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1. INTRODUCTION

This Deliverable 12.1 is on the Water budget and climate change impact assessment of selected MARSOL case-study areas (Portugal, Greece, Israel and Spain). Complementary information on the water budget and on modelling is being presented in the specific case-study and/or Thematic Deliverables, e.g. for Portugal in Deliverable 4.2 on “South Portugal MARSOL demonstration sites characterisation” and Deliverable 8.1 on “DSS with integrated modeling capabilities”, with modelling examples from Italy, e.g. “first application of the FREEWAT preliminary scripts on the S. Alessio site”.

This Deliverable will be further extended and reviewed during the progress of the project as, e.g., Task 12.3: Climate change impact (Task Leader: as in D12.1 also LNEC) is due by month 17, to include a summary of the Water budget and climate change impact assessment of Demo Site 4: Llobregat River Infiltration Basins, Sant Vicenç Dels Horts, Catalonia, Spain; Demo Site 5: River Brenta Catchment, Vicenza, Italy; And, Demo Site 6: Serchio River Well Field, Tuscany, Italy.
2. DEMO SITE 1: LAVRION TECHNOLOGICAL & CULTURAL PARK, LAVRION, GREECE

2.1 CONCEPTUAL MODEL

The region of Lavreotiki was first mined for silver and lead at the late Neolithic period (4200-3100 B.C.) Mining and metallurgical activities ceased after Classical times, for hundreds of years, until 1865 when an Italian-French company founded a metallurgical factory, utilizing the ancient slugs. In 1873 silver and lead exploitation restarted: for a short period they were carried out by “The French Company of Lavrin Mines” and then by the Hellenic state, until the early 20th century.

The area (Figure 1) is characterized by a complicated geological structure with many controversies. The climate is a typical Mediterranean climate with dry and hot summer periods.

![Geographical locations and points of interest the Lavrion basin.](image)

The major feature is a large scale detachment fault (Marinos and Petrascheck, 1956), with an N-S general trend, present both to the eastern and western parts of the study area. That fault connects the two major geological units in the area: the Lower Tectonic Unit, (LTU), and the Upper Tectonic Unit, (UTU), which is also known in the literature as the “phyllitic nappe” (Figure 2).

In the alluvial plain there is a dense network of shallow wells extending in an area of about 5 km². These wells have been sampled for heavy metals and major ions and where the water level depth was also measured in May and August 2014. This network is completed by a cluster of deeper drills placed in the Lavrion Technological and Cultural Park (Figure 2).
Figure 2 - Geological map of the wider area of Lavrion, including the network of deep (more than 15 m) and shallow wells in the alluvial (~3-10 m deep).

The various formations present in the area have the following hydrogeological behavior:

- In the Quaternary deposits an unconfined porous aquifer is developed, widely exploited for agricultural purposes mainly by shallow wells. Those shallow wells get dry during the warmer period of the hydrological cycle (Stamatis et al., 2001). The water table in the area is not deep (1-5 m in most cases). The unconfined granular part of this coastal aquifer system is being recharged from percolating groundwater within the watershed of Thoricos ephemeral stream as well as from lateral inflows originated from the surrounding karstified marbles. The piezometric map shows distinct recharge axes (NE-SW) within the boundaries of the alluvial aquifer, which is an outcome of the hydraulic gradients suggesting lateral inflows from the upper marble unit (Figure 3). Finally, a discharge axes with a NW-SE trend is present in the alluvium. Field investigations show that there is also seawater intrusion in that formation.

- The karstic aquifer is fed in most places only by rainwater, with the high degree of karstification being an enhancing factor for deep percolation of water. Where the karstic aquifer is in direct contact with the alluvial aquifer, lateral contribution from the karst to the alluvial are expected. Apart from that, water from this aquifer is discharged to the sea through coastal and submarine springs, which also show high levels of electrical
conductivities. This aquifer was used for drinking water supply but this stopped due to high salinities found in the water.

