF\) water table fluctuation

Greek symbols

\(\Delta S\) change in soil-water storage, L
\(\Delta h\) change in water table elevation, L
\(\varepsilon_R\) function of the errors in water budget, L
\(\theta_{FC}\) water content at field capacity, m\(^3\) m\(^{-3}\)
\(\theta_r\) residual water content, m\(^3\) m\(^{-3}\)
\(\theta_s\) Saturated water content, m\(^3\) m\(^{-3}\)
List of symbols

\[ \theta_v \] volumetric water content, \( m^3 \text{ m}^{-3} \)

\[ \sum ET \] cumulative \( ET \), mm

Chapter 4

Roman symbols

\[ GHz \] giga hertz, \( 10^9 \text{ s}^{-1} \)

\[ K_a \] dielectric permittivity

\[ K_{aa} \] dielectric permittivity of air

\[ K_{ab} \] dielectric permittivity of bulk soil

\[ K_{ai} \] dielectric permittivity of the component \( i \) in stony soils

\[ K_{as} \] dielectric permittivity of fine soil (<2 mm particles)

\[ K_{ast} \] dielectric permittivity of stones (>2 mm particles)

\[ K_{aw} \] dielectric permittivity of water

\[ MHz \] mega hertz, \( 10^6 \text{ s}^{-1} \)

\[ V_i \] fraction of the \( i \) component in stony soils

\[ V_s \] fraction of fine particles in stony soils (<2 mm)

\[ V_{st} \] fraction of stone particles in stony soils (>2 mm)

\[ v_j \] weighting coefficient in objective function

\[ V_r \] volume fraction of stones in stony soils in Bauwer equation

Greek symbols

\[ \beta \] empirical constant summarizing the geometry of the medium in stony soil

\[ \theta_b \] water content of bulk material (fine soil+stone), \( m^3 \text{ m}^{-3} \)

\[ \theta_f \] water content of the fine soil, \( m^3 \text{ m}^{-3} \)

\[ \theta_r \] residual water content, \( m^3 \text{ m}^{-3} \)

\[ \theta_s \] saturated water content, \( m^3 \text{ m}^{-3} \)

\[ \theta_v \] volumetric water content, \( m^3 \text{ m}^{-3} \)

\[ \theta_v, Tp \] volumetric water content by Topp equation, \( m^3 \text{ m}^{-3} \)
$\theta, m_x$ volumetric water content by mixture equation, m$^3$ m$^{-3}$

$\rho_b$ bulk density, kg m$^{-3}$

$\varphi$ soil porosity

**Chapter 5**

**Roman symbols**

$AR$ artificial recharge, L

$C_s$ concentration of suspended material when infiltrating, g L$^{-1}$

$F$ flooding height, L

$FC$ field capacity

$h$ matric pressure head, m

$i$ infiltration rate, L T$^{-1}$

$i_0$ initial infiltration rate, L T$^{-1}$

$i_s$ steady state infiltration rate, L T$^{-1}$

$I_t$ cumulated infiltration at the time $t$, L

$K_{fs}$ and $K_s$ field saturated hydraulic conductivity, L T$^{-1}$

$K_{fs}$ hydraulic conductivity=$K(\theta_k)$ at some water content $\theta_k$, L T$^{-1}$

$K(\theta)$ hydraulic conductivity at the volumetric water content $\theta$, L T$^{-1}$

$l$ pore-connectivity parameter in van Genuchten equation

$m_q$ number of different sets of measurements in objective function

$n_{q_i}$ number of measurements in a particular measurement set in objective function

$n$ pore-size distribution index in van Genuchten equation

$NR$ natural recharge in water budget, L

$O_i$ observed value at time $i$ in $r^2$ calculation

$\bar{O}_i$ mean of observed data

$q_j(x, t, b)$ corresponding model predictions for the vector of optimized parameters $b$ (e.g., $\theta_r$, $\theta_s$, $\alpha$, $n$, $K_s$, $l$, ...) in objective function
x - List of symbols

\( q^*_j (x,t_i) \) represents specific measurements at time \( t_i \) for the \( j \)th measurement set at location \( x(r,z) \) in objective function

\( r^2 \) coefficient of determination

\( R \) recharge, L

\( S_e \) effective saturation, m\(^3\) m\(^{-3}\)

\( S_{ek} \) effective saturation at \( \theta_k \) corresponding to \( K_k \), m\(^3\) m\(^{-3}\)

\( S_i \) sensitivity value in sensitivity analysis

\( S(h) \) a sink function in Buckingham-Darcy equation, m

\( v_j \) weighting coefficient in objective function

\( V_r \) volume fraction of stones in stony soils in Bauwer equation

\( z \) vertical distance, positive downward in Buckingham equation, m

**Greek symbols**

\( \alpha \) inverse of the air-entry value in van Genuchten equation, m\(^{-1}\)

\( \alpha_i \) coefficient describing clogging properties of system, L g\(^{-1}\) day\(^{-1}\)

\( \Delta \) slope of vapor pressure curve, kPa °C\(^{-1}\)

\( \Delta S \) change in soil-water storage, L

\( \theta_a \) extrapolated parameter \( \leq \theta_s \), m\(^3\) m\(^{-3}\)

\( \theta_b \) water content of bulk material (fine soil+stone), m\(^3\) m\(^{-3}\)

\( \theta_{FC} \) water content at field capacity, m\(^3\) m\(^{-3}\)

\( \theta(h) \) water content at pressure head \( h \), m\(^3\) m\(^{-3}\)

\( \theta_m \) extrapolated parameter slightly larger than \( \theta_s \), m\(^3\) m\(^{-3}\)

\( \theta_r \) residual water content, m\(^3\) m\(^{-3}\)

\( \theta_s \) saturated water content, m\(^3\) m\(^{-3}\)

\( \theta_v \) volumetric water content, m\(^3\) m\(^{-3}\)

\( \lambda_c \) microscopic capillary length, L

\( \lambda E \) latent heat flux of evaporation, W m\(^{-2}\)

\( \rho_b \) bulk density, kg m\(^{-3}\)
### List of Abbreviations

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>ARG</td>
<td>artificial recharge of groundwater</td>
</tr>
<tr>
<td>ASTER</td>
<td>advanced space borne thermal emission and reflection radiometer</td>
</tr>
<tr>
<td>BC</td>
<td>boundary condition</td>
</tr>
<tr>
<td>BS</td>
<td>bare soil</td>
</tr>
<tr>
<td>BZ</td>
<td>Bisheh Zard</td>
</tr>
<tr>
<td>KFCV</td>
<td>k-fold cross validation</td>
</tr>
<tr>
<td>COSRI</td>
<td>combination of spectral responses of bare soil and vegetation</td>
</tr>
<tr>
<td>CROPWAT</td>
<td>a software designed for the calculation of the right amount of water needed for the irrigation of crop fields</td>
</tr>
<tr>
<td>CSC</td>
<td>conveyor-spreader channel</td>
</tr>
<tr>
<td>CV</td>
<td>coefficient of variation</td>
</tr>
<tr>
<td>DEM</td>
<td>digital elevation model</td>
</tr>
<tr>
<td>DN</td>
<td>digital number</td>
</tr>
<tr>
<td>EW</td>
<td>Experimental well</td>
</tr>
<tr>
<td>FAO P-M</td>
<td>FAO Penman-Monteith</td>
</tr>
<tr>
<td>FRWO</td>
<td>Fars Regional Water Organization</td>
</tr>
<tr>
<td>FWH</td>
<td>floodwater harvesting</td>
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<tr>
<td>FWS</td>
<td>floodwater spreading</td>
</tr>
<tr>
<td>GBP</td>
<td>Gareh Bygone Plain</td>
</tr>
<tr>
<td>GIS</td>
<td>geographic information system</td>
</tr>
<tr>
<td>GRASS</td>
<td>geographic resources analysis support system (RS-GIS software)</td>
</tr>
<tr>
<td>GW</td>
<td>groundwater</td>
</tr>
<tr>
<td>H1D</td>
<td>Hydrus one dimensional</td>
</tr>
<tr>
<td>IC</td>
<td>irrigated crops</td>
</tr>
<tr>
<td>ILWIS</td>
<td>integrated land and water information system</td>
</tr>
<tr>
<td>KFCV</td>
<td>k-fold cross validation</td>
</tr>
<tr>
<td>LEACHM</td>
<td>a code for water and solute transport modeling in soil</td>
</tr>
<tr>
<td>LST</td>
<td>land surface temperature</td>
</tr>
<tr>
<td>MAP</td>
<td>mean annual precipitation</td>
</tr>
<tr>
<td>METRIC</td>
<td>mapping evapotranspiration at high resolution with internalized calibration</td>
</tr>
<tr>
<td>MODIS</td>
<td>moderate resolution imaging spectro-radiometer</td>
</tr>
<tr>
<td>MRI</td>
<td>magnetic resonance imaging</td>
</tr>
<tr>
<td>NDVI</td>
<td>normalized difference vegetation index</td>
</tr>
<tr>
<td>NDVI_soil</td>
<td>NDVI of bare soil surface</td>
</tr>
<tr>
<td>NDVI_veg</td>
<td>NDVI of vegetated surface</td>
</tr>
<tr>
<td>OECD</td>
<td>organization for economic co-operation and development</td>
</tr>
<tr>
<td>Symbol</td>
<td>Definition</td>
</tr>
<tr>
<td>--------</td>
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</tr>
<tr>
<td>OW</td>
<td>observation well</td>
</tr>
<tr>
<td>PI</td>
<td>pastures inside the FWS systems</td>
</tr>
<tr>
<td>P-M</td>
<td>Penman-Monteith</td>
</tr>
<tr>
<td>PO</td>
<td>pastures outside the FWS systems</td>
</tr>
<tr>
<td>RA</td>
<td>Rahim Abad</td>
</tr>
<tr>
<td>RETC</td>
<td>a code for Soil-water retention curve fitting model</td>
</tr>
<tr>
<td>RMSE</td>
<td>root mean square error</td>
</tr>
<tr>
<td>RQ</td>
<td>research question</td>
</tr>
<tr>
<td>RS</td>
<td>remote sensing</td>
</tr>
<tr>
<td>RSE</td>
<td>relative standard error</td>
</tr>
<tr>
<td>RSR</td>
<td>RMSE-observations standard deviation ratio</td>
</tr>
<tr>
<td>SAR</td>
<td>sodium absorption ratio</td>
</tr>
<tr>
<td>SEBAL</td>
<td>surface energy balance model</td>
</tr>
<tr>
<td>SEBI</td>
<td>surface energy balance index</td>
</tr>
<tr>
<td>SEBS</td>
<td>surface energy balance system</td>
</tr>
<tr>
<td>SEE</td>
<td>standard error of estimate</td>
</tr>
<tr>
<td>SEP</td>
<td>standard error of prediction</td>
</tr>
<tr>
<td>SO</td>
<td>specific objective</td>
</tr>
<tr>
<td>S-SEBI</td>
<td>simplified surface energy balance index</td>
</tr>
<tr>
<td>SUMAMAD</td>
<td>sustainable management of marginal drylands</td>
</tr>
<tr>
<td>SWB</td>
<td>soil-water budget</td>
</tr>
<tr>
<td>SWC</td>
<td>soil water content</td>
</tr>
<tr>
<td>TDR</td>
<td>time domain reflectometry</td>
</tr>
<tr>
<td>TM</td>
<td>thematic mapper (Landsat sensor)</td>
</tr>
<tr>
<td>TP</td>
<td>tree plantation</td>
</tr>
<tr>
<td>TQ</td>
<td>Tchah Qootch</td>
</tr>
<tr>
<td>TSM</td>
<td>two source energy balance model</td>
</tr>
<tr>
<td>UNU-INWEH</td>
<td>United Nations university-institute for water environment and health</td>
</tr>
<tr>
<td>WR</td>
<td>water reservoir</td>
</tr>
<tr>
<td>ZFP</td>
<td>zero flux plan</td>
</tr>
</tbody>
</table>
# Table of contents

Summary................................................................................................................................. i 
List of symbols .......................................................................................................................... iii 
List of Abbreviations ................................................................................................................ xi 
Table of contents ..................................................................................................................... xiii 
List of tables ............................................................................................................................. xviii 
List of figures ........................................................................................................................... xxi 
List of photos .......................................................................................................................... xxvi 

## Chapter 1. General introduction ................................................................. 1

1.1 GLOBAL WATER CRISIS ........................................................................... 2
1.2 FLOODWATER HARVESTING ................................................................. 6
1.3 SCOPE OF THE PROBLEM ....................................................................... 8
1.4 THESIS FRAMEWORK ........................................................................... 9
1.5 RESEARCH OBJECTIVES AND QUESTIONS ............................................ 11
1.6 STRUCTURE OF THE THESIS ................................................................. 13

## Chapter 2. Improving the actual evapotranspiration estimation and its spatial distribution with limited available data using remote sensing ....... 15

2.1 INTRODUCTION ....................................................................................... 16
2.2 MATERIAL AND METHODS...................................................................... 18

### 2.2.1 Study site ......................................................................................... 18
### 2.2.2 Model description ............................................................................ 20
### 2.2.3 Remote sensing data and basic remote sensing products .......... 25
### 2.2.4 Vegetation products ......................................................................... 29
### 2.2.5 Ground observed weather data ......................................................... 31
### 2.2.6 Sensitivity of SEBS to the input parameters .................................... 32
### 2.2.7 Model calibration ............................................................................ 33

### 2.2.8 Evaluation of the results ................................................................. 35

#### 2.2.8.1 Global radiation ........................................................................... 35
#### 2.2.8.2 Justification of estimated ETa among the land uses .................. 35
#### 2.2.8.3 Cross validation of the estimated ETa ........................................ 35
#### 2.2.8.4 Water balance budgeted ET cross validation ............................... 36
#### 2.2.8.5 Justification of the modelled Kc .................................................. 40
## Table of contents

### 2.3 RESULTS AND DISCUSSION

<table>
<thead>
<tr>
<th>Subsection</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.3.1 Model sensitivity to input parameters</td>
<td>41</td>
</tr>
<tr>
<td>2.3.2 Model calibration</td>
<td>43</td>
</tr>
<tr>
<td>2.3.3 Justification of ET$_{a}$ among the land uses</td>
<td>47</td>
</tr>
<tr>
<td>2.3.4 Validation of the results</td>
<td>51</td>
</tr>
<tr>
<td>2.3.4.1 Global radiation</td>
<td>51</td>
</tr>
<tr>
<td>2.3.4.2 Cross validation of modelled ET$_{a}$</td>
<td>52</td>
</tr>
<tr>
<td>2.3.4.3 Water balance budgeted ET cross validation</td>
<td>52</td>
</tr>
</tbody>
</table>

### 2.4 CONCLUSIONS

<table>
<thead>
<tr>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>59</td>
</tr>
</tbody>
</table>

### Chapter 3. Evaluating the effect of floodwater spreading on groundwater recharge by combining water table fluctuation (WTF) and water budget methods in saturated zone

<table>
<thead>
<tr>
<th>Subsection</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.1 INTRODUCTION</td>
<td>64</td>
</tr>
<tr>
<td>3.2 MATERIALS AND METHODS</td>
<td>67</td>
</tr>
<tr>
<td>3.2.1 Site description</td>
<td>67</td>
</tr>
<tr>
<td>3.2.2 Floodwater spreading systems at the Kowsar Station</td>
<td>71</td>
</tr>
<tr>
<td>3.2.3 Water table level and weather data</td>
<td>74</td>
</tr>
<tr>
<td>3.2.4 Runon and runoff data</td>
<td>75</td>
</tr>
<tr>
<td>3.2.5 Determining the effect of FWS on recharge</td>
<td>79</td>
</tr>
<tr>
<td>3.2.6 Quantification of recharge</td>
<td>81</td>
</tr>
<tr>
<td>3.2.6.1 Amount of extraction from the aquifer</td>
<td>81</td>
</tr>
<tr>
<td>3.2.6.2 Change in groundwater storage (saturated pore volume)</td>
<td>82</td>
</tr>
<tr>
<td>3.2.6.3 Agricultural return flow and ET from the groundwater</td>
<td>84</td>
</tr>
<tr>
<td>3.2.7 Quantification of artificial recharge</td>
<td>85</td>
</tr>
<tr>
<td>3.2.7.1 Application of flow data</td>
<td>85</td>
</tr>
<tr>
<td>3.2.7.2 Water budget method for the FWS systems influenced area</td>
<td>86</td>
</tr>
<tr>
<td>3.3 RESULTS AND DISCUSSION</td>
<td>88</td>
</tr>
<tr>
<td>3.3.1 Water level data quality</td>
<td>88</td>
</tr>
<tr>
<td>3.3.2 Piezometric level and flow directions</td>
<td>89</td>
</tr>
<tr>
<td>3.3.3 Spatial and temporal changes in water level</td>
<td>89</td>
</tr>
<tr>
<td>3.3.4 Substantiating the effect of the FWS systems</td>
<td>97</td>
</tr>
<tr>
<td>3.3.5 Dewatering of the aquifer</td>
<td>100</td>
</tr>
<tr>
<td>3.3.6 Recharge estimation</td>
<td>102</td>
</tr>
<tr>
<td>3.3.7 Artificial recharge estimation</td>
<td>102</td>
</tr>
<tr>
<td>3.4 CONCLUSION</td>
<td>107</td>
</tr>
</tbody>
</table>
Chapter 4. Improved calibration of time domain reflectometry (TDR) for soil water content measurements in stony soils ........................................ 115

4.1 INTRODUCTION ................................................................. 116
4.2 MATERIALS AND METHODS ............................................. 118
  4.2.1 Soil sampling................................................................. 118
  4.2.2 Experimental setup....................................................... 118
  4.2.3 Data analysis................................................................. 120
  4.2.4 Validation of the results ............................................... 121
  4.2.5 Statistical indices for model validation ......................... 121
4.3 RESULTS AND DISCUSSION ............................................. 122
  4.3.1 Soils with original stoniness......................................... 122
  4.3.2 Stone-free soils........................................................... 124
  4.3.3 Effect of extension cables length ................................. 130
4.4 CONCLUSIONS................................................................. 133

Chapter 5. Evaluation of recharge by a modelling approach based on the measured soil water content in a multi layered vadose zone ......... 135

5.1 INTRODUCTION ................................................................. 136
5.2 MATERIALS AND METHODS ............................................. 139
  5.2.1 Site description............................................................. 139
    5.2.1.1 Vegetation ............................................................... 142
    5.2.1.2 Soils ................................................................. 142
    5.2.1.3 Weather data ......................................................... 143
  5.2.2 Experimental wells ..................................................... 143
  5.2.3 Water ponding measurement ...................................... 143
  5.2.4 Determination of hydraulic properties ......................... 144
    5.2.4.1 Field saturated hydraulic conductivity ..................... 144
    5.2.4.2 Bulk density ......................................................... 146
    5.2.4.3 Particle size analysis ............................................. 146
    5.2.4.4 Identifying and correlating the wells’ profile layers ....... 146
    5.2.4.5 Water retention data .............................................. 147
  5.2.5 Long term data collection .......................................... 147
    5.2.5.1 Soil water content measurement ............................. 147
    5.2.5.2 Insulating the well wall ........................................ 147
    5.2.5.3 Installing the TDR probes .................................... 147
    5.2.5.4 SWC measurements .............................................. 148
5.2.5.5 Climatic and flooding data collection.................................148
5.2.5.6 ET data .............................................................................149
5.2.6 Data quality assessment.......................................................149
5.2.7 Assessment of recharge.......................................................150
  5.2.7.1 Soil-water budget approach ............................................. 150
  5.2.7.2 Water flux modelling with Hydrus 1D .................................150
  5.2.7.3 Calibration of the H1D through parameter optimization .......154
  5.2.7.4 Model evaluation and statistical analysis ..............................156
  5.2.7.5 Sensitivity of the simulated flux to the input parameters .......157
  5.2.7.6 Validation of the H1D results .............................................157
5.3 RESULTS AND DISCUSSION ..................................................158
  5.3.1 Layers specification ..........................................................158
  5.3.2 Distribution of layers ........................................................160
  5.3.3 Assigning measured $K_{fs}$ to the layers...............................162
  5.3.4 SWC data quality ...........................................................164
    5.3.4.1 Temporal and vertical distribution of SWC data ...............164
    5.3.4.2 Empirical quantile-quantile (q-q) Plot .............................169
  5.3.5 Effect of sedimentation on infiltration .................................171
  5.3.6 Assessment of recharge ...................................................172
    5.3.6.1 SWB method ................................................................172
    5.3.6.2 Water flux Modelling by H1D .......................................175
    5.3.6.3 Sensitivity analysis .......................................................178
    5.3.6.4 Evaluation of the H1D result .........................................180
5.4 CONCLUSIONS ......................................................................182

Chapter 6. General discussion and conclusions .............................193
  6.1 RESPONSES TO THE RESEARCH OBJECTIVES ......................194
    6.1.1 Overall objective ............................................................194
    6.1.2 Specific objectives (SO) ....................................................196
  6.2 CONTRIBUTIONS OF THE THESIS .......................................200
  6.3 PROSPECTIVE WORKS .........................................................202

References ....................................................................................204

Appendix 1. The scripts for automotive processing of the Landsat TM5 inside the GRASS in Linux for ET calculation ........................................227
Appendix 2. The 1:100,000 Geology Map of the Fasa region which includes the Gareh Bygone Plain (is circled). ..............................................234
Appendix 3. The log of the observation wells (OW) ................................. 236
Appendix 4. Discharge rate measurements of pumped wells .................. 237
Appendix 5. Particle size distribution (lab analysis) of the soil samples and
field saturated hydraulic conductivity (field measured) of experimental well
number1. ......................................................................................................... 239
Appendix 6. Specific yield calculation based on weighted average of different
layers. ............................................................................................................. 240
Appendix 7. Detailed flowchart of the thesis structure. Eq. is equation, irr. is
irrigation,  Exp. is experimental,  Rep. is representative,  ret. is retention and
measur. is measurement. ................................................................................ 241
Acknowledgments ........................................................................................ 242
Curriculum vitae ........................................................................................... 245
List of tables

Table 1-1. Flood exposure by regions. ............................................................ 4
Table 2-1. Representative weather data of Gareh Bygone station, 1996-2011. ........................................................................................................................................ 20
Table 2-2. Vegetation characteristics of the studied land use types .......... 22
Table 2-3. Specification of the downloaded Landsat TM5 images. ............ 27
Table 2-4. Input parameters of SEBS sensitivity for sensible heat flux (H). 42
Table 2-5. Sensitivity of sensible heat flux to global radiation in different land uses........................................................................................................................................ 43
Table 2-6. Sources of products and parameters for calibrating SEBS. ...... 44
Table 2-7. Selected scenarios for calibrating SEBS based on improvement in approaching the estimated $ET_a$ to calculated $ET_o$ in water reservoir pixels... 46
Table 2-8. Descriptive statistics of the actual evapotranspiration computed from SEBS averaged for image dates and land-use types.......................... 50
Table 2-9. Comparing global radiation estimates from the “r.sun” GRASS module and the Fasa Weather Station.......................................................... 52
Table 2-10. Water consumption of the farm fields in the cropping season Dec. 2009 to Oct. 2010........................................................................................................ 53
Table 2-11. Water budget components of cropping season Dec. 2009 to Oct. 2010 for farm fields........................................................................................................ 54
Table 2-12. Water budget components of the image time series May 2009 to Oct. 2010 for pastures and bare soils outside the FWS................................ 54
Table 2-13. Daily estimates of midseason crop coefficients obtained from SEBS derived $ET_a$ and FAO P-M $ET_o$. ................................................................. 56
Table 2-14. Descriptive range of $K_c$ resulted by SEBS relative to the published $K_c$. ...................................................................................................................... 59
Table 3-1. The stratigraphy of the Gareh Bygone Plain and its upland Basins. .............................................................................................................................. 69
Table 3-2. Size and the implementation year of FWS systems. ............... 72
Table 3-3. Flow calculation of a typical broad crested weir in the Gareh Bygone Plain. .............................................................................................................. 79
Table 3-4. Retention curve data and input parameters for $S_y$ calculation..... 85
Table 3-5. Influence of the rainfall and number of dug wells on the GW level as indicated by multiple linear regression ......................................................... 92
Table 3-6. Groundwater level response to flood events for observation well (OW)2. .............................................................................................................. 94
Table 3-7. Calculations of water extraction in the hydrological year October 2010 to September 2011. ................................................................. 103
Table 3-8. Water budget of the total recharge in the hydrological year 2010-2011. ...................................................................................................... 103
Table 3-9. Flow rates of floods diverted into the FWS systems in 2011.... 104
Table 3-10. Components of the recharge calculation based on the flow data. ........................................................................................................ 104
Table 3-11. Extraction from the aquifer during the period of February to April 2011. .................................................................................. 105
Table 3-12. Water budget of the January to April 2011 for the area under the immediate influence of FWS systems. ........................................ 105
Table 3-13. Evaluation of the artificial to total recharge ratio. ................. 105
Table 4-1. Particle size distribution of the sampled layers....................... 118
Table 4-2. Statistical indices for model performance evaluation in stony samples................................................................................................. 125
Table 4-3. Layer specific calibration equations for estimation of $\theta v$ in stone-free samples. .......................................................... 127
Table 4-4. Statistics of SWC in sieved soil samples. ................................. 129
Table 4-5. Calibration equations for the stone-free samples used in validation. ................................................................................................. 130
Table 4-6. Regression coefficients for correction of $K_a$ readings for any cable length longer than 5 meters. ......................................................... 132
Table 5-1. General description of experimental wells (EW). .................... 144
Table 5-2. Some physical properties measured for representative layers. ................................................................. 159
Table 5-3. Statistics of the modified van Genuchten model for the representative layers.................................................................................... 160
Table 5-4. Vertical distribution of the representative layers (RL) in the experimental wells based on arbitrary coding of the pre-defined characteristics of each RL................................................................. 161
Table 5-5. Statistics of measured field saturated hydraulic conductivity ($K_{fs}$) of representative layers........................................................................ 162
Table 5-6. Statistics of measured soil-water content at different depths of W1 (period between 21 August 2010 to 1 December 2013)........................ 166
Table 5-7. Soil-water budget data from the 400 cm top layers, the flooding period 16 January to 23 July 2011. ...................................................... 174
Table 5-8. Characteristics of the different scenarios for running the H1D in inverse mode. ................................................................. 176
Table 5-9. Modelling information of the final H1D running results. ........ 178
Table 5-10. The error in the simulated recharge by the H1D as compared with
the soil-water budget method for the period between 16 January to 23 July
2011. ......................................................... 181
List of figures

Fig. 1-1. Every summer the Arctic sea ice melts down to its minimum in mid-September, before colder weather rebuilds the ice cover. The figure shows the 2012 minimum (recorded on 16 September), compared with the average minimum extent between 1979 and 2010 (curved line) (UNEP, 2013). .................. 2

Fig. 1-2. Map of global threats from floods and droughts (UNEP, 2012a). Iran (circled) is in the location of high/highest priority of droughts......................... 3

Fig. 1-3. Global physical and economical water scarcity (WWDR, 2014). Iran (circled) is approaching physical water scarcity. .................................................. 5

Fig. 1-4. 1960–2000 trends in total global water demand (right axis; indexed for the year 2000), global groundwater (GW) withdrawal (left axis) and global GW depletion (left axis) (Wada et al., 2010). Indexed for the year 2000 means that the amount of proportional demand in 1960 is around a factor of 0.45 based on the 2000 demand and is changed increasingly to the factor of 1.0 in the year 2000. ............................................................................................................... 6

Fig. 1-5. Schematic of floodwater spreading system (After Hashemi (2014)). 7

Fig. 1-6. Classification of techniques for assessing groundwater recharge (based on Scanlon et al. 2002). ............................................................................ 10

Fig. 1-7. Flowchart of the thesis framework. Cal. Eq. is calibrated equation, Cal. & eval. is calibration and evaluation, Rep. is representative. ............... 14

Fig. 2-1. Location of the study site in Iran (A), landscape of the Gareh Bygone Plain with location of the floodwater spreading project (RGB742 Landsat TM5 21/10/2010) (B), and simplified map of main land uses (C). The scale bar is accurate for the map c. The map A is modified based on a published map on the internet, maps b and c are produced in this study.............................. 19

Fig. 2-2. NDVI map generated from Landsat 5 on 18/03/2010 image with indication of main land uses and the locations which are visited at the field for determining the present land uses. ......................................................... 23

Fig. 2-3. Mean values of top of atmosphere radiation for thermal band images of non-cloudy dates during the study period (scale bars are standard deviations)... .......................................................... 26

Fig. 2-4. Presentation of cultivated farm fields maps for wheat (left) as main winter crop and forage corn (right) as main summer crop. .................. 37

Fig. 2-5. Relationship between $ET_a$ modelled by SEBS for water reservoir pixels and $ET_a$ derived by FAO P-M (fitted regression lines) when SEBS parameterized as: using radiation observed at weather station, ToA reflectance, internal LAI and $d_o$ and $z_{om}$ of SEBS and instantaneous wind speed, and (SEBS
adjusted), using r.sun GRASS-radiation, at-surface reflectance, LAI with Xavier and Vettorazzi (2004) equation, $d_o$ and $z_{om}$ of attribute table and adjusted wind speed. ................................................................. 47
Fig. 2-6. Comparing daily $ETa$ computed with SEBS for water reservoir pixels (based on at-surface reflectance products) with FAO PM $ETo$ when using: instantaneous wind speed ($u_{\text{inst}}$), and adjusted wind speed ($u_{\text{adj}}$) incorporating maximum daily wind speed. ................................................................. 47
Fig. 2-7. Actual ET ($ETo$) maps of the Gareh Bygone Plain generated by SEBS for (A) cold season 14/02/2010 and (B) warm season, 07/08/2010. ............... 49
Fig. 2-8. Global radiation estimates by r.sun module in GRASS vs. the measured data at Fasa Weather Station................................................................. 51
Fig. 2-9. Measured vs. predicted $ETo$ (1:1 line) of the cross validated data. 52
Fig. 2-10. Time series of $K_c$ values calculated for tree plantations with varying vegetation density (A) and pastures inside and outside of the FWS project (B). ..................................................................................................................................... 58
Fig. 3-1. Schematic of water budget for the saturated zone. Extraction is any type of direct withdrawal of groundwater; e.g. pumping, $ET$ is evapotranspiration and RF is agricultural return flow.............................. 66
Fig. 3-2. Location of the study site in Iran (A) and on the Mond Basin and the Shur sub-basin (B). Location of floodwater spreading (FWS) systems on the Gareh Bygone Plain (GBP) and its upland Basins (C); and sketch map of FWS systems and the Kowsar Station on the GBP (D). The image is Landsat TM5 false color composite of bands 7-4-2 dated 21/10/2010. BZ is Bisheh Zard, RA is Rahim Abad and TQ is Tchah Qooch. ...................................................... 68
Fig. 3-3. Schematic cross section of the Gareh Bygone Plain. The Agha Jari Formation consists of rhythmically interbedded brown to grey, calcareous, feature-forming sandstones and low weathering, gypsum-veined, red marls and grey to green siltstones. The Bakhtyari Formation mainly consists of pebbles and cobbles (conglomerate) of limestones and dark brown, ferruginous cherts. The diagram is drawn based on the Geological map of 1:100,000-scale published by National Oil Company of Iran and the logs of observation wells. Thickness of the layers is approximate. ............................................................ 70
Fig. 3-4. Generalized diagram of a typical floodwater spreading system. The scales are approximate. .................................................................................. 73
Fig. 3-5. Boundaries of Bisheh Zard and Tchah Qooch aquifers (A), detailed map of floodwater spreading (FWS) systems, observation wells, experimental
wells and distribution of some of the important operational wells in the Gareh Bygone Plain (B). ................................................................. 75
Fig. 3-6. Schematic map of BZ1, BZ2 and BZ3 and location of gauged gates for flow measurements. Dimensions are not to scale. Numbers show placement of Photo 3-2a (1), of Photo 3-2b (2), of Photo 3-2c and Photo 3-3a (3), of Photo 3-3b (4), and of Photo 3-3c (5)................................................................. 77
Fig. 3-7. Schematic of broad crested weir (Montes, 1998). ........................... 79
Fig. 3-8. Empirical q-q plots for the water level data of the observation wells (OW) 1 to 6 ................................................................. 88
Fig. 3-9. Piezometric map (A) and the digital elevation model (DEM) map (B) of the Bisheh Zard aquifer................................................................. 90
Fig. 3-10. A: Temporal trend of rainfall and groundwater hydrograph. B: Temporal trend in the number of wells in the region. GBP is Gareh Bygone Plain and BZ is Bisheh Zard aquifer. WL is groundwater level height. ....... 93
Fig. 3-11. Changes in WL (groundwater level) height in the observation well (OW)s. OW5 and OW6 were installed in 2005........................................... 96
Fig. 3-12. Location of the observation well 4 (OW4) in accordance with the Tchah Qooch River and the proportional ground and groundwater elevation. Picture is taken from the Google Earth of the date 6-9-2011. Corresponding groundwater level of the OW4 is taken from the year 2011. ......................... 97
Fig. 3-13. Response of water table to the flooding events. The FWS systems were not functioning in the first event of December 2003 to January 2004 and functioning in the second event of December 2004 to January 2005. WL is groundwater level and Mm$^3$ is million m$^3$. ........................................ 99
Fig. 3-14. The 3 dimensional presentation of the two overlaid groundwater level surfaces of the year 1997 over the 1993 which have been used to calculate the change in aquifer storage in these two successive years. ................... 101
Fig. 3-15. Variation in average water level in the observation wells before and after flooding for four consecutive months during the period of minimum water withdrawal. WL is groundwater level. 1st to 4th month refer to water level changes at the first to fourth month after the flooding events. ......................... 101
Fig. 3-16. Temporal changes in storage volume of the aquifer. Mm$^3$ is million m$^3$. .................................................................................. 101
Fig. 3-17. Change in GW level contour lines in the period of 3 months after the flooding event of 28-1 to 2-2-2011. ........................................... 106
Fig. 4-1. Observed vs. estimated water content $\theta_v$ based on measured $K_s$, of stony samples with Topp et al. (1980) equation (Topp’s Eq.; Eq. (4-1)) and
mixture equation (Mix Eq.; Eq. (4-4)). Full and dashed lines represent the regression by Topp’s Eq. and Mixture Eq., respectively, for (a) soil layer 1, (b) soil layer 2 and (c) soil layer 3. Number of samples are different because of non equal outlier data removal. Results are only shown for the capture window 20 ns and the probe type 15-cm two-rod (see table 4-2). ........................................ 126

Fig. 4-2. Measured $\theta_v$ of non-stony samples (after sieving) relationship to the dielectric permittivity ($K_a$) for the capture window 20 ns. Lines are regressions of layers 1 to 3 from A to C, respectively. RMSE values are based on two pairs of data calculated by Topp Eq. and new Eq. for each individual layer. ..... 128

Fig. 4-3. Change in dielectric permittivity ($K_a$) due to the extended TDR cable length with $\theta_v$. Graphs are related to soil types 1 to 3 from a to c, respectively. .................................................................................................................... 131

Fig. 5-1. (A) Floodwater spreading project in March 1988, Gareh Bygone Plain; ARG is artificial recharge of groundwater, FWS is floodwater spreading, BZ is Bisheh Zard, RA is Rahim Abad and TQ is Tchah Qootch. (B) Location of BZ$_1$ FWS system and land cover of the 2$^{nd}$ basin; TP is tree plantations, PI is pastures inside the floodwater spreading system and BS is bare soils. W1 to W3 are experimental wells (1-3). ................................................................ 141

Fig. 5-2. 3D interpolated map of the sediment depth (cm) of the study area using simple kriging with locally varying mean method (Esmaeili-Vardanjani et al., 2013). ................................................................................................ 142

Fig. 5-3. Mean RMSE statistics for the hydraulic models fitted to the layers by RETC. VGM is van Genuchten Mualem, VGM-2 is van Genuchten with -2 cm air entry value, MVG is modified van Genuchten, BC is Brooks and Corey, LNK is log normal Kosugi, DPD is dual porosity Durner. RMSE refers to the root mean squared difference between the H1D simulated water contents and the TDR observed ones at all depths of Well 1 in BZ$_1$. .............................. 159

Fig. 5-4. Retention curves for the representative layers of Well 1 in BZ$_1$. A, soil surface loam 7% fine gravel; B, loamy sand 22% gravel; C, sandy loam 54% medium gravel; D, sandy loam 67% coarse gravel; E, sandy loam 7% fine gravel; F, sand 53% medium gravel; G, loamy sand 0.3% medium gravel (red color to be distinguished). BZ is Bisheh Zard................................. 160

Fig. 5-5. Change in the mean field saturated hydraulic conductivity ($K_{fs}$) of the layers due to change in soil texture and stoniness. The lines are regressions between $K_{fs}$ and corresponding parameters clay+silt and stoniness. .......... 162

Fig. 5-6. Diagram of layer distribution and correlation of experimental wells’ layers. Alphabetic characters in left diagram (A to G) are summarized in three
groups in left diagram (A to C) as described in it legend. Dimensions in x direction are not to scale. ................................................................. 163

Fig. 5-7. Soil-water content time series of some selected depths measured by the calibrated TDR probes. Depths are noted at the center of the graphs. .. 167

Fig. 5-8. Change in the SWC of layers before and after the two flooding events 28 Jan. 2011 (A and C) and 1 Aug. 2013 (B and D). The graphs A and B show the change in the SWC over the entire 30 m profile, while graphs C and D zoom into the to 5 of the profile to better illustrate the redistribution of the SWC in the top layers. ................................................................. 168

Fig. 5-9. The empirical q-q plot for the soil-water content at depths of 10 to 400 cm of experimental well. The number of samples is similar for all of the plots and the probability values are greater than 0.01. ......................... 170

Fig. 5-10. Changes in soil-water content (SWC) at different depths of the experimental well as influenced by the rainfall and flooding events. The number of layers has decreased to better differentiate the changes between the layers................................................................. 172

Fig. 5-11. The observed and simulated soil-water content (SWC) data for the layers at 10 cm (A), 60 cm (B), 180 cm (C) and 400 cm (D). $R^2$ is determination coefficient, RMSE is root mean square error. The mass balance error of optimal run of the model was calculated as 0.6%. The optimal hydraulic model was determined to be the modified van Genuchten (MVG) model......... 177

Fig. 5-12. Simulated flux (A) and the cumulative recharge (B) by H1D for the period between 16 January to 23 July 2011 period. Dashed line in figure B shows the amount of total recharge in the entire period......................... 179

Fig. 5-13. Sensitivity of simulated flux by Hydrus 1D to the deviations in hydraulic parameters of modified van Genuchten model $\theta_r$, $K_s$, and $l$ (A) and $\theta_s$, $\alpha$ and $n$ (B). Parameters are presented in two graphs because of different magnitudes. $S_i$ is sensitivity. The initial value for the $n$ parameter was 1.0 and as the $n$ value should not be smaller than 1.0, the deviation of 0.5 and 0.75 was resulted in the model failure in convergence................................. 180
List of photos

Photo 2-1. The deep roots of *Eucalyptus camaldulensis* which is cut during the excavation inside the 28 m deep well inside floodwater spreading systems in Gareh Bygone Plain, Iran................................................................. 61
Photo 3-1. The observation well inside the study area (OW2) and OW4 (B) and the water level measurement tools. ................................................................. 109
Photo 3-2. Hydrometric station on the Bisheh Zard Ephemeral River in the study area during the flooding event of 01/02/2011 (A), a downstream view of the diversion dam of the same River showing the remainder of the diverted flood which flows within the River (B) and a drop which delivers part of floodwater to the FWS systems (C). See Fig. 3-6 for the placement of the structures ................................................................. 110
Photo 3-3. Measurement structures; (A) the drops which divert the floodwater from the Bisheh Zard River to the FWS systems, (B) and (C) two types of broad crested weirs for delivering and measuring the flow to the FWS systems basins at the study site (see Fig. 3-7 for the placement of the structures) .. 111
Photo 3-4. Operational well and connecting pipe with special Try square device for measuring well discharge in an agricultural field at the study site. It should be noted that the photo is taken to show the main settings and device but the discharge which is shown in the photo is a sample of partially full flow and was measured by California pipe method (see Appendix 4). ............... 112
Photo 3-5. Applied irrigation water measurements in an agricultural field inside the study area by cut throat flume; by side view of the flume (A) and top view with Carpenters level measuring tool for leveling the flume (B). ...... 113
Photo 5-1. Floodwater spreading in the studied basin (A) and the height of ponding water (water mark) in flooding events around the experimental well (B). .............................................................................. 185
Photo 5-2. Stoniness of the experimental well layers................................. 186
Photo 5-3. Infiltration measurement by double ring method inside the experimental well at a depth of 3 m................................. 187
Photo 5-4. Insulating the experimental well by concrete tiles (A) and the insulated top view of the experimental well (B). ................................. 188
Photo 5-5. The TDR equipped experimental well (A), the TDR cable insulating pipes (B), and the lowest installed TDR probe at a depth of 28 m (C). ...... 189
Photo 5-6. Inserting the TDR probes. A polystyrene guide and a steel template for preparing the holes for the pipes containing the TDR probe (A), the place for the TDR probe inside the well (B), the inserted polystyrene template (C),
the place for inserting the TDR probe (D and E), and the successive places for
the TDR probes (F). .................................................................................... 190
Photo 5-7. Arrangement of the TDR cables in the multiplexer in the
experimental well. ....................................................................................... 191
Chapter 1. General introduction
1.1  **Global water crisis**

Global population is projected to reach 9.3 billion in 2050 (UNDESA, 2012). Population growth leads to increased water demand, reflecting growing needs for drinking water, health and sanitation, as well as for energy, food and other goods and services that require water for their production and delivery (WWDR, 2014).

Climate change plays a diversity of roles all affecting the water cycle and availability (UNEP, 2014). The world is warming, and with it the Arctic. The size of sea ice was at a record low in recent years. In July 2012, 97% of the Greenland ice sheet surface was melted (UNEP, 2013) (Fig. 1-1).

![Fig. 1-1. Every summer the Arctic sea ice melts down to its minimum in mid-September, before colder weather rebuilds the ice cover. The figure shows the 2012 minimum (recorded on 16 September), compared with the average minimum extent between 1979 and 2010 (curved line) (UNEP, 2013).](image)

This region plays an important role in the climate system and ocean circulation. On the other hand, climate change is accused as the main cause of climate fluctuations. The year 2011 was a record-breaking year among the recent three decades for extreme climate and weather events. According to the latest scientific insights, climate change is leading to changes in the frequency, intensity, length, timing and spatial coverage of extreme weather events (UNEP, 2012b). This means that droughts and flash floods which are the characteristic phenomena of dry regions are going to be more modal with higher peaks as compared to the past. However, countries located in the dry regions of the earth are most subjected to risks of droughts, floods or both (Fig. 1-2).
According to WWDR (2012), more than 90% of the global population exposed to floods lives in South and East Asia and in the Pacific, and exposure is growing most rapidly in sub-Saharan Africa, the Middle East/North Africa (260 and 150% increase from 1970 to 2010 respectively). In contrast, exposure of population to floods is stable in countries of the Organization for Economic Co-operation and Development (OECD), while it is starting to trend downwards in eastern and southeastern Europe and Central Asia (Table 1-1).

Fig. 1-2. Map of global threats from floods and droughts (UNEP, 2012a). Iran (circled) is in the location of high/highest priority of droughts.

Water-related hazards account for 90% of all natural hazards, and their frequency and intensity is generally rising. Some 373 natural disasters killed over 296,800 people in 2010, affecting nearly 208 million others and costing nearly US$110 billion (UN, 2011). One water-related hazard that seldom makes it into the impacts statistics is drought. According to the United Nations Global Assessment Report, since 1900 more than 11 million people have died as a consequence of drought and more than 2 billion have been affected by drought, more than any other physical hazard (UNISDR, 2011). Drought is causing acute water shortages in large parts of Australia, Africa, Asia and the USA (Morrison et al., 2009). Regardless, an urban Australian on the average consumes 300 L water daily and an European 200 L, while in sub-Saharan Africa an individual makes do with less than 20 L per day (Natarajan, 2007). Besides droughts, river flows and water supplies are being reduced by shrinking snow caps across China, India and Pakistan; countries where more
than 1 billion people already lack access to safe drinking water and adequate sanitation (Morrison et al., 2009).

In addition to the internationally recognized role of human activities on climate and the associated consequences on water availability, human activities also lead to growing water consumption. Globally, the list of countries away from water scarcity is getting shorter year by year. The demand for freshwater and energy will continue to increase significantly over the coming decades to meet the needs of increasing populations, growing economies, changing lifestyles and evolving consumption patterns. This will greatly amplify pressures on limited natural resources and ecosystems. The challenge will be most acute in countries undergoing accelerated transformation and rapid economic growth, especially where water resources are scarce or where water-related infrastructure and services are inadequate (Connor and Winpenny, 2014).

Table 1-1. Flood exposure by regions.

<table>
<thead>
<tr>
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<th></th>
<th></th>
<th></th>
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</thead>
<tbody>
<tr>
<td>East Asia and Pacific</td>
<td>9.4</td>
<td>11.4</td>
<td>13.9</td>
<td>16.2</td>
<td>18.0</td>
<td>91</td>
</tr>
<tr>
<td>Europe and Central Asia</td>
<td>1.0</td>
<td>1.1</td>
<td>1.2</td>
<td>1.2</td>
<td>1.2</td>
<td>20</td>
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<tr>
<td>Latin America and the Caribbean</td>
<td>0.6</td>
<td>0.8</td>
<td>1.0</td>
<td>1.2</td>
<td>1.3</td>
<td>117</td>
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<tr>
<td>Middle East and North Africa</td>
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<td>0.3</td>
<td>0.4</td>
<td>0.5</td>
<td>0.5</td>
<td>150</td>
</tr>
<tr>
<td>OECD countries</td>
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<td>1.5</td>
<td>1.6</td>
<td>1.8</td>
<td>1.9</td>
<td>36</td>
</tr>
<tr>
<td>South Asia</td>
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<td>24.8</td>
<td>31.4</td>
<td>38.2</td>
<td>44.7</td>
<td>132</td>
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<tr>
<td>Sub-Saharan Africa</td>
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<td>1.4</td>
<td>1.8</td>
<td>260</td>
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<tr>
<td>World</td>
<td>32.5</td>
<td>40.6</td>
<td>50.5</td>
<td>60.3</td>
<td>69.4</td>
<td>114</td>
</tr>
</tbody>
</table>

Source: WWDR (2012). The OECD are the member countries of Organization for Economic Co-operation and Development. The last updated list of OECD countries includes majority of European territory, South Korea, Australia and USA (http://www.oecd.org).

As depicted in Fig. 1-3, almost all of the Middle East and North Africa, including Iran, are facing with or approaching physical water scarcity (WWDR, 2014). The increasing demand for food and fiber as a consequence of population growth as well as urbanization has formed the other side of the water crisis in dry lands, i.e., intensification of pressure on groundwater (GW) storage.
Fig. 1-3. Global physical and economical water scarcity (WWDR, 2014). Iran (circled) is approaching physical water scarcity.

According to Wada et al. (2010), the estimated extent of GW depletion in sub-humid to dry regions has grown from 50 to 115 (km$^3$ ha$^{-1}$) between 1960 and 2000, whereas the withdrawal of water from the GW has increased unequally from 125 to 295 km$^3$ ha$^{-1}$ in the same period (Fig. 1-4). This means an increase at factor 2.25 and 2.38 for depletion and withdrawal, respectively. It reveals that a major part of withdrawn GW is being compensated by recharge (natural and artificial), but there is no guarantee that the proportion of recharge/withdrawal will remain stable to keep the current trend of depletion.

However, projected changes in the global water cycle regime toward more extreme events suggest more runoff and less infiltration (e.g., Dai, 2011), which would result in a decreasing GW recharge rate all around the world (Wada et al., 2010).

Furthermore, land use change, which results from industrial development especially in emerged economies in developing countries, is another cause of decreasing GW recharge. In addition to loss of GW storage as the mainstay resource in many societies, worsening its quality is a further consequence of depletion. This is the most important cause of emerging salinization in water and soil resources, mostly in dry regions (D’Odorico et al., 2013).
To mitigate over-exploitation of GW resources, increasing magnitudes and number of flash floods, and further climatic change impacts, appropriate technologies are needed to balance destructive floods with optimally utilizing floodwater, particularly in dry regions where it is a major source of water besides precipitation.

### 1.2 Floodwater harvesting

Traditional floodwater utilization, also called floodwater harvesting (FWH), is a successful approach and an effective tool to deal with water scarcity and to alleviate the destructive flash floods in drought-prone areas (Hashemi, 2014). Flood irrigation dates back to 3400 B.C. when the Nile's overflow was used to irrigate some 1.2 million ha west of that lifeline (Lu et al., 2012). A tremendous volume of floodwater is available annually in arid and semi-arid regions; and in some cases, economic damages and loss of human life strongly affect residents of the regions (NRCS, 2002). In general, FWH is the process of collecting and storing runoff for livestock and agricultural use during the rainfall event and utilizing it instantaneously or later (Mbilinyi et al., 2005). Periods of drought are often punctuated by flood-producing downpours that
devastate the drought stricken people and their livestock. However, if harnessed, these very floods can bring life back to the desert. It is therefore essential to upgrade the status of floods from a curse to a blessing and to ameliorate drought conditions with the wise use of floodwaters (Robertson et al., 2013). From a perspective of artificial recharge of groundwater (ARG), FWH can be defined as a type of water-spreading release over the ground surface in order to increase the quantity of water infiltrating into the ground and then percolating to the water table (Todd and Mays, 2005). Among FWH techniques, the most widely practiced surface method is floodwater spreading (FWS), which involves the surface spreading of water, useful in areas with abundant land availability, highly permeable soils, and a shallow unconfined aquifer (O’Hare et al., 1986). ARG is defined as the recharge at a rate greater than natural, resulting from deliberate or incidental human activities (Wels, 2012). Artificial recharge systems are engineered systems where surface water is put on or in the ground for infiltration and subsequent movement to aquifers to augment GW resources (Bouwer, 2002). Water spreading, the lynchpin of aquifer management through ARG, is characterized by a system of dams, dikes, ditches, or other means of diverting or collecting runoff from natural channels, gullies, or streams, and its spreading over relatively flat areas (NRCS, 2002) (Fig. 1-5). Aquifer management, the prudent harvesting and using of floodwater without stressing the environment is a key concept to drought mitigation in dry regions.