- The major river in the area is coming from the north, discharging water from a large area (about 35 km$^2$). In its N-S path the valley is not very wide and the surrounding formation is schist, so all the rainwater in the valley is discharged by the river. The alluvial thickness in the valley is expected to be very small (up to 2 meters). The river is actually an ephemeral stream and it is dry in the summer. Before the river reaches the sea, it flows through a region where the bedrock is marble and recharge from the karstic aquifer is happening.

![Piezometric map of the alluvial area, including the discharge axes of the alluvium (NW-SE) and the lateral contribution from the marble (NE-SW).](image)

The completed and ongoing field activities are:

- Data collection from previous studies
- Survey of existing groundwater wells
  - Shallow wells within the alluvial formation
  - Deep wells within the marble layer (ONGOING)
- Geological mapping for model groundwater model boundaries definition
- Series of infiltration tests (double ring infiltrometer) within the alluvial aquifer
- Geophysical investigations:
• GPR campaigns: mapping of existing seawater intrusion
• Geo-electrical soundings for structures identification
• Geo-electrical soundings for identification of submarine groundwater discharge
• Groundwater sampling campaigns
  • In the alluvial aquifer
  • In the karst aquifer (ONGOING)

The wells survey that has been done resulted in identifying 66 shallow wells in the alluvium, along with 13 wells in the karstic system (up until now). The data gathered include locations, water table measurements, sampling for chemical analyses (major ions, heavy metals) and indication about which aquifer each well or drill exploits.

2.2 WATER BUDGET

The average annual precipitation in the Lavrion area is around 360 mm/year (Figure 4), although there are even drier periods.

![Figure 4 - Monthly and annual rainfall fluctuation for the period 1970-1996.](image)

Data is being prepared to run the daily sequential soil water balance model, BALSEQ_MOD. The description of this method can be found in Annex 1.
3. DEMO SITE 2: ALGARVE AND ALENTEJO, SOUTH PORTUGAL

3.1 INTRODUCTION TO DEMO SITE 2

The demo site 2: Algarve and Alentejo, South Portugal, is composed of three DEMO sites (Figure 5):

- PT1: Rio Seco and Campina de Faro aquifer system (Algarve).
- PT2: Querença-Silves limestone karstic aquifer system (Algarve).
- PT3: Melides aquifer, river and lagoon (Alentejo).

These Demo sites are thoroughly characterised in WP4.

Figure 5 - Location of the PT MARSol DEMO sites
3.2 CONCEPTUAL MODEL

The conceptual model gathers all the information required to understand the situation under demo. A comprehensive analysis of the hydrology of the watershed upstream each MAR site is provided, as well as of the underlying hydrogeology.

GIS layers of information are prepared, including the DRASTIC vulnerability to pollution index, parameters related to the unsaturated zone capacity for incorporating MAR water (e.g. depth to the water table), or the GABA-IFI index to define the most appropriate areas for MAR.

3.2.1 Large diameter infiltrating wells in Campina de Faro aquifer system

The conceptual model of the Campina de Faro aquifer system is detailed on Deliverable 4.2 from where this text was withdrawn. Figure 6 (Stigter, 2005) presents a hydrogeological section of Campina de Faro area. The oldest formations belong to the Jurassic gypsiferous material that locally outcrops near Faro and is related to diapiric activity.

According to Stigter (2005), the oldest aquifer system that occurs in the Campina de Faro area is the Cretaceous, formed by limestone layers separated by marls (Figure 6). They dip to the south (20°-30°) and crop out in the NW part of the area. The top of the sediments is found at a depth below 200 m, near the city of Faro (Stigter, 2000). According to Manupella (1992, in Stigter, 2005) the thickness of Cretaceous aquifer is larger than 1000 m. A grabben-like structure was formed at the end of the Cretaceous where Miocene limestones and, later, sands and marls were deposited in discordance (Silva, 1988, in Stigter, 2000).