![Fig. 1-5. Schematic of floodwater spreading system (After Hashemi (2014)).](image)

FWH has become an increasingly important method and recent decades have shown a renewed interest in research and implementation of ancient FWH systems used for ARG (Evenari et al., 1971; Kowsar, 2011). This was
addressed in the extensive review of Boers and Benasher (1982) that revealed an awareness of increasing need for FWH and recognition of its potential. Despite the long and successful history of these systems, little is still known about their function and effect on hydrological processes in dry areas (Ouessar et al., 2009).

FWH systems have been mainly assessed in terms of supporting rainfed farming rather than their contribution to ARG. As new applications of FWH techniques aim at GW recharge, GW studies seem to be a new area of research for FWH evaluation and assessment (Sanford, 2002). However, adequate estimation of recharged water is often difficult due to complex geophysical features and the large temporal and spatial variability of rainfall and runoff (Rushton and Ward, 1979; Sophocleous, 1991; Flint et al., 2002).

1.3 Scope of the problem

In Iran, mean annual precipitation (MAP) is 226 mm (Rahnemaei et al., 2013) while average rainfall in the planet Earth is reported by FAO (2003) as 806. According to Rahnemaei et al. (2013) MAP ranges from 2000 mm in the North to less than 51 mm in Central Iran. There is no rain in most parts of the country between May and October. Only ~10% of the annual precipitation occurs in the warm and dry seasons (May to August) whereas ~90% occurs in the cold and humid seasons. About 52% of the annual precipitation occurs in 25% of the Iran territory only; hence, some parts of the land suffer from a lack of water resources, where an imminent water crisis is a certainty in the near future (see Fig. 1-2 and 1-3).

The increased pressure on GW and the high rate of subsidence (the gradual caving in or sinking of an area of land) resulting from over-exploitation are likely to become a serious challenge for future development of the GW basins in Central and Northeast Iran (Motagh et al., 2008). Hence, total water resources per capita in Iran have plunged by more than 65% during the last four decades, and are expected to decrease by another 16% by 2025 (Sarraf et al., 2005).

FWS is practiced in several dry regions of Iran as from the late 60s to mitigate water scarcity and minimizing the damages of the flooding events. The details of their objectives and the technical aspects of installation are presented in several publications (Kowsar, 1991; Kowsar and Zargar, 1991; Kowsar and
The remarkable initial success of an ARG project at the Kowsar Floodwater Spreading and Aquifer Management Research, Training and Extension Station (Kowsar Station) in a desert called Gareh Bygone Plain in southern Iran resulted in a substantial increase of irrigated farm fields downstream of the infiltration basins.

However, there are some concerns about the FWS systems in the Gareh Bygone Plain because of following observations:

- the GW table level seems to decline which resulted in the abandonment of some wells;
- the infiltration rate at the soil surface seems to decrease in all FWS systems due to siltation and subsequent clogging, from which might lower their efficiency;
- dense plantations of water demanding trees, mainly *Eucalyptus camaldulensis*, in some parts of the FWS systems might have contributed to increased total water consumption as evapotranspiration.

The present research framework was setup to address these issues.

**1.4 Thesis framework**

According to the diverse objectives and methods of implementing FWS systems, various factors need to be considered when choosing a method for quantifying recharge. Therefore, the rate of aquifer recharge is one of the most difficult items to measure in GW resource evaluation (Sophocleous, 1991). Scanlon et al. (2002) categorized the techniques used in quantifying recharge in three main groups: unsaturated zone, saturated zone and surface water techniques (summarized in Fig. 1-6).
Subdivisions of these techniques are somewhat arbitrary. Regarding the unsaturated (vadose) zone methods, Allison et al. (1994) subdivided the physical methods in direct (i.e., lysimetry) and indirect ones. In turn, indirect methods involve Zero Flux Plane (ZFP), soil-water budget (SWB), and water flux methods based on solving the well-known Buckingham-Darcy equation. Numerical modelling in the unsaturated zone refers to solving the so-called Richards equation with finite differences or finite elements methods. Tracing techniques that use inert tracers have been excluded from physical methods in their classification.

The physical methods pertaining to the saturated zone comprise of Darcy’s low, water table fluctuation, water budget and tracing methods.

In this dissertation, we applied both methods of the unsaturated zone and the saturated zone approaches separately. Surface water techniques are out of the scope of this dissertation and were therefore not discussed above.

In all of the techniques, the governing equation is a general mass balance equation which can be written as:
\[ \text{Input} = \text{output} + \text{accumulation} \quad (1-1) \]

Depending on the zone under the study (saturated, unsaturated and surface water zone) different components are defined as input, output and accumulation. For instance, the output part, evapotranspiration \((ET)\), in saturated zone studies is separated to \(ET\) from the soil surface and the GW while in unsaturated studies it defined as \(ET\) from soil surface merely.

In all of the recharge assessment methods used in this study, \(ET\) with its temporal and spatial variation from different land uses in the Gareh Bygone Plain was an important component (chapters 3 and 5). Therefore part of this dissertation was devoted to \(ET\) mapping with a remote sensing model that was calibrated based on available data.

In applying the saturated zone approach over a span of the entire Gareh Bygone Plain, the water table fluctuation (WTF) and water budget (WB) methods were applied to solve Eq. (1-1) in order to quantify and differentiate natural and artificial recharge.

In the unsaturated zone study conducted at the point-scale within the FWS systems, the methods employed were soil-water budget (SWB) and numerical modelling using soil water content measured along a 30-m deep experimental well. Results of the various saturated and unsaturated methods were then used to evaluate the FWS installed at the Gareh Bygone Plain for recharging the GW table.

### 1.5 Research objectives and questions

**Overall objective**

This dissertation has arisen from concerns regarding ARG through FWS in combating water scarcity at the study site. The overall objective of this dissertation is defined as “evaluation of FWS systems for recharging the GW table”.

**Specific objectives**

The specific objectives are focused essentially on different aspects to clarify its effectiveness and on developing methods needed to provide the required input for the evaluation techniques used. The specific objectives (SO) and their corresponding research questions (RQ) can be defined as below:
12 - General introduction

- **SO1**: Improved estimation of ET; Evaluation and calibration of surface energy balance system (SEBS) model to maximize the reliability of the ET and crop coefficient (Kc) estimations in the study site as a hot and dry region.

RQ1: How can the SEBS model parameterization be improved with limited available data through calibration?

RQ2: What is the agreement between SEBS modeled ETa and ETa calculated by water budget method?

RQ3: Are the Kc values estimated by SEBS comparable to those published in literature for the different land uses and crops?

RQ4: What is the yearly water consumption of the main land uses by means of summed up ET in growing season?

- **SO2**: Assessing the recharge by saturated zone methods; Assessing the impact of FWS systems on recharge by saturated zone methods while minimum sources of GW data are available by combining the water table fluctuation (WTF) and water budget.

RQ5: What is the immediate effect of flooding events on GW level of the main aquifer?

RQ6: What is the proportion of natural and artificial recharge through water budget in a hydrological year when full input data are available?

- **SO3**: Assessing the recharge by unsaturated zone methods; Calibration of Hydrus 1D model through hydraulic parameter optimization for assessing the recharge by FWS systems based on soil water content measurement over the vadose zone.

RQ7: How reliable are soil-water content data acquired by the TDR method in stony soils?

RQ8: Does the clogging due to long term sedimentation after flooding events affects actual infiltration rate in the FWS systems in a real flooding event?
RQ9: What is the influence of flooding event on soil-water budget components (including the net recharge) in aquifer profile?

RQ10: What is the reliability of the direct measurements of hydraulic parameters in disturbed/undisturbed soils as compared to the optimized hydraulic parameters with calibrated Hydrus 1D (H1D) model based on the measured soil-water contents?

RQ11: How sensitive is the H1D model to input hydraulic parameters?

RQ12: To what extent is the simulated recharge by the calibrated H1D model reliable for different aquifer layers?

1.6 Structure of the thesis

As shown in Fig. 1-7 the thesis consists of a chapter of general introduction, four research chapter and seventh chapter of general conclusion. A detailed explanation of the inputs and outputs of each chapter and the interactions is illustrated in Appendix 7. The outlines of the chapters are as follow.

Chapter 1 contemplates the concerns which have motivated this dissertation, referring to the water crisis on a global scale and that of Iran in particular. The objectives, research questions and framework are described as well.

Chapter 2, reports the application of remote sensing for the estimation of ET by using the SEBS model after evaluating and calibrating the model to maximize its results in the study area. The resulted ET time series maps and data were used in further chapters.

Chapter 3 evaluates the effect of flooding events on recharging of the GW using the water table fluctuation and water budget methods, which can both be categorized as saturated zone approaches. This study spans the entire Gareh Bygone Plain.

In Chapter 4, the application of the TDR method for measuring water content in stony soils is evaluated, new calibration curves that minimize errors in measuring soil water content are presented and the impact of extension cables, as probes were installed until ~30 m depth, is discussed as well.
Chapter 5 goes into the evaluation of ARG using unsaturated zone techniques. This study was conducted at a point located in one of the water spreading basins inside the FWS systems, using soil water content measured with TDR probes installed until ~30 m depth before and after flooding events.

Chapter 6 integrates the findings of the thesis and addresses whether FWS systems is effective for ARG. It presents answers to the various research questions, the contributions of this dissertation toward floodwater harvesting context and suggests recommendations for further studies.

Fig. 1-7. Flowchart of the thesis framework. Cal. Eq. is calibrated equation, Cal. & eval. is calibration and evaluation, Rep. is representative.
Chapter 2. Improving the actual evapotranspiration estimation and its spatial distribution with limited available data using remote sensing

This chapter is based on a modified published article:

2.1 Introduction

Knowledge of actual evapotranspiration ($ET_a$) is essential for solving the water budget, particularly in arid areas when water saving practices such as floodwater harvesting basins for groundwater (GW) recharge need to be evaluated. $ET_a$ varies regionally and seasonally according to environmental conditions, mainly climate, land cover and land use, soil water availability and crop management. Conventional techniques using point observations for estimating the $ET$, produce information that is representative of local scales only (Allen et al., 2011), and difficult to extend to large areas due to the heterogeneity of land surfaces and the dynamic nature of heat transfer processes (Su, 2002). Contrarily to local scale methods, remote sensing (RS) provides large scale information and data of hydrological, vegetation, soil and topographic nature that help overcoming limitations relative to ground observation networks (Sun et al., 2009). Hence, many efforts have focused on $ET$ estimation and mapping at various scales using RS data, particularly during the last decade (Su et al., 2005; Gowda et al., 2008; Sun et al., 2009).

Various RS-based $ET$ algorithms are available for estimating $ET$ at various time and space scales and varied complexity. Comprehensive reviews of various land surface energy balance models have been published, e.g., Courault et al (2005), Overgaard et al. (2006), Kalma et al. (2008) and Gowda et al. (2008). Widely tested land surface energy balance models include SEBAL (Bastiaanssen et al., 1998), SEBI (Menenti and Choudhury, 1993), S-SEBI (Roerink et al., 2000), SEBS (Su, 2002), TSM (Norman et al., 1995; Chehbouni et al., 2001), and METRIC (Allen et al., 2007).

Estimation accuracy of various models/algorithms, as reported by Gowda et al. (2008), varied from 67 to 97% for daily $ET$ and above 94% for seasonal $ET$ indicating good potential to estimate regional $ET$ accurately.

The ultimate aim of deriving the surface energy budget (and specifically evapotranspiration) by remote sensing methods is to reach good operational utility under different land surface conditions (Meijerink et al., 2005; Norman et al., 2006). However, RS-based surface energy balance models can be problematic, especially for sparsely vegetated and (occasionally) dry areas. Some studies (Lubczynski and Gurwin, 2005; van der Kwast et al., 2009) indicate that RS-based solutions of the surface energy balance are showing
weak estimates of $ET$ in such areas by an error of 1.5-3.0 mm day$^{-1}$ due to complexity in estimating sensible heat flux $H$ in dry areas.

The one-dimensional surface energy balance system (SEBS) model has been tested over agricultural, grassland and forested sites, and across several spatial scales and with Landsat, ASTER and MODIS satellite-acquired data (Su et al., 2005; McCabe and Wood, 2006; Su et al., 2008; van der Kwast et al., 2009; Lu et al., 2013). A good consistency was observed between flux retrievals from Landsat and ASTER data, but MODIS data were unable to discriminate the influence of heterogeneity in land use at field scale while results were comparable at catchment scale (McCabe and Wood, 2006). Among the widely used models, the SEBS model is less demanding in terms of parameterization than SEBAL or METRIC, but is less tested than these algorithms, particularly for arid landscapes.

The SEBS model is available as part of the open source freeware ILWIS (www.52north.org), and can therefore be used by practitioners with remote sensing knowledge who may not necessarily have the micrometeorological expertise and data to develop an $ET$ estimation model themselves. Whilst the open-source format of SEBS is very useful and can speed up the research process, there are some instances where specialist knowledge is required to implement the model correctly to derive the most accurate results. Therefore, particular attention has to be given to its parameterization (Lu et al., 2013). As reported in some studies, the SEBS model underestimated $H$. van der Kwast et al. (2009) evaluated the SEBS model with flux measurements over different land covers and found that SEBS underestimated $H$ in some places. Gokmen et al. (2012) developed a scaling factor related to soil moisture to correct the SEBS-calculated $k_b^{-1}$ (a dimensionless parameter, which relates the roughness lengths of heat and momentum transfer) so that the underestimation of $H$ can be reduced. However, the method may not perform well under high soil moisture or small temperature gradient conditions.

Furthermore, the diversity of the SEBS model parameters together with the inherent sources of error in using remote sensing derived products as input implies that a number of sources of lack in accuracy may exist and should be properly understood and addressed (Gibson et al., 2011).
The present research refers to a floodwater spreading (FWS) project area located in the Gareh Bygone Plain, south of Zagros Mountains, in southern Iran. $ET$ from agricultural crops and tree plantations play a major role in water use in this area. Currently, there are no reliable data on $ET$ but, because it is a major component of the water budget of the plain, its knowledge and estimation is crucial. Due to the complex and heterogeneous vegetation cover of the project site and difficulties in assessing $ET$ at local scale, the remote sensing based energy balance approach was selected to estimate $ET_a$. Hence, the estimation accuracy for the selected model needs to be evaluated for the conditions of this study and particularly when limited data are available.

The main objectives of this remote sensing study are a) to assess the validity of the SEBS model in a hot and dry region for estimating actual $ET$ for various land cover types, as compared to ground-based reference $ET$, b) assessing the sensitivity of outputs to changes in input parameters c) to assess the effect of parameterization processes on improving its validity, d) to evaluate results based on cross checking the total seasonal $ET$ with irrigation and rainfall and e) to derive appropriate crop coefficients for the vegetation of the Gareh Bygone Plain, thus to map both $ET$ and crop coefficients for this target area.

The results will be used to perform a water budget study in the area to evaluate the floodwater spreading system in the next chapters and to support water management by the practitioners in the region.

2.2 Material and methods

2.2.1 Study site

The Gareh Bygone Plain has an area of 18000 ha. The FWS project area comprises 2033 ha (area under water spreading). Very little freshwater resources were available before the artificial recharge of the GW through the FWS activities, which started there in 1983.

It is located south of Zagros Mountains, Southern Iran, between 28° 35' to 28° 41'N and 53° 53' to 53° 57'E. The altitude of the plain ranges between 1120 and 1160 m.a.s.l. (Fig. 2-1). This is a dry region, with mean annual precipitation of 211 mm, having high inter-annual variability. Rainfall mainly occurs from December to March, with few exceptional events in summer (June-August). The absolute maximum temperature (40-46°C) occurs in July-August and the corresponding minimum (-1 to -6°C) in January–February.
(Table 2-1). Weather data used in this study were taken from the Gareh Bygone and the Fasa weather stations. The Gareh Bygone weather station is part of the national weather stations network (OFCM, 2005) and is located inside the study site (28° 36'N, 53° 55'E and 1162 m of altitude).

![Map of Iran with Study Site Location](image1)

**Fig. 2-1.** Location of the study site in Iran (A), landscape of the Gareh Bygone Plain with location of the floodwater spreading project (RGB 742 Landsat TM5 21/10/2010) (B), and simplified map of main land uses (C). The scale bar is accurate for the map C. The map A is modified based on a published map on the internet, maps B and C are produced in this study.

The Fasa weather station is a synoptic station (28° 58'N and 53° 41'E and 1288 m of altitude) located at 45 km North-East of the study site, whose hourly data was available through “http://www.mundomanz.com/meteo_p/main” but global radiation, was received through personal correspondence. Global radiation measured at the Fasa station was used to compare calculated and measured radiation.

Irrigated crop fields in the study area have an annual crop-fallow rotation system covering nearly 2200 ha and are mostly located in the vicinity of the FWS project. Wheat is the main winter crop, which is sometimes replaced by barley or cotton. Watermelon, melon and cantaloupe are the main mid-summer
crops, which alternate with wheat or barley in the rotation; forage corn also grown in summer. Six species of eucalypt and 6 species of acacia were planted on 132 ha during the 1983-87 period. However, *Eucalyptus camaldulensis* Dehnh., covers some 90% of the plantation (Kowsar, 1991). Although the trees are different in water use strategy, as the tree community mostly consists of *E. camaldulensis*, it is considered homogeneous. The remaining part of the FWS system is covered with pasture crops (Mesbah and Kowsar, 2010).

### Table 2-1. Representative weather data of Gareh Bygone station, 1996-2011.

<table>
<thead>
<tr>
<th>Rain mm</th>
<th>Temperature ˚C</th>
<th>Frost days</th>
<th>RH %</th>
<th>Wind m.s(^{-1})</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>Max.</td>
<td>Mean</td>
<td>Min.</td>
<td>Mean</td>
</tr>
<tr>
<td>Jan</td>
<td>50.2</td>
<td>24.5</td>
<td>8.0</td>
<td>-6.5</td>
</tr>
<tr>
<td>Feb</td>
<td>44.5</td>
<td>26.5</td>
<td>9.4</td>
<td>-6.0</td>
</tr>
<tr>
<td>Mar</td>
<td>34.1</td>
<td>32.0</td>
<td>13.1</td>
<td>-2.5</td>
</tr>
<tr>
<td>Apr</td>
<td>28.2</td>
<td>34.5</td>
<td>17.0</td>
<td>1.0</td>
</tr>
<tr>
<td>May</td>
<td>4.0</td>
<td>40.5</td>
<td>23.3</td>
<td>5.0</td>
</tr>
<tr>
<td>Jun</td>
<td>1.3</td>
<td>44.5</td>
<td>28.1</td>
<td>9.5</td>
</tr>
<tr>
<td>Jul</td>
<td>0.6</td>
<td>46.0</td>
<td>31.9</td>
<td>16.0</td>
</tr>
<tr>
<td>Aug</td>
<td>6.1</td>
<td>45.5</td>
<td>31.2</td>
<td>15.0</td>
</tr>
<tr>
<td>Sep</td>
<td>3.6</td>
<td>43.0</td>
<td>28.0</td>
<td>12.0</td>
</tr>
<tr>
<td>Oct</td>
<td>0.4</td>
<td>38.5</td>
<td>22.7</td>
<td>3.5</td>
</tr>
<tr>
<td>Nov</td>
<td>7.7</td>
<td>33.0</td>
<td>16.3</td>
<td>-2.0</td>
</tr>
<tr>
<td>Dec</td>
<td>38.5</td>
<td>27.0</td>
<td>10.7</td>
<td>-6.5</td>
</tr>
<tr>
<td>Yearly</td>
<td>219.2</td>
<td>38.5</td>
<td>16.6</td>
<td>-6.5</td>
</tr>
</tbody>
</table>

Temperature maximum and minimum are absolute recorded data.

A surface water reservoir covering 62 ha and created by a small earth dam was selected to extract the $ET_a$ data of open water pixels. The distribution of main land uses is presented in Fig. 2-2. The points which are visited at the field for determining the present land uses are overlaid on the land use map. Information on vegetation and its general conditions is summarized in Table 2-2. The soils are classified as Torriorthents, Haplocalcids and Haplocambids (Kowsar and Pakparvar, 2003). More detail on soils and description of the aquifer is presented by Hashemi et al. (2012).

#### 2.2.2 Model description

SEBS uses satellite and commonly available meteorological data to estimate the surface energy balance (Su, 2002):
\[ \lambda E = R_n - G - H \]  

where \( \lambda E \) is latent heat flux of evaporation, \( R_n \) is net radiation, \( G \) is soil heat flux, and \( H \) is the sensible heat flux (all in W m\(^{-2}\) units) (Allen et al. 1998).

Detailed SEBS parameterization can be found in Su et al. (2005). The input data of SEBS can be grouped into three sets:

1) Basic RS products: albedo, emissivity, thermal band \((T_{rad})\) relative to the Top of Atmosphere (ToA), which can be used as LST, and Normalized Difference Vegetation Index NDVI.

2) Vegetation products: LAI, fraction of ground covered by vegetation \((f_v\), non-dimensional\), and vegetation height \((h)\). When vegetation information is not available, NDVI is used as an independent variable to generate the vegetation products.

3) Ground observed weather data: air temperature, wind speed observed at a given height \((u, \text{ m s}^{-1})\), actual vapor pressure \((e_a)\) at a given reference height, atmospheric pressure \((P)\), sunshine duration (daily hours), all for satellite overpass time, and mean daily temperature \((T_a, ^\circ C)\) and downward solar radiation data.
Table 2-2. Vegetation characteristics of the studied land use types.

<table>
<thead>
<tr>
<th>Land use</th>
<th>Main crops</th>
<th>Agronomic dates</th>
<th>Irrigation</th>
<th>Average yield, T ha⁻¹</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Sowing</td>
<td>Harvesting</td>
<td>Max. green canopy</td>
</tr>
<tr>
<td>Farm land(1)</td>
<td>Winter crops</td>
<td>20-25 Nov.</td>
<td>5-10 Jun.</td>
<td>10-20 Apr.</td>
</tr>
<tr>
<td></td>
<td>Late summer crops</td>
<td>15-20 Aug.</td>
<td>5-10 Nov.</td>
<td>25 Sep.-20</td>
</tr>
<tr>
<td>Main plants</td>
<td></td>
<td>Germination</td>
<td>Fall</td>
<td>Max. green</td>
</tr>
<tr>
<td>Pasture(2)</td>
<td>Native perennials</td>
<td>20 Feb</td>
<td>20 Oct.</td>
<td>20 Apr.-5 Jun.</td>
</tr>
<tr>
<td></td>
<td>Native annuals</td>
<td>10-20 Jan.</td>
<td>20 May</td>
<td>20 Mar.-20 Apr.</td>
</tr>
<tr>
<td></td>
<td>Planted pastures</td>
<td>Ever green</td>
<td>20 May</td>
<td>20 Mar.-20 Apr.</td>
</tr>
<tr>
<td>Forests(3)</td>
<td>Native trees</td>
<td>Ever green</td>
<td>20 Jan.-20 May</td>
<td>&lt;1</td>
</tr>
<tr>
<td></td>
<td>Planted eucalypt</td>
<td>Ever green</td>
<td>20 Jan.-20 May</td>
<td>80-100</td>
</tr>
<tr>
<td></td>
<td>Planted acacia</td>
<td>Ever green</td>
<td>20 Jan.-20 May</td>
<td>5-10</td>
</tr>
</tbody>
</table>

1- Winter crops: wheat *Triticum aestivum* (or barley *Hordeum vulgare*); mid-summer crops: water melon *Citrullus lanatus*, melon *Cucumis melo*, cantaloupe *Cucumis melo* var. cantalupensis; late summer crop: forage corn *Zea mays*.
3- Forests, native: *Ziziphus nummularia*; planted trees: mainly *Eucalyptus camaldulensis* and *Acacia victoria*.
4- Based on the measurements in this study.
5- As reported by (Mesbah and Kowsar, 2010). FWS is floodwater spreading.
Fig. 2-2. NDVI map generated from Landsat 5 on 18/03/2010 image with indication of main land uses and the locations which are visited at the field for determining the present land uses.

Further details on the particular techniques used to separate the sensible and latent heat fluxes from the available energy are given by Su (2002).

Integrated land and water information system (ILWIS) Open 3.8.3 and the SEBS Toolbox plug-in (http://www.itc.nl/Pub/WRS/WRS-GEONETCast), which were used in this study, provide a set of routines of SEBS for biogeophysical parameter extraction. It uses satellite-based earth observation data in combination with ground-based meteorological information as inputs to produce SEBS results in the form of different maps including the actual evapotranspiration ($ET_a$).

In SEBS, in order to determine the evaporative fraction (partition of the available energy between sensible and latent), the energy balance at limiting cases are considered. Below the dry-limit, latent ($\lambda E$) heat becomes zero due to the limitation of soil moisture and sensible heat flux is at its maximum value (called $H_{dry}$).

\[
\lambda E_{dry} = R_n - G_0 - H_{dry} = 0, \text{ or } H_{dry} = R_n - G_0 \tag{2-2}
\]
At the wet-limit, the internal resistance (resistance of vapour flow through the transpiring crop and evaporating soil surface) is zero and the $H$ is minimal ($H_{\text{wet}}$).

$$\lambda E_{\text{wet}} = R_n - G_0 - H_{\text{wet}},$$

or

$$H_{\text{wet}} = R_n - G_0 - \lambda E_{\text{wet}} \quad (2-3)$$

The term relative evaporation ($EF_r$) can be then evaluated as:

$$EF_r = \frac{\lambda E}{\lambda E_{\text{wet}}} = 1 - \frac{\lambda E_{\text{wet}} - \lambda E}{\lambda E_{\text{wet}}} \quad (2-4)$$

The Eq. (2-4) might be rewritten based on the Eqs. (2-1), (2-2) and (2-3) as:

$$EF_r = 1 - \frac{H - H_{\text{wet}}}{H_{\text{dry}} - H_{\text{wet}}} \quad (2-5)$$

$H_{\text{dry}}$ is obtained by the Eq. (2-2) while $H_{\text{wet}}$ is given by substitution of Eq. (2-3) to a combination equation similar to the Penman-Monteith equation proposed by (Menenti, 1984) as:

$$\lambda E = \frac{\Delta r_e (R_n - G) + \rho C_p (e_s - e_a)}{r_e (\Delta + \gamma) + \gamma r_i} \quad (2-6)$$

were, $e_s$ is saturation vapor pressure (kPa), $e_a$ is actual vapor pressure (kPa), $\Delta$ is slope of vapor pressure curve (kPa °C$^{-1}$), $\rho$ is the mean air density at constant pressure (kg m$^{-3}$), $C_p$ is the specific heat of the air (kJ kg$^{-1}$ K$^{-1}$), $\gamma$ is psychrometric constant (kPa °C$^{-1}$), $r_e$ is external or aerodynamic resistance (function of surface characteristics and wind speed) (s m$^{-1}$), $r_i$ is internal or surface resistance (s m$^{-1}$). Considering the internal resistance $r_i=0$ by definition in wet limit in Eq. (2-6) and changing the subscripts respectively, the sensible

heat flux at the wet-limit is obtained as:

$$H_{\text{wet}} = \frac{\left[ (R_n - G) - \frac{\rho C_p (e_s - e_a)}{r_e} \right]}{r_e (\Delta + \gamma)} \quad (2-7)$$

The $r_e$ is evaluated as a function of friction velocity parameters (related to aerodynamic characteristics of the wind and height of plant cover) by numerical solution.

The evaporative fraction is eventually obtained as:
\[
EF = \frac{\lambda E}{(R_n - G)} = \frac{E_{Fr} \cdot \lambda E_{wet}}{R_n - G}
\]  \hspace{1cm} (2-8)

The actual latent heat flux can be obtained by turning over the Eq. (2-8).

The evaporative fraction \( EF \) is assumed conservative during the day, which means that the instantaneous values of the \( EF \) are identical to the 24 hours values. If the net radiation for 24 hours (\( R_n \)) is estimated from average daily meteorological parameters, then the average daily evapotranspiration can be calculated as:

\[
ET_{daily} = 8.64 \times 10^7 \times \frac{L \times R_n - \bar{G}}{\lambda \rho_W}
\]  \hspace{1cm} (2-9)

where \( \lambda \) is latent heat of vaporization (~2.45 J kg\(^{-1}\)) and \( \rho_W \) is density of water (~1000 kg m\(^{-3}\)). Moreover the soil heat flux for 24 hours (\( \bar{G} \)) is normally assumed negligible (Su, 2002).

2.2.3 Remote sensing data and basic remote sensing products

Remote sensing (RS) data used in this study were ToA radiance values acquired by Landsat Thematic Mapper (TM) sensors. The RS products are the secondary results of any level of processing made by the end users for their objective applications (e.g., albedo). The TM sensor was carried onboard of Landsat 4 and 5 from July 1982 to May 2012 with a 16-day repeat cycle, referenced to the Worldwide Reference System-2. Very few images were acquired from November 2011 to May 2012. The satellite began decommissioning activities to un-launch (make it out of the orbit) in January 2013. Landsat 4 and 5 TM image data files consist of seven spectral bands that cover the visible (bands 1-3), near-infrared (band 4), shortwave (bands 5 and 7), and thermal infrared (band 6) spectral ranges of the electromagnetic spectrum. The resolution is 30 m for bands 1 to 5 and 7. Thermal infrared band 6 was collected at 120 m, but resampled to 30 m. The approximate scene size is 170 km north-south by 183 km east-west (https://lta.cr.usgs.gov/TM). The TM images of Landsat 5 (TM5) were used as the source of RS products in this study.

A unique number is assigned to every repeat cycle of satellite as path and every captured scene along the path as row. The study site is located in an area where two adjacent Landsat paths, 161 and 162, overlap; thus, potentially successive intervals of 7 and 9 days of images are accessible. Available images (42 scenes)
from May 2009 to October 2010 for both mentioned adjacent paths of row 40 were downloaded from Glovis website (http://glovis.usgs.gov/). Time series of 32 scenes was analyzed by excluding eight cloudy and two low quality images (Table 2-3).

A representation of the changes in radiances for the thermal bands of the analyzed images is shown in Fig. 2-3. Extent of the mean and standard deviations are in harmony with the seasons. The higher values (both mean and standard deviation) are acquired in summer times and lower values in winter times. It proves that the extents of radiances are logic.

All downloaded scenes for this study have been corrected at level 1T which provides systematic radiometric and geometric accuracy by incorporating ground control points while employing a Digital Elevation Model (DEM) for topographic accuracy (more details can be found in: http://landsat.usgs.gov/Landsat_Processing_Details.php). The radiometric correction process including the conversion from digital numbers (DN) to radiance, was applied using the equations proposed by Chander and Markham (2003). The resulting values are radiances at the ToA. The images were converted to earth skin reflectance to minimize the effects of atmospheric interference sources.

![Fig. 2-3. Mean values of top of atmosphere radiation for thermal band images of non-cloudy dates during the study period (scale bars are standard deviations).](image)

It was performed through the 6S algorithm (Second Simulation of Satellite Signal in the Solar Spectrum) inside GRASS-GIS applying the “i.atcorr” module (Neteler and Mitasova, 2008). A script was written for Linux users to run the atmospheric correction module automatically for multiple Landsat TM5 images under analysis (Appendix 1).
Table 2-3. Specification of the downloaded Landsat TM5 images.

<table>
<thead>
<tr>
<th>No.</th>
<th>Date</th>
<th>Paths of Row=40</th>
<th>Quality</th>
<th>Intervals (day)</th>
<th>No.</th>
<th>Date</th>
<th>Paths of Row=40</th>
<th>Quality</th>
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<td>12</td>
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<td>161</td>
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<td>162</td>
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<td>Cloudy</td>
<td>9</td>
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<td>161</td>
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</tbody>
</table>

Some of the interval images were not available to download.

Among the RS products, albedo, emissivity and NDVI were generated by using both ToA and at-surface images to test the impact of atmospheric correction on $ET_a$ production. Thermal band ToA product ($T_{rad}$) was used as LST source.

Albedo is an integration of the surface reflectance of all of the reflective bands assuming a coefficient for every band. Weighing coefficients proposed by Tasumi et al. (2008) were used in this study. Land surface emissivity ($\varepsilon$) was produced by the Thresholds Method NDVI $^{\text{THM}}$ proposed by Sobrino et al. (2004). This method determines the emissivity values from NDVI considering three different cases:

NDVI<0.2, in which the pixel is considered as sparse vegetation or bare soil and then the emissivity is obtained from:

$$\varepsilon = 0.98 - 0.042\rho_r \quad (2-10)$$

where $\varepsilon$ is the emissivity and $\rho_r$ is reflectivity values in the red band (0.63-0.68 $\mu m$)
NDVI > 0.5, in which pixels are considered as fully vegetated, and then a constant value for the emissivity is assumed, typically of 0.99;

0.2 ≤ NDVI ≤ 0.5, in which the pixel is composed of a mixture of bare soil and vegetation, and the emissivity is then calculated as:

\[ \varepsilon = \varepsilon_v P_v + \varepsilon_s (1 - P_v) + d_\varepsilon \]  \hspace{1cm} (2-11)

where \( \varepsilon_v \) is the emissivity of the vegetation, \( \varepsilon_s \) is the soil emissivity, and \( P_v \) is the vegetation proportion obtained according to Carlson and Ripley (1997):

\[ P_v = \left[ \frac{NDVI - NDVI_{\text{min}}}{NDVI_{\text{max}} - NDVI_{\text{min}}} \right]^2 \]  \hspace{1cm} (2-12)

where \( NDVI_{\text{max}} = 0.8 \) and \( NDVI_{\text{min}} = 0.05 \), which refer to observed NDVI values for fully vegetated and bare soil pixels respectively in our study site. The \( \varepsilon_v \) and \( \varepsilon_s \) are measured respectively as 0.96 and 0.99 by Sobrino et al. (2001). The term \( d_\varepsilon \) in Eq. (2-11) includes the effect of the geometrical distribution of the natural surfaces and the internal reflections. According to Sobrino et al. (1990) this term is negligible for plain surfaces, but for heterogeneous and rough surfaces it can reach a value of 2% of the calculated emissivity. An approximation for \( d_\varepsilon \) is obtained from:

\[ d_\varepsilon = 0.014 - 0.01P_v \]  \hspace{1cm} (2-13)

The term global radiation is used in this study as instantaneous solar radiation value (W m\(^{-1}\)) which is a parameter used by SEBS for calculation of net radiation (\( R_n \)).

Global radiation at the satellite overpass time is a SEBS input parameter which can be introduced as a spatially distributed map, or as a measured value for a point which would be assigned to the whole area. Spatial distribution of the global radiation to every location depends mainly on the day of year, latitude, longitude, elevation, slop and aspect. A module inside the GRASS-GIS software is developed to map the global radiation at the basis of the mentioned physical criteria (Neteler and Mitasova, 2008). Potentially, the spatially distributed global radiation can improve the ET\(_a\) map generation but lacking in accuracy of the calculated radiation can be a source of error in final ET estimation. Therefore, two types of input global radiation data were used in this study to test which type could improve the ET\(_a\) estimation: (a) a fixed value of global radiation recorded at the satellite overpass time at Fasa Weather.
Chapter 2 - 29

Station, and (b) values generated as a map by “r.sun” module inside GRASS-GIS (GRASS, 2010) is also including the automotive running the r.sun). A manual was written to explain all procedures of downloading, processing the images and generating the above mentioned products which can be reached at: http://svn.osgeo.org/grass/grass-promo/tutorials/grass_landsat_ETa/main_document.pdf.

2.2.4 Vegetation products

LAI was produced in this study by two approaches to test their suitability for ETa estimation:

i) Using the equation proposed in SEBS in ILWIS 3.8.3 Open,

\[
LAI = \left[ \frac{NDVI(1 + NDVI)}{1 - NDVI} \right]^{0.5}
\]  

ii) Employing equations proposed by other authors for relating the LAI to the remotely sensed vegetation indices.

Many equations are proposed by different authors for a variety of climates and plant conditions (Qi et al., 2000; Colombo et al., 2003; Haboudane et al., 2004; Xavier and Vettorazzi, 2004; Yi et al., 2008; Hasegawa et al., 2010; Wittamperuma et al., 2012). The proposed equations were used separately to prepare the LAI maps for each image corresponding to the dates. The generated LAI maps were then selected for each model run to find out the best equation in accordance with improvement in estimated ETa.

Percentage of vegetation (Pv), was calculated with Eq. (2-12), whereas vegetation height (h), was calculated from NDVI (Su, 2002) when field measurements are not available:

\[
h = 1 - \left[ h_{\text{min}} + \frac{(h_{\text{max}} - h_{\text{min}})}{(NDVI_{\text{veg}} - NDVI_{\text{soil}})(NDVI - NDVI_{\text{soil}})} \right]
\]  

where \( h_{\text{min}} \) and \( h_{\text{max}} \) are the minimum and maximum vegetation heights in the region which is suggested by SEBS as 0.0012 and 2.5 m respectively. \( NDVI_{\text{veg}} \) and \( NDVI_{\text{soil}} \) are NDVI of the fully vegetated and bare soil pixels, and were set to 0.8 and 0.05, respectively. The maximum plant height in the study area refers to Eucalyptus trees, which is 5.0 m; however, using \( h_{\text{max}} = 5.0 \) when
incorporating to the height map resulted in erroneous $h$ maps. For instance, the height of crop plants was then calculated as few centimeters if the $h_{\text{max}}$ was set to 5.0 m. Therefore, $h_{\text{max}}$ in this study was set to 2.0 m corresponding to the plant height of *Atriplex* sp. stocks in the pasture area which resulted in correct values for plant height of crop lands and pastures but incorrect values for tree plantations. In order to solve this problem, vegetation height ($h$) for the tree plantations was separately determined by field measurements and through a digitized map of $h$ generated for the area with trees. The polygonized map was then rasterized and merged to the $h$ maps produced by the NDVI method for every image. SEBS was run with and without adding the tree height on the generated height map (a merged map consisting of one made with Eq. (2-15), and a digitized plant height in tree plantations area) to check the effect of tree height on $ET_a$ calculation.

The zero plane displacement height ($d_o$) and the surface roughness height for momentum ($Z_{om}$) are generally computed as a fraction of the plant height $h$. Many studies have explored the nature of the wind regime in plant canopies. Zero displacement heights and roughness lengths have to be considered when the surface is covered by vegetation. The factors depend on the crop height and architecture (Allen et al., 1998). Several empirical equations for the estimate of $d_o$ and $Z_{om}$ have been developed. In SEBS, the values for $d_o$ and $Z_{om}$ can be either entered externally as an input map or as an attribute table associated with the land use map. If no map is entered, then $d_o$ and $Z_{om}$ are calculated by the model.

A series of calculation is followed in SEBS to evaluate $Z_{om}$ and $d_o$ based on plant height, LAI, and wind speed parameters.

The tabulated values proposed by Wierenga (1993) can be used to assign appropriate values to the land uses. Considerable effort was made for producing the real time land use maps to assure the correspondence of the assigned $d_o$ and $Z_{om}$ to the right land use on the ground. A detailed classified land use map was prepared for every image on the basis of crop type and stage of growth based on available information and visual interpretation of that particular image acquisition date.
2.2.5 Ground observed weather data

The Gareh Bygone Weather Station data of temperature, humidity and wind were used as weather inputs. Two types of temperature values are requested by model, the instantaneous and the mean daily. The mean daily temperature was calculated by averaging three standard times and absolute minimum and maximum data. Air temperature and relative humidity for the time synchronized with the satellite overpass were calculated by interpolating the recorded data of 6:30 and 12:30. Wind speed measurements were used as instantaneous and maximum daily wind parameters as described in Eq. (2-19). The Fasa Weather Station data were used for sunshine hours and air pressure. All parameters were entered as unique values in SEBS as described below.

Specific humidity \( q \) (kg kg\(^{-1}\)), defined as the mass of water vapour per unit mass of moist air, was calculated as (Brutsaert, 1982):

\[
q = \frac{0.622 e_a}{P - (1 - 0.622)e_a}
\]  
(2-16)

where 0.622 is the ratio of the molecular weights of water and dry air, \( P \) is the total air pressure (hPa), and \( e_a \) is the actual vapour pressure of the air (hPa) calculated as:

\[
e_a = e_s \cdot RH / 100
\]  
(2-17)

where RH is the relative humidity (%) and \( e_s \) is the saturated vapor pressure (hPa), which can be estimated based on temperature \( (T \, ^\circ C) \), (Monteith and Unsworth, 2008):

\[
e_s = \left[10^{-8.07 - \left(\frac{0.63}{7 + 233.43}T\right)}\right]
\]  
(2-18)

values of \( q \) in this study varied from 0.003 to 0.014 kg kg\(^{-1}\) which are in the range of published values, e.g., 0.004 to 0.013 (Ma et al., 2012) or 0.009 (Su, 2002).

The wind speed parameter in SEBS, alike the other remote sensing based methods, is based on an instantaneous measurement at the time of the satellite overpass. As inferred by Allen et al. (2005) the assumption of constant evaporative fraction can underestimate 24-h \( ET \) in arid climates where advection often increases with wind and may increase \( ET \) in proportion to \( R_n \). In order to test the importance of maximum daily rather than instantaneous...
wind speed on daily $ET_a$, the following wind speed adjusted $u_{adj}$ function was adopted in this study:

$$u_{adj} = u_{inst} + \left( \frac{u_{max}}{u_{inst}} \right)$$

(2-19)

where $u_{inst}$ is instantaneous wind speed (m s$^{-1}$) at the time of satellite overpass, and $u_{max}$ is maximum daily wind speed (m s$^{-1}$).

Incorporating maximum daily wind speed rather than using average values might be justified owing to the nature of wind occurrence in the study area. A tornado type of wind normally occurs in summer, mostly in the afternoon. Its effect on displacing water vapor is not considered when the instantaneous wind speed is not incorporated in the model. Average daily wind speed cannot truly reflect the impact of high wind speed occurrence. Thus Eq. (2-19) adds a weighted value to the $u_{inst}$ which directly reflects high wind speed if it has occurred. Wind speed data measured at a height of 5 m were converted to that at 2 m height using a logarithmic wind speed profile (Allen et al., 1998). Both sets of data ($u_{inst}$ and $u_{adj}$) were used as wind parameters separately to test the effect of maximum daily wind speed for improving $ET_a$ estimation.

### 2.2.6 Sensitivity of SEBS to the input parameters

Analysis of sensitivity helps to study how the uncertainty in the model output can be allocated to uncertainty in the model inputs by quantifying the sensitivity of the model output to systematic changes in the model input (Loosvelt, 2013).

It is hypothesized in this study that SEBS results can be improved by replacing some of the values derived from different sources of products or from the literature. Before doing so, it is needed to determine to which parameters the model output is most sensitive and thus should be changed accordingly. Therefore a sensitivity analysis was performed on the input maps (surface temperature, emissivity, NDVI, albedo, DEM, $d_o$ and $Z_{0m}$) and weather data (temperature, wind speed, air pressure, relative humidity, global radiation). The reason for selecting the mentioned input parameters is that they are directly incorporated to the model, or are used to generate further dependent maps. Sensitivity of the SEBS model was tested for the sensible heat flux as its most important output product.
As proposed by van der Kwast et al. (2009), the sensitivity \( (S_i) \) for a SEBS input is defined as:

\[
S_i(H \pm) = \left( \frac{H \pm - H_0}{H_0} \right) \times 100
\]

\( S_i \) is calculated for a positive or a negative deviation of an input of the SEBS model. \( H_0, H^+ \) and \( H^- \) are the sensible heat flux predicted by SEBS when the input equals its reference value \( i_0 \), \( 1.25 \times i_0 \) and \( 0.75 \times i_0 \), respectively, with reference values used for all other inputs. A deviation of 25\% was chosen according to the observed coefficient of variations (not shown data) of the input parameters to assure the coverage of the errors which may arise in input data collection. The range of 25\% was also selected by van der Kwast et al. (2009) for all of the SEBS input parameters and a range of 50\% was selected for \( z_{om} \) by Wang et al. (2012). For mean daily air temperature, however, a deviation of 1\% was used, since a 25\% deviation exceeds its natural limits which normally happens in the study area. The sensitivity was analyzed for all land use types; irrigated crops (IC) forage corn and wheat, tree plantation (TP), pasture inside the FWS (PI), pasture outside the FWS (PO), water reservoir (WR), and bare soil (BS) separately and the average effect on each land cover class.

### 2.2.7 Model calibration

Calibration is a process of tuning the model for the particular problem by changing the sources of the input parameters and initial or boundary conditions within reasonable range until the simulated variable closely matches the observed variable (Šimůnek et al., 2013). In order to calibrate the model through improving the results of SEBS, different types of input maps and values were prepared and the model was run for every scenario of parameter combination. When examining the different sources of a particular parameter, all of the other parameters were kept unchanged to see the improvement in the model output due to that particular source.

The sources of parameters and products used in this study are listed in Table 2-6. A massive number of different possible parameters and analyzed images (32) were used for every model run.

Since this study was performed in a site where ground truth \( ET_a \) data for comparison to the simulated \( ET_a \) was lacking, we used the \( ET_o \) calculated by
FAO Penman Monteith (FAO P-M) as the ‘observed’ ET based on following reasons. The philosophy of FAO in selecting the Penman-Monteith (P-M) methods as globally applicable reference method was that “physics are physics everywhere”. Thus, if the primarily physics-based Penman-Monteith method is set up correctly using high quality climatic measurement data from a handful of locations, it should sufficiently serve as a basis for crop ET globally (Pereira et al., 2015). Various sensitivity analyses and regional studies confirm its applicability to a large variety of environments (Gong et al., 2006; Nandagiri and Kovoor, 2006; Estévez and Gavilán, 2009; Ye et al., 2009; Raziei and Pereira, 2013). Using FAO P-M calculated \( E\text{To} \) as observed data was used in many researches in data-scarce regions (Jabloun and Sahli, 2008; Landeras et al., 2008; Estévez and Gavilán, 2009; Maeda et al., 2011; El Tahir et al., 2012).

Besides representing the climate through grass reference evapotranspiration, \( ET_\text{a} \), \( ET_\text{o} \) includes the effects of the crop type through the crop coefficient (\( K_c \)), and of soil-water stress through a stress coefficient (\( K_s \)) (Allen et al., 1998):

\[
ET_\text{a} = K_s K_c E\text{To}
\]

(2-21)

\( K_c \) varies with crop variety, crop growth stage, crop density and management and, only to a limited extent, with climate (Allen et al. 1998) and is influenced by surface residue and mulching (Martins et al., 2013). Therefore, the estimated \( ET_\text{a} \) of pixels located on the water reservoir (WR) inside the study area was selected for comparison against calculated \( ET_\text{o} \) considering that a \( K_c \) of free water with a depth larger than 2 m approaches 1.05 (Allen et al., 1998). The \( K_s \) in open water reservoir can be assumed as the unit due to absence of any type of stress.

Daily \( ET_\text{o} \) (mm day\(^{-1}\)) was calculated with variables observed at Gareh Bygone weather station with the FAO P-M method (Allen et al., 1998):

\[
ET_\text{o} = \frac{0.408\Delta(R_n - G) + \gamma \frac{900}{T+273} u_2 (e_s - e_a)}{\Delta + \gamma(1 + 0.34u_2)}
\]

(2-22)

where \( R_n \) is net radiation at the reference crop surface (MJ m\(^{-2}\) day\(^{-1}\)), \( G \) is soil heat flux (MJ m\(^{-2}\) day\(^{-1}\)), \( T \) is mean daily air temperature at 2 m height (°C), \( u_2 \) is wind speed at 2 m height (m s\(^{-1}\)), \( e_s \) is saturation vapor pressure (kPa), \( e_a \) is actual vapor pressure (kPa), \( \Delta \) is slope of vapor pressure curve (kPa °C\(^{-1}\)), \( \gamma \) is psychrometric constant (kPa °C\(^{-1}\)). The \( e_s \) and \( e_a \) were calculated in this step as
proposed by Allen et al. (1998). Differently as described in section 2.2.5, the parameters $e_r$ and $e_a$ were calculated based on Eq. (2-17) and (2-18) for the SEBS application.

The coefficient of determination ($r^2$) and the root mean square error (RMSE) were used to compare the SEBS estimated $ET_a$ against FAO P-M calculated $ET_o$. The relative standard error (RSE), ratio of the standard error to the mean expressed as a percentage (Sokal and Rohl, 1995) was used to assess the global radiation which is obtained by the two sources; GRASS-GIS module “r.sun” and observed weather station. Estimates with a RSE of 25% or greater are subject to high sampling error and should be used with caution (ABS, 2014).

2.2.8 Evaluation of the results

2.2.8.1 Global radiation
The values assigned to the pixels of the global radiation map generated by GRASS module “r.sun” located in the place of the Fasa Weather Station were extracted to compare them with the measured radiation. Since this station appeared only in the 162-40 satellite scenes, only 18 pairs were compared.

2.2.8.2 Justification of estimated $ET_a$ among the land uses
The pixels located in different land uses were extracted from the time series of $ET_a$ maps and the weighted average of $ET_a$ was calculated for every land use. The $ET_a$ time series were then plotted with regard to the land uses in order to examine how the temporal changes follow the seasons in a rational way. The tree plantations (TP) were separated to dense, semi dense and sparse vegetation to check the variation of estimated $ET_a$ in accordance with the vegetation density of forest area. The estimated $ET_a$ for the variety of TP vegetation and the other land uses were compared to the published values of $ET_a$ for that particular land use and vegetation density.