Miocene fossil-rich sandy-limestone deposits constitute the second aquifer. It deeps to the East but, because of the presence of several N-S faults, a stepwise structure is present (Silva et al., 1986 and Silva, 1988, in Stigter, 2000). There are few outcrops of Miocene limestones in Campina de Faro, also because they are covered by fine sand deposited during Miocene. The depth of the top of Miocene formations varies between 3 and 25 m below surface and the presence of marls seems to be very irregular (Silva, 1988 in Stigter, 2000). The topography of the top of the Miocene is irregular. Despite Miocene outcrops are very few and small, Miocene formations thickness is very large. It increases from north to south and exceeds 200 m near the coast. According to Antunes and Pais (1987, in
INAG, 2000) a deformation might have affected the Miocene deposits and that could explain the apparent thicknesses, larger than the real ones. The carbonated fraction varies from 60% to 95% with a decreasing trend from the base to the top (Silva, 1988).

Covering the Miocene deposits, sands, clayey sandstones, gravels and conglomerates of the Plio-Quaternary are found (“Areias e Cascalheiras de Faro-Quarteira” formation), with a thickness very variable. They crop in the NW and also near the city of Faro. Stigter (2000) refers that its thickness varies between 8 and 50 m. According to Moura and Boski (1994, in INAG, 2000) this formation has a maximum thickness of 30 m. Silva (1998, in INAG, 2000) refers that the thickness of these deposits can reach, in some places, a thickness of 60 m.

The third aquifer system is formed by the fine sand of Miocene and also the Plio-Quaternary sand and gravels. This aquifer presents an average thickness of 50 m (Stigter, 2004). Despite being partly covered by Holocenic materials, this aquifer is still considered phreatic, because their thickness is often too small to give confined characteristics to the underlying aquifer. According to Silva (1988) the average thickness of this aquifer is 25 m, but reaching maximum values of 60 m and 65 m near Galvana and in Quinta do Lago, respectively.

Some authors refer the existence of a confining layer between the second and the third aquifers. According to Silva et al. (1986), that separation is made by several silty-clayey-sandy layers (Figure 7) with variable thickness and apparently with some lateral continuity. Nevertheless, one cannot exclude the possibility of some hydraulic connection in sectors where that confining layers are absent (INAG, 2000).

![Figure 7 – Campina de Faro hydrogeological model, presented in JK15 well log (Silva et al., 1986)](image-url)
Furthermore, in some situations there is a hydraulic connection artificially established due to new wells built within old large wells with the aim of extracting water from the Miocene and confined aquifer. This connection facilitates the confined aquifer contamination coming from the overlying phreatic aquifer.

According to Stigter (2005), the general direction of groundwater flow is N-S (cf. Figure 6). There are preferential flow paths formed by the N-S trending faults but the NW-SE trending fault acts like a barrier, which is indicated by the steeper hydraulic gradient of the water table in the north.

### 3.2.2 Rio Seco infiltration basins

Rio Seco MAR facilities are located in the Rio Seco water course. These facilities stand in the northern part of the Campina de Faro aquifer system and so only slightly are influenced by the hydrodynamics of this system. However MAR at this point will locally influence the hydrogeology of this system by enabling fresh water to renew polluted water existing in the system. This will mainly influence a north-south strip of the aquifer system along Rio Seco.

The general conceptual model of Campina de Faro aquifer system is provided in section 3.2.1. For the area of the hydrographic basin contributing with flow to the MAR facilities a general description is here provided. The area of the hydrograph basin of the infiltration basins shown in Figure 8 is 62.7 km$^2$. The hydrographic basin develops over five groundwater bodies identified also on Table 1.
Table 1 – Groundwater bodies intersected by the hydrographic basin above Rio Seco MAR facilities

<table>
<thead>
<tr>
<th>Groundwater body</th>
<th>Area (km$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>M03RH8 - Orla Meridional Indiferenciado</td>
<td>37.12</td>
</tr>
<tr>
<td>das Bacias das Ribeiras do Sotavento</td>
<td></td>
</tr>
<tr>
<td>M10 - S. Joao da Venda - Quelfes</td>
<td>8.60</td>
</tr>
<tr>
<td>M11 - Chao de Cevada - Quinta de Joao de Ourem</td>
<td>1.28</td>
</tr>
<tr>
<td>M12 - Campina de Faro</td>
<td>1.59</td>
</tr>
<tr>
<td>M8 - S. Bras de Alportel</td>
<td>14.06</td>
</tr>
</tbody>
</table>

As a general rule it is expected that precipitation falling inside the hydrographic basin will originate surface runoff to Rio Seco or will infiltrate in the soil, recharge the aquifer systems and eventually discharge in direction to the Rio Seco water course or its tributaries. This means that in a natural dynamic equilibrium surface-ground water system, all the water that recharges the aquifer system, if not used by plants in shallow systems will eventually be part of the surface flow that passes in the MAR facilities. However two different processes may inhibit this equilibrium, first natural groundwater flow may be in a direction outward of the hydrographic basin, and this could be mainly expected in the case of the presence of karstic aquifer systems with regional flow directions different from the surface ones, and secondly anthropogenic abstraction of groundwater may lower the value of the discharge groundwater to the surface medium. It must be stated that the opposite from the first situation may also occur, i.e. water that recharges the aquifer system outside the hydrographic basin and that flows and discharges into the river network of the hydrographic basin (inward flow).

The São Brás de Alportel aquifer system is unconfined to confined, karstic, 34.4 km$^2$ area, where according with Almeida et al. (2000) natural recharge may be concentrated in sinkholes, or diffuse in epikarst structures (lapias fields) that occur in some areas; this aquifer system is divided into independent blocks. Almeida et al. (2000) refer to an ephemeral resurgence in Rio Seco, which is an indication of groundwater flow discharging into the hydrographic basin. It will be assumed that the groundwater flow divides coincide with the hydrographic basin meaning that all the precipitation water that falls inside the hydrographic basin and infiltrates will not flow outward and that there is no water infiltrating outside the hydrographic basin that will flow inward.

The Chão da Cevada-Quinta de João de Ourém aquifer system is also unconfined to confined, karstic, 5.3 km$^2$ area (Almeida et al., 2000), where the natural replenishment seems to equilibrate the groundwater abstractions, which is demonstrated by the groundwater level behaviour. Accordingly with Almeida et al. (2000) it is not possible to define groundwater flow directions. The western edge of the aquifer system coincides with the Rio Seco river. To the east of the hydrographic basin, other water courses cross this aquifer system, also in a north-south direction. It is assumed that the water that infiltrates in the hydrographic basin will flow to the Rio Seco river.
The São João da Venda-Quelfes aquifer system is multilayer, with two sequences, the lower one mainly detrital and the upper one consisting of marly limestone (Almeida et al., 2000). The system extends over an area of 113 km² and its central part is crossed by the Rio Seco hydrographic basin. It is assumed that water that infiltrates in the hydrographic basin will flow to the Rio Seco river.

Finally, the “Orla Meridional Indiferenciado das Bacias das Ribeiras do Sotavento” groundwater body consists of different hydrogeological materials, with no significant hydrogeological importance that would have allowed any of them to be individualised as an aquifer system. This groundwater body extends over an area of 409 km², and is mainly composed of detrital and carbonate materials of Meso-Cenozoic age occurring in the western hydrographic basins of the Algarve. It is assumed that water that infiltrates in the hydrographic basin will flow towards the Rio Seco river.

3.2.3 Querença-Silves aquifer system

The conceptual model of Querença-Silves aquifer system is detailed on Deliverable 4.2 from where this synthesis text was withdrawn.

The Querença-Silves aquifer system (Figure 9) is the largest aquifer in Algarve, located in the center of the Algarve region, in south Portugal, a region characterized by a Mediterranean climate with dry and warm summers and cool wet winters. It is considered a karst aquifer formed by Jurassic (Lias-Dogger) carbonate sedimentary rocks covering an irregular area of 324 km² from the Arade River (at the west) to the village of Querença (at the East) (Monteiro et al., 2006 and Monteiro et al., 2007). The system is delimited south by the Algibre thrust, which is the main onshore thrust in the Algarve Basin, separating the Lower/Early and the Upper/Late Jurassic and to the north by the Triassic-Hettangian rocks (Terrinha, 1998). The Estômbar springs on the west limit of QS aquifer constitute the main discharge area of the system towards the Arade River, supporting several important groundwater dependent ecosystems.