2.2.8.3 Cross validation of the estimated $ET_a$
The robustness of the modeled results was further tested through cross validation of the estimated $ET_a$. In cross validation, a model is usually given a dataset of known data on which training is run, and a dataset of unknown data against which the model is tested. One round of cross validation involves partitioning a sample of data into complementary subsets, performing the analysis on one set (called the training set), and validating the analysis on the other subset (called validation or test set). To reduce variability, multiple
rounds of cross-validation are performed using different partitions, and the validation results are averaged over the rounds. A k-fold cross validation (KFCV) was used in this study as described by (Arlot and Celisse, 2010). KFCV relies on a preliminary partitioning of data into K subsamples. Each subsample successively plays the role of validation sample and the remainder samples as training group. The set of 32 time series of $ET_a$ was split into flour partitions (25%), every set used to derive the regression equation and the rest of data set (75%) to test its validity and the procedure was repeated in the other rounds to involve all of the partitions as training set. The Standard Error of Estimate (SEE) for the full data set and the Standard Error of Prediction (SEP) and bias values for each cross validation analysis were calculated by using the actual and predicted values according to formulas in Meek et al. (1999) and Montgomery and Peck (1982):

$$SEE = \left( \frac{1}{n-1} \sum_{i=1}^{n} e(i)^2 \right)^{0.5}$$  \hspace{1cm} (2-23)

$$SEP = \left( \frac{1}{n-1} \sum_{i=1}^{n} \left[ (y_{\text{measured}} - y_{\text{predicted}} - \text{bias})(i) \right]^2 \right)^{0.5}$$  \hspace{1cm} (2-24)

$$Bias = \frac{1}{n} \sum_{i=1}^{n} (y_{\text{measured}} - y_{\text{predicted}})$$  \hspace{1cm} (2-25)

For SEE, $n$ is number of observations (32), $i$ is the observation number, and $e(i)$ is the residual term (measured-predicted) at each observation. For SEP, $y_{\text{measured}}$ is the measured ET value from data set 2, while $y_{\text{predicted}}$ is calculated ET from the regression of data set 1; bias is the mean deviation of these values from zero over the original data set.

### 2.2.8.4 Water balance budgeted ET cross validation

The SEBS $ET_a$ is also cross validated with the water balance budgeted ET under the main land uses, farm land, tree plantations, pastures and bare soils. The water budget can be written as Liu et al. (2012):

$$\Delta S = P + I + F - Ro - RF - ET_a$$  \hspace{1cm} (2-26)

where $\Delta S$ is the change in storage, $P$ is precipitation, $I$ is irrigation applied water, $F$ is floodwater entered to the systems, $Ro$ is runoff, $RF$ is return flow, and $ET_a$ is actual ET. When assuming that there is no significant change in storage from year to year ($\Delta S = 0$), $ET_a$ can be computed by reverting the Eq.
(2-26). The $ET_a$ calculated in this way therefore provides an independent benchmark to be compared with modelled seasonal $ET_a$ by SEBS.

i) Farm fields

Two main crops, i.e., wheat and forage corn, were accounted for water budget cross validation. On farm lands, there is no runoff as the irrigation system is close-ended and no excess water exits from individual fields. Floodwater is not a source of applied water, as the fields are located outside the FWS boundary.

The volume of irrigation water ($I$) was measured in 12 farm fields of wheat and 10 forage corn cultivations in the area during the growth season of the hydrologic year 2010-2011. To do so, irrigation discharge was measured in the head ditch above each field using a cut-throat flume (Walker, 1989). This was then converted to seasonal volume of applied water per hectare using the time of application for each turn, the number of irrigations during the season, and the surface area irrigated. To determine the size of area of irrigated farm fields, maps called “cultivated farm field map” were produced based on farm field map for each image by employing NDVI. The NDVI value of cultivated farm fields (non-fallow) of every image date was found based on cultivation time table and expert knowledge. This value was used as a criterion for excluding the non-cultivated pixels of farm fields (i.e. fallow). Resulted maps were used for determining the size of cultivated area of wheat and forage corn based on their cultivation calendar (Fig. 2-4).

![Fig. 2-4. Presentation of cultivated farm fields maps for wheat (left) as main winter crop and forage corn (right) as main summer crop.](image)

The size of the cultivation area for each crop type was determined and then multiplied by the amount of applied water in m$^3$ per hectare (m$^3$ ha$^{-1}$) for that
crop type to find out the total volume of irrigation applied water (i.e. Mm$^3$) in the whole area for that particular year.

Effective rainfall during the cropping season was accounted for precipitation ($P$) component. Effective rainfall refers to the percentage of rainfall which becomes available to plants and crops excluding losses by evaporation of intercepted rainwater, and runoff. As runoff did not occur in the farm fields, effective rainfall was determined by the proposed method by Smith (1992) as implemented in CROPWAT model:

\[
P_{eff} = 125 \cdot \frac{(125 - 0.2P)}{125} \quad \text{for } P \leq 250
\]

\[
P_{eff} = 125 + 0.1P \quad \text{for } P > 250
\]  

(2-27)

where $P_{eff}$ is effective precipitation and $P$ is total precipitation in the cropping season).

As for the return flow ($RF$), because of lack of data, values published in literature for similar crops and irrigation system were considered. For corn, Sabol et al. (1987) measured a RF of 21% in the dry states of the USA under surface irrigation. For wheat, Jafari et al. (2012) reported an average of 24% for a semi-arid region of central Iran under the border irrigation and for soils similar to the farm fields of our study site.

In order to calculate the total $ET_a$ modelled by SEBS, in a cropping season, the $ET_a$ maps generated by SEBS were intersected by each cultivated farm fields’ map, to calculate the weighted average of $ET_a$ for every image. Since the $ETo$ was available as daily base, the ratio $ETo/ET_a$ for the dates of available images were used as a multiplier to corresponding $ETo$ to calculate the $ET_a$ of the interval days. Then the daily $ET_a$ values were summed up for the cropping season for wheat and corn separately.

ii) Land uses outside the FWS

In pastures outside the FWS (PO) and bare soils (BS), the only input water is precipitation ($P$); therefore, the input components $I$ and $F$ are not involved in the water budget Eq. (2-26). The runoff ($R$) was neglected for the reason that the soil surface of the area out of FWS system is loamy sand in texture characterized by high rate of infiltration. According to the measured data
reported by (Kowsar and Pakparvar, 2003) in this study area, the average infiltration rate of natural pastures and bare soils is 7.7 cm hr\(^{-1}\) and maximum rainfall intensity is measured by Dr. S.A Kowsar (personal conversation) as 1.3 cm hr\(^{-1}\), which is by far (7 times) less than the infiltration rate. Besides, the surface sloping of the study site is flat or gentle (less than 1%), which does not generate enough potential energy for producing the runoff. Therefore, no runoff occurs during the rainfall events.

Amount of recorded rainfall during the period of the SEBS modelled \(ET_a\) time series was assigned to the \(P\) component. The effective precipitation was considered as equal to the recorded rainfall due to the sparse vegetation of these land uses. The average canopy cover of >5% and >1% for PO and BS, respectively, is reported by Mesbah and Kowsar (2010) (Table 2-2); therefore, the effect of plant foliage on water interception is negligible.

The \(RF\) is assumed to be 30% of the rainfall based on the reports of Ahmadi et al. (2013) in the Neishaboor Plain in Iran. Their study area is similar in climate (annual precipitation and temperature of 234 mm and 13 °C, respectively) and plant canopy cover as our study site. Then the \(ET_a\) was calculated as the residual of the Eq. (2-26).

Similar procedure, which was used for the farm fields, is applied to the PO and BS land uses to determine the summed up SEBS model \(ET_a\). The only difference was that the period of summation for farm fields was the cropping season for every crop (wheat and forage corn) and the entire \(ET\) maps time series for PO and BS land uses.

**iii) Land uses inside the FWS**

The main difference between the land uses inside the FWS, tree plantations (TP) and pastures inside the FWS (PI), in terms of water budget is presence of floodwater as the source of input water. These land uses are inundated with the floodwater and remain pounded until infiltrating the water to the soil profile.

As the pastures and tree plantations are mixed together inside the FWS systems, the amount of floodwater \((F)\) which is reached to PI and TP area cannot be separated. Therefore the data for main input water in these land uses for the water budget was missing.
Furthermore, *E. camaldulensis* trees which cover more than 90% of TP, are the phreatophyte plants. A phreatophyte is a deep-rooted plant that obtains a significant portion of the needed water from the phreatic zone (zone of saturation) or the capillary fringe immediately above the phreatic zone.

The field observations in this study showed that the roots of these trees explore as deep as the GW table (Photo 2-1). Consequently, the other component of water budget exists for TP land uses as GW uptake. Making separation between the surface and GW sources of water consumption of these trees is complicated and a reliable assessment can be hardly achieved.

Hence, the validation of the SEBS $ET_a$ for the PI and TP land uses was limited to justification of resulted $K_c$ through the published values (following section).

### 2.2.8.5 Justification of the modelled $K_c$

The calibrated model was used for checking the $K_c$ of different land use types inside the region. The $K_c$ values were the ratios between $ET_a$ estimated by SEBS and $ETo$. They were computed for several land uses including irrigated crops, *E. camaldulensis* plantations and pastures, and were compared with $K_c$ published in literature. The average $ET_a$ estimations for the land uses were converted to $K_c$ and the resulted temporal changes in $K_c$ were justified in comparison with published $K_c$ for similar conditions as in our study area.

Wheat was selected as winter crop and forage corn as summer crop because of their dominance in the study area. Other summer crops (Table 2-2) were ignored due to their reduced acreage. During the initial stage of crop growth $K_c$ mostly depends upon soil water and wetting events. In final (end) stages $K_c$ depends on crop management, e.g., purpose of harvesting. Thus initial and end $K_c$ ($K_{c,ini}$ and $K_{c,end}$) cannot easily be compared with published $K_c$. The satellite overpass dates which coincided with the maximum canopy were selected to calculate midseason $K_c$ ($K_c,mid$).

Where rainfall or irrigation is low, water stress might be induced and the evapotranspiration will drop below the standard crop evapotranspiration. The reduction in the value for $K_c$ under conditions of low soil water availability is determined using the stress coefficient $K_s$ (Eq. (2-21)). In absence of measured $K_s$, as an approximation, the $K_c$ during the mid-season stage is for crops that usually nearly completely shade the soil under pristine conditions, but where plant cover is reduced due to disease, stress, pests, or planting density ($K_s$ less
than unit), the values for $K_{c \mid mid}$ can be reduced by a factor depending on the actual vegetation development (Allen et al., 1998):

$$K_{c \text{ adj}} = K_{c \mid mid} - A_{cm}$$

(2-28)

with the empirical adjustment coefficient $A_{cm}$, given by

$$A_{cm} = 1 - \left( \frac{LAI_{\text{stress}}}{LAI_{\text{dense}}} \right)^{0.5}$$

(2-29)

where $K_{c \text{ adj}}$ is the adjusted $K_{c \mid mid}$, $LAI_{\text{dense}}$ is LAI for a crop having appropriate ground cover density, or maximum density, and $LAI_{\text{stress}}$ is the actual LAI of the crop when submitted to stress. Using this approach, $A_{cm} = 0$ when a crop is not stressed and the $A_{cm} > 0$ means the $K_c$ less than unit (Allen et al. 1998). In this study LAI in Eq. (2-29) was calculated based on NDVI data by using Eq. (2-30).

Values for NDVI$_{\text{dense}}$ were extracted from a well-managed irrigated farm belonging to the Agricultural Research Station of Darab, (28º 47’N, 54º19’E) located at 44 km northwest from the study area, at the same altitude and with similar climate and cropping season. Soil conditions and water quality are non-limiting at this farm, where it produces the highest wheat and corn yield in the region.

Crop coefficients for each crop type were finally compared with published $K_c$ to validate the SEBS model estimation similarly to procedures followed by Pôças et al. (2013). When the midseason crop coefficients estimated by SEBS are close to the published $K_c$ it means that $ET_a$ estimation is adequate.

2.3 Results and discussion

2.3.1 Model sensitivity to input parameters

Table 2-4 shows the parameters for the SEBS sensitivity. The parameters to which the model is low sensitive ($S_i < 10\%$) are surface elevation (DEM), emissivity ($\varepsilon$), NDVI, albedo ($\alpha$), relative humidity (RH), $d_o$ and $Z_{0m}$ and the global radiation. It was expected, because the derivation of sensible heat flux requires only meteorological parameters at reference height (2 m) and surface temperature. This means that the calculation of $H$ in SEBS is independent of other surface energy balance terms in contrast with most other models (Su, 2002). Although the basic RS products are not directly measured and are
subjected to some processing errors, they showed minimum impact to the model output error. The parameters to which the $H$ modelled by SEBS is sensitive ($S_i > 10\%$) are wind speed ($u$), air temperature ($T_a$) and air pressure ($P$). These parameters measured at the weather station, directly used in the calculation of sensible heat flux and are subjected to minimum measuring errors.

Most parameters show a comparable sensitivity at different land uses. Exception to this is the global radiation (Table 2-5). Sensible heat flux at the pastures outside the FWS and bare soils are especially sensitive to a small error in global radiation. Global radiation is pertinent to the calculation of $R_n$ through separation of shortwave and longwave. Because of spars vegetation in those mentioned land uses, the absorbance of the incoming shortwave radiation ($K_{down}$) by the green canopy is its minimum. The values of $H$ outside the dry and wet limits ($H_{dry}$ and $H_{wet}$) occur when the iteration in the submodel for the derivation of stability parameters (Su, 2001) does not converge. This occurs with some bare surfaces when varying the $K_{down}$. The $H$ calculation is less sensitive to global radiation in rich vegetation land uses such as forage corn and tree plantations.

Table 2-4. Input parameters of SEBS sensitivity for sensible heat flux ($H$).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>$S_i(H+)$%</th>
<th>$S_i(H-)$%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Albedo ($\alpha$)</td>
<td>-0.6</td>
<td>0.1</td>
</tr>
<tr>
<td>DEM</td>
<td>-0.7</td>
<td>0.7</td>
</tr>
<tr>
<td>Emissivity ($\varepsilon$)</td>
<td>-0.4</td>
<td>0.1</td>
</tr>
<tr>
<td>Global radiation</td>
<td>0</td>
<td>-8.7</td>
</tr>
<tr>
<td>NDVI</td>
<td>0.2</td>
<td>-0.4</td>
</tr>
<tr>
<td>Relative humidity (RH)</td>
<td>-0.3</td>
<td>0.4</td>
</tr>
<tr>
<td>$d_0$ and $Z_{0m}$</td>
<td>-0.9</td>
<td>0.9</td>
</tr>
<tr>
<td>Wind ($u$)</td>
<td>-12.7</td>
<td>12.7</td>
</tr>
<tr>
<td>Air temperature ($T_a$)$^{11}$</td>
<td>-13.2</td>
<td>13.4</td>
</tr>
<tr>
<td>Air pressure ($P$)</td>
<td>-18.3</td>
<td>17.2</td>
</tr>
</tbody>
</table>

$^{1}$-The deviation is set to 25% for all parameters except the $T_a$ as 1%. H+ indicates a positive deviation and H− indicates a negative deviation applied to the input parameter. DEM is Digital Elevation Model, NDVI=Normalized Difference Vegetation Index.
Table 2-5. Sensitivity of sensible heat flux to global radiation in different land uses.

<table>
<thead>
<tr>
<th>Land uses</th>
<th>$S_i(H^+)$%</th>
<th>$S_i(H^-)$%</th>
</tr>
</thead>
<tbody>
<tr>
<td>IC (wheat)</td>
<td>0</td>
<td>-5</td>
</tr>
<tr>
<td>IC (forage corn)</td>
<td>3</td>
<td>0</td>
</tr>
<tr>
<td>TP</td>
<td>4</td>
<td>0</td>
</tr>
<tr>
<td>PI</td>
<td>0</td>
<td>-8</td>
</tr>
<tr>
<td>PO</td>
<td>0</td>
<td>-17</td>
</tr>
<tr>
<td>BS</td>
<td>0</td>
<td>-21</td>
</tr>
<tr>
<td>WR</td>
<td>0</td>
<td>-3</td>
</tr>
</tbody>
</table>

$S_i$ is sensitivity, $H^+$ indicates a positive deviation (25%) and $H^-$ indicates a negative deviation (25%) applied to the input parameter, IC is irrigated crops, TP is tree plantations, PI and PO are pastures inside and outside the floodwater spreading respectively, BS is Bare soils and WR is water reservoir.

2.3.2 Model calibration

The sources of different input parameters which were chosen to run the model and test the improvement in results due to that particular change are summarized in Table 2-6.

The goodness of fit when predicting the $ET_a$ was not substantially improved in most scenarios. Those scenarios that result in noticeable improvement are presented in Table 2-7.

Comparisons were made using the data relative to pixels located on the water reservoir, thus where the estimated $ET_a$ approximates $ET_o$. In Table 2-7, the values of $r^2$ and RMSE illustrate the extent of agreement between the estimated $ET_a$ relative to the water reservoir pixels with the calculated FAO P-M $ET_o$ using different sources of parameterizations. The change in sources of parametrization was made to generate all of the possible scenarios which include of all possible combination of the various sources of the input parameters. According to Table 2-7, the first scenario resulted in the $r^2$ of 0.53 and RMSE of 1.26 mm day$^{-1}$. By replacing global radiation sources from weather station point to GRASS generated map in scenario 2 the $r^2$ is improved to 0.62 and RMSE to 1.13 mm day$^{-1}$. Improvements in prediction were obtained when replacing ToA with at-surface products, leading to an increase of 0.01 in the $r^2$ and a decrease in RMSE as 0.01 mm day$^{-1}$ (scenario 3 in Table 2-7), so inferring a minor positive impact of atmospheric correction in this study. In scenario 4 an improvement is shown by replacing the $d_o$ and $z_{om}$ source from SEBS imbedded equations to tabular method proposed by Wiernga (1993).
Table 2-6. Sources of products and parameters for calibrating SEBS.

<table>
<thead>
<tr>
<th>Products and Parameters</th>
<th>Sources</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basic RS products</td>
<td></td>
</tr>
<tr>
<td>Albedo</td>
<td>ToA</td>
</tr>
<tr>
<td>Emissivity</td>
<td>ToA</td>
</tr>
<tr>
<td>LST</td>
<td>ToA</td>
</tr>
<tr>
<td>NDVI</td>
<td>ToA</td>
</tr>
<tr>
<td>Vegetation products</td>
<td></td>
</tr>
<tr>
<td>LAI</td>
<td>SEBS eq.</td>
</tr>
<tr>
<td>Height</td>
<td>SEBS eq.</td>
</tr>
<tr>
<td>$f_c$</td>
<td>SEBS eq.</td>
</tr>
<tr>
<td>$d_0$ and $Z_0m$</td>
<td>Tabular</td>
</tr>
<tr>
<td>Climatic parameters</td>
<td></td>
</tr>
<tr>
<td>Wind</td>
<td>$U_{inst}$</td>
</tr>
<tr>
<td>Pressure</td>
<td>Instantaneous</td>
</tr>
<tr>
<td>Temp</td>
<td>Instantaneous</td>
</tr>
<tr>
<td>$q$</td>
<td>Instantaneous</td>
</tr>
<tr>
<td>Global radiation</td>
<td>W. station</td>
</tr>
</tbody>
</table>

LST is land surface temperature, ToA is top of atmosphere reflectance, at-surface is atmospherically corrected (at-surface) reflectance, LAI is leaf area index, SEBS eq. is the embedded equation in model, pub.eq. is generated LAI map with using different equations proposed in the published works, TP h is the measured tree heights of tree plantation area, $f_c$ is vegetation fraction, $d_0$ and $Z_0m$ are the displacement height and the roughness height for momentum, tabular is tabulated values by Wiernga (1993) for the different land uses, $U_{inst}$ is instantaneous wind speed, $U_{adj}$ is adjusted wind speed, $q$ is specific humidity. W. station is weather station, GRASS is the GRASS-GIS r.sun module.

Enhancement in the $r^2$ and RMSE in scenarios 5 and 6 shows the improvement due to replacement of the SEBS equation by the one proposed by Wittamperuma et al. (2012) and by Xavier and Vettorazzi (2004), respectively. Among the equations proposed in the literature (Qi et al., 2000; Colombo et al., 2003; Haboudane et al., 2004; Xavier and Vettorazzi, 2004; Yi et al., 2008; Hasegawa et al., 2010; Wittamperuma et al., 2012), the one published by Xavier and Vettorazzi (2004) resulted in maximum enhancement in the $r^2$ and RMSE.

$$LAI = a \times NDVI^b$$  \hspace{1cm} (2-30)

where a and b are regression parameters, which varied for the months under investigation.

Major improvements (changes in the $r^2$ and RMSE as 0.09 and -0.1 respectively) were further obtained in scenario 7 by replacing the height maps...
from SEBS eq. +TP h (merging the map generated by SEBS equation and tree plantation height map) with the map that was generated only by SEBS equation (Eq. (2-15)). Finally, the highest improved values are obtained in scenario 8 by substituting the wind parameter from $U_{\text{inst}}$ to $U_{\text{adj}}$ (Eq. (2-19)) with the final $r^2$ of 0.91 and RMSE of 0.76 mm day$^{-1}$. The best scenario was therefore achieved by application of GRASS r.sun for global radiation, at-surface product0, LAI derived by Eq. (2-30), tabular method for $d_0$ and $Z_0$m, h map by SEBS equation and using $u_{\text{adj}}$ as wind parameter (Fig. 2-5).

In order to show the impact of incorporating the maximum daily wind speed (in terms of $u_{\text{adj}}$), the results of SEBS with $u_{\text{inst}}$ and $u_{\text{adj}}$ are presented in Fig. 2-6. The maximum differences between graphs of SEBS $ET_a$ (extracted from water reservoir pixels) and FAO-PM $ET_o$ occurred in July and August, mainly in 2009.

According to Fig. 2-6, after incorporating maximum daily wind speed the differences became substantially smaller for these months (what is also inferred from $r^2$ and RMSE changes from scenarios 7 and 8 in Table 2-7). The largest deviation between SEBS $ET_a$ and FAO P-M $ET_o$ was observed on 6th July 2009 (7.29 vs. 9.32 mm day$^{-1}$ respectively). The maximum wind speed on this day was 12.3 m s$^{-1}$ as compared to the instantaneous satellite overpass time wind speed of 3.5 m s$^{-1}$. After considering the maximum wind speed, identical corresponding results (9.15 vs. 9.32 mm day$^{-1}$) are obtained.
Table 2-7. Selected scenarios for calibrating SEBS based on improvement in approaching the estimated \( \text{ET}_a \) to calculated \( \text{ETo} \) in water reservoir pixels.

<table>
<thead>
<tr>
<th>Statistics</th>
<th>\text{FAO P-M ETo, mm day}^{-1}</th>
<th>SEBS ( \text{ET}_a ) of water reservoir pixels, mm day(^{-1})</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1</td>
<td>2</td>
</tr>
<tr>
<td>Mean</td>
<td>6.5</td>
<td>5.7</td>
</tr>
<tr>
<td>Std. dev.</td>
<td>2.4</td>
<td>2.2</td>
</tr>
<tr>
<td>Minimum</td>
<td>2.2</td>
<td>1.7</td>
</tr>
<tr>
<td>Maximum</td>
<td>9.8</td>
<td>8.8</td>
</tr>
<tr>
<td>No of samples</td>
<td>32</td>
<td>32</td>
</tr>
<tr>
<td>( R^2 )</td>
<td>0.53</td>
<td>0.62</td>
</tr>
<tr>
<td>RMSE (mm day(^{-1}))</td>
<td>1.3</td>
<td>1.1</td>
</tr>
</tbody>
</table>

Sources for scenarios

- Global radiation: W. station GRASS GRASS GRASS GRASS GRASS GRASS GRASS
- RS product: ToA ToA at-surf at-surf at-surf at-surf at-surf at-surf
- \( d_0 \) and \( Z_0 \): SEBS eq. SEBS eq. SEBS eq. Tabular Tabular Tabular Tabular Tabular
2.3.3 Justification of $ET_a$ among the land uses

Descriptive data of temporal and spatial changes in the $ET_a$ for various land uses are presented in Table 2-8. The $ET_a$ values are in the range of reported $ET_a$ values for arid lands. For example, a thorough investigation was made by Raziei and Pereira (2013) for Iran in which the location of study area on their
resulted map has an average of 1800-2000 mm yr\(^{-1}\) or 5-5.5 mm day\(^{-1}\). Our \(ET_o\) values shows an average of 5.9 and a range between 2.2 to 9.5 mm day\(^{-1}\) in respect with the seasons.

Temporal changes in predicted \(ET_a\) from all main land uses logically follow the seasons. A minimum \(ET_a\) in January is followed by an increasing trend until maximum values occur in June-October, which is followed by a decreasing trend until the next year's minimum in January. As example, two representatives final \(ET_a\) maps of winter and summer conditions are presented in Fig. 2-7.

Variations of the \(ET_a\) for various land use types are also shown in Table 2-8. \(ET_a\) of tree plantations remains higher than for the other land use types in all dates except those corresponding to high demand periods of irrigated crops, e.g., by 18/03/2010 and 03/04/2010. The \(ET_a\) values from pastures located inside the FWS system (PI) are generally higher than \(ET_a\) from pastures outside FWS (PO).
Fig. 2-7. Actual ET ($ET_a$) maps of the Gareh Bygone Plain generated by SEBS for (A) cold season 14/02/2010 and (B) warm season, 07/08/2010.
Table 2-8. Descriptive statistics of the actual evapotranspiration computed from SEBS averaged for image dates and land-use types.

<table>
<thead>
<tr>
<th>Dates</th>
<th>FAO-PM ET₀</th>
<th>IC</th>
<th>TP</th>
<th>PI</th>
<th>PO</th>
<th>WR</th>
<th>BS</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(mm day⁻¹)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>11/05/09</td>
<td>7.9</td>
<td>5.2</td>
<td>4.9</td>
<td>3.5</td>
<td>2.7</td>
<td>8.1</td>
<td>1.3</td>
</tr>
<tr>
<td>18/05/09</td>
<td>7.2</td>
<td>4.5</td>
<td>4.1</td>
<td>2.7</td>
<td>2.0</td>
<td>7.9</td>
<td>1.0</td>
</tr>
<tr>
<td>27/05/09</td>
<td>7.7</td>
<td>5.0</td>
<td>4.8</td>
<td>3.5</td>
<td>2.8</td>
<td>8.8</td>
<td>1.4</td>
</tr>
<tr>
<td>03/06/09</td>
<td>7.6</td>
<td>3.5</td>
<td>2.9</td>
<td>1.8</td>
<td>1.0</td>
<td>8.3</td>
<td>0.5</td>
</tr>
<tr>
<td>19/06/09</td>
<td>8.4</td>
<td>5.0</td>
<td>4.7</td>
<td>3.4</td>
<td>3.0</td>
<td>8.3</td>
<td>1.5</td>
</tr>
<tr>
<td>28/06/09</td>
<td>8.6</td>
<td>4.9</td>
<td>4.1</td>
<td>3.0</td>
<td>2.4</td>
<td>9.3</td>
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<tr>
<td>14/07/09</td>
<td>9.8</td>
<td>5.9</td>
<td>6.2</td>
<td>4.6</td>
<td>4.1</td>
<td>8.6</td>
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</tr>
<tr>
<td>21/07/09</td>
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<td>5.7</td>
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<td>0.9</td>
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<td>1.2</td>
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<td>0.4</td>
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<td>1.8</td>
<td>1.4</td>
<td>3.7</td>
<td>0.7</td>
</tr>
<tr>
<td>13/01/10</td>
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<td>2.4</td>
<td>2.8</td>
<td>2.3</td>
<td>2.1</td>
<td>3.0</td>
<td>1.0</td>
</tr>
<tr>
<td>22/01/10</td>
<td>2.4</td>
<td>1.2</td>
<td>1.9</td>
<td>1.7</td>
<td>1.4</td>
<td>3.1</td>
<td>0.7</td>
</tr>
<tr>
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<td>1.3</td>
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<td>1.5</td>
<td>2.7</td>
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<tr>
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<td>1.5</td>
<td>4.2</td>
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<tr>
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<td>5.3</td>
<td>2.9</td>
<td>1.6</td>
<td>0.6</td>
<td>6.1</td>
<td>0.3</td>
</tr>
<tr>
<td>03/04/10</td>
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<td>6.1</td>
<td>3.3</td>
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<td>1.2</td>
<td>6.8</td>
<td>0.7</td>
</tr>
<tr>
<td>19/04/10</td>
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<td>2.4</td>
</tr>
<tr>
<td>28/04/10</td>
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<td>0.4</td>
<td>7.7</td>
<td>0.3</td>
</tr>
<tr>
<td>08/07/10</td>
<td>8.6</td>
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<td>6.3</td>
<td>4.8</td>
<td>4.4</td>
<td>9.4</td>
<td>2.2</td>
</tr>
<tr>
<td>12/10/10</td>
<td>6.1</td>
<td>2.7</td>
<td>2.5</td>
<td>1.4</td>
<td>1.0</td>
<td>5.9</td>
<td>0.5</td>
</tr>
<tr>
<td>21/10/10</td>
<td>5.9</td>
<td>2.1</td>
<td>1.9</td>
<td>1.0</td>
<td>0.6</td>
<td>5.0</td>
<td>0.3</td>
</tr>
<tr>
<td>Mean</td>
<td>6.5</td>
<td>4.0</td>
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<td>2.8</td>
<td>2.3</td>
<td>6.7</td>
<td>1.2</td>
</tr>
<tr>
<td>St. dev.</td>
<td>2.6</td>
<td>1.5</td>
<td>1.7</td>
<td>1.4</td>
<td>1.4</td>
<td>2.2</td>
<td>0.7</td>
</tr>
</tbody>
</table>

Statistics are the average data of all representative points of each land use (illustrated in Fig. 2-2). FAO P-M ET₀ is FAO Penman-Monteith reference crop ET, IC is irrigated crops; TP istree plantations, PI is pastures inside the floodwater spreading (FWS); PO is pastures outside the FWS, WR is water reservoir and BS is bare soils. St. dev. Is standard deviation.

Our results show a range of 1.2 to 7.4 and an average of 3.6 mm day⁻¹ for Eucalyptus tree plantations. Edraki et al. (2007) observed ET₀ for the same tree plantations site in Gareh Bygone Plain (E. camaldulensis Dehnh.) in three plots, each having 9 six years old trees. They performed a soil-water budget for 180 cm depth using a neutron probe in the period from 31 March to 27 August 1991. The reported daily ET₀ ranged 0.2 to 6.1 mm day⁻¹ and 0.1 to 7.5 mm day⁻¹ for two plots; results for the third one show large errors. The SEBS calculated daily ET₀ values for the tree plantations ranged between 1.2 to 7.4
mm.d-1 (Table 2-8), thus in agreement with referred results by Edraki et al. (2007). In a more extended overview made by Albaugh et al. (2013) from the works in Africa, a range of 1.5 to 7.5 and an average yearly of 3.2 mm day\(^{-1}\) is reported for different Eucalyptus species. Roberts and Rosier (1993) reported transpiration rates of 1.0 to 5.5 mm day\(^{-1}\) for two year old \(E. \textit{camaldulensis}\) trees in Bangalore, India. Cramer et al. (1999) reported transpiration measured with the sap flow method in Queensland, Australia, in 4-6 years old \(E. \textit{camaldulensis}\) trees. Reported results ranged from 1.0 to 4.5 mm day\(^{-1}\) depending on the season and density of plantations. Since results refer to transpiration only, their lower value as compared to our SEBS derived data is pertinent to being transpiration merely.

### 2.3.4 Validation of the results

#### 2.3.4.1 Global radiation

Since Fasa Weather Station was located out of the 161-40 scene, the number of available observations of radiation was reduced to the images with non-cloudy conditions in the 162-40 scene. Thus 21 scenes were available for radiation analysis (Fig. 2-8).

![Fig. 2-8. Global radiation estimates by r.sun module in GRASS vs. the measured data at Fasa Weather Station.](image)

An overview of the statistics is presented in Table 2-9. The GRASS estimates of global radiation show a good agreement with the measured radiation at Fasa Weather Station, by a coefficient of determination \(r^2\) of 0.84 (P<0.0001) and RSE of 6.40%. RSE is by far less than the maximum acceptable (RSE=25%) and the \(r^2\) is meaningful at the P<0.01 thus, both statistics demonstrate the good performance of GRASS r.sun module to estimate global radiation.
52 - Remote sensing of evapotranspiration

Table 2-9. Comparing global radiation estimates from the “r.sun” GRASS module and the Fasa Weather Station.

<table>
<thead>
<tr>
<th>Statistics</th>
<th>Global radiation, W m(^{-2})</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Observed at Fasa weather station</td>
</tr>
<tr>
<td>Mean</td>
<td>863</td>
</tr>
<tr>
<td>Maximum</td>
<td>1016</td>
</tr>
<tr>
<td>Minimum</td>
<td>615</td>
</tr>
<tr>
<td>Standard Deviation</td>
<td>129</td>
</tr>
<tr>
<td>No. samples</td>
<td>18</td>
</tr>
<tr>
<td>Standard Error</td>
<td>53</td>
</tr>
<tr>
<td>Relative Standard Error</td>
<td>6.40%</td>
</tr>
<tr>
<td>(R^2)</td>
<td>0.84</td>
</tr>
</tbody>
</table>

2.3.4.2 Cross validation of modelled \(ET_a\)

Actual vs. predicted values of cross validation pointed close to the 1:1 line (Fig. 2-9). SEE of the full data set is calculated as 0.75 mm day\(^{-1}\) and the SEPs were 0.72 and 0.83 mm day\(^{-1}\) for the split data sets, comparable to the SEE for the complete data set, and bias values were low, -0.12 and -0.18 mm day\(^{-1}\) (<2% of mean values) for the split data sets. Therefore, modelled \(ET_a\) shows enough robustness for predicting the \(ET_a\) close to the calculated \(ET_a\) by FAO P-M method.

Fig. 2-9. Measured vs. predicted \(ET_a\) (1:1 line) of the cross validated data.

2.3.4.3 Water balance budgeted \(ET\) cross validation

i) Farm fields
Amount of water consumption is presented in Table 2-10. In the cropping season of December 2009 to October 2010, from the 2200 ha of farm fields in the study area, 1234 ha was used as winter crops and 716 ha as summer crops and the rest was remained fallow. Total applied water for the cultivated farms is calculated as 14.17 million m³ (Mm³) and the summed up ETₐ for this period as 10.97 Mm³. The components of water budget for the mentioned cropping season are presented in Table 2-11. All components are quantitatively determined except for the RF, which is based on the published work in similar study sites (Jafari et al., 2012) to be 24% of total input water (P+I). Therefore, the net input water is determined as 10.77 Mm³, which accounts for the ETₐ (see Eq. (2-26)). In this way, the difference between the modelled ETₐ by SEBS (10.97 Mm³) and the water balance budgeted ETₐ (10.77 Mm³) is 0.26 Mm³, a difference as 2.3%. Hence, the modelled ETₐ is close enough to be defined as an adequate estimation.

Table 2-10. Water consumption of the farm fields in the cropping season Dec. 2009 to Oct. 2010.

<table>
<thead>
<tr>
<th>Crops</th>
<th>area (ha)</th>
<th>Irrigation (m³/ha⁻¹)</th>
<th>Effective rainfall (mm)</th>
<th>Total (Mm³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Winter crops</td>
<td>1234</td>
<td>4800</td>
<td>100</td>
<td>5.92</td>
</tr>
<tr>
<td>Summer crops</td>
<td>716</td>
<td>9800</td>
<td>0.0</td>
<td>7.02</td>
</tr>
<tr>
<td>Sum</td>
<td></td>
<td></td>
<td></td>
<td>14.17</td>
</tr>
</tbody>
</table>

SEBS ETₐ

<table>
<thead>
<tr>
<th>Area (ha)</th>
<th>Seasonal ETₐ (m³/ha⁻¹)</th>
<th>Total (Mm³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Winter crops</td>
<td>1234</td>
<td>3900</td>
</tr>
<tr>
<td>Summer crops</td>
<td>716</td>
<td>8600</td>
</tr>
<tr>
<td>Sum</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

1-Measured applied irrigation water in the field by cut throat flume.
2-Estimated, based on recorded rainfall using the Smith (1992) method. Amount of rainfall is recorded as 123 mm.
3-Mm³ is million m³.

<table>
<thead>
<tr>
<th>P (mm)</th>
<th>I (mm)</th>
<th>F (mm)</th>
<th>Ro (mm)</th>
<th>RF (mm)</th>
<th>WB ETₐ (mm)</th>
<th>SEBS ETₐ (mm)</th>
<th>Difference WB-SEBS (mm)</th>
<th>Ratio WB-SEBS to ETₐ</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.23</td>
<td>12.94</td>
<td>0</td>
<td>0</td>
<td>3.40</td>
<td>10.76</td>
<td>10.97</td>
<td>0.21</td>
<td>1.02</td>
</tr>
</tbody>
</table>

P is effective precipitation estimated based on recorded rainfall using Smith (1992) method, I is irrigation, F is floodwater, Ro is runoff, RF is return flow calculated as 24% of total input water (here; P+I) based on the published data in the similar condition of the study site (Jafari et al., 2012), WB is water budget and Mm³ is million m³.

**ii) Land uses outside the FWS**

The pastures and bare area outside the FWS systems (PO and BS) covered 6430 ha of the study area where the only input water was annual precipitation and no runoff occurred in this area. A total amount of rainfall of 124 mm was recorded in the period when the time series of images were processed from May 2009 to October 2010; this was considered as effective precipitation. Hence, 6.43 Mm³ of rain water, P, component was received in this area. The RF is defined to be the 30% of input water based on the reported values in the Neishabooor Plain in Iran, which is similar to our study site (Ahmadi et al., 2013). Consequently, the net input water of 5.58 Mm³ remained to be consumed as ETₐ. Besides, the summation of SEBS ETₐ for the same period is calculated, based on weighted average of every ETₐ map, as 292 mm. According to the size of area it is equal to 18.77 Mm³. Water budget components can be summarized as Table 2-12. It is clear that the SEBS has overestimated the ETₐ in these land uses. This can be ascribed to a lack of input water as the source to be used as ET. This fact is reported by recent works (Gibson et al., 2011; Lu et al., 2013).

Table 2-12. Water budget components of the image time series May 2009 to Oct. 2010 for pastures and bare soils outside the FWS.

<table>
<thead>
<tr>
<th>P (mm)</th>
<th>I (mm)</th>
<th>F (mm)</th>
<th>R (mm)</th>
<th>RF (mm)</th>
<th>WB ETₐ (mm)</th>
<th>SEBS ETₐ (mm)</th>
<th>Difference WB-SEBS (mm)</th>
<th>Ratio WB-SEBS to ETₐ</th>
</tr>
</thead>
<tbody>
<tr>
<td>6.43</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>1.93</td>
<td>5.58</td>
<td>18.77</td>
<td>13.19</td>
<td>3.4</td>
</tr>
</tbody>
</table>

P is effective precipitation estimated based on recorded rainfall using Smith (1992) method, I is irrigation, F is floodwater, R is runoff, RF is return flow calculated as 30% of total input water (here; P) based on the published data in the similar condition of the study site (Ahmadi et al., 2013), WB is water budget and Mm³ is million m³.
As described in the section 2.2.8.4, the water budget approach could not be followed for the landuses inside the FWS because of the uncertainties concerning with the separation of the tree plantations and pastures and due to water consumption of the \textit{E. camaldulensis} as a phreatophyte plant. Hence, validation of the SEBS $ET_a$ for the PI and TP land uses was limited to justification of the resulted $K_c$ through the published values.

\textbf{iii) Crop coefficients}

While $ET_a$ varies enormously in time and space and is not comparable among different locations, average crop coefficients may be transferable from one site to another. Thus, they are comparable when averaged (Allen et al., 1998). In this study, the daily $K_c$ values were obtained but comparisons are only valid when referring to the averaged values. However, there is the need to understand when extreme values may be justified by local or regional advection or if they result from SEBS overestimation of the $ET_a$. That possible overestimation is recently noted by some authors, e.g. Lu et al. (2013).

Results of the daily $K_c$ calculations for winter wheat and forage corn are presented in Table 2-13. The calculated $K_c$ in last column is obtained from the RS data using Eq. (2-28). The daily values for winter wheat $K_{c \text{mid}}$ ranged 1.13-1.34, averaging 1.21. The maximum daily $K_{c \text{mid}}$ value was 1.34, which is above the expected averaged $K_{c \text{mid}}$ for wheat, but is likely to occur in a day when atmospheric demand is high, mainly if advection occurs.

Zaitchik et al. (2007) reported positive advective heating rates in plateaus around the Zagros Mountain Ranges, where our study site is situated. Advection was also reported by Malek (1987) relative to the Fars Province, the region where the study area is located, particularly from April to September. The study area has hot, dry and windy days during the midseason of wheat (Table 2-13).

It is surrounded by large areas of sparse and dry natural vegetation where the sensible heat flux, $H$, largely dominates over the latent heat flux, $\lambda E$, thus contributing to produce heated air transported by wind into the cultivated areas. Assuming the advection definition of McIlroy and Angus (1964), and considering the information reported by Malek (1987) and Zaitchik et al. (2007), advection is likely to occur in the area. Because the $ET_o$ is computed considering only vertical energy fluxes, advection is considered in $K_c$ values.
Remote sensing of evapotranspiration

(Hence, the averaged \( K_{c \text{ mid}} \) value of 1.21 compares well with the typical values reported in the literature; ranges 1.08 to 1.19 which are reported in several studies referring to a variety of climates (Bandyopadhyay and Mallick, 2003; Li et al., 2003; Kjaersgaard et al., 2008; Gao et al., 2009; Ko et al., 2009; López-Urrea et al., 2009; Zhao et al., 2013).

Table 2-13. Daily estimates of midseason crop coefficients obtained from SEBS derived \( E_{Ta} \) and FAO P-M \( E_{To} \).

<table>
<thead>
<tr>
<th>Crops</th>
<th>Dates</th>
<th>FAO P-M ( E_{To} ), mm day(^{-1})</th>
<th>SEBS ([2]) ( E_{Ta} ), mm day(^{-1})</th>
<th>( K_c )</th>
<th>NDVI</th>
<th>NDVI(_{\text{dens}})</th>
<th>( A_{cm} )</th>
<th>( K_{c \text{ adj}} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wheat</td>
<td>23/02/10</td>
<td>3.89</td>
<td>4.21</td>
<td>1.0</td>
<td>0.74</td>
<td>0.82</td>
<td>0.05</td>
<td>1.13</td>
</tr>
<tr>
<td></td>
<td>18/03/10</td>
<td>5.58</td>
<td>5.91</td>
<td>1.0</td>
<td>0.7</td>
<td>0.88</td>
<td>0.11</td>
<td>1.17</td>
</tr>
<tr>
<td></td>
<td>3/4/10</td>
<td>5.84</td>
<td>6.78</td>
<td>1.1</td>
<td>0.83</td>
<td>0.86</td>
<td>0.02</td>
<td>1.18</td>
</tr>
<tr>
<td></td>
<td>19/04/10</td>
<td>5.22</td>
<td>6.88</td>
<td>1.3</td>
<td>0.75</td>
<td>0.79</td>
<td>0.03</td>
<td>1.34</td>
</tr>
<tr>
<td></td>
<td>Average</td>
<td>5.13</td>
<td>5.95</td>
<td>1.1</td>
<td>0.76</td>
<td>0.84</td>
<td>0.05</td>
<td>1.21</td>
</tr>
<tr>
<td></td>
<td>StDe</td>
<td>0.22</td>
<td>0.45</td>
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</tr>
<tr>
<td>Forage corn</td>
<td>7/9/2009</td>
<td>8.06</td>
<td>6.63</td>
<td>0.8</td>
<td>0.66</td>
<td>0.78</td>
<td>0.08</td>
<td>0.91</td>
</tr>
<tr>
<td></td>
<td>16/09/09</td>
<td>6.85</td>
<td>5.89</td>
<td>0.8</td>
<td>0.67</td>
<td>0.82</td>
<td>0.10</td>
<td>0.96</td>
</tr>
<tr>
<td></td>
<td>2/10/09</td>
<td>5.03</td>
<td>6.30</td>
<td>1.2</td>
<td>0.72</td>
<td>0.85</td>
<td>0.08</td>
<td>1.33</td>
</tr>
<tr>
<td></td>
<td>9/10/09</td>
<td>5.69</td>
<td>5.29</td>
<td>0.9</td>
<td>0.67</td>
<td>0.81</td>
<td>0.09</td>
<td>1.02</td>
</tr>
<tr>
<td></td>
<td>18/10/09</td>
<td>4.82</td>
<td>5.37</td>
<td>1.1</td>
<td>0.62</td>
<td>0.74</td>
<td>0.08</td>
<td>1.19</td>
</tr>
<tr>
<td></td>
<td>Average</td>
<td>6.09</td>
<td>5.896</td>
<td>0.9</td>
<td>0.66</td>
<td>0.8</td>
<td>0.08</td>
<td>1.08</td>
</tr>
<tr>
<td></td>
<td>StDe</td>
<td>1.36</td>
<td>0.58</td>
<td>0.1</td>
<td>0.04</td>
<td>0.04</td>
<td>0.01</td>
<td>0.17</td>
</tr>
</tbody>
</table>

1- Averaged data from 10 fields with proper management.

FAO P-M is FAO Penman-Monteith reference crop \( E_T \), \( K_{c \text{ adj}} \) is \( K_c \) adjusted for water stress, NDVI\(_{\text{dens}}\) is Normalized Difference Vegetation Index for non-stress dense vegetation, \( A_{cm} \) is correction factor for adjustment of the \( K_c \).

Niazi et al. (2005) reported \( K_{c \text{ mid}} \) values of 1.09 to 1.13 at Zarghan, also in Fars Province, but with a less dry climate and located at a higher altitude (1621 m.a.s.l.), which justifies the small \( K_c \) difference (in addition to adopted crop varieties and adopted practiced cropping techniques).

The \( K_{c \text{ mid}} \) for forage corn averaged 1.08 while daily estimated values varied 0.91-1.33 (Table 2-13). Similar to what occurred with the wheat \( K_{c \text{ mid}} \), the high daily value of 1.33 is likely to be due to advection, which is stronger during the maize crop season than that occurring during the wheat season. Several authors have reported similar but generally larger values than the estimated \( K_{c \text{ mid}} = 1.08 \) for grain maize; e.g., Allen et al. (1998) proposed a value of 1.20,
Gao et al. (2009), Martinez-Cob (2008) and Piccinni et al. (2009) reported 1.19-1.20, while Zhao et al. (2013) obtained a value of 1.15. Values similar to the one obtained in this study are reported by Rosa et al. (2012) and Liu and Pereira (2000), 1.10 and 1.08, respectively. Gheysary et al. (2006) reported $K_{c\, mid}$ of 1.13 for a study implemented at Varamin, south of Tehran with the same mean temperature and annual rainfall. It is therefore likely that the values obtained are appropriate. As the extent of seasonal modelled $ET_a$ was validated by adapted water budget cross validation, minor differences in $K_c$ term may also be due to the crop variety and management practices used.

The time series of $K_c$ for TP land uses presented in Fig. 2-10a refers to four locations having different density of vegetation, from very dense to sparse. As mention in study area description, *Eucalyptus camaldulensis* is predominant species in FWS project, which covers more than 90% of forested area. Furthermore, as Kowsar (1991) has explained in detail, majority of the trees are planted in 1983-1984. The main important difference is related to the accessibility of different sites to the floodwater and plantation density. The minimum $K_c$ derived by SEBS is ranged from 0.10 to 0.43 in cold season, maximum from 1.13 to 1.6 in warm season and the average of 0.42 to 0.83. The $K_c$ values increase with higher plant density. The pixels of very dense vegetation refer to an area where water is available for most part of the year due to flooding. The same extent of $K_c$ is reported for Eucalyptus species for instance, in non-stressed environments, the literature reports the maximum $K_c$ for a full cover of Eucalyptus trees between 1.2 to 1.5 in dry areas (Stibbe, 1975; Sharma, 1984; Grattan et al., 1996; Myers et al., 1999). $K_c$ between 0.1 to 1.2 is reported by Alves et al. (2013) in Brazil for Different species of Eucalyptus. However, a maximum $K_c$ of 0.83 is reported in an arid area with 12 years old *E. camaldulensis* irrigated by saline water (EC=10 dS m$^{-1}$ and concentration of B as 12 mg L$^{-1}$). This lower $K_c$ value was attributed to combined salt and B stress (Dong et al., 1992). In most studies the $K_c$ peaks is simultaneous with the irrigation times and an abrupt fall occurs after irrigation. In contrary, the peak of $K_c$ in our study was occurred in summer seasons when the access to soil surface water content is in minimum. It proves that the Eucalyptus trees in our study site have access to the permanent source of water (i.e. GW table) and their transpiration continued in warm seasons. Nevertheless, the maximum peak of $K_c$ (1.13 to 1.6) is obtained on the January 2010 when a flooding event has occurred showing that the access to readily
available water has intensified the transpiration. Referring to the above comparisons, it can be concluded that the modelled $K_c$ values for tree plantations by SEBS in our study are in the range of published one.

Temporal changes in $K_c$ values for pasture lands are presented in Fig. 2-10b. Peaks are happened for the same dates as eucalyptus trees. Maximum $K_c$ values of 1.27 and 1.10 and average of 0.47 and 0.39 for PI and PO, respectively are obtained. For irrigated pastures (similar to PI), the $K_c$ values found by Pôças et al. (2013) using RS data were 0.88–0.89, thus lower than those obtained in this study. Various studies reported by these authors also show $K_c$ lower than most of values shown in Fig. 2-10.

Fig. 2-10. Time series of $K_c$ values calculated for tree plantations with varying vegetation density (A) and pastures inside and outside of the FWS project (B).