Manuppella et al., (1993) presents a cross-section of the central Algarve region (Figure 10). This cross-section allows a synthetized visualization of the geometric relations of the Early/Lower Jurassic lithology which support the Querença-Silves aquifer system identified in blue, in Figure 10.
Figure 9 – Querença-Silves aquifer system’s geology. Underlined features are the ones identified within the aquifer limits. Almeida et al. (2000).

Figure 10 - Geometry of the carbonated rocks of early Jurassic which constitute the most important support of the aquifer system Querença-Silves (dark blue colour, to the left of the Algibre thrust). Adapted from Manuppella et al., (1993).
Accordingly to previous studies, the hydrogeological setting of the Querença-Silves karstic aquifer, has a complex compartmented structure, with two distinct domains separated by a fault: a western domain and an eastern domain. Its western domain has a well-developed karst, westward flow direction, with the main discharge areas along the Arade river, with particular relevance to Estômbar springs (westernmost point). Its eastern domain has more random flow directions, less regular piezometric surfaces (Figure 11) and a lower karst development. The tectonic activity of this region results in its widespread fracturing, defining a significant number of semi-independent aquifer blocks, with more or less constrained and restricted hydraulic links between them. Such hydraulic restrictions are more expressive in the eastern domain, because in the western domain the pervasive karstic network largely obliterates such tectonic setting (Mendonça and Almeida, 2003; Monteiro et al., 2006).

The Ribeiro Meirinho stream is located in the central-western area of Querença-Silves aquifer and its upper reaches are located outside the aquifer, in Serra Algarvia. The latter are Palaeozoic terrains, composed mainly of schist and graywakes, essentially impervious lithologies, being therefore, the main source of water for this stream until it reaches the Jurassic limestones, dolomites, dolomitic limestones and other, less important, calcareous formations composing the karst aquifer of Querença-Silves.

3.2.4 Querença-Silves infiltration basins
The areas draining to the location of the three infiltration basins are represented in Figure 12. The river basin of the WWTP of São Bartolomeu de Messines (area = 13.6 km²) is included in the river basin of Ribeiro Meirinho (total area = 57.6 km²). The river basin of Cerro do Bardo (29.4 km²) is west of the previous one. The Cerro do Bardo river basin is almost completely installed on the Querença-Silves aquifer system (unless a small part on the north that belongs to the “Orla Meridional indiferenciado das Bacias das ribeiras do Sotavento” groundwater body (GWB)). The WWTP of São Bartolomeu de Messines river basin is almost exclusively formed by the “Orla Meridional
indiferenciado das Bacias das ribeiras do Sotavento” GWB and, in a very small area of 2.0 km$^2$, by the “Zona Sul Portuguesa das Bacias das ribeiras do Sotavento” GWB. Concerning the Ribeiro Meirinho river basin, apart the area included in the WWTP of São Bartolomeu de Messines river basin, is almost exclusively developed on the Querença-Silves aquifer system.

The conceptual model of the Querença-Silves aquifer system is described in section 3.2.3, as well as a general characterisation of the “Orla Meridional indiferenciado das Bacias das ribeiras do Sotavento” GWB is presented in section 3.2.2. Concerning the “Zona Sul Portuguesa das Bacias das ribeiras do Sotavento”, this GWB has a total area of 293 km$^2$ and is comprised of Paleozoic geological formations that compose the part of the geostructural unit of the Portuguese South Zone located in the western hydrographic basins of the Algarve. It is a low productivity area mainly composed of schist and greywackes. In this last area it is also assumed that water that infiltrates in the hydrographic basin will flow towards the correspondent water course.

![Figure 12 – Hydrograph basin above selected MAR facilities in PT2 and related groundwater bodies](image)

### 3.3 WATER BUDGET

#### 3.3.1 Natural water budget

For the demo sites a natural water budget is provided. This includes a review of ancient studies plus a new characterisation specific of the demo sites. Water budget is carried out using the daily sequential soil water balance model BALSEQ_MOD.