However, values for non-irrigated pastures are smaller, for instance, (Aase et al., 1973) report $K_c$ of 0.55 native vegetation with sparse cover ground in
Montana. A $K_c$ of 0.45 is reported for steppe grasslands of Mongolia (Yang and Zhou, 2011). The $K_c$ for the pasture outside the FWS also seems to be overestimated. Some recent investigations have similarly proposed overestimation of $ET_a$ by SEBS in the conditions of water deficit in soil surface as the source of $ET$ (Gibson et al., 2011; Lu et al., 2013). An overview of the resulted $K_c$ relative to the land uses is presented in Table 2-14.

### Table 2-14. Descriptive range of $K_c$ resulted by SEBS relative to the published $K_c$.

<table>
<thead>
<tr>
<th>Plant specification</th>
<th>$K_c$ this study</th>
<th>$K_c$ Published in dry regions</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wheat</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Forage corn</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tree plantations</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pastures</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Inside FWS</td>
<td>0.09</td>
<td>0.47</td>
</tr>
<tr>
<td>Outside FWS</td>
<td>0.05</td>
<td>0.39</td>
</tr>
</tbody>
</table>

#### 2.4 Conclusions

The use of SEBS model for the arid landscape of southern Iran was performed with calibration of the model through improved parameterization. Due to the impact of high velocity winds in the area an adjustment was done to consider the maximum daily wind speed. After improved parameterization, SEBS $ET_a$ relative to water reservoir pixels compared well with the reference $ETo$.

The model showed minor sensitivity ($S_i < 10\%$) to the RS products, the vegetation input parameters, DEM and relative humidity and high sensitivity ($S_i > 10\%$) to weather data $T_a$, $u$, $P$. This is in the agreement with other reports (Su et al., 2005; van der Kwast et al., 2009)

Cross validation analyses showed that the predictions made by modelled $ET_a$ is statistically robust with the comparable SEE and SEPs and with low bias. The cross validation of seasonal $ET_a$ showed minor difference between modelled and water budget $ET_a$ for wheat and forage corn crops (ratio of 1.02), but major discrepancy for the pastures and bare soils outside the FWS systems (ratio of 3.4). Therefore, an overestimation in $K_c$ value for sparse vegetation pastures and bare soils area might be inferred from this study.
While $ET_a$ varies in time and space and is not comparable among different locations, crop coefficients may be transferable from a site to another one. Thus, time averaged $K_c$ are comparable. Daily $K_c$ values were computed and therefore, only their averaged values for the mid-season could be compared with those reported in the literature.

Moreover, extreme $K_c$ might be considered non-erroneous since they could be due to local and regional advection. Water consumption by cultivated crops based on SEBS results is compared well with that calculated by measured applied water consequence a reasonable calculated irrigation returned flow. Thus, the SEBS estimation of $ET_a$ and consequent $K_c$ is useful for performing an adequate assessment of the floodwater spreading project as well as to be used further in water management in the region. SEBS can now be applied as a tool for monitoring the impact of various land use scenarios in the study area. However, it is advisable that ground observations be further developed to better assess SEBS and control uncertainties in its parameterization.

In the hydrological year 2010-2011, according to cross validated results the two methods; SEBS and water budget, main crops wheat and forage corn have consumed ~11 million m$^3$ (Mm$^3$) both from GW and rainfall. An amount of 1 Mm$^3$ has been used by means of evapotranspiration by the TP land use area. As the major sources of both land uses is the GW, therefore it can be inferred that the agricultural activity is by predominant water consumer in the study area and the role of forested area (mainly by *E. camaldulensis*) plays minor role in water consumption.
Photo 2-1. The deep roots of *Eucalyptus camaldulensis* which is cut during the excavation inside the 28 m deep well inside floodwater spreading systems in Gareh Bygone Plain, Iran.
Chapter 3. Evaluating the effect of floodwater spreading on groundwater recharge by combining water table fluctuation (WTF) and water budget methods in saturated zone
3.1 Introduction

Iran is a land of floods and droughts. The ancient desert-dwelling Persians had discovered that floodwater is a vital source, which, if harvested, could greatly enhance the yield and quality of their rainfed crops. Further, they had realized that debris cones and coarse-grained alluvial fans, which abound in Iran, were the best places to build their living quarters and develop their farm fields and orchards. It has been hypothesized that the seepage face development downstream of the flood-irrigated farms led to the invention of “Qanat”, the most important contribution of Persians in developing water harvesting techniques (Kowsar and Kowsar, 2012). As modern technology, rotary drilling machines and powerful pumps have caused a drastically recession of water tables, and since groundwater (GW) supplies some 60% of water requirements in Iran (Mohammadnia and Kowsar, 2003), resorting to artificial recharge of groundwater (ARG) is of utmost importance as the country’s very survival is at stake. The increased demand for GW and over-exploitation of this vital resource is likely to become a serious challenge for future development at GW basins of central and northeast Iran (Motagh et al., 2008). Hence, total water resources per capita in Iran have plunged by more than 65% during the last four decades, and are expected to decrease by another 16% by 2025 (Sarraf et al., 2005). Land subsidence, which is defined as a gradual settling or sudden sinking of the Earth's surface owing to subsurface movement of earth materials, resulting from high rates of GW depletion is the other side of emerging concerns in arid regions of Iran (Davoodijam et al., 2015).

According to the diverse objectives and methods of implementing floodwater spreading (FWS) system, various factors need to be considered when choosing a method of quantifying recharge. Therefore, the rate of aquifer recharge is one of the most difficult elements to measure in the evaluation of GW resources (Sophocleous, 1991). Classification of the techniques used to quantify recharge is somewhat arbitrary. Scanlon et al. (2002) divided the techniques in three main groups, unsaturated zone, saturated zone and surface water (see chapter 1). The saturated zone techniques are subdivided in Darcy’s law, tracing (physical methods), numerical modelling, water table fluctuation (WTF), and water budget methods (Scanlon et al., 2002). The latter two methods will be applied in this chapter. The unsaturated zone techniques will be evaluated in chapter 5. Surface water techniques fall beyond the scope of this dissertation.
Chapter 3 - 65

The WTF method has been used by many authors (Weeks and Sorey, 1973; Gerhart, 1986; Hall and Risser, 1993). Evaluation of the influence of GW on plant cover (Danyar et al., 2004), the effect of urbanization on GW dynamics (Hoque et al., 2007; Naik et al., 2008), optimization of water supply and demand (Ahmad et al., 2010), and recharge volume identification (Crosbie et al., 2005; Yu and Chu, 2012) are recent examples of using the WTF method. The attractiveness of the WTF method lies in its simplicity and the ease of use. As it measures water level in an observation well and is thus representative of an area of at least several square meters, the WTF method can be addressed as an integrated approach and less a point measurement than those methods that are based on data in the unsaturated zone (Healy and Cook, 2002). This method is strictly dependent on the specific yield ($S_y$) assigned to the aquifer which can be assumed as a source of error in the estimated recharge. Specific yield of a rock or soil, is defined as the ratio of the volume of water that, after saturation, can be drained by gravity to its own volume (Todd and Mays, 2005). The method is based on the premise that the rises in GW level in unconfined aquifers is due to recharge water arriving at the water table. Change in GW storage is calculated as:

$$\Delta S = S_y \frac{\Delta h}{\Delta t} \quad (3-1)$$

where $\Delta S$ is change in GW storage, $S_y$ is specific yield, $h$ is water-table height, and $t$ is time (Scanlon et al., 2002). Eq. (3-1) assumes that the water arriving at the water table goes immediately into storage and that all other components in the GW budget including $ET$, and in and out flow from the water table, are zero (Healy and Cook, 2002). This typically occurs during the period of recharge so, Eq. (3-1) is applicable for short windows of time when the magnitude of recharge is high, the vertical flow is significant in comparison with the lateral underground flow, and for each individual water table rise. For long term identification of recharge (e.g. a hydrological year), the long term change in GW budget components must be considered when using WTF (Healy and Cook, 2002). A general budget equation for the saturated zone which is modified based on Hoque et al. (2007) can be written as:

$$E = -\Delta S + R + RF - ET^{gw} + (Q_{on}^{gw} - Q_{off}^{gw}) \quad (3-2)$$

where $E$ is extraction from the aquifer, $\Delta S$ is change in GW storage (or change in saturated pore volume), $R$ is recharge of GW, RF is agricultural return flow, $Q_{on}^{gw}$ and $Q_{off}^{gw}$ are subsurface flow on and off and $ET^{gw}$ is $ET$ from the GW.
table, all on a volume basis. $ET^{gw}$ can be is neglected as the deep water tables in arid regions do not allow direct $ET$ from the aquifer. Unless deep roots take up the water from GW table. Similar to the short term assumption in Eq. (3-1), the components of subsurface flow $Q_{on}^{gw}$ and $Q_{off}^{gw}$ can be neglected in the long term as well because of a balance between the lateral input and output water to the aquifer. Recharge can be natural and artificial. Natural recharge pertains to diffusion from upland and adjacent aquifers and water infiltration in river beds whilst artificial recharge is that influenced by an engineered structure to generate or intensify the recharge of GW). With $\Delta S$ calculated by WTF (Eq. (3-1)), recharge can then be determined as the remainder of Eq. (3-2). The saturated zone water budget components are illustrated in Fig. 3-1.

Fig. 3-1. Schematic of water budget for the saturated zone. Extraction is any type of direct withdrawal of groundwater; e.g. pumping, $ET$ is evapotranspiration and $RF$ is agricultural return flow.

The initial success of the ARG project by FWS at the Kowsar Floodwater Spreading and Aquifer Management Research, Training and Extension station (Kowsar Station) in a desert in the Gareh Bygone Plain in southern Iran resulted in a magnificent increase in the area of irrigated farm fields downstream of the infiltration basins. However, over-exploitation of GW caused an decline of the water table level which resulted in the abandonment of some wells. In this study, the impact of the FWS on water level fluctuation was therefore investigated through comparisons of observation wells inside and outside of the FWS system during different flooding events. The objectives of this study were: a) to assess the temporal and spatial trend of GW level fluctuation as influenced by the FWS system of interest; b) to evaluate its impact on the GW level during individual events, and c) to quantify the proportion of natural and artificial recharge.
3.2 Materials and Methods

3.2.1 Site description

General specifications of the study area are presented in chapter 2. The FWS systems are implemented in Gareh Bygone Plain on a debris cone deposited by the ephemeral streams Bisheb Zard (Yellow Marsh) and Tchah Qooch (the Ram of the Well), that are tributaries to the Shur (Salty) River and that drains a sub-basin of the Mond River Basin (Fig. 3-2a and b). The upland basins of these two rivers are 194 ad 177 km² in extent, respectively, both comprise 6% of the Shur River contributing area of about 6400 km², or only 0.7% of the 48,400 km² of the Mond River Basin. The upland Basins are located on a northwest to southeast syncline formed by the tectonic movements of the Zagros Mountain Ranges during the Mio-Pliocene time in the Agha Jari Formation (James and Wynd, 1965). As described by Kowsar (1991) the size of Quaternary coarse alluvium sediment deposits that formed the aquifer ranges from coarse (debris cone) to small grains (fan). The debris cone is terminated on its western extremity by the Shur River, an effluent, perennial stream which flows southward in the thalweg of the Gareh Bygone Plain (Fig. 3-2c). The Agha Jari Formation, one of the most widespread geologic Formations in south–southwestern Iran, ranges in age from Late Miocene to Pliocene (Table 3-1). This formation consists of rhythmically interbedded brown to grey, calcareous, feature-forming sandstones and low weathering, gypsum-veined, red marls and grey to green siltstones. The Agha Jari Formation lies over the grey marls and limestones of the Mishan Formation, which is of Early to Middle Miocene in age. Although the Agha Jari Formation is usually capped by the Plio-Pleistocene Bakhtyari Formation, severe erosion during the Quaternary period has left only small, scattered patches of the Bakhtyari Formation on the upland Basins.
Fig. 3-2. Location of the study site in Iran (A) and on the Mond Basin and the Shur sub-basin (B). Location of floodwater spreading (FWS) systems on the Gareh Bygone Plain (GBP) and its upland Basins (C); and sketch map of FWS systems and the Kowsar Station on the GBP (D). The image is Landsat TM5 false color composite of bands 7-4-2 dated 21/10/2010. BZ is Bisheh Zard, RA is Rahim Abad and TQ is Tchah Qooch.
Table 3-1. The stratigraphy of the Gareh Bygone Plain and its upland Basins.

<table>
<thead>
<tr>
<th>Formation Name</th>
<th>Age</th>
<th>Descriptions</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lower portion of Gachsaran</td>
<td>Miocene</td>
<td>Ferrous and ferric multicolored low weathering marls, marly limestone, superficial evaporate sulfates (more gypsum and less anhydrite) and gypseous dolomite</td>
</tr>
<tr>
<td>Upper portion of Gachsaran</td>
<td>Miocene</td>
<td>Superficial chemical deposits gypsum and anhydrite, silty and marly limestone, ferrous and ferric marl and gypseous bearing limestone</td>
</tr>
<tr>
<td>Agha Jari</td>
<td>Miocene-Pliocene</td>
<td>Feature forming, fine- to coarse-grain sandstone, diamictic and grit stone and mainly marls, marly sandstone, siltstone polymictic micro-conglomerate and conglomerate in top of portion.</td>
</tr>
<tr>
<td>Mishan</td>
<td>Miocene</td>
<td>Ferruginous marls with inter-bedded sandstone and sandy limestone.</td>
</tr>
<tr>
<td>Bakhtyari</td>
<td>Pliocene-Pleistocene</td>
<td>Polymictic, proximal, heterogeneous imbricate cherty conglomerate and low weathering lutite.</td>
</tr>
<tr>
<td>Quaternary system</td>
<td>Pleistocene-Holocene</td>
<td>Detrital fan, polygenetic, containing piedmont colluvial deposits consisting immature and clastic boulder to gravel-size clastic, clay and sand, polymictic conglomerate, alluvial and colluvial deposits of major intermittent watercourses and sheet washes.</td>
</tr>
</tbody>
</table>


The Bakhtyari Formation, which mainly consists of pebbles and cobbles (conglomerate) of Cretaceous, Eocene, and Oligocene limestone and dark brown, ferruginous chert (James and Wynd, 1965), has provided the bulk of the alluvium in the debris cone; the Agha Jari Formation has contributed the rest. The Agha Jari Formation forms the major bedrock on which the alluvium has been deposited. The westward flowing Bisheh Zard and Tchah Qooch Rivers have deposited the debris cone in such a way that it slopes from east to southwest/northwest. The Agha Jari Formation forms the major bedrock on which the alluvium has been deposited (Kowsar, 1991). The bedrock of the margins of the Gareh Bygone Plain consists of Gachsaran Formations in west to east of Rud Shur River and Mishan Formation in the east part of the Plain. The thickness of the aquifer is decreased from west to east in such a way that the Mishan outcrops can be found on the eastern part of the Plain (Kowsar and Pakparvar, 2003). A schematic cross section (Fig. 3-3) is provided by using the Geologic map of 1:100,000 of the region (Appendix 2), data collected from the three experimental wells excavated in this study (Fig. 5-6) and the log of the
Fig. 3-3. Schematic cross section of the Gareh Bygone Plain. The Agha Jari Formation consists of rhythmically interbedded brown to grey, calcareous, feature-forming sandstones and low weathering, gypsum-veined, red marls and grey to green siltstones. The Bakhtyari Formation mainly consists of pebbles and cobbles (conglomerate) of limestones and dark brown, ferruginous cherts. The diagram is drawn based on the Geological map of 1:100,000-scale published by National Oil Company of Iran and the logs of observation wells. Thickness of the layers is approximate.
four observation wells installed on the study site (Appendix 3). The logs of the two other observation wells were not drawn satisfactorily by the installers so were not used. The study site includes two aquifers namely Bisheh Zard aquifer in the West (7600 ha) and Tchah Qootch aquifer in the East (2000 ha), both named after their corresponding rivers (Fig. 3-5a).

The study area is a dry region of a Mediterranean type of climate of wet winters and dry summers, with mean annual precipitation of 211 mm, which shows high inter-annual variability. Rainfall mainly occurs from December to March, with few exceptional events in summer (June-July). The absolute recorded maximum temperature (40-46 ºC) occurs in July-August and the corresponding minimum (-6 to -1 ºC) in January–February (Table 2-1). Average annual class-A pan evaporation was recorded as 2555 mm.

3.2.2 Floodwater spreading systems at the Kowsar Station

The Kowsar Station was established in 1983 with the aim of desertification control through FWS (Kowsar, 1991). The FWS systems were planned following the methods pioneered by Philip (1957), improved by Newman (1963) and Quilty (1972), and modified by Kowsar (1991). Eight FWS systems with an extent of 1236 ha were constructed during the 1983-1985 period. Expansion of the FWS systems to 2033 ha was performed from 1996 to 2003 (Table 3-2). The names and locations of the FWS systems on the Gareh Bygone Plain are presented in Fig. 3-2d. However, the entire FWS systems are not covered by floodwater in all of the events, at least one event occurs in non-drought periods which results in full coverage of the systems. Our experience shows that the flooding events of flow rate higher than 100 m³ s⁻¹ in Bisheh Zard River cause a complete coverage of the systems.

A diagram is presented in Fig. 3-4 to illustrate a typical FWS system and its essential components. A FWS system starts with a diversion dam which is constructed on the river bed widthwise with a spillway to deliver a pre-defined discharge of the flow to the diversion gap. Flow in turn, reaches the beginning of the FWS system via a conveyor canal. As this canal acts simultaneously as a sedimentation basin, does not include the upslope bank and the flow inundates the upstream of the bank which causes a decrease in flow velocity, it is called as inundation canal. The flow then enters to the most important element of the system that is a dual purpose conveyor-spreader channel which
transfers water to the head of the spreading area and, ideally, distributes it evenly onto the land.

This channel is actually a long, shallow stilling basin which converts some of the kinetic energy of the flowing water into potential energy, thus elevating the level of the water surface a few cm above that of the lower sill of the channel, with the resultant flow of a shallow sheet of water on a very long front (Fig. 3-4). The excavated soil forms a bank on the upslope side of the channel. Openings or weirs are provided in the bank to facilitate entrance of water into the channel during major floods when flows, larger than the capacity of the conveyor-spreader channel, bypass the junction of the diversion canal and the channel and flow along the topside of the bank. Water is supplied to the conveyor-spreader channel by a diversion, or an inundation canal. It is imperative to realize that the conveyor-spreader channel is the only structure in the system which is connected to the river.

Table 3-2. Size and the implementation year of FWS systems.

<table>
<thead>
<tr>
<th>FWS system</th>
<th>Implementation year</th>
<th>Size, ha</th>
<th>No. of weirs</th>
<th>Ungauged</th>
<th>Gauged</th>
</tr>
</thead>
<tbody>
<tr>
<td>BZ₁</td>
<td>1983</td>
<td>196.1</td>
<td>28</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>BZ₂</td>
<td>1983</td>
<td>199.4</td>
<td>14</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>BZ₃</td>
<td>1984</td>
<td>29.5</td>
<td>2</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>BZ₄</td>
<td>1984</td>
<td>26.3</td>
<td>2</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>RA₁</td>
<td>1984</td>
<td>200.0</td>
<td>18</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>RA₂</td>
<td>1987</td>
<td>98.6</td>
<td>12</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>RA₃</td>
<td>1987</td>
<td>139.9</td>
<td>17</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>TQ₁</td>
<td>1987</td>
<td>346.1</td>
<td>28</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>350ha</td>
<td>1996</td>
<td>272.5</td>
<td>23</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>60ha</td>
<td>1996</td>
<td>55.9</td>
<td>6</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>Ahrrar</td>
<td>1996</td>
<td>82.3</td>
<td>14</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>Old airport</td>
<td>1997</td>
<td>156.1</td>
<td>5</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>RA₁ continued</td>
<td>1998</td>
<td>128.3</td>
<td>12</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>TQ₂</td>
<td>2003</td>
<td>101.8</td>
<td>8</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td>2032.8</td>
<td>189</td>
<td>26</td>
<td></td>
</tr>
</tbody>
</table>

BZ and RA are named after Bisheh Zard and Rahim Abad villages, respectively. TQ is named after Tchah Qooch River. The location of the floodwater spreading (FWS) systems is shown in Fig. 3-2d.
The spillage from the conveyor-spreader channel flows over the land in a sheet whose depth and velocity depend on the flow rate, slope, infiltration capacity, sediment concentration, soil and water temperature, ground cover, etc.

![Diagram of a typical floodwater spreading system](image)

**Fig. 3-4. Generalized diagram of a typical floodwater spreading system. The scales are approximate.**

The depth of water on the sill of the conveyor-spreader channel, which also depends on the foregoing factors, is usually 3-5 cm, rarely exceeding 10 cm. The terminal velocity of a 10 cm sheet of water on a 2% slope on denuded land seldom exceeds 60 cm s\(^{-1}\). At this depth and velocity the flow is non-erosive for all practical purposes. The flow of water on the land is regulated by the level-silled channels which are closed at both ends. These channels, which are located at 140-250 m spacing downstream of the conveyor-spreader channel,
function as described for the conveyor-spreader channel with two exceptions. First, they receive water only through the gaps provided in their banks at 100-400m intervals; thus in low flows or short duration floods they may receive no water to spread. Second, they are level along their entire lengths so they spread the water more evenly. When floodwater reaches the end of a floodwater spreading system it has lost most of its sediment load and is suitable for artificial recharge, or filling up surface reservoirs. Construction of recharge ponds, returning the rather clear water to alluvial streambeds or, in rare cases, injecting the water into aquifers through recharge wells are three methods of replenishment of groundwater of which the first two are employed in the Gareh Bygone Plain. As the upslope of the banks of both conveyor-spreader and level-silled channels are wet for a long time after flooding events, they are appropriate places for planting vegetation (trees or shrubs) in order to increase the stability of the embankments and to ameliorate the environment. More details can be found in Kowsar (1991, 1995).

3.2.3 Water table level and weather data

Installation of observation wells (OW) in the study area started in 1992 (Fig. 3-5b and Photo 3-1). Well Nos. 1 and 4 (OW1 and OW4) are located outside of the FWS systems. OW2, OW5, and OW6 are situated inside, and OW3 immediately downstream of the FWS systems. OW5 and OW6 were installed in 2005. None of the OWs are located within 200 m distance of operational wells so, there was no effect of lateral in and out flow to the OW as influenced by active pumping zones of adjacent wells. The criteria of 200 m distance is defined by the responsible authority (Water Resources Research Organization) in Iran as a guideline for observation wells installation.

Observation wells data include the height of the measuring point (MP) from the mean sea level, and depth of the WT relative to the MP during 1993-2012. The depth to WT was measured on a monthly basis from 1993 to 2012 by the Fars Regional Water Organization (FRWO) using an ordinary water level meter (Photo 3-1) with light indicator and a resolution of 0.01 m. Missing values of water level in the OWs were estimated employing regression equations between the water level of the adjacent OWs.

Weather data were available at the Gareh Bygone Weather Station from 1996 to 2012. Rainfall data for the 1992-1996 period were collected from the nearest climatologic station (Baba Arab), 15.75 km from the Kowsar Station.
3.2.4 Runon and runoff data

The volume of floodwater diverted to the FWS systems from January 1983 to November 2002 has been estimated and reported by the responsible authority.
of the Kowsar Station (internal technical reports). Due to a complete lack of data and instrumentation before 2002, the peak flow rates have been calculated by using empirical slope-area method as discussed in the following section.

From November 2002, there have been two types of flow measurements in the FWS systems in the study area. First, peak flow was recorded at the hydrometric station and second, diverted flow inside the FWS systems was measured by broad-crested weirs. In order to explain how the flow measurement was performed in our study site an example of the BZ1 and BZ3 (Bisheh Zard 1 and 3) FWS systems is given in the following. As depicted in Fig. 3-6, a standard hydrometric station is located upstream of the diversion dam in the BZ Ephemeral River. The dam diverts nearly one sixth of the flow to the diversion canal. The diverted flow in turn, is divided into two parts by two drops (gauged gates) (Photo 3-2c and Photo 3-3a). The wider drop directs the flow to the first (conveyor-spreader) channel of the BZ3 through a conveyor canal. The narrower drop conveys the flow to the BZ1 FWS systems. Surplus flow from BZ3 joins BZ1 through three weirs installed on its upslope bank. The flow rate and duration of floodwater diverted into BZ1 and BZ3 is measured. There are two other weirs, one installed at the left corner of the first basin and one at the lower end of the sixth basin of BZ1 which deliver the surplus flow to Ahrar and BZ2 FWS systems, respectively (Fig. 3-6). The excess of flow from the BZ2 which is returned back to the main river through tail drain is eventually measured in a weir installed at its junction with the river. All of the weirs are constructed based on broad crested weir design (Photo 3-3). Thus, the flow which exits through the outlets is also measured and consequently the volume of floodwater which is harvested by the three systems (BZ1 to BZ3) is determined for all events. The flow measurement which is explained for BZ1 to BZ3 as a part of FWS systems is a good representative of all systems. The flow rate of water entering and leaving the FWS systems in the study area is similarly measured. Therefore the volume of water that retained in the individual FWS systems and in the entire project is calculated in every event. To avoid excessive details, the other parts of the FWS systems are not explained here.

The flow rates are being measured by educated and practically trained technicians and published as technical reports of the Kowsar Station. The reported data were used in this study after a thorough investigation on its reliability. The simultaneously recorded rainfall data were used to check the
consistency of, and agreement between, measured flows in comparison with the amount of recorded floodwater. In addition a comparison was made between archived hand written notes and the published technical reports to remove human error in data transfer.

![Diagram of BZ1, BZ2, and BZ3 with gauged gates for flow measurements. Dimensions are not to scale. Numbers show placement of Photo 3-2a (1), of Photo 3-2b (2), of Photo 3-2c and Photo 3-3a (3), of Photo 3-3b (4), and of Photo 3-3c (5).]

**Fig. 3-6.** Schematic map of BZ1, BZ2, and BZ3 and location of gauged gates for flow measurements. Dimensions are not to scale. Numbers show placement of Photo 3-2a (1), of Photo 3-2b (2), of Photo 3-2c and Photo 3-3a (3), of Photo 3-3b (4), and of Photo 3-3c (5).

**Flow calculations**

From 1983 to 2002 the maximum upstream flow of the ephemeral rivers was calculated by the slope-area method. The continuity equation, the most basic formula of water flow (Israelsen and Hansen, 1965), can be applied in estimating the maximum discharge of an ephemeral river:

\[ Q = AV \]  
(3-3)

where \( Q \) is maximum discharge \((m^3 s^{-1})\), \( A \) is flow cross-sectional area at the maximum discharge \((m^2)\) and \( V \) is flow velocity at the highest peak \((m s^{-1})\). Estimation of \( V \) is done by applying Manning's formula when the channel slope is less than 10% (Linsley et al., 1975) and for conditions of uniform flow in which the water surface profile and energy gradient are parallel to the streambed and the area, hydraulic radius, and depth remain constant.
throughout the reach (Dalrymple and Benson, 1968). As these two main criteria were true for the flooding events in the two ephemeral rivers (Bisheh Zard and Tchah Qooch), flow rate was determined using Manning’s equation (Powell, 1960):

\[ V = \left( \frac{3}{n} \left( \frac{R_h^2 S_h^{\frac{1}{3}}} {\frac{1}{3} g^{\frac{1}{2}}} \right) \right) \]

where \( V \) is velocity (m s\(^{-1}\)), \( R_h \) is the hydraulic radius (m), \( S_h \) the slope of the hydraulic grade line or the linear hydraulic head loss (m m\(^{-1}\)), which is the same as the channel bed slope when the water depth is constant and \( n \) is Gauckler–Manning coefficient (s m\(^{-1/3}\)). Flow rate was estimated based on the slope measured along the energy line of the river, calculated water surface profile and cross-sectional area based on the flood marks and an estimate of Gauckler–Manning coefficient.

From 2002 the upstream flow level of the rivers is being continuously determined using data recorded by a lymnograph at a hydrometric station and a rating curve available for the river section. The observed mean upstream flow with river cross section profiles was used to determine the cross-sectional area, the wetted perimeter of the flow, as well as the top and bottom widths of the flow area located 5 m upstream and downstream of the lymnograph.

The flow rate for broad-crested weirs inside the FWS systems was calculated as follows. A broad-crested weir (Fig. 3-7) is a flat-crested structure with a crest length that is large compared to the flow thickness for the streamlines to be parallel to the crest invert and the pressure distribution to be hydrostatic (Montes, 1998; Chanson, 2004). For an ideal fluid flow above a rectangular weir, the Belanger principle yields the relationship between the flow rate \( Q \) and upstream head above crest \( h_w \) (Sargison and Percy, 2009) as:

\[ Q = C_w \left( \frac{2}{3} \left( \frac{2}{3} g \right) \right) L h_w^{3/2} \]

where \( Q \) is flow (m\(^3\) s\(^{-1}\), \( C_w \) is discharge coefficient (m\(^{1/2}\) s\(^{-1}\), \( g \) is gravitational acceleration (m s\(^{-2}\)), \( L \) is weir length (m), \( h_w \) is hydraulic head (m).

The \( C_w \) can be defined based on the \( h_w/L \) ratio as proposed by Beirami et al. (2008).
\[ C_w = 0.604 \left( \frac{h_w}{L} \right)^{0.0619} \]  

(3-6)

The equation is a variant of the equation proposed by Bos (1984) which is valid in the range of \(0.1 < \frac{h_w}{L} < 1.0\).

As an example the flow calculation of a typical event over a broad crested weir with the length of 0.6 m and its resulted volume of water is presented as Table 3-3. A volume of 191 m³ is passed through this weir during a 505 minutes of time.

Table 3-3. Flow calculation of a typical broad crested weir in the Gareh Bygone Plain.

<table>
<thead>
<tr>
<th>(h_w), m</th>
<th>(L), m</th>
<th>(C_w)</th>
<th>(Q), m³ s⁻¹</th>
<th>(T), min</th>
<th>(V), m³</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.02</td>
<td>0.6</td>
<td>0.49</td>
<td>0.001</td>
<td>30</td>
<td>2.5</td>
</tr>
<tr>
<td>0.04</td>
<td>0.6</td>
<td>0.51</td>
<td>0.004</td>
<td>45</td>
<td>11.3</td>
</tr>
<tr>
<td>0.06</td>
<td>0.6</td>
<td>0.52</td>
<td>0.008</td>
<td>360</td>
<td>170.0</td>
</tr>
<tr>
<td>0.03</td>
<td>0.6</td>
<td>0.50</td>
<td>0.003</td>
<td>40</td>
<td>6.4</td>
</tr>
<tr>
<td>0.01</td>
<td>0.6</td>
<td>0.47</td>
<td>0.000</td>
<td>30</td>
<td>0.9</td>
</tr>
</tbody>
</table>

The \(h_w\) is the flow head above crest, \(L\) is the length of weir in parallel direction to the flow, \(C_w\) is discharge coefficient, \(Q\) is flow rate, \(T\) is time and \(V\) is volume.

Fig. 3-7. Schematic of broad crested weir (Montes, 1998).

3.2.5 Determining the effect of FWS on recharge

To determine how the FWS systems function as a potential source of recharge of GW, water level responses of wells within and outside of the FWS systems to the flooding events were compared. Comparison was made between the GW
level before the occurrence of each particular flood and that measured successively for four months after the flooding event. The period of four months was selected to cover the maximum change in GW level as influenced by the flooding event. As the entire dataset shows, in most cases the increase in GW level is continued until the third month and sometimes the fourth month. To minimize the effect of abstractions on water level changes, respective data related to the months of minimum water withdrawal (October through February) were used. The above analysis was carried out for all flooding events during the study period 1993 to 2012.

The response of the wells in terms of water level was also compared with that of two consecutive major floods which occurred in 2003-2004 and 2004-2005. In the first event in December 2003 and January 2004, the FWS systems did not function due to a major repair and maintenance which deprived the aquifer from the generous floods of that event. In the subsequent event in December 2004 and January 2005, the FWS systems were operating properly. This provided an opportunity to compare the effect of the FWS systems on aquifer recharge.

Infiltrating rainfall or flooding is not the only cause of changes in water table level. Water table fluctuations are also due to other reasons: ocean tides; earth tides (caused by the forces exerted on the Earth's surface by the Moon and the Sun); barometric pressure changes; increases in gas pressure in the unsaturated zone; pumping; and lateral flow. These effects need to be avoided, minimized or removed from the water level signals (Crosbie et al., 2005).

Unconfined aquifers are commonly insensitive to changes in barometric pressure. The effect of ocean tides is not applicable here as the study area is an inland location. Lateral flow is insignificant as the wells are located in a relatively flat plain with low hydraulic gradient. Pumping by production wells is a major source of water level drop after recharge events which must be taken in to account in all the analyses in the study area. A hydrograph of GW in the study area was then plotted based on the mean GW level for each month and employing the grid layers generated by the Surfer software (Surfer 13, Golden software, LLC). Rainfall depth and simultaneous mean GW level of the study area were also plotted concurrently to inspect the effect of drought on hydrographs.
3.2.6  Quantification of recharge

The method used by Hoque et al. (2007) was applied for estimating recharge in a particular duration of time (Eq. (3-2)). This analysis was performed for the hydrological year 2010-2011 as the complete data necessary for the analysis was only available for that hydrological year. Eq. (3-2) was solved to find the recharge \( R \) as a remainder when the other components such as extraction from the aquifer \( E \), change in GW storage \( \Delta S \) and return flow \( RF \) are known.

\[
R = E + \Delta S - RF + ET^{gw} \tag{3-7}
\]

\( ET \) from the GW \( ET^{gw} \) was considered as the water consumption by the tree plantations which have access to GW by their deep roots. The subtraction of in and out flow to the groundwater \( (Q_{in}^{gw} - Q_{off}^{gw}) \) in Eq. (3-2) was assumed as zero in a hydrological year time period. Thereafter, the artificial recharge \( AR \) was calculated (as explained in the next section 3.2.7). As a result, the proportion of \( AR \) to total recharge \( R \) was determined. The remainder of subtraction of \( R \) and \( AR \) was ascribed to natural recharge \( NR \). The determination of different sources of \( NR \) (river bed infiltration, rainfall infiltration in the uplands, diffusion from the adjacent aquifers, etc.) and its quantification is out of the objectives of this study so, is left to be studied in future researches.

3.2.6.1  Amount of extraction from the aquifer

Extraction from the aquifer was estimated by two methods, total water withdrawal and the total applied irrigation water. First, the annual volume of water withdrawal from the BZ aquifer that was studied by Hosseinimarandi et al. (2011) through a field survey was employed. These include the location of irrigation wells, their discharge rate (Photo 3-4), the daily hours of operation and the number of operation days per year. The field data which was recorded by Hosseinimarandi et al. (2011) was used to check and recalculate (the flow discharge) based on the methods mentioned in Appendix 4. In addition, the result of a survey conducted by FRWO (see section 3.2.2) containing the number of all pumping wells installed in the study area, the installation year, the depth of well and some pumping rates were used to cross check the data. These two sets of data were compared and checked in order to minimize the errors in the wells’ location and discharge rate. Those wells located inside and nearby the BZ aquifer were extracted from the whole dataset. The total
extraction by water withdrawal was calculated for every well’s discharge and working hours and summed up for the all wells.

Second, the volume of applied irrigation water was measured in 12 farm fields of wheat and 10 farm field of forage corn cultivations in the area during the growing season of the hydrologic year 2010-2011. To do so, well discharge was measured in the head ditch above each field using a cut-throat flume (Walker, 1989) (Photo 3-5). This was then converted to seasonal volume of applied water per hectare using the time of application for each turn and the number of irrigations during the season. The total volume of applied water for the cultivated farm fields was then calculated using the size of area provided by remote sensing (see chapter 2, section 2.2.8.4).

3.2.6.2 Change in groundwater storage (saturated pore volume)

The depleted volume was calculated employing the volume calculation facility of Surfer software (surfer 13, Golden software, LLC). It finds the volume between the two surfaces given as grid files. As described by Hoque et al. (2007), the volume is defined by a double integral:

$$V=\int_{x_{min}}^{x_{max}} \int_{y_{min}}^{y_{max}} f(x,y)dx\,dy$$  \hspace{1cm} (3-8)

Surfer computes this by first integrating over $x$ (the columns) to get the areas under the individual rows, and then integrating over $y$ (the rows) to get the final volume.

The depth to the impermeable layer in lithological logs of the six OWs and those reported by Hosseinimarandi et al. (2011) for irrigation wells was used as the lower boundary of the aquifer. The GW levels in 1993, 1997, 2003, 2005, and 2012 were considered as the upper boundary. Dewatering volume was calculated by subtracting aquifer volume for each year from its previous one. Selection of the above mentioned years was based on significant changes in the hydrograph inclination in those years.

Saturated pore volume was calculated as the product of specific yield ($S_r$) and dewatered volume of the aquifer. Values of specific yield reported for the BZ aquifer are 0.05 (averaged) by Hashemi et al. (2013), and 0.008 by Hosseinimarandi et al. (2011) employing the pumping test. Values as low as 0.008 obtained in some of the measurements performed by the two mentioned authors in our study area. The under-estimation of $S_r$ with pump test has also
been experienced by some authors (Nwankwor et al., 1984; Moench, 1994). Crosbie et al. (2005) showed 2-3 times under-estimation of $S_y$ by pump test as evaluated by a mass balance study by chlorine. Therefore, the method used in this study is based on the physical concept of water retention in porous media using a soil moisture retention curve (Crosbie et al., 2005):

$$S_y = \theta_s - \theta_r$$  \hspace{1cm} (3-9)

where $\theta_s$ is the saturated soil water content and $\theta_r$, the residual water content, is defined as the water content for which the gradient ($d\theta/dh$) becomes zero (excluding the region near zero matric potential that also has a zero gradient). However, as the specific yield is water content of the porous media (here the aquifer) which can be easily drained, moved and pumped out, the lower boundary of water content was set to field capacity (FC) which was considered as soil-water potential head as -330 cm (Brady and Weil, 1996; Kolay, 2008). The term $\theta_r$ in Eq. (3-9) was therefore changed to $\theta_{FC}$. The amount of soil-water content which is assigned to the FC depends on the hydraulic properties specially the pore size distribution. As the soil potential in which the water is retained by the soil particles can better explain this term, the FC has been defined as soil-water corresponding to a particular soil potential and normally considered as -1/3 atmosphere or -330 cm.

Three experimental wells were dug (28 to 32 meters in depth) in this study from May to September 2008 and explored to describe the depth and distribution of the layers. Seven representative layer (RL)s were recognized which were repeated throughout the experimental well’s profile. Identifying the RLs and their explanations is presented in chapter 5 (sections 5.2.4.3 to 5.2.4.5) and the particle size distribution as the basis for RLs differentiation is given in Appendix 5. Soil samples from the seven RLs of these experimental wells were taken and physical properties including texture, fraction of stones and water content were measured. Complete water retention curves including water content at saturation and -330 cm were established for <2 mm fraction (see chapter 5) by sand box apparatus and pressure plate (Table 3-4). Due to the presence of stone fractions, water contents were adjusted employing the Bouwer and Rice (1984) equation:

$$\theta_b = (1 - V_f)\theta_f$$  \hspace{1cm} (3-10)
where $\theta_b$ is the water content of bulk material (undisturbed fine soil plus stones), $V_r$ is the volume fraction of stones, and $\theta_f$ is the water content of the fine soil. The values calculated as specific yield of RLs were then employed to assign a value as the aquifer’s $S_y$. To do this, a weighted average was calculated for the entire well’s profile considering the occurrence of each RL through the profile, multiplying its thickness by its pre-defined $S_y$ value, summing up the products and dividing by the depth of all the summed up layers. The two surface layers were excluded from the weighted average because of their absence in sub-surface profile of the aquifer. Details of the data and the calculations are presented in Appendix 6.

### 3.2.6.3 Agricultural return flow and ET from the groundwater

In chapter 2 (section 2.3.4.3) the RF was assumed as 24% of the applied water based on published results in a region similar to our study site in order to cross check the modeled $ET_a$ by the water balance $ET_a$. As the values obtained by the two methods were close enough, we decided here to assign the SEBS calculated $ET_a$ as a reliable value.

Therefore, RF was calculated as a remainder of subtracted actual evapotranspiration from the applied water. In chapter 2, the part of applied water consumed by crops and tree plantations was determined by evapotranspiration ($ET$) mapping using the SEBS model and 32 successive Landsat images. Parameterization of SEBS was primarily optimized by assessing against observed $ET$ data and then used for calculating the yearly $ET_a$. According to the availability of Landsat images at the time of study, $ET_a$ calculation was made for the period of October 2009 to December 2010. Since the cropped area in the two successive hydrological years 2009-2010 and 2010-2011 was practically the same, the amount of return flow was used for the hydrological year 2010-2011 for which the full input data of the water budget was available. In the same way the $ET^w$ component was calculated based on the summed up annual $ET_a$ data from the $ET$ maps and the size of area of tree plantations.
Table 3-4. Retention curve data and input parameters for $S_y$ calculation.

<table>
<thead>
<tr>
<th>Layers</th>
<th>Stoniness</th>
<th>Depth, m</th>
<th>Potentials, cm</th>
<th>$S_y$, $\theta_v$, m^3 m^{-3}</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>0</td>
<td>330</td>
</tr>
<tr>
<td>A</td>
<td>0.07</td>
<td>0-0.1</td>
<td>0.45</td>
<td>0.42</td>
</tr>
<tr>
<td>B</td>
<td>0.22</td>
<td>0.1-0.42</td>
<td>0.44</td>
<td>0.34</td>
</tr>
<tr>
<td>C</td>
<td>0.54</td>
<td>0.42-1.0</td>
<td>0.38</td>
<td>0.17</td>
</tr>
<tr>
<td>D</td>
<td>0.66</td>
<td>1.0-1.3</td>
<td>0.35</td>
<td>0.12</td>
</tr>
<tr>
<td>E</td>
<td>0.07</td>
<td>1.3-1.8</td>
<td>0.39</td>
<td>0.36</td>
</tr>
<tr>
<td>F</td>
<td>0.53</td>
<td>1.8-2.25</td>
<td>0.38</td>
<td>0.18</td>
</tr>
<tr>
<td>G</td>
<td>0.11</td>
<td>2.25-3.35</td>
<td>0.4</td>
<td>0.36</td>
</tr>
</tbody>
</table>

These representative layers are repeated along the aquifer profile in a 30 meter depth experimental well except A and B. The $\theta_v$ is volumetric soil-water content, m is measured $\theta_v$, c is corrected $\theta_v$ based on the stoniness (using Eq. (3-10)), $S_y$ is specific yield calculated as the difference between $\theta_v$ of the saturated soil and the field capacity (potentials of 0 and 330 cm, respectively).

3.2.7 Quantification of artificial recharge

Contribution of the FWS systems to total recharge was assessed by two methods in order to cross check the results as described below.

3.2.7.1 Application of flow data

The part of water in a recharge pound consumed by total losses (open water evaporation plus $ET$), when subtracted from the amount of pounded water, is net total infiltration (Hendrickx et al., 1991). Because of the deep GW table and due to the coarse textured stony soils in the vadose zone, the positive flux (toward the soil surface) resulting from capillary movement after the event was assumed negligible. Therefore, the net infiltration to the deep layers was counted as the net recharge and the net recharge was considered as artificial recharge based on the flooding data inside the FWS systems. Indeed, the lateral movement will cause a part of the infiltrated water to be moved to the natural drainage (the Shur River) but the proportion of vertical to the lateral movement is not large enough to substantiate the lateral movement. To justify this, it should be noted that the horizontal hydraulic conductivity of aquifers similar to the aquifer in our study (medium to coarse gravelly material) is reported as 0.02 to 1.02 cm day$^{-1}$ (Domenico and Schwartz, 1990) whereas the vertical hydraulic conductivity measured in this study varied between 17 to 2840 cm day$^{-1}$ (Appendix 5) which is by far higher than the expected horizontal
hydraulic conductivity. Therefore, the assumption of neglecting lateral movement at least when vertical infiltration due to a flooding event is in progress is reasonable.

The evaporation losses during a given period depend on the duration of infiltration and, therefore, on the infiltration rate. To evaluate the recharge, the equation derived by Hendrickx et al. (1991) was used:

\[ AR = D - \frac{eD}{i + e} - \sum ET \]  \hspace{1cm} (3-11)

where, \( i \) is the infiltration rate (mm day\(^{-1}\)), \( e \) is the open water evaporation during the period of infiltration (mm day\(^{-1}\)), \( \sum ET \) is the cumulative \( ET \) after all water has infiltrated (mm), \( AR \) is the net artificial recharge (mm), and \( D \) is the depth of water applied for recharge (mm).

Infiltration rates of the soil surface in the FWS systems of the study area measured at several points by Rahbar (2010), were used to prepare an infiltration map by Surfer software (Surfer 13, Golden software LLC). Normally the floodwater which is retained within the FWS systems sustains less than 48 hours. The FWS systems dry out in summer times after 10-15 days and in winter times after 30-40 days. Open water evaporation for a duration of 6 days of the events was taken from weather data. The cumulative \( ET \) was calculated based on \( ET \) maps generated in a chapter 2. As some of the infiltrated water is lost by \( ET \), the time after flooding events, when \( ET \) is taking place was considered as 30 days at the basis of the observations of soil-water content in different depths (see section 5.2.5.4). The long term data showed that the soil-water content of the top soils (to the depth of 60 cm) is returned back to the level it had before a rainfall or flooding event in a period of maximum one month.

### 3.2.7.2 Water budget method for the FWS systems influenced area

Recharge determined in the section (3.2.6) is based on water budget for the entire hydrological year (2010-2011) and the whole Bisheh Zard aquifer and therefore is considered as total recharge. In order to focus on artificial recharge, the same water budget (Eq. (3-7)) was employed but the time was restricted to the flooding period and the place was limited to the location of FWS systems. Therefore, the water budget can be written as:

\[ AR = E + \Delta S - RF + ET^{gw} \]  \hspace{1cm} (3-12)
With the terms being defined as in Eq. (3-11) but for a different time period and size of the area. The location of the area which is under the immediate influence of artificial recharge (AR) was separated by making polygon surrounding the FWS systems which includes the location of the six OWs. Then the last GW levels before the flooding event of 28-1-2011 to 2-2-2011 and the GW levels of the further months of maximum increase were collected. For instance, the GW level of OW1 at 15-1-2011 is measured as 1158.54 (before the event) and 1158.63, 1158.72, 1158.86 m.a.s.l for the months February, March and April, respectively. As the level then decreased to 1158.68 m.a.s.l in May, March was selected as the month of maximum change and the rise in GW level for the other OWs was similarly calculated. The two sets of GW level data for the OWs corresponding to before the flooding event and the third month after the event were used to make the two piezometric maps by Surfer software (Surfer 13, Golden software, LLC). The area outside the boundary of the FWS systems were excluded by the Blank module in Surfer. The maps of April and January were introduced to the Volume module as the top and the bottom layers, respectively in the same software to calculate the volume of the aquifer in which the recharged water resided as influenced by the FWS systems. The resulting volume of the aquifer was then multiplied by $S_y$ to obtain the change in GW storage ($\Delta S$) in the water budget equation.

The $E$ component is considered as equal to agricultural water extraction for the similar calculation period of $\Delta S$ and for the farmlands located inside the predefined boundary of FWS systems. To calculate $E$, the amount of applied water for the cultivated (not fallow) part of farmlands inside the FWS systems boundary was calculated based on the measured applied water per hectares (see chapter 2 section 2.2.8.4) multiplied by the size of the area. The $RF$ component was then calculated by subtracting $E$ from total $ET$ of the corresponding farmlands. Total $ET$, in turn, was calculated by summing up the $ET$ values extracted from $ET$ maps generated by remote sensing (see chapter 2 the same section). The $ET^{gw}$ component was calculated based on the summed up annual $ET_a$ data from the $ET$ maps and the size of area of tree plantations.
3.3 Results and discussion

3.3.1 Water level data quality

As inferred from the empirical q-q plots (Fig. 3-8), the tendency of scatter points to the trend line is different for the OWs, though, meaningful correlation in q-q plot is seen in all of the OWs, at least to one adjacent OW.

Fig. 3-8. Empirical q-q plots for the water level data of the observation wells (OW) 1 to 6.
When a q-q plot is reasonably linear, one may conclude that the two distributions involved have similar shapes. When there are marked departures from linearity, the character of those departures can reveal the ways in which the shapes differ (EPA, 2000). When the several data sets with similar temporal measurement intervals show linearity in cross validation by empirical q-q plots, it is an adequate evidence of reliability of the measurements (Popham and Sirotnik, 1992). Therefore, it can be concluded that the probability of the water level measurements are similar for the OWs of interest in this study.

3.3.2 Piezometric level and flow directions

The GW level of the Bisheh Zard aquifer and the GW flow direction is presented in Fig. 3-9a. The water level data of the six OWs together with the collected water level of the operational wells in the year 2012 corresponding to the Bisheh Zard aquifer were used to prepare the maps.

The piezometric contour lines start at its highest value of 1165 m.a.s.l at the Northeast, decline gradually to 1115 m.a.s.l under the zone of FWS systems activity and show a concave shape in this area which is a sign of influence of artificial recharge. The contours continue to descend to 1105 to 1100 m.a.s.l at the South to Southwest part in the vicinity of Shur River. The rate of decline is proportional to the surface topography (Fig. 3-9b) which means that the GW topography follows the soil surface topography. The GW flow in the aquifer (Fig. 3-9a) shows a direction from upland to lowland and eventually to the Southwest and West where the Shur River drains the aquifer. A different minor local directions is also seen in Southern part from the Tchah Qooch River to the North direction which corresponds to the recharge from the mentioned River.

3.3.3 Spatial and temporal changes in water level

Fig. 3-10a depicts the synchronized rainfall and GW hydrograph for the study area. The generalized hydrograph shows an increase in water level from 1993 to 1997, a steep decrease from 1997 to 2004, and a gradual decrease from 2005 to 2012. The difference between the initial and final GW level is about 6 m.