Water budget of MARSOL demo sites where MAR techniques are to be implemented is a crucial issue for the success of the improvement of aquifer management.
3.3.1.1 **Campina de Faro**

Considering the aquifer area equal to 86 km$^2$, the average annual precipitation equal to 550 mm and a recharge rate (of the phreatic aquifer) between 15 and 20% of the precipitation, direct recharge has an approximate value of 10 hm$^3$/year (INAG, 2000) or 116 mm/year. Oliveira and Lobo-Ferreira (1994) calculated the potential recharge of hydrogeological system ALG-5, that includes also Campina de Faro aquifer, and obtained a value of 36.3 hm$^3$/year (143 mm/year).

A daily sequential water balance was developed during the GABARDINE EU project using the BALSEQ_MOD numerical model developed in LNEC. A description of this model can be found in Annex 1.

The period analysed ranged from 1981/10/01 to 1991/09/30. Daily precipitation was taken from the INAG 31K-02 – Quelfes rain gauge station (average precipitation in this period: 611 mm/year). Monthly reference evapotranspiration (ETo) was estimated in Ribeiras do Algarve Watershed Plan for the Tavira meteorological station, using the FAO Penman-Monteith method (average ETo 1235 mm/year). Information on soil parameters and land use was taken from the Soil Map of Portugal at the 1:50000 scale from IHERA and from 1:100000 scale 2000-Corine Land Cover.

Average aquifer recharge in The Campina de Faro aquifer system (Figure 13) was estimated in Lobo-Ferreira *et al* (2006) as 139 mm/year, 23% of the precipitation, despite, depending on the land use and the soil properties, the recharge distribution is highly variable. Lower values are in the order of 5 mm/year in the alluvium formations and larger values can go up to 380 mm/year in sandy outcrops.

---

**Figure 13 – Distribution of recharge in the Campina de Faro aquifer system (Lobo Ferreira *et al*, 2006)**
For instance Figure 14 shows the recharge distribution along the years in these types of soil/land use combinations.

Figure 14 – Yearly variation of the modelled parameters (including aquifer recharge) in the artificial recharge area in an alluvium soil (Soil = “Atac”) and in a regossoil (Soil = “Rgc”) (Lobo Ferreira et al., 2006)

3.3.1.2 Rio Seco hydrographic basin upstream the infiltration basins
A study for the estimation of water available for infiltration under natural conditions was developed using 10 year daily data from Rio Seco station (Figure 15).

Figure 15 - River flow cumulative curve and infiltration capacity of the IB (depending on the area of the IB), considering the infiltration capacity = 1.2 m/d
Estimation of the relation between infiltration capacity of the IB and river surface flow (Figure 15) determining the average number of days per year in which river flow is higher than the infiltration capacity of the IB. The value of 200 m² corresponds to the old existing IB and the value 400 m² corresponds to the actually existing infiltration basins capacity.

The BALSEQ_MOD model was also run for the area of the hydrographic basin upstream Rio Seco river infiltration basins in order to determine the available amount of water for the infiltration basins. The period of analysis is of 13 years, from 1/10/2001 until 30/9/2014. Precipitation was taken from the meteorological gauge station 31J/01C – São Brás de Alportel belonging to the Portuguese Environment Agency (data accessible from http://snirh.pt/index.php?idMain=2&idItem=1) that collected reliable data until 31/05/2009. Lacking data was estimated using other rain gauge stations, namely in first place 31J/04UG – Estói, also from the Portuguese Environment Agency, and, in second place, the meteorological station of Patacão, belonging to the Algarve Regional Directorate of Agriculture and Fisheries with data starting from 01-01-2006 (http://www.drapalg.min-agricultura.pt/ema/pat.htm). The relation of average precipitations in these stations was used to fill the gaps.