The annual trend of rainfall during the study period revealed an irregularity, though a general declining trend can be perceived (Fig. 3-10a).
Fig. 3-9. Piezometric map (A) and the digital elevation model (DEM) map (B) of the Bisheh Zard aquifer.
As inferred from Fig. 3-10b, the number of irrigation wells started to increase from the beginning of the FWS systems construction in 1983, but an expedited increase is shown from 1989 to 1994 (72 wells in 5 years). The increase in number of wells continued until 2005, after which it remained stable up to 2012. The recession trend in GW from 1997 to 2004 coincided with recurrent droughts and increase in the number of irrigation wells. The increase in the amount of annual rainfall from 2000 to 2005 slowed down the rate of decrease in water level. The recent drought from 2005 to 2012 has increased the rate of drop in water level again.

To quantify the influence of rainfall and number of dug wells on the GW level change, a multiple linear regression was made between the annual rainfall and number of dug wells in Bisheh Zard aquifer as independent and the mean GW level as dependent variables (Table 3-5). The model itself has an F value of <0.001 and an adjusted $r^2$ of 0.79 which is significant at <0.01 level. Incorporation of the independent variable “number of wells” that has a low p-value (here, <0.01) is likely to be a meaningful addition to the model because changes in the predictor's value are related to changes in the dependent variable. Conversely, a larger insignificant (here, 0.81) p-value of the “rainfall” variable suggests that changes in this variable are not associated with changes in the response.

Although some studies have shown high correlations between water table fluctuation and precipitation (Crosbie et al., 2005; Rimon et al., 2007; Yu and Chu, 2012), a statistically significant correlation could not be established between the two sets of data. Therefore, the decreasing trend of GW level cannot be statistically ascribed to the temporal trend of rainfall. However, timely influence of rainy years on the GW hydrograph can be implied. For instance, the increase in the GW level during 1993-1997 is concurrent with large storms in 1993 and 1996. A drought period from 1997 to 2000 coincided with severe recession in the GW level.

To demonstrate the role of the FWS systems in the above setting, the behavior of individual wells located inside and outside of the FWS systems was examined. Water level change shows no trend in OW1 (Fig. 3-11). This could be due to the location of the well, which is in close vicinity of a fault supplying water to the plain (Hashemi et al., 2013). OW2, OW3 and OW4 reveal the same temporal trend as the general hydrograph of the study area. OW2 being
located inside the FWS systems shows a greater resistance to the drought periods and extractions than OW3 and OW4. The drop in the water level in OW2 during 1997-2004 was 2.07 m, while OW3 and OW4 showed a drop of 9.0 and 11.26 m in the same period, respectively. This clearly demonstrates the impact of the FWS systems on the recharge of GW.

Table 3-5. Influence of the rainfall and number of dug wells on the GW level as indicated by multiple linear regression.

<table>
<thead>
<tr>
<th>Summary of the model</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>R²</td>
<td>0.81</td>
</tr>
<tr>
<td>Adj. R²</td>
<td>0.79</td>
</tr>
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<td>Observations</td>
<td>20</td>
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</table>

<table>
<thead>
<tr>
<th>ANOVA</th>
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<tr>
<td>DF</td>
<td>SS</td>
<td>MS</td>
<td>F</td>
</tr>
<tr>
<td>Regression</td>
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<td>101.94</td>
<td>50.97</td>
</tr>
<tr>
<td>Residual</td>
<td>17</td>
<td>23.89</td>
<td>1.41</td>
</tr>
<tr>
<td>Total</td>
<td>19</td>
<td>125.83</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Variables</th>
<th>Coeff.</th>
<th>SE</th>
<th>t Stat</th>
<th>P-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Intercept</td>
<td>1161.38</td>
<td>2.87</td>
<td>404.23</td>
<td>2.67E-35</td>
</tr>
<tr>
<td>Wells BZ</td>
<td>-0.1419</td>
<td>0.0186</td>
<td>-7.62</td>
<td>7.01E-07</td>
</tr>
<tr>
<td>Rain</td>
<td>-0.0007</td>
<td>0.0029</td>
<td>-0.24</td>
<td>0.81</td>
</tr>
</tbody>
</table>

Multiple linear regression between the GW level as dependent (Y) and the number of dug wells in Biheh Zard aquifer (Wells BZ) and annual rainfall (Rain) as independent (X1 and X2) variables. Adj. is adjusted, ANOVA is analysis of variances, DF is degree of freedom, SS is sum of squares, MS is mean of squares and Sig. is significance.

As shown in Table 3-6, an anomaly is observed in the amount of rainfall and the volume of flood in which the extent of rainfall and its subsequent flooding are not always proportional (some higher rainfall resulted in lower volume of flooding). The reason for this diversity is the disparity in the location of rainfall. When the rainfall covers both the highland basins as well as the Gareh Bygone Plain, maximum runoff occurred and the volume of resulting flood flow was relatively high. In contrast, when the rainfall was limited to within the Plain, the resulting flow in the Ephemeral Rivers is not enough to produce an extreme flooding. The season of the flood occurrence is the other source of variation. In summer time, when the soil surface is barren due to drought, runoff is much higher than in the winter. In spite of the low rainfall during this period, numerous flooding events occurred and the FWS systems were operating (Table 3-6).
During the 2005-2012 period the hydrograph for the three wells show a similar mild drop. During this period the number of floods had decreased due to recent drought and therefore, the role of the FWS systems diminished. Hence, it is expected that the observation wells show similar behavior. The drop in OW2, OW3, and OW4 were 2.9, 3.4, and 0.7 m, respectively. The lower drop in OW4 could be due to recharge from the Tchah Qootch River located south of the OW4 (Fig. 3-12) as a supplemental source of recharge. The GW flow direction from the river towards the OW4 is obvious in the piezometric map (Fig. 3-9a). This fact is evident by comparing the elevation of Tchah Qooch River at the closest location to the GW level of the OW4 (1145.5 m.a.s.l) which is higher than the GW elevation of this OW (1125.5 m.a.s.l) (Fig. 3-12). Eight years of data collection for the other two OWs (2005-2012) show a recession of 1.4 and 3.2 m for OW5 and OW6, respectively. OW5 and OW6 show the same trend as OW2 and OW3.

![Fig. 3-10. A: Temporal trend of rainfall and groundwater hydrograph. B: Temporal trend in the number of wells in the region. GBP is Gareh Bygone Plain and BZ is Bisheh Zard aquifer. WL is groundwater level height.](image-url)
Table 3-6. Groundwater level response to flood events for observation well (OW)2.

<table>
<thead>
<tr>
<th>Date of flooding</th>
<th>Flood, Mm³&lt;sup&gt;(1)&lt;/sup&gt;</th>
<th>Rain, mm</th>
<th>Depth to groundwater level and date of measurement&lt;sup&gt;(2)&lt;/sup&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Before flooding</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Date</td>
</tr>
<tr>
<td>8.02.93</td>
<td>NR</td>
<td>15.01.93</td>
<td>19.1</td>
</tr>
<tr>
<td>27.03.94</td>
<td>NR</td>
<td>9.03.94</td>
<td>17.7</td>
</tr>
<tr>
<td>3.12.94</td>
<td>NR</td>
<td>17.11.94</td>
<td>17.2</td>
</tr>
<tr>
<td>8.03.95</td>
<td>1.5</td>
<td>14.02.95</td>
<td>16.7</td>
</tr>
<tr>
<td>24.07.95</td>
<td>NR</td>
<td>15.07.95</td>
<td>16.5</td>
</tr>
<tr>
<td>6.12.95</td>
<td>NR</td>
<td>18.11.95</td>
<td>16.3</td>
</tr>
<tr>
<td>25.03.97</td>
<td>NR</td>
<td>5.03.97</td>
<td>14.6</td>
</tr>
<tr>
<td>2.01.98</td>
<td>NR</td>
<td>17.12.97</td>
<td>14.9</td>
</tr>
<tr>
<td>2.03.99</td>
<td>NR</td>
<td>15.02.99</td>
<td>15.4</td>
</tr>
<tr>
<td>8.07.99</td>
<td>NR</td>
<td>16.06.99</td>
<td>15.2</td>
</tr>
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<td>16.12.01</td>
<td>NR</td>
<td>15.12.01</td>
<td>17.1</td>
</tr>
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<td>12.01.02</td>
<td>NR</td>
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<td>17.1</td>
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<td>22.12.02</td>
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<td>17.7</td>
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<td>17.4</td>
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<td>16.08.03</td>
<td>0.4</td>
<td>15.08.03</td>
<td>17.3</td>
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</table>

1-The volume of retained flow in the system (diverted inflow minus outflow). NR=not recorded. Mm³ is million m³.
2-1<sup>st</sup>, 2<sup>nd</sup>, 3<sup>rd</sup> and 4<sup>th</sup> are the first, second, third and fourth times of measurements of groundwater level after the occurrence of a flooding event.
Table 3-6. Continued.

<table>
<thead>
<tr>
<th>Date of flooding</th>
<th>Flood, Mm³</th>
<th>Rain, mm</th>
<th>Depth to groundwater level and date of measurement(2)</th>
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</thead>
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<tr>
<td></td>
<td></td>
<td></td>
<td>Before flooding</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Date</td>
</tr>
<tr>
<td>6-7.12.03</td>
<td>3.14</td>
<td>66</td>
<td>13.11</td>
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<td>27.01.04</td>
<td>13.58</td>
<td>225.5</td>
<td>15.01</td>
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<td>211</td>
<td>11.01</td>
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<td>30.03.06</td>
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<td>2.03</td>
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<td>5.09.06</td>
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<td>61.5</td>
<td>14.08</td>
</tr>
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<td>4.02.07</td>
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<td>28</td>
<td>11.01</td>
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<td>24.07.07</td>
<td>0.18</td>
<td>11.5</td>
<td>12.07</td>
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<td>9.09.08</td>
<td>1.26</td>
<td>91</td>
<td>11.08</td>
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<td>30.03.09</td>
<td>0.24</td>
<td>23.5</td>
<td>12.03</td>
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<td>8.12.09</td>
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<td>18</td>
<td>11.11</td>
</tr>
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<td>5-27.02.10</td>
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<td>54.5</td>
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<td>0.84</td>
<td>20.5</td>
<td>11.11</td>
</tr>
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<td>26.02.12</td>
<td>1.65</td>
<td>23.2</td>
<td>9.02</td>
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</table>

1-The volume of retained flow in the system (diverted inflow minus outflow). NR=not recorded. Mm³ is million m³.
2-1st, 2nd, 3rd and 4th are the first, second, third and fourth times of measurements of groundwater level after the occurrence of a flooding event.
Fig. 3-11. Changes in WL (groundwater level) height in the observation well (OW)s. OW5 and OW6 were installed in 2005.

The line which named as “All” represents the generalized hygrograph of the Gareh Bygone Plain.
Fig. 3-12. Location of the observation well 4 (OW4) in accordance with the Tchah Qooch River and the proportional ground and groundwater elevation. Picture is taken from the Google Earth of the date 6-9-2011. Corresponding groundwater level of the OW4 is taken from the year 2011.

3.3.4 Substantiating the effect of the FWS systems

Among the relatively large number of flood occurrence data and the corresponding values of the GW levels, only the representative data of OW2 are shown in Table 3-6. Rainfall data were accessible from 1993 and the floodwater diversion data have been collected since December 2002.

To demonstrate the effect of the FWS systems on the aquifer recharge, the behavior of OWs 1, 2, 3, and 4 is examined more closely in response to the flooding events (Fig. 3-13). In the year 2003 the FWS systems were not functioning due a necessary repair and again started to act in 2004. The hydrograph of OW2 shows that the response to the flood in 2004-2005 resulted in a considerable rise (2.05 m) after two months. The change in water table level due to the previous flood in 2003-2004 shows a common trend, which was due to natural recharge (river bed and other sources), whereas in 2004-2005, the sudden rise in water level was the direct response of the FWS systems’ operation. The other wells located in the vicinity of the FWS systems (OW3 and OW4) exhibit different behavior than OW2. In 2003-2004, when the FWS systems were not functioning, OW3 showed a 0.3 m rise, nearly
similar to that of the OW2; however, in the following flood in 2004-2005 when the FWS systems were functioning, the rise in water level of OW3 (0.5 m) after three months was much less than that of OW2 (2.05 m). The sustained flooding on the FWS systems are the same in both of the events because they occurred in the same season of the two years (December and January). This difference is expected to occur for two reasons. Firstly, due to the location of OW3, which is out of the FWS systems, a delay in reaching to the peak in water level as compared to OW2 is anticipated; secondly, because of being in close vicinity of the farming wells, the water level did not rise in the following months.

As described by Bouwer (2002), if no clogging layer exists on the bottom of an infiltration basin and the basin is “clean”, the water table would rise to the water level in the basin, and the water in the basin and in the aquifer would then be in direct hydraulic connection. In this case, a mound underneath the infiltration basin is expected as is evidenced by OW2 data of 2004-2005. Therefore, it proves the hydraulic connection between the FWS systems and the aquifer, and infers insignificant clogging effect on its surface after a long period (32 years) of functioning.

The rise in the water level in OW4 (1.20 m) was much more noticeable than at OW2 and OW3 in the first flood event. This supports the claim mentioned above that there was some recharge from the Tchah Qootch River during the flooding events. A comparable rise to that of OW2 is seen in the subsequent flood (1.90 m).

In addition to the above analysis on two successive floods, long term comparison is made between wells in response to all of the flooding events. The mean difference between water level in the wells before the flooding events and the corresponding values for four months after the events is depicted in Fig. 3-15. As expected, OW2 shows a positive change up to four months after the storms. This is true for OW1 as well, but to a lesser extent. OW3 shows the least amount of rise in water level compared to the other wells. OW4, located outside of the FWS systems, shows the highest rise in water level up to the third month after the events. It is seen that the long term behavior of wells supports the determinant role of the FWS systems on aquifer recharge.
Fig. 3-13. Response of water table to the flooding events. The FWS systems were not functioning in the first event of December 2003 to January 2004 and functioning in the second event of December 2004 to January 2005. WL is groundwater level and Mm³ is million m³.
3.3.5 Dewatering of the aquifer

The data from which \( S_y \) was calculated are presented in Table 3-4. The mean \( S_y \), calculated for the entire aquifer layer, was 0.18 (Appendix 6). The specific yield, as reported by several authors in similar lithological materials to our study area are: 0.18-0.36 (Crosbie et al., 2005), 0.063 (Moon et al., 2004), 0.2-0.25 (Gutentag et al., 1984), and 0.15 (Hoque et al., 2007). In addition, as indicated by Bouwer (2002), the fillable porosity to be used in the equations for mound rise is usually larger than the specific yield of the aquifer, because vadose zones often are relatively dry, especially in dry climates, and if they consist of coarse materials like sands and gravels. The fillable porosity should be taken as the difference between the existing and saturated water contents of the material outside the wetted zone below the infiltration system. Therefore, according to the literature, the value of \( S_y \) which is determined in this study as 0.18, is justified, and was chosen as the multiplier to convert the dewatered volume to the dewatered pore volume. An estimate of the change in storage volume is presented in Fig. 3-16.

The piezometric maps which were employed to calculate the amount of change in water storage of the two successive years of the 1992 and 1993 are presented in Fig. 3-14 as a sample of the whole procedure of the 1992 to 2012. A noticeable volume change of 39.6 million m\(^3\) (Mm\(^3\)) was observed during the 1993-1997 period, which is \(~8.0\ Mm^3\) per year. The second increase \(~3.3\ Mm^3\), although a minor one, occurred during the 2004-2005 period. Apart from this, continuous dewatering has occurred from 1997 to 2012 with an estimated depletion of 113.9 Mm\(^3\) giving a net result of 71.0 Mm\(^3\) dewatering for the 1993-2012 period. This is a stern warning that the limited natural and induced recharge cannot supply unlimited withdrawal; therefore, given the shallow depth of the aquifer, its probable drying up in a very near future is imminent.
Fig. 3-14. The 3 dimensional presentation of the two overlaid groundwater level surfaces of the year 1997 over the 1993 which have been used to calculate the change in aquifer storage in these two successive years.

Fig. 3-15. Variation in average water level in the observation wells before and after flooding for four consecutive months during the period of minimum water withdrawal. WL is groundwater level. 1st to 4th month refer to water level changes at the first to fourth month after the flooding events.

Fig. 3-16. Temporal changes in storage volume of the aquifer. Mm³ is million m³.
3.3.6 Recharge estimation

Results of extraction are presented in Table 3-7. Applied agricultural water and withdrawal estimation were 13.82 and 15.27 Mm³, respectively. These two methods of extraction estimation resulted in different but close values as some of the extracted water is not used as irrigation application and is lost during transport within the earth canals between the farm fields. The amount of loss is 1.45 Mm³ or 9.5% of the withdrawn water. Therefore extraction volume based on the withdrawal water (15.27 Mm³) was chosen as the basis for further calculations.

Of the 15.27 Mm³ extracted water from the aquifer, 3.2 Mm³ was estimated as the irrigation return flow, which is assumed to return to the aquifer (chapter 2). As the extracted water for irrigation is 14.24 of 15.27 Mm³ then the return flow of 3.2 Mm³ is 22.5% of irrigation applied water. This percentage is close to those values reported for the main crops in the dry states of the USA, 15-55% (average 24%) (Sabol et al., 1987) and to 24% which is reported by Jafari et al. (2012) for a similar environment and management settings to our study area. There is an uncertainty about the specific time when the return flow reaches the GW (Bouwer, 1987). However, continuation of farming activities in successive years guarantees a permanent irrigation return volume of what is calculated for a particular year.

Considering a calculated dewatered pore volume as 22.95 and a $S_r$ of 0.18, the depleted water from the aquifer ($\Delta S$) was obtained as 4.13 Mm³ during the hydrological year 2010-2011. The return flow was assumed as 3.2 Mm³ hence, the recharge was calculated at 7.94 Mm³, which was a consequence of both artificial recharge and natural replenishment (Table 3-8).

3.3.7 Artificial recharge estimation

i) Based on the flow data

The artificial recharge data in 2010-2011 are presented in Table 3-9. During the flooding events from 28 January to 2 February 2011, a total volume of 6.92 Mm³ of flood water was retained in the BZ FWS systems. These events ponded the entire FWS systems of 2033 ha (22.33 Mm²) and resulted in an average depth of 0.34 m on the FWS surface. The duration of the infiltration was counted as 6 days, evaporation rate ($e$) as 0.024 m day⁻¹, the sum of ET ($\Sigma ET$) after the event for the 30 days was calculated as 0.075 m, weighted average of
infiltration rate \( (i) \) was calculated as \( 0.28 \text{ m day}^{-1} \) and therefore, the recharged water was estimated as \( 0.24 \text{ m} \) using Eq. (3-11) (Table 3-10).

Table 3-7. Calculations of water extraction in the hydrological year October 2010 to September 2011.

<table>
<thead>
<tr>
<th>Consumers</th>
<th>Area, ha</th>
<th>Ave. applied, m³ ha⁻¹</th>
<th>Consumption, Mm³</th>
</tr>
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<tr>
<td>Winter crops</td>
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<td>4800</td>
<td>5.68</td>
</tr>
<tr>
<td>Summer crops</td>
<td>725</td>
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<td>7.11</td>
</tr>
<tr>
<td>Tree plantations</td>
<td>132</td>
<td>7450</td>
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</tr>
<tr>
<td>Domestic use</td>
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</tr>
<tr>
<td><strong>Sum</strong></td>
<td></td>
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<td><strong>13.82</strong></td>
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</table>

Based on measured applied water

<table>
<thead>
<tr>
<th>Sources</th>
<th>No. wells</th>
<th>Mean discharge, L s⁻¹</th>
<th>Mean working times, Hr yr⁻¹</th>
<th>Abstraction, Mm³</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wells</td>
<td>Total 148</td>
<td>Active 88</td>
<td>7.95</td>
<td>5655</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>14.24</td>
</tr>
<tr>
<td>Tree plantations</td>
<td></td>
<td></td>
<td></td>
<td>0.98</td>
</tr>
<tr>
<td>Domestic use</td>
<td></td>
<td></td>
<td></td>
<td>0.05</td>
</tr>
<tr>
<td><strong>Sum</strong></td>
<td></td>
<td></td>
<td></td>
<td><strong>15.27</strong></td>
</tr>
</tbody>
</table>

| Transport losses%    | 10.3      |

\(^1\)Applied water for tree plantations is calculated by the ET₀ mapping. The water consumption of tree plantations is considered as a source of GW extraction due to their roots access to the GW table (observed in this study at the depth of 28 m. Mm³ is million m³. The total extraction by water withdrawal was calculated for every well’s discharge and working hours and summed up for the all wells. The means are shown here to be indicative of the wells’ discharge and working times. Transport losses is calculated as the ratio irrigation applied/withdrawn water (5.68+7.11)/14.24.

Table 3-8. Water budget of the total recharge in the hydrological year 2010-2011.

<table>
<thead>
<tr>
<th>Inputs, Mm³</th>
<th>Outputs, Mm³</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \Delta S )</td>
<td>4.13</td>
</tr>
<tr>
<td>RF</td>
<td>3.20</td>
</tr>
<tr>
<td>R</td>
<td>7.94</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>15.27</strong></td>
</tr>
<tr>
<td>E</td>
<td>15.27</td>
</tr>
</tbody>
</table>

\( \Delta S \) is change in GW storage, R is recharge, RF is return flow, E is total extraction from the GW, Mm³ is million m³. R is calculated as the remainder of the water budget Eq. (3-7).

These events ponded the entire FWS systems of 2033 ha (22.33 Mm²) and resulted in an average depth of 0.34 m on the FWS surface. The duration of the infiltration was counted as 6 days, evaporation rate \( (e) \) as 0.024 m day⁻¹, the sum of \( ET (\Sigma ET) \) after the event for the 30 days was calculated as 0.075 m, weighted average of infiltration rate \( (i) \) was calculated as 0.28 m day⁻¹ and
therefore, the recharged water was estimated as 0.24 m using Eq. (3-11) (Table 3-10).

**Table 3-9. Flow rates of floods diverted into the FWS systems in 2011.**

<table>
<thead>
<tr>
<th>Flooding date</th>
<th>Mean Discharge, m³ s⁻¹</th>
<th>Duration, hr</th>
<th>In-flow</th>
<th>Out-flow</th>
<th>Retained flow</th>
</tr>
</thead>
<tbody>
<tr>
<td>28-Jan-11</td>
<td>115</td>
<td>3</td>
<td>1.24</td>
<td>0.25</td>
<td>0.99</td>
</tr>
<tr>
<td>1-Feb-11</td>
<td>159</td>
<td>13</td>
<td>7.44</td>
<td>4.62</td>
<td>2.82</td>
</tr>
<tr>
<td>2-Feb-11</td>
<td>203</td>
<td>8.5</td>
<td>6.21</td>
<td>3.10</td>
<td>3.11</td>
</tr>
<tr>
<td>sum</td>
<td></td>
<td></td>
<td>14.90</td>
<td>7.97</td>
<td>6.92</td>
</tr>
</tbody>
</table>

In-flow and out-flow are the volume of incoming and outgoing flood flow to the floodwater spreading system. Retained flow is the volume of flood flow retained by the system.

**Table 3-10. Components of the recharge calculation based on the flow data.**

<table>
<thead>
<tr>
<th>Retained flow, Mm³</th>
<th>Area, ha</th>
<th>D, m</th>
<th>e, m day⁻¹</th>
<th>i, m day⁻¹</th>
<th>ΣET, m</th>
<th>AR, m</th>
<th>AR, Mm³</th>
</tr>
</thead>
<tbody>
<tr>
<td>6.92</td>
<td>2033</td>
<td>0.34</td>
<td>0.024</td>
<td>0.28</td>
<td>0.075</td>
<td>0.24</td>
<td>4.84</td>
</tr>
</tbody>
</table>

*D is depth of ponded water, e is evaporation rate, i is infiltration rate, ΣET is the sum of ET in the 30 days after flooding event, AR is recharge.*

As a result, the volume of artificial recharge due the FWS on the 20.33 Mm² of the FWS systems was estimated as 4.84 Mm³. The ratio of artificial recharge to the depth of ponded water (0.24 and 0.34, respectively) was 0.7. Hendrickx et al. (1991) reported similar values (0.6 to 0.8) for an alluvial stony fan resembling our study site.

**ii) Based on the water budget**

The contour lines of change in GW level for the duration of 3 months after the flooding events which is placed on the surrounding area of the immediate influence of FWS systems is presented in Fig. 3-17. A mound of GW rise was formed in the Southeast part of the FWS systems which shows the place of maximum influencing area of the system. It then gradually decreased in the West and Southwest direction. The surrounding area has a size of 3232 ha and volume of GW rise was calculated at the 9.03 Mm³. According to the determined $S$, as 0.18, the volume of change in GW storage due to the artificial recharge ($ΔS$) was obtained as 1.625 or ~1.63 Mm³. During the three months after the event (the period which was considered for GW rise), some part of recharged water was used as irrigation water and consumed by tree plantations (Table 3-11). The size of cultivated area of the
farmlands located inside the FWS systems influencing boundary was measured as 705 ha, the rate of applied irrigation was considered as 4800 m$^3$ ha$^{-1}$ hence, the extracted volume of irrigation water was obtained as 3.38 Mm$^3$. The amount of extracted water by tree plantations was calculated based on summed up $ET_a$ in the same period as 2300 m$^3$ ha$^{-1}$. Hence, the extracted volume of irrigation water was obtained as 0.30 Mm$^3$. As the amount of $RF$ was obtained as 22.5%, the volume of $RF$ was calculated as 0.73 Mm$^3$ (22.5% of the 2.95 Mm$^3$ crops irrigation + 0.31 Mm$^3$ transport losses) and therefore, the artificial recharge is estimated as 4.46 Mm$^3$ (Table 3-12).

Table 3-11. Extraction from the aquifer during the period of February to April 2011.

<table>
<thead>
<tr>
<th>Land uses</th>
<th>Area, ha</th>
<th>Extracted water, m$^3$ ha$^{-1}$</th>
<th>Mm$^3$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Irrigated crops</td>
<td>615</td>
<td>4800</td>
<td>2.95</td>
</tr>
<tr>
<td>Transport Losses (10.3%)</td>
<td></td>
<td></td>
<td>0.31</td>
</tr>
<tr>
<td>Tree plantations</td>
<td>132</td>
<td>2300</td>
<td>0.30</td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td></td>
<td>3.56</td>
</tr>
</tbody>
</table>

The size of irrigated crops is measured for the cultivated part of the farms located in the FWS systems influencing area.

Table 3-12. Water budget of the January to April 2011 for the area under the immediate influence of FWS systems.

<table>
<thead>
<tr>
<th>Inputs, Mm$^3$</th>
<th>Outputs, Mm$^3$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$AR$</td>
<td>4.46</td>
</tr>
<tr>
<td>$RF$</td>
<td>0.73</td>
</tr>
<tr>
<td>Total</td>
<td>5.19</td>
</tr>
<tr>
<td>$E$</td>
<td>3.56</td>
</tr>
<tr>
<td>$\Delta S$</td>
<td>1.63</td>
</tr>
</tbody>
</table>

$\Delta S$ is change in GW storage, $AR$ is artificial recharge, $RF$ is return flow, $E$ is total extraction from the aquifer, Mm$^3$ is million m$^3$. AR is calculated as the remainder of the water budget Eq. (3-12).

Table 3-13. Evaluation of the artificial to total recharge ratio.

<table>
<thead>
<tr>
<th>$R$, Mm$^3$</th>
<th>$AR$, Mm$^3$</th>
<th>$NR$, Mm$^3$</th>
<th>$AR/R$</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.94</td>
<td>4.84*</td>
<td>2.87</td>
<td>0.61</td>
</tr>
<tr>
<td>7.94</td>
<td>4.46**</td>
<td>3.25</td>
<td>0.56</td>
</tr>
</tbody>
</table>

$R$ is recharge, $AR$ is the artificial recharge calculated by flow data (*) and water budget (**), respectively, $NR$ is natural recharge and Mm$^3$ is million m$^3$. NR is calculated as the remainder of the A-AR.
Comparison of the two methods of artificial recharge estimation and their ratio to total recharge is shown in Table 3-13. The contribution of the artificial recharge is calculated as 4.84 and 4.46 Mm$^3$ by the flow data and the water budget, respectively. The lower value obtained by the water budget method is expected as it is essentially based on the change (rise) in GW storage as influenced by the net recharge of the groundwater and the one obtained by the flow data is based on the total infiltrated water to the deep layers. Consequently, the difference between the two values appear to be the portion of the net infiltration (7% of 4.84 Mm$^3$) which had not contributed to recharge and is moved horizontally. Conditional to selecting one of the resulted values to the artificial recharge, 56 to 61% of the total recharge could be attributed to the impact of the FWS systems for that hydrological year. Hashemi et al. (2013) found an average ratio of 61% by a modeling approach in the same study area with 80% for extreme events and 41% for normal events. As they did not define the criteria for differentiating between extreme and normal events, we might not judge whether the studied event in this study is considered
as extreme so that it can be compared with their results. However, looking to the range of the flooding event volumes (Table 3-6) in the study site, the studied event can hardly be assumed as extreme. The study area received 276 mm of precipitation in the hydrological year 2010-2011 hence, considering the BZ aquifer’s areal extent as 7600 ha, the equivalent height of rainfall is 20.9 Mm$^3$. The remainder of the total and artificial recharge subtraction is counted as natural recharge therefore, and calculated natural recharge is 2.87 to 3.25 Mm$^3$ and the share of precipitation to the natural recharge is 14-15%. The magnitude of natural recharge has been studied worldwide by many authors. A review undertaken by Bouwer (2002) revealed that natural recharge is typically about 30–50% of precipitation in temperate humid climates, 10–20% of precipitation in Mediterranean type climates, and about 0–2% of precipitation in extremely dry climates. The low rate of direct rainfall recharge in dry regions can be attributed to high runoff amounts associated with the high rainfall intensity and poor vegetation cover or barren soil surfaces and to deep water table levels. Thus, GW recharge takes place mainly through runoff infiltration process (Bedinger, 1987; Bouwer, 1996; Bouwer, 2000). Scanlon et al. (2006) proposed a worldwide contribution of precipitation to natural recharge as 0.1 to 5%. Therefore, the proportion of artificial to natural recharge obtained in this study seems within the expected range for the Mediterranean type climate of our study area.

### 3.4 Conclusion

Spatial changes in the GW level indicated that the highest drop in the water level has occurred in places where irrigated fields were concentrated. On the other hand, the lower recession in OW2, OW5 and OW6 located inside the FWS systems proved that the area under the direct impact of the FWS systems is less susceptible to water withdrawal.

Aquifer storage volume showed a noticeable increase (40 Mm$^3$) in the initial years of the study (1993-1997), when the number of irrigation wells were limited. On the other hand, a greater decrease in aquifer storage (114 Mm$^3$) was observed during 1997 to 2012 coinciding with the boom of water extraction in the entire study area.

In the particular hydrological year of 2010-2011, when reliable data to evaluate abstraction and depletion were available, the total recharge was calculated as 7.94 Mm$^3$ and the artificial recharge as 4.84 and 4.46 Mm$^3$ equal to 56 to 61%.
The difference between the values resulted from the two methods of artificial recharge used and might be attributed to the part of net infiltration which moves horizontally and determined as 7% of the net infiltration.

Decrease in infiltration rate and clogging due to siltation in all artificial recharge systems, especially those utilizing turbid floodwater, is unavoidable and lowers their efficiency, therefore, the system studied here was assumed to have a lifetime of 12 to 15 years initially (Kowsar, 1991). But as our study shows, the system was still functioning properly in 2010-2011 after about 32 years from its initiation. Experience has shown that regular maintenance of such systems, particularly after major events, is essential and elongates their lifetime and efficiency as it has happened in the study area. This requires small yearly investments, which yields conspicuous returns. Artificial recharge of GW through FWS, undoubtedly, is an activity that may sustain desert-dwellers if accompanied with a prudent water withdrawal.
Photo 3-1. The observation well inside the study area (OW2) and OW4 (B) and the water level measurement tools.
Photo 3-2. Hydrometric station on the Bisheh Zard Ephemeral River in the study area during the flooding event of 01/02/2011 (A), a downstream view of the diversion dam of the same River showing the remainder of the diverted flood which flows within the River (B) and a drop which delivers part of floodwater to the FWS systems (C). See Fig. 3-6 for the placement of the structures.
Photo 3-3. Measurement structures; (A) the drops which divert the floodwater from the Bisheh Zard River to the FWS systems, (B) and (C) two types of broad crested weirs for delivering and measuring the flow to the FWS systems basins at the study site (see Fig. 3-7 for the placement of the structures).
Photo 3-4. Operational well and connecting pipe with special Try square device for measuring well discharge in an agricultural field at the study site. It should be noted that the photo is taken to show the main settings and device but the discharge which is shown in the photo is a sample of partially full flow and was measured by California pipe method (see Appendix 4).
Photo 3-5. Applied irrigation water measurements in an agricultural field inside the study area by cut throat flume; by side view of the flume (A) and top view with Carpenters level measuring tool for leveling the flume (B).
Chapter 4. Improved calibration of time domain reflectometry (TDR) for soil water content measurements in stony soils

This chapter is based on a modified article:

4.1 Introduction

Time domain reflectometry (TDR), which exploits dielectric properties of the medium under study, is essentially a guided wave technology that has been widely used in electrical engineering for detection of cable breakages (Yu and Yu, 2011). Using the dielectric constant (or permittivity) for moisture measurement in soils and other porous media was first reported by Thomas (1966). Several studies showed that the dielectric permittivity of porous media is primarily dependent on its water content (Lundien, 1971; Cihlar and Ulaby, 1974; Hoekstra and Delaney, 1974; Selig and Mansukhani, 1975; Okrasinski et al., 1978; Topp et al., 1980). Relatively simple but reliable measurements of dielectric permittivity in the frequency domain of 1 MHz to 1 GHz is an applicable and effective way of soil water content measurement (Topp et al., 1982). Several methods were discussed by Selig and Mansukhani (1975) for dielectric permittivity measurement. In applications, the water content $\theta_v$ is related to the dielectric permittivity, $K_a$, using the following empirical relation (Topp et al., 1980):

$$\theta_v = -5.3 \times 10^{-2} + 2.92 \times 10^{-2} K_a - 5.5 \times 10^{-4} K_a^2 + 4.3 \times 10^{-6} K_a^3 \quad (4-1)$$

This relation is reported by the authors as holding for most mineral soils and was verified by various researchers, leading to suggestions that it may generally be adopted (Coppola et al., 2013). However, it has been noted that soil-specific calibration equations can significantly deviate from the universal calibrations for various reasons.

Dirksen and Dasberg (1993) examined changes in the dielectric permittivity for particular $\theta_v$ values in 11 soils with different mineral and clay contents. They showed that the coefficients of the Topp et al. (1980) equation cannot be considered constant for all soil types. Deviations from Topp’s equation appear to be rather due to low bulk density and thus higher air volume fraction at the same $\theta_v$ associated with fine-textured soils than to tightly bound water with low dielectric permittivity. Furthermore, Malicki et al. (1996) showed that bulk density, and thus porosity, substantially affects the relationship between dielectric permittivity and water content. They presented two equivalent, empirical, normalized calibration functions, one accounting for bulk density and the other for porosity to reduce the root mean square error of the dielectric TDR determinations of moisture. As indicated by Cataldo et al. (2009), the individuation of an optimal calibration procedure for each of the materials
under investigation is a key point to improve the accuracy of the results. Bittelli et al. (2008) noticed that TDR always overestimated soil water content (SWC) in clay soils as compared to standard oven-dry gravimetric measurements, up to 20% of soil saturation. They derived a new calibration curve for replacing existing calibrations curves, like the widely-used Topp’s equation.

Traditionally, experimental measurements and different procedures are used to extrapolate empirical calibration curves, which, as a result, are not readily comparable. Arsoy et al. (2013) proposed an artificial neural network to incorporate dry bulk density, specific gravity and fines content to minimize the errors arising from using the universal TDR equation. Besides, Hignett and Evett (2008) alert that sensors of SWC measurements generate results that are barely related to the authentic field settings if used for unsuitable purposes. Numerous studies have indicated that the performance of soil-water content sensors, which have achieved widespread use and provide instant measurements, should be assessed before being used in specific soils (Varble and Chávez, 2011). The reliability of the TDR method for measuring water content in stony soils, as they prevail in our study site, is rarely reported in the literature. Coppola et al. (2013) obtained good relationships between the TDR-measured and observed $\theta_v$ for different stone fractions. They used 2-5 mm stones to prepare different stone fractions, leading to homogeneous stony soil mixtures which differ from non-homogeneous and larger in size (2-250 mm) mixtures of our soils.

A potential error also arises when long cables are used as cable length affects the waveform of a given probe due to degraded rise time of the reflected pulse as cables become longer (Tektronix, 1987). (Logsdon, 2000) and Thomsen et al. (2000) showed that long cables influence the soil water content reading from TDR. When using long cables, as was necessary in this study, effect of cable length on the dielectric permittivity and associated water content must therefore be determined.

The objectives of this chapter are a) to test available calibration equations for accurately deriving $\theta_v$ from dielectric permittivity in coarse textured, stony soils and to develop new equations, which would improve the relationship, b) to examine the effect of selected capture windows, probe type and cable length on the measurement error and providing an improved calibration equation that
reduces additional error, and c) to find out the practical implications of using the TDR method in these soil types.

### 4.2 Materials and methods

The Gareh Bygone Plain is a debris cone and an alluvial fan which was formed by the eroded material delivered from the Agha Jari Formation (inter-bedded sandstones, siltstones and marls) and the Bakhtyari Formation (mainly conglomerate). Therefore, the soils formed on this materials are characterized as multi layered stony soils which are classified according to Soil Taxonomy (Soil-Survey-Staff, 2010) as Torriorthents, Haplocalcids and partly Haplocambids (Kowsar and Pakparvar, 2003).

#### 4.2.1 Soil sampling

Soil samples were taken from an experimental well of 30 m depth, located at 28° 36' 37" N and 53° 56' 02" E in a recharge basin of the FWS systems. Details about the study area are given in chapter 2. The profile of the well consists of multiple layers with distinct features, but commonly belong to three main layers, i.e., a top layer of recently deposited sediment (loam), and two types of sub-surface layers, sandy loam and sand having different amount of stones (Table 4-1). These main layers are repeated along the vadose zone (see chapter 5). Three representative disturbed samples of 50 kg were taken from depths of 0-10, 10-109 and 109-150 cm (layers 1 to 3 respectively).

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>0-10</th>
<th>10-109</th>
<th>109-150</th>
</tr>
</thead>
<tbody>
<tr>
<td>Depth (cm)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-10</td>
<td>0.0</td>
<td>4.3</td>
<td>15.7</td>
</tr>
<tr>
<td>10-109</td>
<td>4.3</td>
<td>10.2</td>
<td>15.6</td>
</tr>
<tr>
<td>109-150</td>
<td>15.7</td>
<td>15.6</td>
<td>21.5</td>
</tr>
</tbody>
</table>

Table 4-1. Particle size distribution of the sampled layers.

- Textural class of below 2 mm is calculated based on no stone percentage. Texture is textural class (USDA) L is loam, SL is sandy loam and S is sand.

#### 4.2.2 Experimental setup

Bulk soils were air dried and their SWC was measured gravimetrically. Two kg of each sample was used to particle size analysis (PSA). This was carried out using the hydrometer method combined with sieve analysis to characterize
the range of particle diameter from 0.002 up to 2 mm (Gee and Or, 2002). The particle density of the samples was determined by measuring the mass and the water displacement.

The remaining material of each layer (1 to 3), including gravel and stones, was divided into three homogenized sub-samples (as replications) and placed into plastic PVC containers of 30×50 cm (diameter, height) whose bottom ends had been removed and covered with plastic mesh. Bulk densities were measured by considering the air dry weight of the bulk sample and the volume of soil filled part of the containers. A 20 cm long three-rod uncoated metal waveguide (home-made) and a 15 cm long two-rod probes connected to waveguide connector were used with a main device of TDR Trase system 6050X1 (Soilmoisture Equipment Corp., USA). Both probes were inserted vertically into each container to a depth of 25 cm. Each container was placed in a drum filled with water to a height of 2 cm above the soil surface to make sure that the water replaces the air in the soil and to avoid entrapped air. The containers were left as they were until the soil showed a shiny surface. The water jacket was then drained from the drums. Gravimetric SWC was calculated based on the air dry SWC and weights before and after saturation. Then bulk densities were used to find out the $\theta_v$ at saturation. Thereafter, dielectric permittivity, $K_a$ and $\theta_v$ of both probes for capture windows of 10, 20 and 40 nano-seconds (ns) were recorded and the containers were weighed until the change in $\theta_v$ was less than 0.01 m$^3$ m$^{-3}$, which occurred after 81 days. The TDR measured $\theta_v$ ranged from 0.032 to 0.385 m$^3$ m$^{-3}$. The period in which the range of soil water content started from the saturation to its minimum level at the laboratory. During the experiment, air temperature was constant at ~22 °C.

In order to minimize the errors incurred from stoniness, the same three sub-samples of previous section were sieved to remove the particles >2 mm. Air dry gravimetric SWC was measured and a method based on Gong et al. (2003) was used for measuring the TDR and the observed $\theta_v$ for the sieved sub-samples. The samples were packed into the PVC tubes of 30×15 cm (height, diameter) and a 20 cm long three-rod uncoated metal waveguide (home-made) was inserted vertically into each one. Bulk densities were set to 1.30, 1.45 and 1.60 g cm$^{-3}$ for the samples of the layer 1 to 3, respectively. The same procedure as the last section, entering the water from the beneath of the soil sample, was employed. However, due to a smaller sample size the samples were moistened by connecting two Mariotte bottles to the tubes’ lower end.
until the soil was saturated (showing shiny surface). Measurements of $K_a$ and $\theta_v$ for capture windows of 10 and 20 ns and changes in tube weights for the tubes were then taken. The volumetric water content was measured using the conventional gravimetric method. Measurements were continued until the final $\theta_v$ reached $0.01 \text{ m}^3 \text{ m}^{-3}$ (14 readings ranging from 0.01 to 0.28 $\text{ m}^3 \text{ m}^{-3}$).

To validate the effect of cable length on the $\theta_v$ measurements, the previous trial was repeated without gravimetric measurements. Samples were again moistened and $K_a$ and $\theta_v$ were measured using the original 2 m long probe cable connected with extension cables of 3, 5, 10, 15, 20, 25 and 30 m length (manufactured by the same company).

### 4.2.3 Data analysis

The size of stones in our soils (2-250 mm) is somewhat larger than the TDR probes’ rod spacing which is not enough wide for involving the contribution of both the fine soil and the stones in the bulk soil. Thus, following the guidelines of Birchak et al. (1974) and Coppola et al. (2013), we distinguished the dielectric permittivity of different components of the soil mixture i.e., fine soil, stones, air and water. A semi empirical model proposed by Birchak et al. (1974) for relating the bulk (the mixture of different components) dielectric permittivity ($K_{ab}$) to the volumetric fraction, $V_i$, of its components and the corresponding dielectric permittivity, $K_{ai}$ was used. The model may be written as:

$$K_{ab}^\beta = \sum_{i=1}^{N} V_i K_{ai}^\beta$$  \hfill (4-2)

The component $\beta$ in Eq. (4-1) is an empirical constant summarizing the geometry of the medium respect to the applied electric field. A value of $\beta = 0.5$ has been proposed for homogeneous soils (Birchak et al., 1974; Ledieu et al., 1986). Coppola et al. (2013) found an average value of $\beta = 0.55$ for different stone fractions. In a stony soil, by neglecting the effect of the bound water and separating the individual contribution of the fine soil particles ($V_i$) and the stones ($V_{st}$), Eq. (4-1) might be expanded as:

$$K_{ab}^\beta = (1 - \varphi - V_{st})K_{as}^\beta + V_{st}K_{ast}^\beta + (\varphi - \theta_v)K_{aa}^\beta + \theta_v K_{aw}^\beta$$  \hfill (4-3)
where φ, θ, and Vṣ are the soil porosity, the volumetric water content and stoniness, respectively, referred to the bulk soil (fine soil plus stones). Kṣ, Kas, Kaa, and Kaw are the dielectric permittivity for stones, fine soil particles, air and water, respectively. By assigning the pre-determined parameters, the TDR Kab field measurements can be converted to the corrected θb(Vṣ) as follows (Coppola et al., 2013):

$$
\theta_v = \frac{\frac{\beta}{K_{ab} - \phi K_{aa}} - (1 - \phi - V_{st}) K_{as} - V_{st} K_{ast}}{K_{aw} - K_{aa}}
$$

(4-4)

In this study the porosity of bulk soils (φ) was determined based on bulk density and particle densities, Vṣ was measured by sieving for three soil layers and dielectric permittivity of water and air (Kaw, Kaa) was measured by TDR as 80 and 1 respectively. The Kṣ and Kas were set to 7.3 and 4.3 as measured by Coppola et al. (2013). This is in the range of Ka for solid materials of 4-16, indicated by Bittelli et al. (2008). To find the optimum value of β, a set of paired θv (observed and calculated by Eq. (4-4)) were used for optimization of β by least square method with solver add-in in excel.

In the following, the TDR-based θv estimated by Eqs. (4-1) and (4-4) as θvTp and θv,mx, respectively (subscript Tp is for Topp’s equation; subscript mx is for mixture equation) are used.

### 4.2.4 Validation of the results

In stony soils, the improvement in θv estimation because of considering the mixture of soil-stone (Eq. (4-4)) was assessed based on comparison between θvTp and θv,mx. In stone-free soils, among the whole data set, two-third was used to generate the new θv-Ka relationships and one-third for evaluating them against Topp’s equation. Polynomial linear regressions were used to find the best equation for our soils under the experimental conditions.

### 4.2.5 Statistical indices for model validation

Pearson’s correlation coefficient (r) and coefficient of determination (r²) describe the degree of colinearity between simulated and measured data (Moriasi et al., 2007). The correlation coefficient, which ranges from −1 to 1, is an index of the degree of linear relationship between observed and simulated data. If r = 0, no linear relationship exists. If r = 1 or −1, a perfect positive or negative linear relationship exists. Similarly, r² describes the proportion of the variance in measured data explained by the model. r² ranges from 0 to 1, with
higher values indicating less error variance (Santhi et al., 2001; Van Liew et al., 2007). Although \( r \) and \( r^2 \) have been widely used for model evaluation, these statistics are over-sensitive to high extreme values (outliers) and insensitive to additive and proportional differences between model predictions and measured data (Legates and McCabe, 1999). The graphic results were also quantified by calculating the root mean squared error (RMSE), which supply a measurement of scatter around the 1:1 line. For an ideal prediction, its value should be close to 0. According to Singh et al. (2004) RMSE values less than half the standard deviation of the measured data may be considered low. As we worked with two estimated data series (uncorrected and corrected for stones data series), characterized by different standard deviations, we opted for a normalized statistic, named the RMSE-observations standard deviation ratio (RSR) (Moriasi et al., 2007). RSR is calculated as the ratio of the RMSE and standard deviation of measured data. Thus, it includes a normalization factor, so that the resulting statistic can apply to different data series. RSR varies from the optimal value of 0, which indicates zero RMSE or residual variation and therefore perfect model simulation, to a large positive value. The lower RSR, the lower the RMSE, the better the model simulation performance. RSR is calculated according to the following equation:

\[
RSR = \frac{\sqrt{\sum_{i=1}^{N}(O_i - S_i)^2}}{\sqrt{\sum_{i=1}^{N}(O_i - \bar{O}_i)^2}}
\]  

(4-5)

where \( O_i \) and \( S_i \) are the observed and the simulated values, \( \bar{O}_i \) is the mean of observed data, and \( N \) is the total number of observations.

### 4.3 Results and discussion

#### 4.3.1 Soils with original stoniness

Soils with original stone volumes showed a deviation of the estimated vs. observed \( \theta_v \) with a declining deviation from layers 1 to 3 for \( \theta_vTp \) (\( r^2 \) of 0.71, 0.88 and 0.91) and \( \theta_vmx \) (\( r^2 \) of 0.71, 0.91 and 0.93) both corresponding to layers 1 to 3, respectively. Textural classes of the soil layers are in the order of fine to coarse (loam, sandy loam and sand, respectively) showing the less deviation in coarser soil textures (Fig. 4-1). Many researchers have considered the effect of soil texture on the reliability of the TDR measurements (Dirksen and Dasberg, 1993; Jacobsen and Schjønning, 1993; Ponizovsky et al., 1999; Gong
et al., 2003; Bittelli et al., 2008; Stangl et al., 2009). As a general conclusion, a higher amount of fine particles (clay sized), which leads to a lower bulk density of soils, causes a higher deviation in the TDR vs. gravimetrically measured water content. Considering the effect of stones by separating the different fractions, reflecting in Eq. (4-4), is tend to the nearly similar $r^2$ in $\theta_{,mx}$ and $\theta_{,Tp}$ however, noticeable decline in RSR and RMSE of $\theta_{,mx}$ vs. $\theta_{,Tp}$ (Table 4-2) indicates better performance of mixture model as compared to conventional Topp's equation in $\theta_{v}$ estimation. Similar $r^2$ infers insensitivity of this statistics to additive and relative differences between model predictions and measured data while the RMSE and RSR are capable to demonstrate the proportional disparities.

Capture windows and the probe types used in the stony soils showed diverse effects on $\theta_{v}$ estimation (Table 4-2). However, capture window 20 ns showed minor improvement in $\theta_{v}$ estimation when compared to 10 ns, major improvements are found by using probe type 15 cm two-rod (connector) as compared to 20 cm three-rod (buriable). The more strength and shorter length of connector type's rods help to minimize lateral deformations of probe rods during the insertion to the stony soil as compared to three-rod probes. Furthermore, a procedure of primarily zero set in TDR measurement, needed for connector type might lead to more precise measured $K_a$.

Therefore, RSR and RMSE resulted from capture window 20 ns and the 15 cm two-rod (connector) probe type was used to make final comparisons (Fig. 4-1). RMSE values of mixture equation (0.04, 0.03 and 0.02 m$^3$ m$^{-3}$, respectively for layers 1 to 3) shows good improvement as compared to Topp's equation (0.07, 0.10 and 0.09, respectively for layers 1 to 3) inferring the final error in $\theta_{v}$ estimation as 2 to 4 percent in volumetric water content when considering the stones. RMSE values of mixture equation are equal or less than the half standard deviation of observed $\theta_{v}$ (Table 4-2) which might be considered as low. The same conclusion is inferred from RSR values showing the less RSR of mixture equation (0.87, 0.32 and 0.71) as compared to Topp’s equation (1.34, 1.07 and 3.05) corresponding to soil layers 1 to 3, respectively. Therefore good improvement is gained by applying the mixture equation, model performance.

Coppola et al. (2013) similarly obtained good improvement in $\theta_{v}$ estimation for bulk soil by employing Eq. (4-4) with better agreement between observed
\( \theta_v \) and \( \theta_v m_x \) in compare with \( \theta_v T_p \) (RSR of 1.13 and 2.94 for \( \theta_v m_x \) and \( \theta_v T_p \), respectively). The distinction between our soils and their studied soils is the stone size and configuration. They used 2-5 mm stones to prepare different stone fractions, leading to homogeneous mixtures. Whereas, the fragmental soils of our study were collected from the research site where a heterogeneous mixture of stones in size (2-250 mm) and shape were naturally distributed. This difference in stoniness is reflected in lower mean error of 0.004, as compared to our final error of 0.02 to 0.04, indicating the better performance of Eq. (4-4) in the stony soils with stone size comparable to the probe rod's spacing.

4.3.2 Stone-free soils

The second experiment, in which the stones were removed, resulted in local calibration curves to be applied for precise water content estimation in individual soil layers. As shown in Fig. 4-2, the plotted points are explicitly closed to the regression lines leading to \( r^2 \) of more than 0.99 and RMSE of 0.002 m\(^3\) m\(^{-3}\) for the three soil layers.

Importance of calibration equations (presented in Table 4-3) in this study refers to the need for minimizing the error in water content detection to an extent of practically 1 percent as the minor changes in soil water content in arid landscapes are important due to lack of accessible water. In addition, as the insertion of the TDR probes in original stony layers are practically impossible due to occasionally presence of large stones (<250 mm), it would be suggested to make the lateral holes filled with original fine soil for placing the TDR probes used for long term monitoring (chapter 5). Furthermore, as our data show, the main shortcoming of the built-in Topp’s equation in the stone-free soils is the failure in properly estimating the \( \theta_v \) below 0.05 m\(^3\) m\(^{-3}\).
Table 4-2. Statistical indices for model performance evaluation in stony samples.

<table>
<thead>
<tr>
<th>Soil layers</th>
<th>Texture</th>
<th>$V_{st}$ %</th>
<th>$\phi$</th>
<th>SD</th>
<th>CW, ns</th>
<th>TDR probe</th>
<th>Mix Topp Mix Topp</th>
<th>$\beta$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>L</td>
<td>6.7</td>
<td>0.49</td>
<td>0.08</td>
<td>10</td>
<td>1</td>
<td>1.24</td>
<td>0.06</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.89</td>
<td>0.04</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.42</td>
<td>0.07</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.87</td>
<td>0.04</td>
</tr>
<tr>
<td>2</td>
<td>SL</td>
<td>22.1</td>
<td>0.43</td>
<td>0.09</td>
<td>10</td>
<td>1</td>
<td>0.44</td>
<td>0.04</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.41</td>
<td>0.04</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.29</td>
<td>0.03</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.32</td>
<td>0.03</td>
</tr>
<tr>
<td>3</td>
<td>S</td>
<td>52.8</td>
<td>0.36</td>
<td>0.07</td>
<td>10</td>
<td>1</td>
<td>2.11</td>
<td>0.06</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.78</td>
<td>0.03</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.96</td>
<td>0.06</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.71</td>
<td>0.02</td>
</tr>
</tbody>
</table>

Texture is textural classes (USDA) L is loam, SL is sandy loam, S is sand. $V_{st}$ is stone (>2 mm particles) fraction and $\phi$ is porosity. SD is standard deviation of observed water contend data. CW is capture window of time in TDR measurement in nano seconds. TDR probe 1 is buriable waveguide with three 20 cm long rods, 2 is connector waveguide with two 15 cm long rods. Mix is the mixture equation defined as Eq. (4-4), Topp is Topp et al. (1980) equation (Eq. (4-1)). $\beta$ is empirical constant summarizing the geometry of the medium in mixture equation.
Fig. 4-1. Observed vs. estimated water content $\theta_v$, based on measured $K_s$, of stony samples with Topp et al. (1980) equation (Topp’s Eq.; Eq. (4-1)) and mixture equation (Mix Eq.; Eq. (4-4)). Full and dashed lines represent the regression by Topp’s Eq. and Mixture Eq., respectively, for (a) soil layer 1, (b) soil layer 2 and (c) soil layer 3. Number of samples are different because of non equal outlier data removal. Results are only shown for the capture window 20 ns and the probe type 15-cm two-rod (see table 4-2).
Calibrating TDR at low $\theta_v$ was attempted by Skierucha et al. (2008). They stated that the convolution effects cause an increase in reflection time and consequently an increase in measured values of permittivity. This is especially evident for low dielectric permittivity and the corresponding low $\theta_v$ values. For soil water content values exceeding 0.2 m$^3$ m$^{-3}$, the gravimetrically measured data were close to Topp’s standard calibration equation. As an alternative to Topp’s equation, new regression curves that relate $K_a$ to gravimetrically measured $\theta_v$ were established (Fig. 4-2). Low $\theta_v$, which is expected in our profiles in an arid area with only a few rain and flooding events per year, can be effectively obtained by these new equations (Table 4-3).

A comparison between $\theta_v$ estimated by the new equations and the Topp et al. (1980) equation is presented in Table 4-4. In all soil types the RMSE of the new equation is lower than Topp’s. Measured $K_a$ in capture windows of 20 ns also resulted in a better estimation of $\theta_v$ with a lower RMSE than the 10 ns window. The capture windows 40 ns showed erroneous data in some measurements; therefore they are not shown here. Estimation of the minimum $\theta_v$ by the new equation is closer to gravimetrically measured $\theta_v$ than that by the Topp et al. (1980) equation in all soil types. The RMSE for the new equation is equal to 0.002 m$^3$ m$^{-3}$ for all soil layers.

Table 4-3. Layer specific calibration equations for estimation of $\theta_v$, in stone-free samples.

<table>
<thead>
<tr>
<th>Soil layers</th>
<th>$y=a+bx+cx^2+dx^3$</th>
<th>$R^2$</th>
<th>RMSE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-2.35E-04 8.08E-03 1.05E-01 2.85E-01</td>
<td>0.992</td>
<td>0.002</td>
</tr>
<tr>
<td>2</td>
<td>-1.69E-04 6.10E-03 8.27E-02 2.24E-01</td>
<td>0.991</td>
<td>0.002</td>
</tr>
<tr>
<td>3</td>
<td>-1.95E-04 6.81E-03 9.04E-02 2.47E-01</td>
<td>0.997</td>
<td>0.002</td>
</tr>
</tbody>
</table>

$Y$ is estimated $\theta_v$ and $x$ is measured $K_a$ by the TDR.
Fig. 4-2. Measured $\theta_v$ of non-stony samples (after sieving) relationship to the dielectric permittivity ($K_a$) for the capture window 20 ns. Lines are regressions of layers 1 to 3 from A to C, respectively. RMSE values are based on two pairs of data calculated by Topp Eq. and new Eq. for each individual layer.
Table 4-4. Statistics of SWC in sieved soil samples.

<table>
<thead>
<tr>
<th>Soil types</th>
<th>CW, ns</th>
<th>$\theta_v$, m$^3$m$^{-3}$</th>
<th>Min</th>
<th>Max</th>
<th>Mean</th>
<th>St.de.</th>
<th>RMSE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Observed</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Topp et al., 1980</td>
<td>10</td>
<td>4.40</td>
<td>37.50</td>
<td>16.53</td>
<td>10.43</td>
<td>0.005</td>
<td></td>
</tr>
<tr>
<td></td>
<td>20</td>
<td>5.00</td>
<td>36.52</td>
<td>17.05</td>
<td>10.22</td>
<td>0.005</td>
<td></td>
</tr>
<tr>
<td>New equation</td>
<td>10</td>
<td>0.96</td>
<td>36.77</td>
<td>16.10</td>
<td>10.64</td>
<td>0.004</td>
<td></td>
</tr>
<tr>
<td></td>
<td>20</td>
<td>1.00</td>
<td>36.56</td>
<td>16.05</td>
<td>10.56</td>
<td>0.002</td>
<td></td>
</tr>
<tr>
<td>Observed</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Topp et al., 1980</td>
<td>10</td>
<td>2.22</td>
<td>37.74</td>
<td>12.90</td>
<td>11.06</td>
<td>0.007</td>
<td></td>
</tr>
<tr>
<td></td>
<td>20</td>
<td>2.01</td>
<td>36.93</td>
<td>12.85</td>
<td>10.92</td>
<td>0.009</td>
<td></td>
</tr>
<tr>
<td>New equation</td>
<td>10</td>
<td>1.90</td>
<td>36.10</td>
<td>14.88</td>
<td>10.47</td>
<td>0.002</td>
<td></td>
</tr>
<tr>
<td></td>
<td>20</td>
<td>1.40</td>
<td>36.70</td>
<td>15.48</td>
<td>10.46</td>
<td>0.002</td>
<td></td>
</tr>
<tr>
<td>Observed</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Topp et al., 1980</td>
<td>10</td>
<td>4.10</td>
<td>32.85</td>
<td>12.21</td>
<td>9.30</td>
<td>0.009</td>
<td></td>
</tr>
<tr>
<td></td>
<td>20</td>
<td>4.70</td>
<td>33.12</td>
<td>12.81</td>
<td>9.19</td>
<td>0.009</td>
<td></td>
</tr>
<tr>
<td>New equation</td>
<td>10</td>
<td>1.05</td>
<td>32.54</td>
<td>11.72</td>
<td>9.88</td>
<td>0.008</td>
<td></td>
</tr>
<tr>
<td></td>
<td>20</td>
<td>0.96</td>
<td>33.20</td>
<td>12.22</td>
<td>10.17</td>
<td>0.002</td>
<td></td>
</tr>
</tbody>
</table>

CW is capture windows of time domain reflection in nanoseconds (ns), $\theta_v$ is volumetric soil water content.

Validation of the model performance of the new equations prepared for stone-free soils is performed by separating the whole data set (soil samples of the three layers) to two groups one for calibration and the other for validation. Several pairs were generated by combination of layers (Table 4-5), for instance, in the first pair, the soil samples of layer 1 were selected as calibration set, and the samples of layers 2 and 3 as validation. As shown in Table 4-5, the best result is obtained by using the soil samples of layers 1 and 2 as calibration and of layer 3 as validation (RMSE=0.014). The new equations for stone-free samples as well as the Topp’s equation are shown in Table 4-5.
### Table 4-5. Calibration equations for the stone-free samples used in validation.

<table>
<thead>
<tr>
<th>Eq.</th>
<th>Soil layers sample set</th>
<th>y=a+bx+cx^2+dx^3</th>
<th>R^2</th>
<th>RMSE</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Cal.</td>
<td>Val.</td>
<td>a</td>
<td>b</td>
</tr>
<tr>
<td>Topp</td>
<td></td>
<td></td>
<td>-5.30E-02</td>
<td>2.92E-02</td>
</tr>
<tr>
<td>This study</td>
<td>1 2&amp;3</td>
<td></td>
<td>-2.85E-01</td>
<td>1.05E-01</td>
</tr>
<tr>
<td></td>
<td>2 1&amp;3</td>
<td></td>
<td>-2.24E-01</td>
<td>8.27E-02</td>
</tr>
<tr>
<td></td>
<td>3 1&amp;2</td>
<td></td>
<td>-2.47E-01</td>
<td>9.04E-02</td>
</tr>
<tr>
<td></td>
<td>2-3 1</td>
<td></td>
<td>-1.97E-01</td>
<td>7.23E-02</td>
</tr>
<tr>
<td></td>
<td>1-2 3</td>
<td></td>
<td>-1.73E-01</td>
<td>6.28E-02</td>
</tr>
<tr>
<td></td>
<td>1&amp;3 2</td>
<td></td>
<td>-2.85E-01</td>
<td>1.05E-01</td>
</tr>
</tbody>
</table>

Eq. is equations, Cal. is sample set used for calibration and val. is sample set used for validation. Y is estimated $\theta_v$ and x is measured $K_a$ by the TDR. Topp is Topp et al. (1980) equation and this study equations are the regression equations prepared based on soil layers sample sets.

#### 4.3.3 Effect of extension cables length

The effect of extension cables on the TDR-measured $\theta_v$ was minor ($\theta_v$ of less than 0.01 m³ m⁻³) for cables of 3 and 5 m length. Increasing the length caused a linear decrease in $K_a$ and $\theta_v$. In addition, deviations from the standard 2 m cable length became larger as water content increased (Fig. 4-3). A sets of linear regression equations were generated for the three soil layers to correct $K_a$ for cable length. Increase in the cable length caused a non-linear decrease in the $K_a$ reading, which fits to a 3rd order polynomial equation. The coefficients for finding the correction factor for $K_a$ readings of any cable length longer than 5 meters are presented in Table 4-6.
Fig. 4-3. Change in dielectric permittivity (Ka) due to the extended TDR cable length with $\theta_v$. Graphs are related to soil types 1 to 3 from a to c, respectively.
Table 4-6. Regression coefficients for correction of $K_a$ readings for any cable length longer than 5 meters.

<table>
<thead>
<tr>
<th>soil types</th>
<th>$y=a+bx+cx^2+dx^3$</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$a$</td>
<td>$b$</td>
</tr>
<tr>
<td>1</td>
<td>9.97E-1</td>
<td>1.31E-3</td>
</tr>
<tr>
<td>2</td>
<td>1.02</td>
<td>-6.81E-3</td>
</tr>
<tr>
<td>3</td>
<td>1.01</td>
<td>-3.41E-3</td>
</tr>
</tbody>
</table>

$y$ is correction factor for dielectric permittivity ($K_a$), $x$ is cable length (m); applicable for extension cables longer than 5 m in length.

These equations can be used for any cable length shorter than 30 m. For instance, $K_a$ reading with 27 m length extension cable for the soil type 1 results in a correction factor of 1.048, which should be multiplied by the TDR reading to find the corrected $K_a$.

Heimovaara (1993) reported that a 0.1 m long probe could be used with a cable shorter than 15 m without losing distinct reflections from the beginning and end of the probe. Comparable results are attained using probes longer than 0.2 m with cables up to 24 m long. Short probes (0.05 m) or longer cables cannot be used when measuring dry soils owing to indistinguishable reflections (Heimovaara, 1993). This occurs due to an increased rise time of the voltage pulse from the cable filtering of the high frequency components. If a TDR instrument with a higher bandwidth is used, shorter probes with longer cables than those described above can be applied (Noborio, 2001). Our results, which have shown a deviation in the measured $K_a$ with cable length longer than 10 m from those of standard 2 m cables do not completely correspond with the findings of Heimovaara (1993). Our probes had similar specifications to those of Heimovaara (1993), but we used a TRASE system, which is different from that of Heimovaara (1993) (1502B/C). The accuracy of the reading unit to produce and transmit the exact frequency, the accuracy of timing, and the quality of manufacturing the cables may have caused the difference. Therefore, calibration of the cables prior to the application seems mandatory. It is also suggested by Logsdon (2000) that for high surface area soils, TDR calibration should be carried out with the coaxial cable length and equipment combination that will be used on site, or calibration should be done on site.
4.4 Conclusions

Application of the TDR technique for measuring SWC is widely accepted for laboratory and experimental purposes. Although some results of using and calibrating TDR in stony soils can be found in the literature, the applicability of TDR in soil with naturally distributed stone fragments has been rarely investigated. In the present study, a multilayer profile was sampled and the natural distribution of soils was created in the laboratory. The main objective was to assess the role of stones on measured and simulated water contents, as well as on the effects of extension cable on the reliability of the results. Based on the main findings of this study, the following conclusions may be drawn:

In order to compare properly measured and simulated water content of stony soils, in situ TDR-based water content measurements on the fine fraction of the bulk soils should correct for stoniness. The latter could be accounted for in the in situ TDR measurements by converting the bulk dielectric permittivity to the bulk water content by adopting an approach explicitly considering the role of volumetric fraction of stone and their dielectric properties. This approach can be applied for converting the in situ measured dielectric permittivity to $\theta_v$ of the bulk soil based on the determined stoniness. The 15 cm two-rod (connector) probe type and capture window 20 ns resulted in better performance than the 20 cm three-rod (buriable) probe type and capture window 10 ns (with final RMSE of 0.02 to 0.04 m$^3$ m$^{-3}$).

The calibration equations prepared for the fine part (stone-free) soils might lead to perfect match between observed and simulated water content for local soils (in our case, the RMSE of 0.002 m$^3$ m$^{-3}$). They are particularly worthy at low water content (e.g. below 0.05 m$^3$ m$^{-3}$).

Noticeable effects of cable length on measured $K_d$ were found for lengths exceeding 10 m. Accurate $K_d$ values (corrected for extension cable length) can be obtained if the suggested correction factors reported in this study are applied.

As a final conclusion related to practical implication of the TDR method in layered soils with large size stone fractions (larger than the TDR probe rod spacing) it is vital to use the connector probe type which is designed for working in undisturbed soils. This is due to the need for strong metal rods, suitable for insertion by hammer, for minimizing lateral deformations of probe
rods during the insertion. Apart from the practical difficulties for slotting the probes, which may cause the probes’ rods not to be inserted in a parallel way, an inadequate contact between the rods and the soil particles will lead to insufficient media for transmission. Owing to these difficulties, if one must decide to work with the disturbed soil, the buriable probe type can be employed. Then, it would be recommended to excavate the desired horizontal hole with enough space lengthwise in the undisturbed soil profile, filling the hole with soil of similar texture, and inserting the probes inside the hole. Thereafter, the hole, including the probe, might be covered with insulating material inside the hole in order to prevent the atmospheric moisture to interfere with the soil water content measurement. In this circumstance, the calibration equations dressed out for the local (stone-free) soil can be used to find the accurate water content of the fine part and then convert it to the water content of bulk soil (including the stone fraction) by applying the approach presented in this study.
Chapter 5. Evaluation of recharge by a modelling approach based on the measured soil water content in a multi layered vadose zone
5.1 Introduction

Floodwater harvesting has become an increasingly important technique to improve water security and caused a renewed interest in research and implementation. This was addressed in the extensive review of Boers and Benasher (1982) that revealed an awareness of increasing the need for recognition of its potential. Despite the long and successful history of these systems, little is known about their function and effect on the hydrological processes in dry areas (Ouessar et al., 2009). Clogging of the infiltrating surface, which results in reduced infiltration rates, are the bane of all artificial recharge systems resulting in unknown number of deficiencies in recharge processes (Bouwer, 2000), which necessitates in-depth and site specific studies.

According to the diverse objectives and methods of implementing artificial recharge of groundwater (ARG) systems, various factors need to be considered when choosing a method for quantifying recharge. Therefore, the rate of aquifer recharge is one of the most difficult items to measure in groundwater (GW) resources evaluation (Sophocleous, 1991).

Scanlon et al. (2002) categorized the techniques used in quantifying recharge in three main groups: unsaturated zone, saturated zone and surface water (summarized in Fig. 1-6). Subdivisions of the techniques are somewhat arbitrary. Allison et al. (1994) subdivided unsaturated (vadose) zone physical methods to direct (i.e. lysimetry) and indirect ones. In turn, indirect methods involve Zero Flux Plane, soil-water budget, and water flux methods based on solving the Buckingham-Darcy’s or Richards’ equation. Tracing techniques have been excluded from physical methods in their classification.

The soil-water budget equation can be written as (Evett et al., 2012):

$$ R = P + I + F - \Delta S - ET_a + \varepsilon_R $$

where $R$ is deep recharge, $P$ is precipitation, $I$ is irrigation, $F$ is flooding which means runoff minus runon, $\Delta S$ is the change in soil-water storage over the considered time period, $ET_a$ is actual $ET$ and $\varepsilon_R$ is a function of the errors in determination of $\Delta S$, $I$, $P$, $R$, and $F$. The soil-water budget technique has been used in many areas mostly in temperate regions. Some authors have criticized this method when used in arid regions, particularly where the magnitude of
recharge is smaller than \( ET \) (Gee and Hillel, 1988; Hendrickx and Walker, 1997). This is a valid criticism when the model fails to represent soil and crop conditions, or uses too long (e.g. monthly) a time interval. The longer the time interval, the greater the numerical similarity between infiltration and actual evapotranspiration, and hence the smaller the difference (deep drainage) between the two, and consequently the greater the possible error (Eilers et al., 2007). Therefore, it is suggested to decrease the time intervals at most to less than 10 days (Howard and Lloyd, 1979). Some reasonable results of using soil-water budget methods in various climates including semi-arid and arid regions have been reported for estimating \( ET \) (Evett et al., 2012; Bellot and Chirino, 2013; Ponti et al., 2013) and assessing recharge (Kendy et al., 2003; Eilers et al., 2007; Jin et al., 2007; Chen et al., 2008; Raes et al., 2009; Ma et al., 2013).

In SWB method (and in Zero Flux Plane method as well), soil water movement is inferred by measuring the changes in water content of the soil profile by gravimetric sampling or automatic devices. These methods have not been proven satisfactory in low flow conditions, as there is often insufficient resolution to detect movement of small quantities of water. Therefore, other methods, based on hydraulic conductivity, potential gradients and directly calculated water fluxes for unsaturated flow were developed (Enfield et al., 1973).

The Buckingham-Darcy law can be used under the steady flow condition where water contents and fluxes change with depth but do not vary as a function of time (Radcliffe and Šimůnek, 2010). It has been employed in arid and semiarid conditions for recharge estimation (Enfield et al., 1973; Stephens and Knowlton, 1986), or for assessing the exchange flow between surface water reservoir and GW (LaBaugh et al., 1995; Rosenberry et al., 2008). The method requires measurements or estimates of the vertical total head gradient and the unsaturated hydraulic conductivity at the ambient soil water content following the Buckingham-Darcy equation (Buckingham, 1907):

\[
R = -\frac{K(\theta)d(h + z)}{dz} = -K(\theta)\left(\frac{dh}{dz} + 1\right)
\]  

(5-2)

where \( R \) is recharge, \( K(\theta) \) is the hydraulic conductivity at the volumetric porous media water content, \( \theta \) is volumetric soil water content, \( h \) is the matric pressure head and \( z \) is vertical distance.
The matric pressure head gradient is often nearly zero, and water movement is essentially gravity-driven for thick unsaturated zones, below the zone of fluctuations related to climate, and in uniform or thickly layered porous media (Hillel, 1998). Under these conditions, little error results in assuming that the total head gradient is equal to 1 (Scanlon et al., 2002). In this case, zero gradient of the pressure head (free drainage) at the bottom of the soil column is taken (van Dam, 2000) and recharge is assumed to be equal to the hydraulic conductivity at the in situ soil water content:

\[
R = -K(\theta) \left( \frac{dh}{dz} + 1 \right) = -K(\theta) (0 + 1) = -K(\theta)
\]

The Richards equation is used under transient flow conditions where at least one of the variables characterizing flow changes as a function of time. The Richards equation is derived from a water conservation equation and the Buckingham-Darcy equation (Radcliffe and Šimůnek, 2010), and can be written for one-dimensional vertical flow as:

\[
\frac{\partial \theta(h)}{\partial t} = \frac{\partial}{\partial z} \left( K(h) \frac{\partial h}{\partial z} \right) + \frac{\partial K(h)}{\partial z} - S(h)
\]

where \( t \) is time, and \( S(h) \) is a sink function (a negative \( S(h) \) represents a source), usually accounting for the root water uptake (transpiration). Process-based models using numerical solutions to the Richards equation have been extensively used to simulate deep flux (Fayer et al., 1996; Bethune and Wang, 2004; Jhorar et al., 2004; Vrugt et al., 2004; Skaggs et al., 2006; Smerdon et al., 2008; Jiménez-Martínez et al., 2009; Selle et al., 2011).

Application of a Richards’ equation-based model requires data on water content as a function of pressure head (the water-retention curve), and relative hydraulic conductivity as a function of pressure head (the unsaturated hydraulic conductivity curve).

Direct measurement of parameters required for these models is very difficult in the field. Therefore, model parameters are usually calibrated or estimated from other observations such as soil water content and soil texture. Field and laboratory measurements often do not produce information content sufficient to define the parameters adequately (Vrugt et al., 2004). Furthermore, complex, parameter-rich models can yield dramatically different predictions for conditions outside the range of the calibration data (Kirchner, 2006).
Therefore, a robust simulation cannot be achieved without improved calibration in local natural settings (Yao et al., 2014).

In chapter 3, recharge was assessed based on a combination of the two saturated zone methods; GW fluctuation and water budget. This chapter focuses on application of other methods of assessing the efficiency of FWS systems for GW recharge, based on soil-water monitoring in the unsaturated zone above the water table, thus providing a new independent window to evaluate the recharge process. To that end, the soil-water budget and water flux methods were applied, after parameter optimization and calibration of the model. The optimized parameters will be employed further in studies with up-scaling objectives.

An integrated field and laboratory investigation was carried out to acquire reliable data on the soil hydraulic parameters of the aquifer layers and the spatial distribution of the layers in a location inside a floodwater spreading (FWS) basin of the study site (Gareh Bygone Plain). In parallel, time domain reflectometry (TDR) probes were installed along the profile of an experimental well to generate a time series of soil water content in a period of three years. The objectives of this study were to employ the collected soil water content data set together with the other input data for: a) assessing the impact of long term spreading of the turbid floodwater on the clogging of soil surface, which might led to decrease in water infiltration to the aquifer; b) solving a water budget equation to assess the recharge due to rainfall and ponded floodwater; c) to use the soil water content data set as a source of observations to optimize the hydraulic parameters and calibrating the HYDRUS 1D (H1D) model (Šimůnek et al., 2008), and d) to simulate the recharge by the calibrated H1D and determining the error of simulation by comparing with the SWB results. The ultimate and overall objective is evaluating the recharge by the FWS system at the study site for the studied flooding event.

5.2 Materials and methods

5.2.1 Site description

The artificial recharge of groundwater (ARG) project in the Gareh Bygone Plain contains three main FWS systems, namely Bisheh Zard (BZ), Rahim Abad (RA), and Tchah Qootch (TQ), as well as some minor and separate parts. These are named after their nearby villages and ephemeral rivers. The BZ and
RA compartments consist of several divisions (called BZ\textsubscript{1} to BZ\textsubscript{4} and RA\textsubscript{1} to RA\textsubscript{3}), while TQ has a single division. Each division is an independently functioning FWS system. They comprise of 3-6 polygons (the basins), which are surrounded by earth dikes, built along the contour lines and bordered by trail walls (Fig. 5-1). The very first upstream basin acts mostly as a sedimentation basin for coarse grains. Floodwater covers all basins only during major flooding events, whereas minor events result in partial coverage only. The ARG project under study was initiated in January 1983 and covered 1365 ha in March 1988. The system was expanded to 2033 ha in the year 2000. Nearly 105 ha are under tree plantations, mainly with *Eucalyptus camaldulensis* Dehnh. the predominant tree species in all of the basins, being mainly planted from 1983 to 1986. Some other eucalyptus and acacia species were also planted and persist in the FWS systems, but cover less than 10\% of the plantations. The majority of these trees were planted by the upslope toe of the channels banks, along the waterline of the diversion canals, and by the inside toe of the trail walls. Tree plantations were later extended in some basins resulting in a dense forested area. More detailed descriptions of the FWS systems in the study site are presented in chapter 2 of this thesis, and can also be found in Kowsar (1991) and Hashemi et al. (2012). BZ\textsubscript{1} FWS system was selected as the main study area being located in the central part of the FWS scheme. It was the first system constructed in January and February 1983 and it has been under operation longer than the other systems. The second basin from the inlet of the BZ\textsubscript{1} FWS system was selected as experimental site in this study (Fig. 5-1), in which among others, three experimental wells were installed until a depth of ~30 m (see section 6.2.2.). This basin receives floodwater in every event while its sediment load is somewhat similar to that of the other basins. The depth of the deposited sediment due to the functioning of FWS from 1983 to 2003 was measured and mapped in a separate study (Fig. 5-2) (Esmaeili-Vardanjani et al., 2013).

They showed that while the first basin is characterized as having high load of siltation (10-30 cm in depth), the second basin shows sedimentation depth as low as 5-8 cm. The surface area of BZ\textsubscript{1} FWS system and its second basin are 189 and 29 ha, respectively.
Fig. 5-1. (A) Floodwater spreading project in March 1988, Gareh Bygone Plain; ARG is artificial recharge of groundwater, FWS is floodwater spreading, BZ is Bisheh Zard, RA is Rahim Abad and TQ is Tchah Qootch. (B) Location of BZ₁ FWS system and land cover of the 2nd basin; TP is tree plantations, PI is pastures inside the floodwater spreading system and BS is bare soils. EW1 to EW3 are experimental wells (1-3).
5.2.1.1 Vegetation
Typical land cover of the study site includes tree plantations, pastures and bare surfaces (Fig. 5-1b). This diversity was another reason for selecting the second basin as the study area. Tree plantations consist of *Eucalyptus camaldulensis* Dehnh. and *Acacia victoriae* Benth. Rangeland plants naturally appeared in pasture areas of all FWS systems. Of the 12 range plants determined in the second basin of the BZ1 system, the most dominant species were *Acantholimon* sp. Boiss., *Aegilop scerasa* Boiss., *Artemisia sieberi* Besser., *Carex stenophylla* L., and *Helianthemum lippii* (L.) Pers. (Mesbah and Kowsar, 2010).

5.2.1.2 Soils
Soils of the study basin, which is formed on a debris cone and the alluvial fan developed by the BZ Ephemeral River, is covered with a layer of drifting fine sand ranging in thickness from a few mm to several cm. A structureless, coarse sandy loam with average sand, silt and clay contents of 73.2, 14.5 and 12.2%, respectively forms the A horizon, 10-20 cm thick. The stony C horizon lies directly under the A horizon. It has been classified as coarse-loamy skeletal,
carbonatic (hyper) thermic, Typic Hapl calcids (Kowsar, 1991; Soil-Survey-Staff, 2010).

5.2.1.3 Weather data
Details of the climatic data collection within the study area have been described in chapters 2 and 3 of this thesis.

5.2.2 Experimental wells
Three experimental wells (EW) were hand dug in locations with different land covers, i.e., tree plantations (TP), pastures inside FWS system (PI) and bare soils (BS) (Fig. 5-1b). The objectives of installing the three wells were: a) to identify profile layers of the basin, b) to determine hydraulic properties of the layers, c) to equip one of the wells with TDR probes. The wells were dug until the water table level (28.8 to 31.6 m) and had a diameter of 120 cm for easy access into the wells. General specifications of the wells are presented in Table 5-1. The difference in depth of the wells is attributed to the difference in the months of excavation (from May to August) as the GW level falls during this period.

The FWS systems used as the research site was constructed in 1983. A top layer of silt loam, 10 to 20 cm thick, had been formed over the original soil surface due to the deposition of the suspended load during 30 years of being used as sedimentation basin in which the recharge also takes place simultaneously. The observation well through which the collected data analyzed in this study is located at 50 m distance from the upstream level-silled channel and 60 m from the downstream one.

5.2.3 Water ponding measurement
The time at which the area surrounding the experimental wells was pounded by floodwater and the height of ponding water were recorded. The height was measured by an in-built ruler in each one hour interval (Photo 5-1).
Table 5-1. General description of experimental wells (EW).

<table>
<thead>
<tr>
<th>No.</th>
<th>Location, m (UTM)</th>
<th>elevation, m.a.s.l.</th>
<th>Surrounding land cover(1)</th>
<th>Depth, m</th>
<th>WT elevation, m.a.s.l.</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>786919 3168378</td>
<td>1157.53</td>
<td>TP (A. victoriae)</td>
<td>28.80</td>
<td>1128.73</td>
</tr>
<tr>
<td>2</td>
<td>786543 3168617</td>
<td>1157.30</td>
<td>BS</td>
<td>30.60</td>
<td>1126.70</td>
</tr>
<tr>
<td>3</td>
<td>786291 3168873</td>
<td>1157.00</td>
<td>PI (different species)</td>
<td>31.60</td>
<td>1125.40</td>
</tr>
</tbody>
</table>

1-TP is tree plantations, BS is bare soils and PI is pastures inside floodwater spreading system.

5.2.4 Determination of hydraulic properties

5.2.4.1 Field saturated hydraulic conductivity

The field saturated hydraulic conductivity $K_{fs}$ was measured by the double ring method (Reynolds et al., 2002) using two rings with inside diameter of 0.30 and 0.59 m, respectively in well number 1 during its construction.

The $K_{fs}$ was measured for each 30 cm increment from the soil surface down to 3 m, and for each 100 cm increment from 3 m down to 30 m (Photo 5-3). The rings, 0.40 and 0.25 m high, were driven into the soil to a depth of 0.03-0.05 m, respectively. The soil surface in the inner ring was covered with a plastic sheet to prevent soil disturbance. Water was added cautiously to both the inner and outer ring until a 20 cm water depth was reached; the plastic sheet was then removed from under the inner ring. The falling water depth in the inner ring was recorded using a ruler after 1, 3, 5, 10, 15, 25, 35, 55 and 75 minutes, i.e., for at least 75 minutes or until the infiltration rates in the successive time intervals remained practically constant. The measurement was interrupted when the water level inside the ring reached a height of 5 cm, and the water level in both rings was adjusted to 20 cm head before the reading was resumed. In all cases, measurements were repeated if the results showed improper magnitude of $K_{fs}$. The main problem with this measurement was the difficulty to embed the rings into the well’s floor due to its stoniness.

The first measurement was done on the soil surface at the location of the well. Then after removing the first 30 cm depth of soil the second measurement was performed on the well’s floor. This was repeated for the depth intervals as described above until the final depth of 28.8 m where the water table was reached (Photo 5-3). Vertical infiltration was calculated using the two-parameter Philip equation (Bouwer, 1986)
\[ I_t = S_t t^{0.5} + At \]  \hspace{1cm} (5-5)

where \( I_t \) is the cumulative infiltration, \( t \) is the cumulative time, \( S_t \) is the sorptivity and \( A \) is a factor related to the soil’s permeability. The least squares method was applied to estimate the unknown parameters \( A \) and \( S_t \). Infiltration rate \( i \), i.e. the derivative of Eq. (5-6), for any given time was then calculated from (Philip, 1957):

\[ i = 0.5 S_t t^{-0.5} + A \]  \hspace{1cm} (5-6)

The \( A \) term in Eq. (5-5) and (5-6) can be taken as \( K_{fs} \) for long-term infiltration, and as \( \frac{1}{2} K_{fs} \) for short term infiltration (Bouwer, 1986 and the references therein). We then used Wooding’s equation (Radcliffe and Šimůnek, 2010):

\[ \lambda_c = \frac{4 \lambda_c}{\pi r} \]  \hspace{1cm} (5-7)

where \( \lambda_c \) is the steady state infiltration rate, \( r \) is the radius of the ring, and \( \lambda_c \) is microscopic capillary length, which represents the effect of capillary. The larger the \( \lambda_c \), the higher the dominance of the capillary force relative to gravity force and vice versa. Elrick and Reynolds (1992) have presented tabular guidelines to assign a value to \( \lambda_c \) for different soil texture/structure categories. According to the known physical characteristic of the layers in this study, the desired \( \lambda_c \) was used for calculating \( K_{fs} \).

The water used in infiltration measurements should have similar properties as those of the actual operating system. As floodwater in FWS systems of the study site contains suspended solids, the reduction in infiltration rate due to deposition of sediments on the soil surface was taken into account, by the exponential decay function of Bernardes (1967) for clogging during infiltration (Bouwer, 1986):

\[ i = i_0 - \alpha I_t \]  \hspace{1cm} (5-8)

where \( i \) is infiltration rate at time \( t \), \( i_0 \) is initial infiltration rate, \( \alpha \) is coefficient describing clogging properties of system, \( C_s \) is concentration of suspended material in infiltrating water that will be retained on the soil surface and \( I_t \) is cumulative infiltration at the time \( t \). Reported values for \( \alpha \) span a range of 0.1 to 5.5 L g\(^{-1}\) day\(^{-1}\) for floodwaters with relatively high sediment contents of 0.14 to 2 g L\(^{-1}\) (Bouwer, 1986). As the effect of sediment due to turbid floodwater affects only infiltration at the soil surface, Eq. (5-8) was only applied for the first layer. Based on periodical measurements of the sediment concentration of...
floodwater, which spreads on the surface of FWS systems in the study site, a mean value of 1.2 g L⁻¹ was assigned as $C_s$.

5.2.4.2  **Bulk density**

Upon completion of each infiltration measurement in well number 1, bulk density was measured inside the inner ring. Because of stoniness of many layers, the core method was not applicable and the excavation method modified based on Grossman and Reinsch (2002) was applied for all of the layers. At each depth, a cavity with a rough volume of 2500 cm³ was dug and insulated with a plastic liner. The excavated material was collected, oven dried and weighed before and after drying. The exact volume of the cavity was measured by adding a measured volume of water into it. Bulk density was then calculated by dividing the dry weight of the bulk materials (soil and stone) by the cavity volume.

5.2.4.3  **Particle size analysis**

Bulk material of every depth was collected and texture of the material <2 mm was determined in the laboratory using the sieving and hydrometer method in order to better distinguish the layers. Two kg of each sample was used to particle size analysis. This was carried out using the hydrometer method combined with sieve analysis to characterize the range of particle diameter from 0.002 up to 2 mm (Gee and Or, 2002). Stone fraction (>2mm) was determined by the sieving method.

5.2.4.4  **Identifying and correlating the wells’ profile layers**

The data set of the hydraulic properties, soil texture and stoniness, of the different depths was used as the basis for differentiation between the layers of the profile of the experimental well number 1. The whole dataset is presented in Appendix 5. Then, a visual description of the profile layers in the three wells was made by descending inside the wells. Alphabetic codes (A to G) were assigned to seven representative layer (RL). Then the distribution (thickness and location) of every code over the wells’ profile was determined to prepare the log of each well. The logs of the three wells were then correlated to match the congruent layers and to find the horizontal distribution of the aquifer layers in the study area (Photo 5-2).

The specified layers in the three wells were analyzed in order to recognize the correspondence between the RLs and their distribution over the wells’ profile.
Hence, a correlation between the horizontal layers was made to produce a 3D imagine (2D vertical distribution of the layers in different wells).

5.2.4.5 Water retention data
Soil water content corresponding to potentials of 0, -100, -200, -330, -1000, -3000, -5000, -15000 cm were determined for the <2 mm fraction for the seven RLs by the sand box apparatus and pressure plate. Due to the presence of stone fractions, water contents were adjusted employing the Bouwer and Rice (1984) equation (Eq. (3-10)) (Photo 5-2). It should be noted that the samples were not repacked and the analysis was done with separated sub samples.

5.2.5 Long term data collection

5.2.5.1 Soil water content measurement
The Trase System 6050X1 (Soilmoisture Equipment Corp., USA) was used for soil water content measurement. It is a TDR-based system which was first calibrated for the main soil types of the study area (described in chapter 4). The main steps in installing the system were as described below.

5.2.5.2 Insulating the well wall
The wall of the experimental well was insulated to avoid lateral flow of soil water into the well over its entire depth. To do so, 20 cm high concrete tiles with a diameter of 110 cm, slightly smaller than the well diameter, were fabricated and placed inside the well. To seal the rings against the well wall, the 5 cm inner space between both was filled with cement mortar. A circular lid made of stainless steel was installed over the well to prevent flow of ponded floodwater into it (Photo 5-4).

5.2.5.3 Installing the TDR probes
To install the three-rod TDR probes, which were manufactured and tested in this study (see chapter 4), small openings were made in the concrete tiles to have access to the vadose zone behind. In order to prevent probe damage, and also to ensure good soil contact, a mold of TDR probe was fabricated, which was first inserted into the soil. This facilitated easy and tight insertion of the TDR probes. Probes were installed approximately at the same depths at which infiltration measurements were made. From top to well bottom, probes were installed at 0.1, 0.4, 0.6, 0.8, 1.1, 1.4, 1.7, 2, 2.3, 2.6, 2.9, 4, 5, 6, 7, 8, 9, 10, 12, 14, 16, 18, 20, 24, 26, 28 m depths (Photo 5-5). They were connected to the desired extension cables (if needed) and each cable was driven to the top
of the well by its unique guidance polyethylene pipe. The pipes were organized and numbered in order to have easy access to the each probe’s depth (Photo 5-6and Photo 5-7).

When stone fragments prevented probe insertion, a small hole was excavated in the soil profile with approximate dimensions of 10 by 25 cm (diameter and depth respectively) and refilled with sandy loam soil. This soil texture showed better correlation when calibrating the TDR probe (see chapter 4). Moderate compaction was applied to assure good contact with the original soil, and at the same time, to avoid over compaction.

5.2.5.4 SWC measurements
The data collection spanned from 21 August 2010 to 1 December 2013. Measurements were taken manually by connecting each probe to the main device and reading the data. Both the measured dielectric permittivity ($K_a$) and the estimated volumetric SWC ($m^3 m^{-3}$) were recorded and stored in the TRASE device. Measurements were performed at 6-8 day intervals before rainfall and/or flood occurrence, twice daily for 30 days during and after such events, and daily thereafter until returning to the prior SWCs of previous events. There were some missing data at some pre-planned time intervals owing to the logistical shortcomings. In addition, there were some missing data for some particular depths due to disconnection of the extension cable. The $K_a$ data were later on corrected to compensate for the effect of extension cables using the equations provided in Table 4-4. The corrected $K_a$ data were then converted to $\theta_v$ based on the proposed equations in this study (Table 4-3).

Missing or erroneous data (zero or near zero numbers) were corrected. In case of one missing data among a series of normal values, it was replaced by the average of the two adjacent before and next after that measuring day. In case of several successive recorded zero values, they were corrected based on the most correlated values of the adjacent layers. Corrected erroneous data consisted of about 1% of the whole recorded data sets. Some layers failed to produce any or some data point during the study period, and such layers were excluded from the analysis.

5.2.5.5 Climatic and flooding data collection
Depth and duration of rainfall, which were recorded at the Gareh Bygone Station, and the depth and duration of the ponded floodwater by the
experimental well that was measured at the experimental site, were organized and synchronized with the SWC measurement. The ponding duration and the measured $Kfs$ after correction for the sediment concentration (Eq. (5-8)), were used to calculate the depth of water infiltrating the soil. Since the corrected $Kfs$ was calculated as 24 cm day$^{-1}$ or 1 cm h$^{-1}$, the time of ponding on an hourly basis was considered as the depth of water (cm) reaching the soil profile. The depth of the recorded rainfall that coincides with the time of floodwater ponding was considered as zero, as the ponding water was assumed as an accumulation of rain and floodwater.

5.2.5.6 ET data

The actual daily $ET$ ($ETa$) was generated by a combination of the FAO Penman-Monteith (FAO P-M) and remote sensing based on the SEBS model. The details are presented in chapter 3 of this thesis. The starting date on which the $ETa$ was measured by the SEBS model was 21 December 2010 (chapter 2). The first important recorded flooding event started on 28 January 2011 and its influence on SWC of the profile continued until 23 July 2011. In order to generate an $ETa$ dataset for this period, image processing and $ETa$ mapping was performed for the flooding period using the same method described in chapter 2. The second important flooding event occurred on 7 August 2013 and its influence on SWC lasted until 2 November 2013. As the climate data synchronic to this period were not available, it was not possible to calculate the $ETa$; consequently, recharge was calculated only for the first flooding event.

5.2.6 Data quality assessment

Rainfall data, values of ponding floodwater, and hydraulic properties of soil samples were supposed to be the directly measured values which can be used in the further analyses without the worries on uncertainty of data acquisition to be assessed.

The $ETa$ data by which the recharge assessment was carried out had been evaluated by means of cross validation in chapter 2 (section 2.3.4).

The long term measured data set of SWC was assessed in this chapter by means of checking the temporal distribution, its response to the flooding and rainfall events, and a statistical method, empirical quantile-quantile plots. The latter statistical method is explained in detail in chapter 3 (section 3.2.2).
5.2.7 Assessment of recharge

5.2.7.1 Soil-water budget approach
SWC measured per depth ($\theta_v \times$ thickness of each layer) was summed up for all of the layers as total storage ($S$) in each day and the change in SWC ($\Delta S$) was calculated by subtracting the $S$ of the successive days. The $\Delta S$ of the layers were then used to calculate the soil-water storage within the vadose zone. Records of the ponded floodwater on the soil surface close to the well during the flooding events were considered as remainder of runoff and run on in the system.

Then Eq. (5-1) was solved for recharge $R$ on a daily basis. The period between the time at which the SWC of any layer started to increase due to the flooding event and the time of returning back to its previous normal level (SWC of all layers before the event) was set as the calculation period. Summation of the daily recharge data during the calculation period was considered as the net recharge due to that particular event.

5.2.7.2 Water flux modelling with Hydrus 1D
To numerically solve Eq. (5-4) and calculate the transient flux at a depth of 4.0 m, hence recharge, Hydrus-1D software package (H1D) version 4.16 (Šimůnek et al., 2013) was used.

A 30 m profile was discretized with 301 nodes and a resolution of 10 cm. Seven RLs were defined and their distribution throughout the profile was introduced. The initial condition of the profile was defined based on the measured initial SWC of the layers and a gradual change of SWC was set between the layers. Rainfall, ponded floodwater and $ET$ were added as the variable boundary conditions (BC).

The top BC was set as atmospheric BC with the surface layer as the BC closest to the condition of this study. The bottom BC was set as free drainage, which corresponds to a zero matric head gradient. Such a situation often occurs in field studies of water flow and drainage in the vadose zone e.g., Hillel (1998). This lower BC is most appropriate for situations where the water table lies far below the domain of interest. Due to the bare soil surface (no planted or natural grasses or shrubs) and the absence of roots in top layers of the profile at the study site, root water uptake was neglected. The acacia trees surrounding the experimental well have a deep rooting system penetrating the aquifer;
therefore, the main water consumption is by the direct uptake from the water table. This process has been proven by ET estimation, which has been reported in the third chapter of this thesis.

Solving Eq. (5-4), and more particularly its matric head form, in which \( \frac{\partial \theta(h)}{\partial t} \) is replaced by \( C(h) \frac{\partial h}{\partial t} \) according to the chain rule (Jury and Horton, 2004), where \( C(h) = \frac{d\theta}{dh} \) is the differential soil-water capacity or the slope of the water retention curve, requires the water retention and hydraulic conductivity curves of the different RLs. The measured values of \( K_f \) and \( \rho_b \) in the whole data set were rearranged to find out which data were correspondent to which layer code. Then the corresponding data to each layer code were averaged and assigned to that code. Some outlier data, which belonged to the margins of adjacent layers, were excluded from the averages.

H1D allows users to select six types of models for the soil hydraulic parameters (Šimůnek et al. (2013), which were all tested:

a) the van Genuchten-Mualem model (VGM) (van Genuchten, 1980) in which van Genuchten (1980) used the statistical pore-size distribution model of (Mualem, 1976) to obtain a predictive equation for the unsaturated hydraulic conductivity function in terms of soil-water retention parameters:

\[
\theta(h) = \begin{cases} 
\theta_r + \frac{\theta_s - \theta_r}{[1 + |\alpha h|^n]^m} & h < 0 \\
\theta_s & h \geq 0 
\end{cases} 
\]  

\( (5-9) \)

\[
K(h) = K_s S_e^l [1 - (1 - S_e^{1/n})^m]^2 
\]  

\( (5-10) \)

\[
m = 1 - 1/n \quad , n>1 
\]  

(5-11)

where \( \theta(h) \) is volumetric soil-water content at pressure head \( h \), \( \theta_r \) and \( \theta_s \) denote the residual and saturated water contents, \( K_s \) is the saturated hydraulic conductivity, \( \alpha \) is the inverse of the air-entry value (or bubbling pressure), \( n \) is a pore-size distribution index, and \( l \) is a pore-connectivity parameter assumed to be 0.5 as an average for many soils, \( S_e \) is effective water content, \( K(h) \) is unsaturated hydraulic conductivity at pressure head \( h \). The parameters \( \alpha, n \) and \( l \) in HYDRUS are considered to be empirical coefficients affecting the shape of the hydraulic functions.
b) the modified van Genuchten type equation (MVG) based on Vogel and Cislerova (1988) who modified the equations of van Genuchten (1980) to add flexibility in the description of the hydraulic properties near saturation:

\[
\theta(h) = \begin{cases} 
\theta_a + \frac{\theta_m - \theta_a}{[1 + (\alpha h)^n]^m} & h < h_s \\
\theta_s & h \geq h_s
\end{cases}
\]

\[
K(h) = \begin{cases} 
K_k K_r(h) & h \leq h_k \\
K_k + \frac{(h-h_k)(K_s - K_k)}{h_s - h_k} & h_k < h < h_s \\
K_s & h \geq h_s
\end{cases}
\]

\[
K_r = \frac{K_k}{S_{ek}} \left( \frac{F(\theta_r) - F(\theta)}{F(\theta_r) - F(\theta_{kr})} \right)^2
\]

\[
F(\theta) = \left[ 1 - \left( \frac{\theta - \theta_a}{\theta_m - \theta_a} \right)^{1/m} \right]^m
\]

\[
S_{ek} = \frac{\theta_k - \theta_r}{\theta_s - \theta_r}
\]

The above equations allow for a non-zero minimum capillary height, \( h_s \), by replacing the parameter \( \theta_s \) in van Genuchten's retention function by an extrapolated parameter \( \theta_m \) slightly larger than \( \theta_s \). While this change from \( \theta_s \) to \( \theta_m \) has little or no effect on the retention curve, the effect on the shape and value of the hydraulic conductivity function might be considerable, especially for fine-textured soils when \( n \) is relatively small (e.g., \( 1.0 < n < 1.3 \)). The \( S_{ek} \) is effective saturation at \( \theta_k \) (volumetric water content corresponding to \( K_k \)) and \( K_k \) is the measured value of the unsaturated soil hydraulic conductivity at \( \theta_k \). To increase the flexibility of the analytical expressions, the parameter \( \theta_r \) in the retention function was replaced by the extrapolated parameter \( \theta_a \leq \theta_r \). The approach maintains the physical meaning of \( \theta_s \) and \( \theta_r \) as measurable quantities. Eq. (5-14) assumes that the predicted hydraulic conductivity function is matched to a measured value of the hydraulic conductivity, \( K_k = K(\theta_k) \), at some water content, \( \theta_k \), less than or equal to the saturated water content, i.e., \( \theta_k \leq \theta_s \) and \( K_k \leq K \). Inspection of Eq. (5-12) through (5-16) shows that the hydraulic
characteristics contain nine unknown parameters: \( \theta_r, \theta_s, \theta_a, \theta_m, \alpha, n, K_s, K_k, \) and \( \theta_k \). When \( \theta_a = \theta_r, \theta_m = \theta_k = \theta_s \) and \( K_k = K_s \), the soil hydraulic functions of Vogel and Cislerova (1988) reduce to the original expressions of van Genuchten (1980).

c) the van Genuchten-Mualem model with an air-entry value of -2 cm (VGM-2) as suggested by Vogel and Cislerova (1988). It is recommended that this model be used for heavy textured soils (e.g., clays).

d) The equations of Brooks and Corey (1964) (BC). The soil-water retention, \( \theta(h) \), and hydraulic conductivity, \( K(h) \), functions according to Brooks and Corey (1964) are given by:

\[
S_e = \begin{cases} 
|ah|^{-n} & h < -1/\alpha \\
1 & h \geq -1/\alpha
\end{cases} \quad (5-17)
\]

\[
K = K_s S_e^{2/(n+2)} \quad (5-18)
\]

respectively, where \( S_e \) is effective saturation:

\[
S_e = \frac{\theta - \theta_r}{\theta_s - \theta_r} \quad (5-19)
\]

The terms \( \theta_r \) and \( \theta_s \), \( K_s \), \( \alpha \) and \( n \) are defined as in the VGM model and \( l \) is a pore-connectivity parameter assumed to be 2.0 in the original study of Brooks and Corey (1964).

e) The lognormal distribution model of Kosugi (1996) (LNK), who suggested the following lognormal distribution model for \( S_e \):

\[
S_e = \frac{\theta - \theta_r}{\theta_s - \theta_r} \left\{ \begin{array}{ll}
\frac{1}{2} \text{erfc} \left( \frac{\ln(h/\alpha)}{\sqrt{2n}} \right) & (h < 0) \\
1 & (h \geq 0)
\end{array} \right. \quad (5-20)
\]

where \( \text{erfc} \) denotes the complementary error function. Application of Mualem's pore-size distribution model (Mualem, 1976) now leads to the following hydraulic conductivity function:

\[
K = \begin{cases} 
K_s S_e \left\{ \frac{1}{2} \text{erfc} \left( \frac{\ln(h/\alpha)}{\sqrt{2n}} + \frac{n}{\sqrt{2}} \right) \right\}^2 & (h < 0) \\
K_s & (h > 0)
\end{cases} \quad (5-21)
\]
A dual-porosity model (Durner, 1994) (DPD). Durner (1994) divided the porous medium into two (or more) overlapping regions and suggested to use for each of these regions a van Genuchten-Mualem type function (van Genuchten et al., 1998) of the soil hydraulic parameters. Linear superposition of the functions for each particular region gives then the functions for the composite multimodal pore system (Durner et al., 1999):

\[ S_e = W_1[1 + (\alpha_1 h)^{n_1}]^{-m_1} + W_2[1 + (\alpha_2 h)^{n_2}]^{-m_2} \]  

(5-22)

Combining this retention model with Mualem’s pore-size distribution model (Mualem, 1976) leads to with:

\[ K(S_e) = \frac{(W_1 S_{e1} + W_2 S_{e2})^l}{(W_1 \alpha_1 + W_2 \alpha_2)^l} \left[ \frac{1 - (1 - S_{e1}^{1/m_1})^{m_1}}{W_1 \alpha_1} + \frac{1 - (1 - S_{e2}^{1/m_2})^{m_2}}{W_2 \alpha_2} \right]^2 \]  

(5-23)

where \( w_i \) (\( w_1, w_2 \)) are the weighting factors for the two overlapping regions, and \( \alpha_i, n_i, m_i (=1-1/n_i) \), and \( l \) are empirical parameters of the separate hydraulic functions (\( i=1,2 \)).

In a forward (direct) solution, H1D was run several times until the best operational criteria, especially those related to time discretization, iteration, and time steps were determined. The hydraulic models (a) to (e) were selected and the desired parameters were introduced in separate runs for each model. Simulated SWCs were compared with measured ones in order to find the most efficient hydraulic model.

5.2.7.3 Calibration of the H1D through parameter optimization

The predictive capability of the unsaturated flow and transport models relies heavily on the accurate estimates of the soil-water retention and unsaturated soil hydraulic characteristics at the application scale of the model (Vrugt, 2004). Parameter optimization is an indirect approach for the estimation of soil hydraulic and/or solute transport parameters from transient flow and/or transport data. Inverse methods are typically based upon the minimization of a suitable objective function, which expresses the discrepancy between the observed values and the predicted system response. The system response is represented by a numerical solution of the flow equation, augmented with the parameterized hydraulic functions, selected transport parameters, and suitable initial and boundary conditions. Initial estimates of the optimized system parameters are then iteratively improved during the minimization process until
a desired degree of precision is obtained. This methodology was originally applied to one-step and multi-step column outflow data generated in the laboratory (Kool et al., 1985; van Dam et al., 1994), and laboratory or field transport data during steady-state water flow (van Genuchten, 1981; Toride et al., 1995). The objective function \( \Phi \) to be minimized during the parameter estimation process was defined as (Šimůnek et al., 2013):

\[
\Phi(b,q,p)=\sum_{j=1}^{m_q} v_q \sum_{i=1}^{n_{qj}} w_{i,j} [q_j(x,t_i) - q_j(x,t_i,b)]^2
\]

\[
+ \sum_{j=1}^{m_p} \tilde{v}_j \sum_{i=1}^{n_{pj}} \tilde{w}_{i,j} [p_j(\theta_j)]^2
\]

\[
- p_j(\theta_j,b)]^2 + \sum_{j=1}^{n_b} \tilde{v}_j [b_j - b_j]^2
\]

where the first term on the right-hand side represents deviations between measured and calculated space-time variables, such as pressure heads, water contents, and/or concentrations at different locations and/or times in the flow domain, or actual or cumulative fluxes versus time across a certain boundary. The water content was the measured variable in our study.

In the first term, \( m_q \) is the number of different sets of measurements, \( n_{qj} \) is the number of measurements in a particular measurement set, \( q_j(x,t_i) \) represents specific measurements at time \( t_i \) for the \( j \)th measurement set at location \( x(r,z) \), \( q_j(x,t_i,b) \) are the corresponding model predictions for the vector of optimized parameters \( b \) (e.g., \( \theta_r, \theta_s, \alpha, n, K_s, \ldots \)), and \( v_q \) and \( w_{i,j} \) are weights associated with a particular measurement set or point, respectively. The second term of Eq. (5-24) represents differences between independently measured and predicted soil hydraulic parameters (e.g., retention, \( \theta(h) \) and/or hydraulic conductivity, \( K(\theta) \) or \( K(h) \) data), while the terms \( m_p, n_{pj}, p_j^* (\theta_j), p_j(\theta_j,b), \tilde{v}_j \) and \( \tilde{w}_{i,j} \) have similar meanings as for the first term but now for the soil hydraulic parameters. The last term of Eq. (5-24) represents a penalty function for deviations between prior knowledge of the soil hydraulic parameters, \( b_j^* \), and their final estimates, \( b_j \), with \( n_b \) being the number of parameters with prior knowledge and \( \tilde{v}_j \) representing pre-assigned weights. Estimates, which make use of prior information (such as those used in the third term of Eq. (5-24)) are known as Bayesian estimates. We note that the covariance (weighting) matrices, which provide information about the measurement accuracy, as well as any possible correlation between measurement errors and/or parameters, are
assumed to be diagonal in this study. The weighting coefficients $v_j$, which minimize differences in weighting between different data types because of different absolute values and numbers of data involved, are given by (Clausnitzer and Hopmans, 1995):

$$v_j = \frac{1}{n_j \sigma_j^2}$$  \hspace{1cm} (5-25)

which causes the objective function to become the average weighted squared deviation normalized by the measurement variances $\sigma_j^2$. Minimization of the objective function $\Phi$ was accomplished by using the Levenberg-Marquardt nonlinear minimization method (a weighted least-squares approach based on Marquardt’s maximum neighborhood method) (Marquardt, 1963). This method combines the Newton and steepest descend methods, and generates confidence intervals for the optimized parameters. The method was found to be very effective and has become a standard in nonlinear least-squares fitting among soil scientists and hydrologists (van Genuchten, 1981; Kool et al., 1985).

H1D was run in inverse mode with different scenarios to find out the optimized values for the hydraulic parameters $\theta_r$, $\theta_s$, $\alpha$, $n$, $K_s$, and $l$. The field measured SWC dataset of different layers corresponding to the period equal to those used for SWB method was employed as the observed $\theta_v$ values and the measured retention data as the observed $\theta(h) - h$ for the RLs. H1D allows users to optimize up to 15 parameters. However, since unsaturated flow problems are inherently ill-posed, it is not recommended to optimize that many parameters simultaneously. When more parameters are optimized, the problem often becomes non-unique (Šimůnek et al., 2013).

5.2.7.4 Model evaluation and statistical analysis

The performance of H1D was evaluated by employing three statistical criteria (Šimůnek and Hopmans, 2002). The root-mean-square errors (RMSE) (Hall, 2001), the coefficient of determination ($r^2$), and the Nash–Sutcliffe coefficient of model efficiency ($Ce$) (Nash and Sutcliffe, 1970) are the most popular and widely used performance criteria to evaluate the difference between the observed and modelled data, and were also employed in our study:

$$Ce=1 - \frac{\sum_{i=1}^{m} (\hat{\theta}_i - \hat{\theta}_i)^2}{\sum_{i=1}^{m} (\theta_i - \hat{\theta}_i)^2}$$  \hspace{1cm} (5-26)
5.2.7.5 Sensitivity of the simulated flux to the input parameters

Analysis of sensitivity helps to study how the uncertainty in the model output can be allocated to uncertainty in the model inputs by quantifying the sensitivity of the model output to systematic changes in the model input (Loosvelt, 2013). To calculate the extent of change in estimated fluxes by the calibrated model that would be caused by change in input parameters for a steady state transient flow system, a sensitivity analysis was performed on all of the input hydraulic parameters $\theta_r$, $\theta_s$, $K_s$, $\alpha$, $n$ and $l$ in VGM model. A modified form of what is proposed by van der Kwast et al. (2009) was used as follows:

$$S_i(F \pm) = \left(\frac{F \pm - F_0}{F_0}\right) \times 100$$  \hspace{1cm} (5-29)

$S_i$ is calculated for a positive or a negative deviation of an input of the hydraulic property. $F_0$, $F^+$ and $F^−$ are the fluxes predicted by H1D when the input equals its reference value $i_0$, and a set of multiplication factors of 1.5, 1.25, 0.75 and 0.5 $i_0$, respectively, when the reference values are used for all other inputs. The variation of 1.5 to 0.5 was chosen to cover the variations is observed in hydraulic parameters measurements. As the field measurements of the stony soils is subjected to practical complexities, the variation in measured properties is high and the tolerance of +/-50% was seen in replications especially for $K_{fs}$, $\theta_r$ and the bulk densities.

5.2.7.6 Validation of the H1D results

The validity of simulated recharge by the calibrated H1D was assessed by means of comparison by calculated recharge based on SWB method as the
observed value. The same procedure as conducted for the entire profile (0-400 cm), which is explained in section 5.2.8.1, was recurred for every layer (e.g. 0-10, 0-60, 0-80, 0-400) to calculate the amount of recharge by SWB method. Similarly the amount of simulated recharge which had been calculated using H1D by considering the observation nodes for the same layer depths was used to prepare a set of concurrent simulated recharge data. The set of simulated was assessed against the set of observed data by means of error percentage.

### 5.3 Results and discussion

#### 5.3.1 Layers specification

The description of the RLs distributed in the profiles of the three wells is summarized in Table 5-2. The layers are arbitrary coded in alphabetic order but their appearance in the profiles does not follow the same order (Table 5-3). The layers can be categorized into fine texture with a low percentage of stones (less than 10% and small size), and coarse texture with a high percentage of stones (over 50% and large size), but also stone size distribution and shape of the stones (platy vs. rounded) resulted in differences in hydraulic properties of the layers. For instance, layers C and F have almost the same amount of stones but differ noticeably in their measured infiltration rates. Layers A and E both are fine textured with low stone content but compaction of layer F has led to a low measured infiltration rate compared to a high rate in layer A. Measured data related to the water-retention curve can better show the difference in hydraulic properties. When running the H1D in forward (direct) solution, the hydraulic models (a) to (e) were selected and the desired parameters were introduced in separate runs for each model. Simulated SWCs were compared with measured ones in order to find the most efficient hydraulic model. As depicted in Fig. 5-3 the MVG model has shown the less RMSE of SWC estimation among the hydraulic models. Hence, the retention curves for the layers presented in Fig. 5-4 are drawn based on the best fitted model using the RETC code version 6.02 (van Genuchten et al., 1998) (Fig. 5-3). The model parameters of the selected model (MVG) and the related statistical criteria are shown in Table 5-3.
Table 5-2. Some physical properties measured for representative layers.

<table>
<thead>
<tr>
<th>Layer code</th>
<th>Considering gravel</th>
<th>Neglecting gravel</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>mm</td>
<td>%</td>
</tr>
<tr>
<td>200-0 0.00</td>
<td>0.00-0.02</td>
<td>0.02-2</td>
</tr>
<tr>
<td>A</td>
<td>17.5</td>
<td>42.5</td>
</tr>
<tr>
<td>B</td>
<td>3.3</td>
<td>16.1</td>
</tr>
<tr>
<td>C</td>
<td>3.1</td>
<td>7.8</td>
</tr>
<tr>
<td>D</td>
<td>2.2</td>
<td>6.3</td>
</tr>
<tr>
<td>E</td>
<td>6.4</td>
<td>13.6</td>
</tr>
<tr>
<td>F</td>
<td>1.2</td>
<td>2.9</td>
</tr>
<tr>
<td>G</td>
<td>1.5</td>
<td>22.4</td>
</tr>
</tbody>
</table>

Texture L is loam, SL is sandy loam, SiL is silty loam, LS is loamy sand, S is sand.

Fig. 5-3. Mean RMSE statistics for the hydraulic models fitted to the layers by RETC. VGM is van Genuchten Mualem, VGM-2 is van Genuchten with -2 cm air entry value, MVG is modified van Genuchten, BC is Brooks and Corey, LNK is log normal Kosugi, DPD is dual porosity Durner. RMSE refers to the root mean squared difference between the HID simulated water contents and the TDR observed ones at all depths of Well 1 in BZ1.
Table 5-3. Statistics of the modified van Genuchten model for the representative layers.

<table>
<thead>
<tr>
<th>Layer code</th>
<th>$R^2$</th>
<th>$C_e$</th>
<th>RMSE, m$^3$m$^{-3}$</th>
<th>$\theta_s$, m$^3$m$^{-3}$</th>
<th>$\theta_r$, m$^3$m$^{-3}$</th>
<th>$\alpha$, m$^{-1}$</th>
<th>$n$</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0.967</td>
<td>0.965</td>
<td>0.030</td>
<td>0.061</td>
<td>0.34</td>
<td>0.01</td>
<td>2.34</td>
</tr>
<tr>
<td>B</td>
<td>0.988</td>
<td>0.987</td>
<td>0.013</td>
<td>0.001</td>
<td>0.34</td>
<td>0.02</td>
<td>1.58</td>
</tr>
<tr>
<td>C</td>
<td>0.972</td>
<td>0.972</td>
<td>0.010</td>
<td>0.003</td>
<td>0.26</td>
<td>0.31</td>
<td>1.18</td>
</tr>
<tr>
<td>D</td>
<td>0.974</td>
<td>0.972</td>
<td>0.006</td>
<td>0.002</td>
<td>0.12</td>
<td>0.04</td>
<td>1.46</td>
</tr>
<tr>
<td>E</td>
<td>0.980</td>
<td>0.979</td>
<td>0.019</td>
<td>0.071</td>
<td>0.42</td>
<td>0.03</td>
<td>1.52</td>
</tr>
<tr>
<td>F</td>
<td>0.970</td>
<td>0.966</td>
<td>0.011</td>
<td>0.001</td>
<td>0.18</td>
<td>0.02</td>
<td>1.67</td>
</tr>
<tr>
<td>G</td>
<td>0.995</td>
<td>0.995</td>
<td>0.007</td>
<td>0.078</td>
<td>0.40</td>
<td>0.05</td>
<td>1.58</td>
</tr>
</tbody>
</table>

A is soil surface loam 7% fine gravel, B is loamy sand 22% gravel, C is sandy loam 54% medium gravel, D is sandy loam 67% coarse gravel, E is sandy loam 7% fine gravel, F is sand 53% medium gravel, G is loamy sand 0.3% medium gravel. The $r^2$ is determination coefficient, $C_e$ is Nash-Satcliffe coefficient, RMSE is root mean square error.

Fig. 5-4. Retention curves for the representative layers of Well 1 in BZ1. A, soil surface loam 7% fine gravel; B, loamy sand 22% gravel; C, sandy loam 54% medium gravel; D, sandy loam 67% coarse gravel; E, sandy loam 7% fine gravel; F, sand 53% medium gravel; G, loamy sand 0.3% medium gravel (red color to be distinguished). BZ is Bisheh Zard.

5.3.2 Distribution of layers

Distribution of different layers in the three wells is shown in Table 5-4. The locations and the depths of appearance of layers in each well do not show any order and follow the natural historical periods of low and high flooding magnitude and intensity during the alluvial cone formation.
The seven RLs were categorized to three main groups in order to correlate the layer codes in the three wells: the surficial deposited sediment, the sub layer with fine texture, and the coarse gravelly sub layer. Therefore, the new layers were illustrated in every well and then linked to the corresponding layers in the other wells. As depicted in Fig. 5-6, however, there is no real correspondence between the layers distribution at identical depths, but the majority of the layers continue in all wells.

The statistics of the three wells (Table 5-4), which were dug and investigated for vertical distribution of the pre-defined RLs, show that identical RLs are distributed along all three wells but in different depths. This challenges the assumption of uniform layering of the aquifers which is normally considered in modelling concepts (Fig. 5-6).

**Table 5-4. Vertical distribution of the representative layers (RL) in the experimental wells based on arbitrary coding of the pre-defined characteristics of each RL.**

<table>
<thead>
<tr>
<th>Layer</th>
<th>EW1 Depth, m</th>
<th>Layer</th>
<th>EW2 Depth, m</th>
<th>Layer</th>
<th>EW3 Depth, m</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0.10</td>
<td>E</td>
<td>24.40</td>
<td>A</td>
<td>0.10</td>
</tr>
<tr>
<td>B</td>
<td>0.93</td>
<td>F</td>
<td>25.40</td>
<td>B</td>
<td>0.60</td>
</tr>
<tr>
<td>E</td>
<td>1.09</td>
<td>G</td>
<td>28.00</td>
<td>E</td>
<td>1.60</td>
</tr>
<tr>
<td>C</td>
<td>3.30</td>
<td>F</td>
<td>28.80</td>
<td>C</td>
<td>1.80</td>
</tr>
<tr>
<td>G</td>
<td>4.20</td>
<td>F</td>
<td>28.00</td>
<td>G</td>
<td>2.15</td>
</tr>
<tr>
<td>E</td>
<td>6.50</td>
<td>E</td>
<td>28.60</td>
<td>E</td>
<td>2.65</td>
</tr>
<tr>
<td>G</td>
<td>7.00</td>
<td>F</td>
<td>30.60</td>
<td>G</td>
<td>3.20</td>
</tr>
<tr>
<td>E</td>
<td>7.40</td>
<td>D</td>
<td>3.70</td>
<td>C</td>
<td>4.40</td>
</tr>
<tr>
<td>G</td>
<td>9.05</td>
<td>F</td>
<td>4.90</td>
<td>E</td>
<td>4.70</td>
</tr>
<tr>
<td>E</td>
<td>9.20</td>
<td>C</td>
<td>7.10</td>
<td>F</td>
<td>6.30</td>
</tr>
<tr>
<td>G</td>
<td>10.7</td>
<td>E</td>
<td>7.50</td>
<td>E</td>
<td>6.60</td>
</tr>
<tr>
<td>D</td>
<td>12.5</td>
<td>F</td>
<td>7.95</td>
<td>F</td>
<td>7.60</td>
</tr>
<tr>
<td>G</td>
<td>13.6</td>
<td>G</td>
<td>11.0</td>
<td>E</td>
<td>8.00</td>
</tr>
<tr>
<td>C</td>
<td>17.5</td>
<td>E</td>
<td>12.2</td>
<td>F</td>
<td>8.40</td>
</tr>
<tr>
<td>G</td>
<td>18.4</td>
<td>G</td>
<td>12.6</td>
<td>E</td>
<td>8.70</td>
</tr>
<tr>
<td>C</td>
<td>19.4</td>
<td>F</td>
<td>13.5</td>
<td>F</td>
<td>9.50</td>
</tr>
<tr>
<td>E</td>
<td>19.9</td>
<td>E</td>
<td>15.0</td>
<td>E</td>
<td>9.80</td>
</tr>
<tr>
<td>G</td>
<td>21.1</td>
<td>F</td>
<td>18.8</td>
<td>F</td>
<td>10.0</td>
</tr>
<tr>
<td>F</td>
<td>21.5</td>
<td>E</td>
<td>19.0</td>
<td>E</td>
<td>10.4</td>
</tr>
<tr>
<td>G</td>
<td>21.8</td>
<td>C</td>
<td>19.3</td>
<td>F</td>
<td>11.2</td>
</tr>
<tr>
<td>F</td>
<td>24.1</td>
<td>F</td>
<td>19.6</td>
<td>D</td>
<td>11.6</td>
</tr>
</tbody>
</table>

The letters A to G are assigned arbitrarily to the representative layers in this study for differentiation and delineation. EW is experimental well.
5.3.3 Assigning measured $K_f$ to the layers

Statistics of measured values of the $K_f$ (Table 5-5) show a distinct variation at different depths for the RLs. All of the measurements performed on a particular RL are not necessarily a good representative of that RL, mainly because of the marginal effects in the transition zone between the two layers, or the local differences of each RL at different depth of a well.

Table 5-5. Statistics of measured field saturated hydraulic conductivity ($K_f$) of representative layers.

<table>
<thead>
<tr>
<th>Layer code</th>
<th>Max. $K_f$, cm day$^{-1}$</th>
<th>Min. $K_f$, cm day$^{-1}$</th>
<th>Geometric Mean$^{(1)}$</th>
<th>Corrected for sediment con.$^{(1)}$</th>
<th>No. of measurements</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>88</td>
<td>66</td>
<td>76</td>
<td>25</td>
<td>2</td>
</tr>
<tr>
<td>B</td>
<td>265</td>
<td>200</td>
<td>230</td>
<td>232</td>
<td>2</td>
</tr>
<tr>
<td>C</td>
<td>897</td>
<td>897</td>
<td>897</td>
<td>897</td>
<td>1</td>
</tr>
<tr>
<td>D</td>
<td>481</td>
<td>481</td>
<td>481</td>
<td>481</td>
<td>1</td>
</tr>
<tr>
<td>E</td>
<td>858</td>
<td>39</td>
<td>240</td>
<td>140</td>
<td>6</td>
</tr>
<tr>
<td>F</td>
<td>832</td>
<td>70</td>
<td>244</td>
<td>553</td>
<td>14</td>
</tr>
<tr>
<td>G</td>
<td>626</td>
<td>17</td>
<td>206</td>
<td>72</td>
<td>9</td>
</tr>
</tbody>
</table>

1- Correction for sediment concentration is made by using the Eq. (5-8).

Fig. 5-5. Change in the mean field saturated hydraulic conductivity ($K_f$) of the layers due to change in soil texture and stoniness. The lines are regressions between $K_f$ and corresponding parameters clay+silt and stoniness.
Fig. 5-6. Diagram of layer distribution and correlation of experimental wells’ layers. Alphabetic characters in left diagram (A to G) are summarized in three groups in left diagram (A to C) as described in it legend. Dimensions in x direction are not to scale.
Although the variation of measured $K_{fs}$ for the RLs is relatively high, when the averaged values are plotted vs. the soil texture and stoniness (Fig. 5-5), a logic trend of change is observed. The finer soil texture and the less stoniness caused the lower $K_{fs}$ and vice versa.

5.3.4 SWC data quality

SWC measurements from 21 August 2010 to 1 December 2013 at different depths in experimental well are given in Table 5-6. This resulted in ~180 measurements per depth. The data related to the depths 0.1 to 12.0 m form a complete set with some minor exception at depths 2.9, 7.0 and 10.0 m. The data of the lower depths from 14.0 to 28.0 m show many missing values mainly due to disconnection of extension cables in some time intervals. According to the close relationship between the measured data of different depths, a corrected set of data was prepared based on generated regression equations.

5.3.4.1 Temporal and vertical distribution of SWC data

The magnitude and change in the SWC data as influenced by rain and flooding are presented in Fig. 5-7. SWC in the layers do not show a gradual change due to their dependence on the physical characteristics of the layers. For instance, the lowest SWC in dry periods was 0.110 m$^3$ m$^{-3}$ at a depth of 10.0 m, whereas it was 0.054 m$^3$ m$^{-3}$ at a depth of 12.0 m. A good consistency in the SWC data is apparent at all depths. The SWC values remained stable during the dry periods and they changed after rainfall and/or flooding events up to the 4.0 m depth; however, the dry period values remained unchanged at a depth of 5.0 m and below. This behavior can be observed more clearly in Fig. 5-8, which illustrates the response of the SWC at different depths due to the two major flooding events. The first date in both events refers to the last measurement of the SWC before the flooding events. The second date is related to the measurement immediately after the first date and the third one refers to the measured data of the maximum change in the SWC of the layers. The SWC noticeably increased in the top layers from 0.1 to 1.6 m after the event (first vs. second date). After redistribution of water the value of SWC in deeper layers (2.0 to 4.0 m) exceeded the SWC of the upper part. No obvious persistent change can be detected from 5.0 m downward during the entire monitoring period. We conclude that the SWC data series are temporally persistent enough and are robustly responsive to the water infiltration as influenced by rainfall and the ponding water.
The reason for restriction of flow below a depth of 4.0 m might be ascribed to the major difference in $K_{fs}$ between the two distinctive layers at this depth. As shown in Table 5-4, the RL of G is located at the depth of 3.3 to 4.2 m. The $K_{fs}$ of the lower and upper layers (RL of E and G) was measured as 857 and 17 cm day$^{-1}$, respectively (see the $K_{fs}$ values in Appendix 5). Our long time data indicate that the water flow in the form of uniform wetting front was interrupted at this depth (Fig. 5-7 and Fig. 5-8).

These observations contradict, at first sight, those made in chapter 3, where the recharge of GW from floodwater spreading was illustrated and calculated based on a water budget. Furthermore, the SWC data of the layers below 5.0 m infers an occasional jump in SWC at some dates, which are not necessarily related to the flooding or rainfall events. Downward movement of water, therefore, seems to occur at lower depths albeit the fact that its occurrence could not be sensed by the TDR probes. A possible explanation for this behavior may be preferential flow of water in the form of fingers in this layer.

Fingering phenomena in layered soils have been studied and formulated by several authors (Hill and Parlange, 1972; Hillel and Baker, 1988; Kawamoto et al., 2004; Posadas et al., 2009; Zhao et al., 2010; Hardie et al., 2013). It typically occurs in layered soil with finer and less conductive upper layer overlaying coarser layers, with the wetting front becoming unstable and breaking into narrow wetting columns, or “fingers”. The fingers move downward at a velocity equal to the saturated conductivity divided by the saturated volumetric water content of the bottom layer (Hill and Parlange, 1972). Fingering columns have been physically sensed by freezing method (Hill and Parlange, 1972), or by magnetic resonance imaging (MRI) (Posadas et al., 2009). Samani et al. (1989) found in their laboratory experiment that up to a hydraulic conductivity ratio of 20 of lower vs. upper layer, the wetting front is one-dimensional, and the movement through the second layer is unsaturated. When the hydraulic conductivity ratio exceeds 20, the wetting front loses its one-dimensionality and the water moves through the second layer in the form of wet columns as a finger. Occurrence of fingering flow in our study is evidenced by the interruption of the wetting front in the layer located over a depth of 400 cm, where a fine layer with low hydraulic conductivity covers the lower gravelly coarse layer starting from a depth 420 cm. The measured hydraulic conductivity of the two layers in this study (17 and 857 cm day$^{-1}$) for the upper and lower layers, respectively, shows a ratio
of 50, which is, by far, more than the critical value reported by Samani et al. (1989) for starting the fingering flow.

Table 5-6. Statistics of measured soil-water content at different depths of W1 (period between 21 August 2010 to 1 December 2013).

<table>
<thead>
<tr>
<th>Depth, cm</th>
<th>No of measurements</th>
<th>Mean</th>
<th>St. dev.</th>
<th>Min</th>
<th>Max</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Original</td>
<td>Corrected</td>
<td>Original</td>
<td>Corrected</td>
<td>Original</td>
</tr>
<tr>
<td>10</td>
<td>180</td>
<td>180</td>
<td>0.108</td>
<td>0.058</td>
<td>0.031</td>
</tr>
<tr>
<td>40</td>
<td>180</td>
<td>180</td>
<td>0.105</td>
<td>0.039</td>
<td>0.047</td>
</tr>
<tr>
<td>60</td>
<td>180</td>
<td>180</td>
<td>0.145</td>
<td>0.059</td>
<td>0.066</td>
</tr>
<tr>
<td>80</td>
<td>180</td>
<td>180</td>
<td>0.076</td>
<td>0.028</td>
<td>0.048</td>
</tr>
<tr>
<td>110</td>
<td>180</td>
<td>180</td>
<td>0.106</td>
<td>0.055</td>
<td>0.051</td>
</tr>
<tr>
<td>140</td>
<td>180</td>
<td>180</td>
<td>0.129</td>
<td>0.062</td>
<td>0.054</td>
</tr>
<tr>
<td>170</td>
<td>180</td>
<td>180</td>
<td>0.086</td>
<td>0.042</td>
<td>0.050</td>
</tr>
<tr>
<td>200</td>
<td>180</td>
<td>180</td>
<td>0.145</td>
<td>0.057</td>
<td>0.079</td>
</tr>
<tr>
<td>230</td>
<td>180</td>
<td>180</td>
<td>0.121</td>
<td>0.042</td>
<td>0.079</td>
</tr>
<tr>
<td>260</td>
<td>180</td>
<td>180</td>
<td>0.111</td>
<td>0.031</td>
<td>0.084</td>
</tr>
<tr>
<td>290</td>
<td>172</td>
<td>180</td>
<td>0.092</td>
<td>0.026</td>
<td>0.072</td>
</tr>
<tr>
<td>400</td>
<td>180</td>
<td>180</td>
<td>0.114</td>
<td>0.031</td>
<td>0.079</td>
</tr>
<tr>
<td>500</td>
<td>180</td>
<td>180</td>
<td>0.066</td>
<td>0.005</td>
<td>0.061</td>
</tr>
<tr>
<td>600</td>
<td>180</td>
<td>180</td>
<td>0.068</td>
<td>0.002</td>
<td>0.063</td>
</tr>
<tr>
<td>700</td>
<td>180</td>
<td>180</td>
<td>0.100</td>
<td>0.004</td>
<td>0.093</td>
</tr>
<tr>
<td>800</td>
<td>180</td>
<td>180</td>
<td>0.139</td>
<td>0.012</td>
<td>0.095</td>
</tr>
<tr>
<td>900</td>
<td>180</td>
<td>180</td>
<td>0.117</td>
<td>0.025</td>
<td>0.087</td>
</tr>
<tr>
<td>1000</td>
<td>177</td>
<td>180</td>
<td>0.165</td>
<td>0.014</td>
<td>0.117</td>
</tr>
<tr>
<td>1200</td>
<td>180</td>
<td>180</td>
<td>0.071</td>
<td>0.018</td>
<td>0.054</td>
</tr>
<tr>
<td>1400</td>
<td>76</td>
<td>180</td>
<td>0.052</td>
<td>0.005</td>
<td>0.034</td>
</tr>
<tr>
<td>1600</td>
<td>123</td>
<td>180</td>
<td>0.071</td>
<td>0.013</td>
<td>0.056</td>
</tr>
<tr>
<td>2000</td>
<td>73</td>
<td>180</td>
<td>0.059</td>
<td>0.004</td>
<td>0.037</td>
</tr>
<tr>
<td>2400</td>
<td>90</td>
<td>180</td>
<td>0.063</td>
<td>0.003</td>
<td>0.039</td>
</tr>
<tr>
<td>2600</td>
<td>56</td>
<td>180</td>
<td>0.076</td>
<td>0.019</td>
<td>0.035</td>
</tr>
<tr>
<td>2800</td>
<td>100</td>
<td>180</td>
<td>0.211</td>
<td>0.012</td>
<td>0.185</td>
</tr>
</tbody>
</table>
Fig. 5-7. Soil-water content time series of some selected depths measured by the calibrated TDR probes. Depths are noted at the center of the graphs.
Fig. 5-8. Change in the SWC of layers before and after the two flooding events 28 Jan. 2011 (A and C) and 1 Aug. 2013 (B and D). The graphs A and B show the change in the SWC over the entire 30 m profile, while graphs C and D zoom into the top 5 of the profile to better illustrate the redistribution of the SWC in the top layers.

Therefore, the placement of the TDR probes at depths of 500 cm downward may intercept the finger columns only by chance, and there is no guarantee to intercept the column, owing to the absence of a uniform, one dimensional wetting front. As shown in Fig. 5-8, some occasional jumps in the SWC of the
lower than 500 cm layers was observed, and may be inferred as interception of the finger columns by the TDR probe installed at that position. Conversely, a minor but gradual increase in SWC at depths of 1000 and 2600 cm is depicted in Fig. 5-7, which is not coincident with the flooding time. These layers, also categorized as RL of G with a low $K_{fs}$, have the capacity to collect the incoming water from their top layers, and initiate fingering flow in their underlying layers.

5.3.4.2 Empirical quantile-quantile (q-q) Plot
As inferred from the empirical q-q plots (Fig. 5-9), the scatter varies for the different depths representing different layers, although meaningful correlations in q-q plots can be seen at all of depths. It can be concluded that the empirical probability of the SWC measurements are comparable for all depths.
Fig. 5-9. The empirical q-q plot for the soil-water content at depths of 10 to 400 cm of experimental well. The number of samples is similar for all of the plots and the probability values are greater than 0.01.
5.3.5 Effect of sedimentation on infiltration

The FWS system used as the research site was constructed in 1983. A top layer of silt loam, 10 to 20 cm thick, had been formed over the original surface due to the deposition of the suspended load during 30 years of being used. Therefore it was subjected to sedimentation and consequently to reduction in infiltration rate.

Fig. 5-10 illustrates the SWC changes at different soil depths from 8 January to 7 February 2011. Before the flooding event of 29 January to 2 February 2011, there had been an increase in the SWC at 10 and 40 cm depths due to the rainfalls of 16 and 22 January (11.5 and 29.0 mm, respectively). The rainfall on 28 January (28.5 mm) had also resulted in a minor increase in the SWC at the depths of 10, 40 and 400 cm. It shows that rainfall of 28.5 mm per day can still produce infiltration in our soils. A major change in the SWC occurred in all depths due to the flooding of 29 January to 1 February. Our data of water ponding in that event showed that all of the ponded water had infiltrated into the soil profile after 30 hours. Likewise, the increase in the SWC of the depths 170 and 400 cm took only 48 hours to reach its maximum. Therefore, it can be concluded that the hydraulic conductivity of the top soil did not hinder the infiltrating water to reach a depth of 400 cm.

The reduction in hydraulic conductivity in FWS systems has been reported by some authors (Boroomand et al., 2004; Sarreshtehdari and Skidmore, 2005; Ghazavi et al., 2010; Mahdian et al., 2011) by comparing measured infiltration rates inside FWS basins and those of control sites. In these group of works the experimental measurements were performed by the double ring method using non-turbulent water movement, while the process of infiltration under natural conditions includes the impact of the turbulent floodwater and the resulting sediment deposition on infiltration rate (Simpson and Meixner, 2012).

The hydraulic conductivity measurements in our study were performed during floodwater spreading events where a turbulent horizontal movement is being take place. It is found by many authors that $K_{fs}$ increased during the flooding event. Doppler et al. (2007) attributed an increase in $K_{fs}$ after a large flood along an impounded reach to erosion of fine surface sediments. Mutiti and Levy (2010) inferred consistent increases in $K_{fs}$ during the rising limbs, and declines in $K_{fs}$ during the falling limbs of a series of floods at a site with induced infiltration due to nearby pumping. The authors associated these
changes to the preferential removal of fine bed sediment particles. A study by (Hatch et al., 2010) also inferred increases in $K_f$ during the rising discharge and declines in it after the flood termination in central coastal California. These types of temporal changes in $K_f$ during flood events could alter the amount and timing of fluxes between the impounded surfaces and their underlying aquifers, both during and after events. These results are in accord with our findings acquired under natural conditions of flood induced infiltration showing minor negative effect of long term sedimentation on soil surface hydraulic conductivity.

![Fig. 5-10. Changes in soil-water content (SWC) at different depths of the experimental well as influenced by the rainfall and flooding events. The number of layers has decreased to better differentiate the changes between the layers.](image)

**5.3.6 Evaluation of recharge**

**5.3.6.1 SWB method**

As inferred from Table 5-7, the rainfall and flooding event that started on 28 January 2011, caused an increase in soil-water storage $S$ of the top 400 cm layers from 10.80 to 19.33 cm. Flooding continued until 1 Feb 2011, and the increase in $S$ of the top 400 cm reached 35.18 cm on that day. The profile's $S$ decreased gradually afterwards until 23 July, when total $S$ had decreased to 11.54 cm, close to its amount before the flooding event. Summation of the difference in soil-water storage between two measurements $\Delta S$ for the entire period should approach zero. $ET_o$ values are daily based for the successive days
of the SWC measurements and the cumulative for the intervals of non-
successive days (Table 5-7).

As shown in Table 5-7, the remainder of $\Delta S$ for the entire period was 2.0 cm, which means that the total water infiltrated into the profile in terms of rainfall and floodwater (51.8 cm) had left the profile as either $ET$ or recharge. As $ET_a$ was calculated as 20.2 cm, the net recharge depth to the aquifer was estimated at 29.6 cm for that particular event.
**Table 5-7. Soil-water budget data from the 400 cm top layers, the flooding period 16 January to 23 July 2011.**

<table>
<thead>
<tr>
<th>Dates</th>
<th>$S$</th>
<th>$\Delta S$</th>
<th>$P$</th>
<th>$F$</th>
<th>$ET$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1/16/11</td>
<td>30.0</td>
<td>0.0</td>
<td>1.2</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>1/23/11</td>
<td>35.4</td>
<td>5.4</td>
<td>2.9</td>
<td>0.0</td>
<td>0.9</td>
</tr>
<tr>
<td>1/28/11</td>
<td>40.1</td>
<td>4.7</td>
<td>2.9</td>
<td>0.0</td>
<td>1.0</td>
</tr>
<tr>
<td>2/1/11</td>
<td>55.5</td>
<td>15.4</td>
<td>5.5</td>
<td>0.0</td>
<td>1.0</td>
</tr>
<tr>
<td>2/2/11</td>
<td>65.8</td>
<td>10.3</td>
<td>0.0</td>
<td>35.0</td>
<td>0.0</td>
</tr>
<tr>
<td>2/3/11</td>
<td>64.8</td>
<td>-1.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.3</td>
</tr>
<tr>
<td>2/4/11</td>
<td>63.2</td>
<td>-1.6</td>
<td>0.0</td>
<td>0.0</td>
<td>0.3</td>
</tr>
<tr>
<td>2/5/11</td>
<td>62.5</td>
<td>-0.7</td>
<td>0.0</td>
<td>0.0</td>
<td>0.3</td>
</tr>
<tr>
<td>2/6/11</td>
<td>61.6</td>
<td>-0.8</td>
<td>1.8</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>2/7/11</td>
<td>60.3</td>
<td>-1.3</td>
<td>0.2</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>2/9/11</td>
<td>59.4</td>
<td>-0.9</td>
<td>0.0</td>
<td>0.0</td>
<td>0.5</td>
</tr>
<tr>
<td>2/10/11</td>
<td>62.9</td>
<td>3.6</td>
<td>0.0</td>
<td>0.0</td>
<td>0.2</td>
</tr>
<tr>
<td>2/11/11</td>
<td>61.2</td>
<td>-1.7</td>
<td>0.0</td>
<td>0.0</td>
<td>0.2</td>
</tr>
<tr>
<td>2/12/11</td>
<td>60.9</td>
<td>-0.3</td>
<td>1.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>2/13/11</td>
<td>61.1</td>
<td>0.2</td>
<td>0.0</td>
<td>0.0</td>
<td>0.2</td>
</tr>
<tr>
<td>2/14/11</td>
<td>60.8</td>
<td>-0.3</td>
<td>0.0</td>
<td>0.0</td>
<td>0.2</td>
</tr>
<tr>
<td>2/15/11</td>
<td>59.4</td>
<td>-1.4</td>
<td>0.0</td>
<td>0.0</td>
<td>0.2</td>
</tr>
<tr>
<td>2/16/11</td>
<td>59.8</td>
<td>0.3</td>
<td>0.0</td>
<td>0.0</td>
<td>0.2</td>
</tr>
<tr>
<td>2/17/11</td>
<td>58.7</td>
<td>-1.1</td>
<td>0.0</td>
<td>0.0</td>
<td>0.2</td>
</tr>
<tr>
<td>2/18/11</td>
<td>59.0</td>
<td>0.3</td>
<td>0.4</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>2/19/11</td>
<td>61.6</td>
<td>2.6</td>
<td>1.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>2/20/11</td>
<td>60.7</td>
<td>-0.9</td>
<td>0.0</td>
<td>0.0</td>
<td>0.2</td>
</tr>
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$S$ is soil-water storage summed up for 400 cm top layers, $\Delta S$, is change in soil storage as subtraction of measured $S$ for each day from the last previous measured $S$, $P$ is precipitation, $F$ is flooding and $ET_a$ is actual $ET$ which is cumulative for the non-consecutive days.
Table 5-7. Continued

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1/16/11 to 7/23/11  2.0  16.8  35.0  20.2

Recharge (cm) in entire period: R=P+F-ΔS-ET  29.6

$S$, is soil-water storage summed up for 400 cm top layers, $\Delta S$, is change in soil storage as subtraction of measured $S$ for each day from the last previous measured $S$, $P$ is precipitation, $F$ is flooding and $ET$ is actual $ET$ which is cumulative for the non-consecutive days.

5.3.6.2 Water flux Modelling by HID

Eleven scenarios were evaluated in which a variety of hydraulic parameters and models as well as boundary conditions were subjected to change in every scenario as explained in Table 5-8. An optimum parameter set (with the highest $r^2$ and the lowest RMSE) was obtained with the hydraulic model MVG (Table 5-8, scenario 11). It is important to know that the statistics $r^2$ and RMSE are
generated based on the whole data set of simulated water content of each scenario and the observed water content corresponding to all of the layers.

Table 5-8. Characteristics of the different scenarios for running the H1D in inverse mode.

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<th>K_f</th>
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R² | 0.73 | 0.76 | 0.81 | 0.82 | 0.83 | 0.86 | 0.89 | 0.91 | 0.92 | 0.93 | 0.96
RMSE | 0.12 | 0.09 | 0.08 | 0.08 | 0.07 | 0.05 | 0.05 | 0.04 | 0.04 | 0.03 |

Hydraulic models: a is van Genuchten Mualem, b is Modified Van Genuchten, c is modified van Genuchten Mualem with air entry -2, d is Brooks and Corey, e is log normal Kosugi, f is dual porosity Durner. BC is boundary condition. Hydraulic parameters optimized n is no and y is yes. θ_s and θ_r are residual and saturated soil-water content, K_f is field saturated hydraulic conductivity; α, n and l are hydraulic models parameters. The layer depths of 10, 60, 180 and 400 cm are equal to the representative layers A, B, C and G respectively. R² and RMSE are the statistical criteria.

The flux calculated at 400 cm depth was considered as the final net recharge due to the rainfall and flooding events during the period between 16 January to 2 February July 2011. A good agreement was achieved between the
observed and simulated SWCs (Fig. 5-11). The optimized hydraulic parameters for the layers are shown in Table 5-9.

![Graphs showing observed and simulated SWCs](image)

Fig. 5-11. The observed and simulated soil-water content (SWC) data for the layers at 10 cm (A), 60 cm (B), 180 cm (C) and 400 cm (D). $R^2$ is determination coefficient, RMSE is root mean square error. The mass balance error of optimal run of the model was calculated as 0.6%. The optimal hydraulic model was determined to be the modified van Genuchten (MVG) model.

The hydraulic parameters, calculated from the soil moisture retention data and applying the RETC model, did not match the results obtained from the inverse modelling in H1D using measured SWC (Table 5-9). It proves that the retention data, which were determined on disturbed samples in the laboratory did not represent the real soil-water relationships in the field. Optimized parameters by H1D are based on the real change in the SWC of the underlying layers, and may be considered as what has really happened in the field after the flooding event. Solone et al. (2012) have also concluded that using retention curves derived from pressure plates causes an error in the recharge calculation.
Table 5-9. Modelling information of the final H1D running results.

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<td>60</td>
<td>180</td>
<td>400</td>
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<td>( \theta_r ), m(^3) m(^{-3})</td>
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<td>0.061</td>
<td>0.001</td>
<td>0.001</td>
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</tbody>
</table>

The ini. and opt. are initial and optimized values of the parameters. Initial values for \( \theta_r \), \( \theta_s \) and \( K_{fs} \) are based on measurements and for \( \alpha \) and \( n \) are generated by RETC parameter fitting on the measured retention data. \( \theta_r \) and \( \theta_s \) are residual and saturated soil-water content, \( K_{fs} \) is field saturated hydraulic conductivity; \( \alpha \), \( n \) and \( l \) are hydraulic models parameters. The optimal hydraulic model was determined to be the modified van Genuchten (MVG) model. The layer depths of 10, 60, 180 and 400 cm are equal to the representative layers A, B, C and G respectively.

Inspection of the SWC data indicated that after the flooding event SWC in all of the layers did not exceed 50% of \( \theta_0 \) in all layers. This is supported by Dahan et al. (2008) who found that water content in the vadose zone with similar soil conditions as our study site did not exceed field capacity.

As inferred from Fig. 5-12, the amount of recharge computed by the calibrated H1D over the entire simulation period was 28.8 cm.

5.3.6.3 Sensitivity analysis

The sensitivity of the simulated flux to the deviation in hydraulic parameters (Fig. 5-13) showed that the parameters can be categorized into three groups. Parameters to which the model is highly sensitive with \( S_i > 50\% \) are \( n \) and \( \theta_s \). Medium sensitivity with \( S_i \) 10-50\% is shown for \( K_{fs} \) and \( \alpha \), while the other parameters showed low sensitivity with the \( S_i \) less than 10\%. Evidently from the last optimum running scenario (Table 5-9, column 11), the H1D run includes the optimization of the parameters \( n \), \( \theta_s \) and \( K_{fs} \) for all layers. Since the model showed to be the most sensitive to these parameters, their optimized values based on measured soil-water content during the calibration procedure, it can guaranty the reliability of the results in this study.

This is in accordance with the other works in which H1D was used under a variety of soil conditions, e.g. (Schoups and Hopmans, 2006; Leão and Gentry, ...
2011; Li et al., 2012). In a water harvesting study in stony soils in drylands, Verbist et al. (2009) found that the Hydrus code was most sensitive to $K_{fs}$, $\theta_s$, $n$ and $\alpha$. Accordingly, excluding the parameters $\theta_r$ and $l$ as insensitive parameters from the parameter optimization did not result in additional error in our simulations.

Fig. 5-12. Simulated flux (A) and the cumulative recharge (B) by H1D for the period between 16 January to 23 July 2011 period. Dashed line in figure B shows the amount of total recharge in the entire period.
5.3.6.4 Evaluation of the H1D result

The recharge simulated by H1D and that calculated by the soil-water budget method for the different layer depths is presented in Table 5-10. The simulations show both over and under estimations on the recharge as compared to the soil-water budget method. The resulted RMSE of 6.6 which is smaller than the standard deviation of measured (soil-water budget method’s) data of 7.1 indicates that the simulated recharge by the calibrated H1D is statistically acceptable. On the other hand, the optimized hydraulic parameters values for the layers under study are robust enough and might therefore be acceptable to
use in future simulations including up-scaling to the larger area of the FWS systems in our study site.

Table 5-10. The error in the simulated recharge by the H1D as compared with the soil-water budget method for the period between 16 January to 23 July 2011.

<table>
<thead>
<tr>
<th>Profile depth, cm</th>
<th>Recharge, cm</th>
<th>SWB</th>
<th>H1D</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-10</td>
<td>39.5</td>
<td>36.1</td>
<td></td>
</tr>
<tr>
<td>0-60</td>
<td>38.2</td>
<td>38.7</td>
<td></td>
</tr>
<tr>
<td>0-80</td>
<td>35.6</td>
<td>40.9</td>
<td></td>
</tr>
<tr>
<td>0-110</td>
<td>33.2</td>
<td>47.6</td>
<td></td>
</tr>
<tr>
<td>0-200</td>
<td>25.2</td>
<td>26.5</td>
<td></td>
</tr>
<tr>
<td>0-230</td>
<td>23.0</td>
<td>30.3</td>
<td></td>
</tr>
<tr>
<td>0-290</td>
<td>43.0</td>
<td>31.3</td>
<td></td>
</tr>
<tr>
<td>0-400</td>
<td>29.6</td>
<td>28.8</td>
<td></td>
</tr>
<tr>
<td>Mean</td>
<td>33.4</td>
<td>35.0</td>
<td></td>
</tr>
<tr>
<td>SD</td>
<td>7.1</td>
<td>7.2</td>
<td></td>
</tr>
<tr>
<td>RMSE</td>
<td></td>
<td>6.6</td>
<td></td>
</tr>
</tbody>
</table>

SWB is soil-water budget method, H1D is Hydrus 1D model, SD is standard deviation and RMSE is root mean squared error.

Consequently we can consider the amount of 29.6 cm resulted by soil-water budget method as the reliable outcome of recharge calculation of this study. Hence, from the 51.8 cm of input water (16.7 and 35.0 cm rainfall and floodwater, respectively), 29.6 cm (57%) moved below 400 cm depth to follow its downward movement to the aquifer. The amount of artificial recharge which was calculated in this chapter is comparable to that was evaluated in the chapter 3. In the both methods the amount of pounded floodwater plus rainfall was considered as the input water and the part of infiltrated water to the deep layers were evaluated. In the chapter 3, conditional to the method used, 56 to 61% of the input water was estimated as artificial recharge. Similarly, in a silty clay loam with a fragmented layer below 60 cm, Soldevilla-Martinez et al. (2013) calculated the drainage from the profile bottom at 170 cm depth as 70% of precipitation and ponding water using the soil-water budget method. On the contrary, some reports show lower recharge percentage; for instance, Touhami et al. (2013), calculated recharge to be 4 to 28% of precipitation in a fine-
textured soil with different land cover types using three soil-water budget models. Bellot and Chirino (2013) calculated recharge due to net precipitation (precipitation minus interception) as 30 to 51% in wet years, and 8 to 30% in dry years in a forest with fine textured (clay loam) soil. Webb et al. (2008) calculated recharge by the LEACHM model as 9 to 13% of the annual precipitation (1200 mm). The main difference between our study and those presenting low recharge percentages can be ascribed to the type of input water. In our study, the summation of rainfall and the ponded floodwater were the sources of input, which play a major role in the recharge efficiency. In addition, the hydraulic properties of the soils under study in our site are characterized by coarse and gravelly texture, and therefore, show higher hydraulic conductivity and lower water retention in comparison with the above mentioned reports.

The ~60% recharge we calculated with both methods indicates that the FWS system of interest is efficient in recharging the aquifer. This finding is valid for the location of the experimental well and cannot be considered as the efficiency of the entire system.

5.4 Conclusions

A unique collection of soil hydraulic properties and SWC data until ~ 30 m depth was achieved in an active basin of the floodwater spreading system used mainly for the artificial recharge of GW.

Seven representative layers (RLs) could be differentiated and recognized among the profile of the experimental well. A range of measured $K_f$ obtained showed a meaningful negative relationship to the fine particles (clay+silt) and positive relationship to the extent of stoniness with $r^2$ of 0.70 and 0.86 respectively.

A thorough investigation on two other wells, which were dug in vicinity of experimental well, showed that identical RLs are distributed along all three wells but with different vertical distribution and depth of appearance. This challenges the assumption of uniform layering of the aquifers, which is normally considered in modelling concepts.

Although SWC data series collected were temporally persistent enough and robustly responsive to infiltration as influenced by rainfall and ponding water,
no noticeable change in the SWC was detected at depths lower than 400 cm. The reason why the flow below a depth of 400 cm was limited might be ascribed to the major difference in $K_{fs}$ between the two distinctive layers at this depth which might cause the fingering flow. Samani et al. (1989) showed that the fingering flow in a vertical water movement in layered column is started if a ratio of >20 is occurred for hydraulic conductivities of two adjacent fine over coarse textured layers. As shown in Table 5-4, the RL of G is located at depths of 3.3 to 4.2 m. The $K_{fs}$ of the two adjacent upper (G) and lower (E) layers, was 17 and 857 cm day$^{-1}$ respectively. Our long time data indicate that the water flow in the form of uniform wetting front was interrupted at this depth. Occurrence of fingering flow in our study is evidenced by the interruption of the wetting front in the layer located below 400 cm, where a fine layer with a low hydraulic conductivity covers the lower coarse gravelly layer starting from a depth of 420 cm. The measured $K_{fs}$ of the two layers in this study shows a ratio of 50, which is by far more than the critical ratio (>20) for initiation of the fingering flow. It can be inferred from the $\theta_v$ data set that the decrease in hydraulic conductivity due to clogging by siltation in the studied artificial recharge system did not affect its efficiency, at least at the experimental site. The fact that the wetting front arrived at a depth of 400 cm after 48 hours from the start of the flooding event clearly challenges the assumption of negative impact of surface clogging on infiltration. As the wall of the well was primarily insulated by concrete tiles, there was no possible lateral movement of water to the well; hence, the change in the SWC can only be attributed to the vertical movement of water as influenced by the vertical infiltration of rainfall and/or ponded water.

The SWC time series, the concurrent collected ponding water, rainfall and the remotely sensed $ET_a$ data were used to solve the soil-water budget approach for a period of 188 days. In this period the initial SWC of the soil layers increased in response to the rainfall and/or flooding events from the top to the underlying layers successively and thereafter started to decrease due to $ET$ and recharge until reverting to its initial value. H1D was run in inverse mode for calibrating and optimization of the modified van Genuchten parameters $\alpha$, $n$, $\theta_s$ and $K_{fs}$, to which the model showed highest sensitivity according to a sensitivity analysis. As the optimized values are
based on the vertical changes in the field measured SWC, it is reasonable to be considered as what has really occurred under the experimental conditions. Measurements of hydraulic parameters related to retention data are based on the disturbed samples, and the optimized values are determined by the dataset of natural undisturbed conditions. Simulations with the calibrated H1D model showed for a 188 day period a 28.8 cm recharge, at 400 cm depth, which was of the same order of magnitude as that calculated with the soil-water budget method of 29.6 cm.

The 29.6 cm of recharge that occurred after a total of 51.8 cm input water from rainfall and floodwater was added to the basin, results in a 57% of efficiency.

Although a reliable set of data is obtained for calculating recharge at the very location of this study, up-scaling of the results for the entire floodwater systems and for the other flooding events with extreme volumes and flow rate needs extended investigation period and thorough identification of the underlying layers. The determined hydraulic properties of the RLs obtained in this study will be utilized in the future research works in the FWS systems in our study site.
Chapter 5 - 185

Photo 5-1. Floodwater spreading in the studied basin (A) and the height of ponding water (water mark) in flooding events around the experimental well (B).
Photo 5-2. Identification of representative layers inside the experimental well.

Photo 5-2. Stoniness of the experimental well layers.
Photo 5-3. Infiltration measurement by double ring method inside the experimental well at a depth of 3 m.
Photo 5-4. Insulating the experimental well by concrete tiles (A) and the insulated top view of the experimental well (B).
Photo 5-5. The TDR equipped experimental well (A), the TDR cable insulating pipes (B), and the lowest installed TDR probe at a depth of 28 m (C).
Photo 5-6. Inserting the TDR probes. A polystyrene guide and a steel template for preparing the holes for the pipes containing the TDR probe (A), the place for the TDR probe inside the well (B), the inserted polystyrene template (C), the place for inserting the TDR probe (D and E), and the successive places for the TDR probes (F).
Photo 5-7. Arrangement of the TDR cables in the multiplexer in the experimental well.
Chapter 6. General discussion and conclusions
6.1 Responses to the research objectives

6.1.1 Overall objective

This dissertation has arisen from concerns regarding the impact of artificial recharge through floodwater spreading (FWS) to groundwater augmentation in combating water scarcity at the study site. The overall objective of this dissertation was “to evaluate FWS systems for recharging the groundwater table”. In general, evaluation of effectiveness is a measurable concept quantitatively determined by the ratio of output to input (Robinson et al., 1991).

In order to evaluate the studied FWS system for artificial recharge of the groundwater table, an extensive multidisciplinary approach was developed to quantify the recharge as influenced by the addition of the captured and diverted floodwater to the system.

In all of the recharge assessment methods used in this thesis, temporal and spatial variation of $ET$ from the different land uses in the study site was an important component. Therefore, part of this study was devoted to the mapping of actual evapotranspiration ($ET_a$) by a pre-calibrated remote sensing model. Then, the recharge rate was assessed through two independent approaches, a saturated and unsaturated zone one.

Saturated zone approach

In the saturated approach, we firstly examined the immediate impact of FWS on the water table level. The observation wells (OW) inside the FWS system showed a twice as high response to the flooding event in comparison with the OW outside the system. Tolerance against groundwater (GW) extraction lasted for OWs inside the FWS much longer than those outside the FWS system. We further employed a combination of water budget and water table fluctuation (WTF) methods to solve for recharge by quantifying all of the involving components. Considering that the calculated 4.13 Mm$^3$ was depleted from the aquifer storage during the hydrological year 2010-2011, and the return flow as 3.20 Mm$^3$, the recharge was calculated at 7.94 Mm$^3$ in that year, which was a consequence of both artificial recharge and natural replenishment.

Artificial recharge data in the same year during the flooding events from 28 January to 2 February 2011, showed a total volume of 6.92 Mm$^3$ of flood water
that was retained in the FWS systems. The calculated artificial recharge was calculated at 4.48 and 4.46 Mm$^3$ by the methods of using flow data and the water budget, respectively, from these events. Therefore, the contribution of the FWS systems to recharge was calculated as 56 to 61% for this hydrological year. This ratio is similar to those reported by Hendrickx et al. (1991) who found values of 60% to 80% for an alluvial stony fan resembling our study site. Moreover, in a parallel study in our study area (Gareh Bygone Plain), Hashemi et al. (2013) showed, using a numerical hydrologic modelling, that the contribution of the FWS systems to the recharge is 80% in normal events and 41% in extreme events.

**Unsaturated zone approach**

In unsaturated approach, after calibrating the TDR method for the stony soils of the studied aquifer layers, a combination of soil-water budget (SWB) and numerical modelling with Hydrus 1D (H1D) was employed after calibration to simulate the recharge in a period between 16 January 2010 to 30 July 2011.

In this period the SWB method showed that the total water infiltrated into the profile in terms of rainfall and floodwater (51.8 cm) had left the profile as either $ET$ or recharge and a soil-water storage of 2.0 cm was retained in the profile layers. The cumulative $ET_a$ was calculated as 20.2 cm and therefore, the net recharge was estimated at 29.6 cm for this period. The concurrent recharge simulated by H1D and calculated by the SWB method for the different layer depths provided the basis for evaluation of the H1D. An acceptable RMSE of 6.6 cm which was smaller than 7.1 cm standard deviation of SWB method showed that, simulated recharge by the calibrated H1D model was acceptable. Therefore, from the 51.8 cm of input water, 29.6 cm was net recharge, which shows an efficiency of 57% for the system. This value is valid for the experimental site, which is considered as being representative for the BZ$_1$ FWS system under study.

**Integration of the results**

In spite of the persistent lowering of the GW level in recent decades, the efficiency of the system was quantitatively evaluated as 58-63% and 57% by the saturated and unsaturated zone approaches, respectively. This is supported by a parallel study by Hashemi (2014), who introduced different water management scenarios to assess the impact of management parameters such as
size of FWS system, GW extraction and climate change on the GW level in the coming years. The scenarios revealed that abstraction most substantially affects the GW level and the continuation of abstraction at the current pumping rate would lead to a GW decline by 18 m up to 2050. Their results show that the recharge volume can be increased by expanding the artificial recharge system, even for small flood events, while recharge through the river channel is only substantial for major flood events.

Therefore, extension and development of FWS systems in the studied area would be a main and permanent solution to increase the floodwater recharge in case of any extreme or small flash flood event.

Apart from the quantified evaluation of the system as such, which was the main objective of the thesis, there were two ambiguities with the system that needed to be resolved i.e. the effect of sedimentation on infiltration rate due to clogging and the consequence of water consumption of dense plantation of water demanding *E. camaldulensis* trees on the water budget. These issues are addressed along with the specific research questions (RQ) as presented below.

### 6.1.2 Specific objectives (SO)

**SO1: Improved estimation of ET**

Evaluation and calibration of surface energy balance system (SEBS) model to maximize the reliability of the ET and crop coefficient (*Kc*) estimations in the study site as a hot and dry region.

*RQ1: How can the SEBS model parameterization be improved with limited available data through calibration?*

Parameterization of the SEBS model which includes selecting the appropriate sources and methods of acquiring the model parameters, was improved through maximizing the agreement between estimated *ETa* and reference *ETo*, for the water reservoir pixels. Global radiation, maximum daily wind speed, *do*, *zom*, and vegetation height were the most important parameters whose adjustment resulted in better predictions. The model showed minor sensitivity (*Si* <10%) to RS products, vegetation input parameters, DEM and relative humidity, and a high sensitivity (*Si* >10%) to air temperature, wind speed and air pressure.
**RQ2:** What is the agreement between SEBS modeled ETa and ETa calculated by water budget method?

The cross validation of seasonal ETa showed a minor difference in ETa simulated with SEBS and that calculated from the water budget for the irrigated crops (ratio simulated/observed of 1.02). However, a major discrepancy for the pastures and bare soils outside the FWS systems (ratio of 3.4) was observed. Because of lack of reliable data on water consumption for the pastures inside the FWS systems and tree plantations, the predicted ETa could not be compared with that from the water budget ETa. Therefore, it can be concluded by this evaluation method that, the calculated water consumption based on summing up the modelled ETa is reliable for irrigated crops, but not reliable for the pastures and bare soils outside the FWS systems. Reliability of the estimated ETa was left to be evaluated by means of the crop coefficient Kc.

**RQ3:** Are the Kc values estimated by SEBS comparable to those published in literature for the different land uses and crops?

The estimated Kc for the main cultivated crops (wheat and forage corn) based on the SEBS acquired results compared well with the Kc published for these crops with similar climatic soil conditions. The Kc values for the tree plantations with different canopy densities compared well with published Kc values. On the contrary, the Kc values estimated for the pastures inside and outside the FWS systems was higher than those published in the literature.

**RQ4:** What is the yearly water consumption of the main land uses by means of summed up ET in growing season?

Results showed that the main crops, wheat and forage corn, consumed ~11 million m$^3$ (Mm$^3$), both from groundwater (GW) and rainfall during the hydrological year 2010-2011, whereas trees used 1 Mm$^3$ in the same period. As the major source of both land uses is GW, it can be inferred that agriculture was a key water consumer and the afforested area (mainly *E. camaldulensis*) only played a minor role in water depletion in the study site.

**SO2:** Assessing the recharge by the saturated zone method
Assessing the impact of the FWS systems on recharge by saturated zone methods with limited sources of groundwater data available by combining the water table fluctuation and water budget methods.

**RQ5: What is the immediate effect of flooding events on the GW level of the main aquifer?**

Spatial changes in the GW position indicated that the highest drop in the water level occurred in places where irrigated fields were concentrated. On the other hand, the lower recession in observation well Nos. 2, 5 and 6, being located inside the FWS systems, proved that the area under the direct impact of the FWS systems was less susceptible to water withdrawal. The observation well located inside the FWS systems revealed a greater resistance to drought and abstractions than the other wells in the area. The hydrograph of observation OW2 displayed a substantial disparity in GW rise (0.5 vs. 2.05 m) in two major floods in comparable months in 2004 and 2005, which additionally demonstrates the effectiveness of the FWS system. In 2003, it was not functioning due to maintenance and repair, which was reflected in a minor GW rise, whereas in 2004, when the system was active again, GW rise was substantial.

**RQ6: What is the proportion of natural and artificial recharge through water budget in a hydrological year when full input data are available?**

In the particular hydrological year of 2010-2011, in which the data for abstraction and depletion were conceivable to evaluate, contribution of the FWS systems to the net recharge was calculated at nearly 60%. Thus, natural recharge contributed for 30% for the same duration.

**SO3: Assessing recharge by the unsaturated zone method**

Calibration of the Hydrus 1D model through hydraulic parameter optimization for assessing the recharge by the FWS systems based on soil-water content measurements over the vadose zone.

**RQ7: How reliable are soil-water content data acquired by the TDR method in stony soils?**

A thorough investigation with the TDR method indicated that the equations developed here for converting measured dielectric permittivity \( K_a \) to soil
water content $\theta_v$, and for compensating for cable length effect, resulted in improving the reliability of the derived $\theta_v$. The $\theta_v$ measurements of multi-layered stony soil might be noticeably erroneous if the TDR probes are directly inserted into the bulk soil. Apart from the practical difficulties for slotting the concrete tile and the soil adjacent to it, it may cause the probe’s rods not to be inserted into the soil in a parallel position; an inadequate contact between the rods and the soil particles will affect the medium for transmission. It would be advantageous to excavate the desired hole with enough space lengthwise in the undisturbed soil profile, filling the hole with soil of similar texture, and inserting the probes inside the hole.

**RQ8: Does the clogging due to long term sedimentation after flooding events affects actual infiltration rate in the FWS systems in a real flooding event?**

It was inferred from the SWC data set that the decrease in hydraulic conductivity due to clogging by siltation in the studied artificial recharge system did not play a role in decreasing its efficiency at the experimental site. The fact that the wetting front arrived at a depth of 400 cm 48 hours after the start of the flooding event, clearly challenges the assumption of negative impact of surface clogging on infiltration. As the wall of the well was primarily insulated by concrete tiles, there was no possible lateral movement of water to the well; hence, the change in the SWC could only be attributed to vertical water movement resulting from infiltration of rainfall and/or ponded water.

**RQ9: What is the influence of a flooding event on the soil-water budget components (including the net recharge) in aquifer profile?**

The remainder of $\Delta S$ for the entire period from 16 January 2010 to 30 June 2011 was 2.0 cm, which means that the total water infiltrated into the profile in terms of rainfall and floodwater (51.8 cm) had left the profile as either $ET$ or recharge. As $ET_a$ was calculated as 20.2 cm, the net recharge depth due to flooding and rainfall was estimated at 29.6 cm for this particular period.

**RQ10: What is the reliability of the direct measurements of hydraulic parameters in disturbed soils as compared to the optimized hydraulic parameters with calibrated Hydrus 1D (H1D) model based on the measured soil-water contents?**
The hydraulic parameters, calculated from the soil moisture retention data and applying the RETC models, do not match the results obtained from the inverse modelling using the measured SWC in the H1D. It proves that the retention data, whose determination is based on the laboratory measurement of disturbed samples, cannot show the real soil-water relationship under the actual situation. Optimized parameters by the H1D are based on the real change in the SWC of the underlying layers, and may be considered as what had really happened in the field after the flooding event.

*RQ11: How sensitive is the H1D model to the input hydraulic parameters?*

The sensitivity of the simulated flux to deviations in hydraulic parameters showed that they can be categorized in three groups. H1D showed high sensitivity to parameters $n$ and $\theta_s$ (the $S_i>50\%$), medium sensitivity to $\alpha$ and $K_{fs}$ (the $S_i 10-50\%$), and the least sensitivity (the $S_i <10\%$) to the other parameters. Since the sensitive parameters have been optimized based on measured soil-water content, reliability of the results in this study are guaranteed.

*RQ12: To what extent is the simulated recharge by the calibrated H1D model reliable for different aquifer layers?*

When comparing recharge simulated with H1D and that calculated by the SWB method for the different layer depths, both over and under estimations were observed. Since the RMSE was as small as 6.6 cm (less than the standard deviation of observed data of 7.1 cm), simulated recharge by the calibrated H1D model is considered as acceptable. On the other hand, the optimized hydraulic parameters values for the layers under study are robust enough to be used in future up-scaling simulations.

### 6.2 Contributions of the thesis

The contributions of this thesis can be summarized as: a) development of approaches for application, calibration and validation of existing models with limited available data, b) incorporation of new concepts into the models used, c) generating a unique and robust field data set to support the modelling approaches, and d) provision of new information in the context of floodwater harvesting and its impact on groundwater recharge.

a) Developed approaches
The SEBS model was calibrated to maximize the reliability of its $ET_a$ predictions in a study site with no access to real field measurements of $ET_a$ by assuming comparability of simulated $ET_a$ for water reservoir pixels and standard FAO P-M $ET_0$. This assumption improved model parameterization. Water budgeted $ET_a$ appeared to be a reliable tool for evaluating modeled $ET_a$. Maps generated to show spatial and temporal distribution of $ET_a$ were a useful source for solving the water budget. In the saturated zone approach that was used for assessing recharge, combining water table fluctuation and water budget methods enabled to quantify the proportion of artificial vs. natural recharge.

In the unsaturated zone approach, an improved optimization of the hydraulic parameters made the local calibration of H1D model achievable. Validation of H1D using recharge values calculated by the SWB method, demonstrated that H1D is a reliable tool for further studies on the whole FWS systems. Improved reliability of the TDR measurements and its implications for stony aquifer layers are the further contributions of this part of the thesis.

b) Incorporating new concepts into the models

New concepts for mapping global radiation and incorporating maximum daily wind speed in the governing equation of the SEBS model (chapter 2), for improved evaluation of specific yield in deep aquifers (chapter 3), new local equations for converting the $K_a$ to $\theta_e$, and for compensating the effect of cable length in the TDR method (chapter 4) were introduced.

c) Generating a unique and robust field data set to support modelling

In this study, aquifer layers were described along three wells inside the FWS system, hydraulic properties of the aquifer layers were determined from the surface to the groundwater level, and long-term time series of SWC over the entire aquifer profile until 30 m depth inside the FWS system, together with a concurrent flooding and rainfall data set, were established.

d) Providing new insights toward floodwater harvesting
Floodwater harvesting, especially in the form of FWS, is an emerging issue in water management in dry regions, which needs better understanding and evaluation of its impact on the surrounding environment. Small scale but nature friendly water management plans, such as FWS systems, are seriously criticized, since there are numerous methods, which are more attractive in terms of investments and money return to investors. However, they are rarely investigated. This study provided quantitative evidences that proves the effectiveness of FWS systems.

6.3 Prospective works

Application of the newly free downloadable Landsat (5 and 8) data has provided an opportunity to map ET. Given that the knowledge about actual ET from different sources of water consumption is absolutely crucial, the validated model (SEBS) in this study can be applied for water management purposes. Furthermore, the data set prepared in this study might be applied for evaluating other more theoretical models such as the METRIC (Allen et al., 2007), which is more widely applied in the world.

In parallel with the use of other physical methods such as direct measurements of water use by sap flow meter and eddy covariance method, ground truth data need to be generated to more accurately verify ET mapping based on the remote sensing.

To prepare intensive spatial data on water table behavior, there is a need to extend the number and to build an adequate network of equipped observation wells inside and around the FWS systems in the study site.

Other methods of assessing recharge, especially those based on using tracers for better differentiating the different recharge sources are to be employed.

More accurate abstraction data from the GW sources in terms of time and the space are needed to solve the water budget in all seasons and in all different scenarios. It necessitates to control the water withdrawal from the operational wells.

The H1D model study provided optimized and measured hydraulic properties and SWC, which can be applied for further up-scaling at this research site. A feasible number of experimental wells, at least to a depth of 400 cm inside the
FWS system is needed to spatially simulate three-dimensional water movement. This would help to better understand the efficiency of the system and determine the less efficient places, which would result in improving its effectiveness to ground water recharge.

The recession trend of the ground water table in the study site has reached a very critical and alarming state as demonstrated in this study. A logical balance between recharge and extraction must be necessarily and urgently developed to prevent the Gareh Bygone Plain to becoming a real desert in a not faraway future.
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Appendix 1. The scripts for automotive processing of the Landsat TM5 inside the GRASS in Linux for ET calculation

Step by step scripts for Linux users. For more detailed explanation please refer to the uploaded manual at:

svn.osgeo.org/grass/grass-promo/tutorials/grass_landsat_ETa/main_document.pdf

1- Unzipping the tar ball compressed files

For Linux users in Terminal and for Windows users in Dos prompt window should be written:

```
for file in *.tar.gz
  do
    tar -xvzf $file
  done
```

Control should be given to run the script in the same directory as the images are in.

2- Renaming and importing

In GRASS Command Line Interface (CLI) write the script:

```
echo "RUN from the MTL.txt directory and within the GRASS environment"
for file in L5*0.TIF
  do
    out=$(echo $file | sed 's/(.*)_\(.*_B(.*).*\)0.TIF/\1_\2/3/g')
    echo $out
    r.in.gdal input=$file output=$out
  done
```

For more information about this script, refer to “sed” program manual in Unix/Linux and to the “r.in.gdal” help in GRASS GIS.

3- Conversion of DN to radiance

Conversion of DN to reflectance at top of atmosphere (TOAR) is done with the script. This script works when the current path is the same as the location of folder containing the images.

```
echo "RUN from the MTL.txt directory and within the GRASS environment"
```
This script requires to change the path and row of the target images you are processing.

4-Atmospheric correction with i.atcorr

Script for atmospheric correction by 6S method for landsat 5 TM. This script uses a base name for a given landsat 5 image. The files DEM (altitude map) must be ready. The script will generate new parameter file for every band automatically.

```bash
#!/bin/bash
# Basic script for i.atcorr for L5 TM

# Geometrical conditions (L5TM)
geom=7

# Sensor height (satellite is -1000)
sens_height=-1000

# Here we suppose you have altitude (DEM)

# Visibility dummy value (overwritten by VIS raster input)
vis=9
r.mapcalc expression="visibility=$vis" --overwrite

# Altitude dummy value (in Km should be negative in this param file)
alt=-1.200

# Please change as you need

# L5 basename as stored in GRASS GIS and used by i.landsat.toar
L5basename=L5162040_04020090705

# Location of parameter file
root=/home/icwater/

# Datetime of satellite overpass (month, day, GMT decimal hour)
mdh="7 5 6.30"
```
# Central Lat/Long
Long=51.410
Lat=24.234

# Atmospheric mode
atm_mode=6 # US standard 62 (for lack of more precise model)

# Aerosol model
aerosol_mode=1 # continental

# Satellite band number (L5TM [25,26,27,28,29,30])
satbandno=25 # Band 1 of L5TM is first to undergo atmospheric correction

for bandno in 1 2 3 4 5 7
   do
      # Generate the parameterization file
      echo "$geom                            - geometrical conditions=Landsat 5 TM"
      > $root/param_L5.txt
      echo "$mdh $Long $Lat   - month day hh.ddd longitude latitude" >> $root/param_L5.txt
      echo "$atm_mode                            - atmospheric mode=tropical" >> $root/param_L5.txt
      echo "$aerosol_mode                            - aerosols model=continental"
      >> $root/param_L5.txt
      echo "$vis                           - visibility [km] (aerosol model concentration), not used as there is raster input" >> $root/param_L5.txt
      echo "$alt                       - mean target elevation above sea level [km] (here 600m asl), not used as there is raster input" >> $root/param_L5.txt
      echo "$sens_height                        - sensor height (here, sensor on board a satellite)"
      >> $root/param_L5.txt
      echo "$satbandno                           - 'i'th band of TM Landsat 5"
      >> $root/param_L5.txt
   done

5- Net radiation and the other input maps for ET calculation

i=0
g=0
  echo "day of year length:" $doy_value_len
  echo "sunzangle length:" $sunzangle_value_len
  echo "act_sunhours length:" $act_sunhours_value_len
  echo "atmosphericpressure length:" $atmosphericpressure_value_len
  echo "relativehumidity length:" $relativehumidity_value_len
  echo "eact length:" $eact_value_len
  echo "windspeed length:" $windspeed_value_len
  echo "Ustar length:" $Ustar_value_len
  echo "tempraturecelsuse length:" $tempraturecelsuse_value_len
  echo "netrad station length:" $netradstation_value_len

for file in $(g.mlist type=rast pattern=*.surf.1)
do
  g=$(echo "$i + 1" | b)
  echo ""
  echo "Image Number" $g
  echo "image name:" $file
  echo ""
  doy_value=${doy_value_list[$i]}
  sunzangle_value=${sunzangle_value_list[$i]}
  act_sunhours_value=${act_sunhours_value_list[$i]}
  atmosphericpressure_value=${atmosphericpressure_value_list[$i]}
  relativehumidity_value=${relativehumidity_value_list[$i]}
  eact_value=${eact_value_list[$i]}
  windspeed_value=${windspeed_value_list[$i]}
  Ustar_value=${Ustar_value_list[$i]}
  tempraturecelsuse_value=${tempraturecelsuse_value_list[$i]}
  netradstation_value=${netradstation_value_list[$i]}
  echo "day of year:" $doy_value
  echo "sunzangle:" $sunzangle_value
  echo "act_sunhours:" $act_sunhours_value
  echo "atmosphericpressure:" $atmosphericpressure_value
  echo "relativehumidity:" $relativehumidity_value
  echo "ustar:" $Ustar_value
  echo "windspeed:" $windspeed_value
  echo "eact:" $eact_value
  echo "tempraturecelsuse:" $tempraturecelsuse_value
  echo "netrad station:" $netradstation_value
  echo "local time:" $localtime_value
"
b1=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1/1\2.2 /')
b2=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1/1\2.3 /')
b3=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1/1\2.4 /')
b4=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1/1\2.5 /')
b5=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1.toar\1/1\2.6 /')
b6=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1/1\2.7 /')
b7=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1/1\2.8 /')
eecho $b1
echo $b2
echo $b3
echo $b4
echo $b5
echo $b6
echo $b7
t0m=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1_t0m /')
rm.mapcalc expression="$t0m=$b6-(0.00627*$dem)" --overwrite
albedo=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1_albedo /')
i.albedo -l --overwrite input=$b1,$b2,$b3,$b4,$b5,$b7 output=$albedo
ndvi=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1_ndvi /')
i.vi --overwrite viname=ndvi red=$b3 nir=$b4 output=$ndvi
emissivity=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1_emissivity /')
i.emissivity --overwrite input=$ndvi output=$emissivity
cropheight=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1_cropheight /')
r.mapcalc expression="$cropheight=if($ndvi <= 0.05, 0.02, if($ndvi <= 0.2, 0.2, if($ndvi <= 0.4, 0.25, if($ndvi <= 0.6, 0.3, if($ndvi > 0.6, 0.5)))))" --overwrite
cropheight1=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1_cropheight1 /')
Z0m=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1_Z0m /')
r.mapcalc expression="$Z0m=$cropheight*0.136" --overwrite
echo $ndvi
echo $t0m
echo $albedo
echo $emissivity
echo $Z0m
netradstation=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1_netradstation /')
r.mapcalc expression="$netradstation=$netradstation_value" --overwrite
localtime=$(echo $file | sed 's/\(.*/\|\.(.*)\|.\.(.*)\)/\1_localtime /')
r.mapcalc expression="$localtime=$localtime_value" --overwrite
airtemperature=$(echo $file | sed 's/\(.*\)\.(.*)\.(.*)\/\1\_airtemperature /')
r.mapcalc expression="$airtemperature=$tair_slope_value*$b6+$tair_intercept_value" --overwrite
doy=$(echo $file | sed 's/\(.*\)\.(.*)\.(.*)\/\1\_doy /')
r.mapcalc expression="$doy=${doy_value_list[$i]}" --overwrite
temperaturecelsuse=$(echo $file | sed 's/\(.*\)\.(.*)\.(.*)\/\1\_temperaturecelsuse\/)'
r.mapcalc expression="$temperaturecelsuse=${temperaturecelsuse_value_list[$i]}" --overwrite
sunhoures=$(echo $file | sed 's/\(.*\)\.(.*)\.(.*)\/\1\_sunhours /')
i.sunhours --overwrite dayofyear=$doy latitude=$latitude output=$sunhoures
tsw=$(echo $file | sed 's/\(.*\)\.(.*)\.(.*)\/\1\_tsw /')
echo "Transmissivity (n/N)"
r.mapcalc expression="$tsw=${act_sunhours_value_list[$i]}/$sunhoures" --overwrite
sunzangle=$(echo $file | sed 's/\(.*\)\.(.*)\.(.*)\/\1\_sunzangle /')
r.mapcalc expression="$sunzangle=${sunzangle_value_list[$i]}" --overwrite
dtair=$(echo $file | sed 's/\(.*\)\.(.*)\.(.*)\/\1\_dtair /')
r.mapcalc expression="$dtair=$b6-$airtemperature" --overwrite
netradiation=$(echo $file | sed 's/\(.*\)\.(.*)\.(.*)\/\1\_netradiation /')
i.eb.netrad --overwrite --quiet albedo=$albedo ndvi=$ndvi temperature=$b6 localutctime=$localtime temperaturedifference2m=$dtair emissivity=$emissivity transmissivitysingleway=$tsw dayofyear=$doy sunzenithangle=$sunzangle output=$netradiation
atmosphericpressure=$(echo $file | sed 's/\(.*\)\.(.*)\.(.*)\/\1\_atmosphericpressure /')
r.mapcalc expression="$atmosphericpressure=${atmosphericpressure_value_list[$i]}"

etpotd=$(echo $file | sed 's/\(.*\)\.(.*)\.(.*)\/\1\_etpotd /')
i.evapo.potrad -r -d --overwrite --quiet albedo=$albedo temperature=$b6 latitude=$latitude dayofyear=$doy transmissivitysingleway=$tsw waterdensity=1005.0 slope=$slope aspect=$aspect output=$etpotd
relativehumidity=$(echo $file | sed 's/\(.*\)\.(.*)\.(.*)\/\1\_relativehumidity /')
r.mapcalc
expression="$relativehumidity=${relativehumidity_value_list[$i]}" --overwrite

windspeed=$(echo $file | sed 's/\(.*\)\.(.*\)\.(.*\)/\1\_windspeed/')
r.mapcalc expression="$windspeed=${windspeed_value_list[$i]}" --overwrite

Mode 1

```
echo "Mode 1 instantaneous rnet (w/m^2)"
glob_rad1=$(echo $file | sed 's/\(.*\)\.(.*\)\.(.*\)/\1\_glob_rad1/')
r.sun --overwrite --quiet elevin=$dem aspin=$aspect slopein=$slope
lin=3.2 albedo=$albedo latin=$latitude longin=$longitude
glob_rad=$glob_rad1 day=$doy_value step=0.5 time=$localtime_value
dist=1.0 numpartitions=1
```

Mode 2

```
echo "Mode 2 daily rnet (w/m^2.day)"
glob_rad2=$(echo $file | sed 's/\(.*\)\.(.*\)\.(.*\)/\1\_glob_rad2/')
r.sun --overwrite --quiet elevin=$dem aspin=$aspect slopein=$slope
lin=3.2 albedo=$albedo latin=$latitude longin=$longitude
glob_rad=$glob_rad2 day=$doy_value step=0.5 dist=1.0 numpartitions=1
```

```
echo "average daily irradiation (w/m^2.hr)"
glob_rad2Day=$(echo $file | sed 's/\(.*\)\.(.*\)\.(.*\)/\1\_glob_rad2Day/')
r.mapcalc expression="$glob_rad2Day=$glob_rad2/$act_sunhours" --overwrite
```
Appendix 2. The 1:100,000 Geology Map of the Fasa region which includes the Gareh Bygone Plain (is circled). Source: Iranian Oil Company.
Close look of the Geology map; extracted from the 1:100,000 Geology map of the Fasa region.
Appendix 3. The log of the observation wells (OW)

<table>
<thead>
<tr>
<th>Number</th>
<th>Date of installation</th>
<th>UTM-X</th>
<th>UTM-Y</th>
</tr>
</thead>
<tbody>
<tr>
<td>OW1</td>
<td>25/11/1992</td>
<td>785500</td>
<td>3169800</td>
</tr>
<tr>
<td>OW2</td>
<td>17/11/1992</td>
<td>788000</td>
<td>3165350</td>
</tr>
<tr>
<td>OW3</td>
<td>10/11/1992</td>
<td>784000</td>
<td>3166600</td>
</tr>
<tr>
<td>OW4</td>
<td>22/11/1992</td>
<td>787100</td>
<td>3166700</td>
</tr>
</tbody>
</table>
Appendix 4. Discharge rate measurements of pumped wells.

The flow (discharge) rate of the pumped wells were measured in this study in three steps as follow.

1. When the pipe was fully discharging.

\[ Q = C_d (\pi D_p^2 / 4) \left( \frac{X^2}{2Y} \right) \]

where \( C_d \) is flow coefficient (~1.1), \( D_p \) is internal pipe diameter (m), \( X \) horizontal and \( Y \) is vertical distance (m) to the center of flow. The formula is valid for the 0.05 to 0.2 m of diameter standard pipes. A manufactured Try square with built-in table for flow calculation based on the \( X \) and \( Y \) was used in the field (Photo 3-4).

2. When the pipe was partially discharging.

\[ Q = 4.685 (1 - a/D_p)^{1.88} D_p^{2.48} \]

where \( a \) is the air filled length in pipe. The flow is reliable when the almost half of the pipe is full \((a/D_p > 0.45)\).
3. When the water discharge is as low as filling less than half of the pipe \((a/D_c)\).

In this case the discharge was measured by filling the water container of pre-defined volume and measuring the filling time (the discharge which is shown in the Photo 3-4).
Appendix 5. Particle size distribution (lab analysis) of the soil samples and field saturated hydraulic conductivity (field measured) of experimental well number 1.

<table>
<thead>
<tr>
<th>Depth, cm</th>
<th>Layers code</th>
<th>Clay %</th>
<th>Sand %</th>
<th>Silt %</th>
<th>USDA texture</th>
<th>Gravel %</th>
<th>Stone %</th>
<th>Gravel + stone %</th>
<th>Meas. $K_s$, cm day$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>A</td>
<td>12.0</td>
<td>56.0</td>
<td>32.0</td>
<td>SL</td>
<td>6.7</td>
<td>0.0</td>
<td>6.7</td>
<td>65.7</td>
</tr>
<tr>
<td>30</td>
<td>B</td>
<td>20.0</td>
<td>54.0</td>
<td>26.0</td>
<td>SL/SCL</td>
<td>22.0</td>
<td>0.0</td>
<td>22.0</td>
<td>199.8</td>
</tr>
<tr>
<td>60</td>
<td>C</td>
<td>5.0</td>
<td>86.0</td>
<td>9.0</td>
<td>LS</td>
<td>24.0</td>
<td>30.0</td>
<td>54.0</td>
<td>264.8</td>
</tr>
<tr>
<td>90</td>
<td>C</td>
<td>3.0</td>
<td>90.0</td>
<td>7.0</td>
<td>S</td>
<td>50.2</td>
<td>6.3</td>
<td>56.5</td>
<td>395.5</td>
</tr>
<tr>
<td>120</td>
<td>D</td>
<td>8.0</td>
<td>71.0</td>
<td>21.0</td>
<td>SL</td>
<td>12.0</td>
<td>54.0</td>
<td>66.0</td>
<td>110.3</td>
</tr>
<tr>
<td>150</td>
<td>E</td>
<td>6.0</td>
<td>90.0</td>
<td>4.0</td>
<td>S</td>
<td>3.2</td>
<td>4.2</td>
<td>7.4</td>
<td>480.9</td>
</tr>
<tr>
<td>180</td>
<td>F</td>
<td>5.0</td>
<td>92.0</td>
<td>3.0</td>
<td>S</td>
<td>22.0</td>
<td>32.0</td>
<td>54.0</td>
<td>545.4</td>
</tr>
<tr>
<td>210</td>
<td>F</td>
<td>4.0</td>
<td>83.0</td>
<td>13.0</td>
<td>LS</td>
<td>20.5</td>
<td>31.5</td>
<td>52.0</td>
<td>764.3</td>
</tr>
<tr>
<td>240</td>
<td>F</td>
<td>4.0</td>
<td>83.0</td>
<td>13.0</td>
<td>LS</td>
<td>30.0</td>
<td>25.3</td>
<td>55.3</td>
<td>72.7</td>
</tr>
<tr>
<td>270</td>
<td>C</td>
<td>14.5</td>
<td>54.5</td>
<td>18.0</td>
<td>SL</td>
<td>9.2</td>
<td>2.1</td>
<td>11.3</td>
<td>217.4</td>
</tr>
<tr>
<td>300</td>
<td>G</td>
<td>16.0</td>
<td>50.0</td>
<td>34.0</td>
<td>L</td>
<td>9.2</td>
<td>2.1</td>
<td>11.3</td>
<td>184.7</td>
</tr>
<tr>
<td>400</td>
<td>G</td>
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<td>83.0</td>
<td>13.0</td>
<td>LS</td>
<td>43.2</td>
<td>5.7</td>
<td>48.9</td>
<td>17.5</td>
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<tr>
<td>500</td>
<td>E</td>
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<td>92.0</td>
<td>2.0</td>
<td>S</td>
<td>27.5</td>
<td>17.0</td>
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<tr>
<td>600</td>
<td>F</td>
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<td>85.5</td>
<td>7.5</td>
<td>LS</td>
<td>33.5</td>
<td>34.2</td>
<td>67.7</td>
<td>311.6</td>
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<tr>
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<td>36.0</td>
<td>L</td>
<td>6.5</td>
<td>1.9</td>
<td>8.4</td>
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<td>F</td>
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<td>86.0</td>
<td>7.0</td>
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<td>34.7</td>
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<td>86.0</td>
<td>6.0</td>
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<td>9.2</td>
<td>1.8</td>
<td>11.0</td>
<td>625.8</td>
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<tr>
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<td>95.0</td>
<td>2.0</td>
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<td>2.0</td>
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<td>15.3</td>
<td>58.0</td>
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</tr>
<tr>
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<td>85.0</td>
<td>9.5</td>
<td>LS</td>
<td>6.0</td>
<td>2.1</td>
<td>8.1</td>
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</tr>
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<td>89.0</td>
<td>6.0</td>
<td>S</td>
<td>32.0</td>
<td>47.4</td>
<td>79.4</td>
<td>831.8</td>
</tr>
<tr>
<td>1700</td>
<td>D</td>
<td>5.5</td>
<td>72.0</td>
<td>22.5</td>
<td>SL</td>
<td>14.5</td>
<td>51.3</td>
<td>65.8</td>
<td>150.9</td>
</tr>
<tr>
<td>1800</td>
<td>D</td>
<td>0.0</td>
<td>97.0</td>
<td>3.0</td>
<td>S</td>
<td>12.0</td>
<td>51.5</td>
<td>63.5</td>
<td>274.2</td>
</tr>
<tr>
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<td>F</td>
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<td>2.0</td>
<td>S</td>
<td>21.0</td>
<td>37.4</td>
<td>58.4</td>
<td>2836.4</td>
</tr>
<tr>
<td>2000</td>
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<td>84.0</td>
<td>11.0</td>
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<td>40.0</td>
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<td>81.0</td>
<td>14.5</td>
<td>LS</td>
<td>43.9</td>
<td>12.8</td>
<td>56.7</td>
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<td>2.2</td>
<td>11.0</td>
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<td>95.0</td>
<td>3.0</td>
<td>S</td>
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<td>16.5</td>
<td>57.5</td>
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</tr>
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<td>95.0</td>
<td>2.0</td>
<td>S</td>
<td>40.0</td>
<td>9.0</td>
<td>49.0</td>
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<td>2600</td>
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</tr>
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<td>2700</td>
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<td>13.0</td>
<td>52.0</td>
<td>35.0</td>
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<tr>
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<td>S</td>
<td>48.5</td>
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<td>1297.8</td>
</tr>
</tbody>
</table>

Textural classes are SL for sandy loam, SCL for sandy clay loam, LS for loamy sand, S for sand and L for loam. Meas. $K_s$ is measured field saturated hydraulic conductivity.
Appendix 6. Specific yield calculation based on weighted average of different layers.

<table>
<thead>
<tr>
<th>Layers code</th>
<th>Depth, cm</th>
<th>Thickness, cm</th>
<th>$S_y$</th>
<th>$S_y$-thick, cm</th>
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<tbody>
<tr>
<td>A</td>
<td>0</td>
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<td>-</td>
<td>-</td>
</tr>
<tr>
<td>B</td>
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<td>-</td>
<td>-</td>
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<td>30</td>
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<td>2.76</td>
</tr>
<tr>
<td>D</td>
<td>120</td>
<td>30</td>
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<td>3.10</td>
</tr>
<tr>
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<td>30</td>
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</tr>
<tr>
<td>F</td>
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<td>4.67</td>
</tr>
<tr>
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<td>4.67</td>
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<td>4.67</td>
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</tr>
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<td>15.58</td>
</tr>
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</tr>
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<td>0.16</td>
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<td>28.54</td>
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</tr>
<tr>
<td>C</td>
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<td>Total</td>
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<td></td>
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</table>

*Weighted average $S_y* 0.18

The top layers A and B are excluded from the weighted average as they were not repeated at the lower depths.
Appendix 7. Detailed flowchart of the thesis structure. Eq. is equation, irr. is irrigation, Exp. is experimental, Rep. is representative, ret. is retention and measur. is measurement.
Acknowledgments

In this rapturous time, sensations are delighted inside but enough words do not come to express. It was just like yesterday, on the 21 June 2000 when I met Prof. Sayyed Ahang Kowsar at his office where we had a motivating discussion about what I can do as a researcher to improve the knowledge about the know-how on floodwater spreading. The dream that is emerged in that occasion is started to be realized in this dissertation: however, the way remained for the future seems quite long.

As the commencing step I would like to reminisce overwhelming hidden and apparent assistance from Allah that steered me to the right ways allowing to have the contribution to the advantageous knowledge to the human being. No doubt, crossing my road with the Prof. Kowsar is the most imperative case. His contribution to my scientific and spiritual life cannot be translated into the words.

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I had an opportunity to meet Prof. Luis S Pereira from the University of Lisbon, Portugal when he had a scientific visit to Iran. Our friendship continued through publishing an article based on my works in Australia. His incredible contribution as co-author had not been ever experienced before by me.

The reviewers either as the known members of the Examination Board of PhD defense or as the unknown referees of the submitted papers spent plenty of their valuable times, made constructive comments and provided useful suggestions which resulted in a noticeable improvement in the research and in the thesis manuscript.

Dr. Sayyed Ali Mohammad Cheraghi, a knowledgeable colleague in Iran, also had significant contribution in this study. Mr. Meisam Rezaei, a PhD student at the same department in Ghent University put his finger print on this research by spending his time for putting me through professional application of the Hydrus1D model. We also had beneficial discussions with Prof. Piet Seuntjens for running and analyzing the results of the model.

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I sincerely wish for all the hands which helped me to be in the place where I am now, the best rewards from our compassionate Allah.

Mojtaba Pakparvar

Gent, Belgium 27/October/2015
Curriculum vitae

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1995-2000: Head of research lab for arid land soils at RIFR.

2000-now: Researcher in the field of floodwater harvesting in the Fars Research Center for Agriculture and Natural Resources (FRCANR), Shiraz, Iran.

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International A1 journals


banks in the Arasbaran Protected Area of Iran and their significance for conservation management. Biological Conservation, Volume 109, Number 3, pp. 425-431(7)

Books


2. Pakparvar, M, Gabriels, D, Rahbar, E, Abtahi, S.M, Ahmadian, M, Dadrasi Sabzevar A, and Abdi, A. 2013. Monitoring of soil salinity in marginal drylands of Iran. P 131-159 in: (Maarten De Boever, Muhammed Khlosi, Nele Delbecque, Jan De Pue, Nick Ryken, Ann Verdoodt, Wim M. Cornelis, Donald Gabriels, eds), Desertification and land degradation processes and mitigation, UNESCO Chair of Eremology, Ghent University, Belgium.

Awards / Honors

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2. Award for the prominent researcher of the Fars Research Center for Agriculture and Natural Resources (FRCANR), year 2004.

3. Prominent research project award, Research Institute of Forests & Rangelands, Iran, year 2000.

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And 10 other national journal papers

International conferences

through vadose zone as influenced by a flooding event in the Gareh Bygone (saturated phase), 2nd International Conference on Hydrology & Groundwater Expo, 27-28 Aug. 2013, Raleigh NC, USA.


And 9 other intl. conference contributions

**National conferences (in Persian language)**


And 10 other national conference contributions

**Scientific reports (projects done)**

1. Evaluation of Capacity of Remotely Sensed Images on Detection of Gully Erosion Rate and its Pattern

2. Classification of Fars Province’s Sand Dune Systems, Using Morphology, Physico-chemical and Mineralogical Characteristics

3. Recognition of the ecological regions of Darab region.

4. Classification of Fars Province’s sand dune systems, using morphology, physico-chemical and mineralogical characteristics.

And 6 other scientific reports
Graduate theses supervisory

1. Heidarnejad, M. 2013. Trend of evapotranspiration in Yazd plain, an energy balance study by SEBS model. Yazd University, Iran.


3. Movasaghi, H.R. 2008. Study on applicability of visual and thermal bands of ASTER sensor in finding the potential of gypsum mines in West Kazeroon. Islamic Azad University, Shiraz, Iran.


And 10 other thesis supervisory

Memberships

Soil Science Society of Iran

International Soil Science Society

Others

Reviewer of scientific international and national journals