Daily reference evapotranspiration (ETo) was estimated using the FAO Penman-Monteith method with data of daily maximum and minimum temperature, daily average relative humidity or hourly relative humidity, daily solar radiation, and daily wind speed, also measured in the meteorological gauge station 31J/01C – São Brás de Alportel. Missing data was estimated by several procedures using measurements in the 30K/02C – Picota meteorological station, also from the Portuguese Environment Agency, and, in second place, using data registered in the above mentioned meteorological station of Patacão.

The procedures used to fill missing data are quite extensive and will be reported in a separate report.

Average yearly precipitation and average yearly reference evapotranspiration were estimated in São Brás de Alportel meteorological station as 638 mm/year and 1023 mm/year, with the monthly and yearly distributions shown in Figure 16 and Figure 17, respectively.
Figure 16 - Monthly rainfall and reference evapotranspiration (ETo) fluctuation for the period 10/2001-09/2014 in São Brás de Alportel meteorological station.

Figure 17 - Yearly rainfall and reference evapotranspiration (ETo) fluctuation for the period 10/2001-09/2014 in São Brás de Alportel meteorological station.

Information on soil parameters and land use was taken from the Soil Map of Portugal at the 1:50000 scale from IHERA and from 1:100000 scale 2006-Corine Land Cover (Caetano et al, 2009).

The results obtained by the BALSEQ_MOD model may be summarised as follows: average surface runoff = 360 mm/year; average groundwater recharge = 28 mm/year; average actual
evapotranspiration = 250 mm/year. The average monthly and yearly distribution of the values is represented in Figure 18 and Figure 19.

Figure 18 - Monthly surface runoff, natural recharge and actual evapotranspiration fluctuation for the period 10/2001-09/2014 in Rio Seco hydrographic basin upstream the infiltration basins.

Figure 19 - Yearly surface runoff, natural recharge and actual evapotranspiration fluctuation for the period 10/2001-09/2014 in Rio Seco hydrographic basin upstream the infiltration basins.

The spatial distribution of the surface runoff, recharge and actual evapotranspiration values can be observed in Figure 20, Figure 21 and Figure 22.
Figure 20 – Average surface runoff distribution in Rio Seco hydrographic basin upstream the infiltration basins

Figure 21 – Average groundwater recharge distribution in Rio Seco hydrographic basin upstream the infiltration basins
Figure 22 – Average actual evapotranspiration distribution in Rio Seco hydrographic basin upstream the infiltration basins

3.3.1.3 Querença-Silves

An extensive literature review allowed the identification of “three generations” for the estimation of the water balance of the Querença-Silves nowadays available as follows:

- Almeida (1985) and Almeida et al. (2000) estimated a total recharge of $70 \pm 17 \, \text{hm}^3\cdot\text{yr}^{-1}$, using the Kessler method (1965) in the areas where carbonated rocks outcrops are present (in which the average recharge varies between 40 and 60% of precipitation) and a sequential water balance in the soil in the areas where carbonate rocks are covered by soils or sedimentary deposits (in which recharge varies between 5 and 18% with an average of 10%).

- Vieira and Monteiro (2003) refined the previous balance with the infiltration values determined for the covered and outcropping areas of the carbonate rocks proposed by Almeida (1985) and Almeida et al., (2000). However in this case it was possible to use a new generation of geological maps in which the percentage of covered and uncovered areas of the aquifer were much more reliable. The diminution of that source of uncertainty allowed to estimate a value in the range of the highest values proposed in the previously mentioned balance. In that case a total annual average recharge of $93.4 \, \text{hm}^3\cdot\text{yr}^{-1}$ was estimated.
- Oliveira et al. (2008) estimated a 100 hm$^3$.yr$^{-1}$ recharge with the sequential daily water balance model BALSEQ_MOD (Annex 1), later updated by Oliveira et al. (2011) to 94 hm$^3$.yr$^{-1}$ (Figure 23).

Figure 23 - Average Querença-Silves aquifer recharge (period 1941-1991)

Due to its karstic properties, there is a strong relationship between the aquifer and the streams with some influent sections that can significantly contribute to its recharge (Monteiro et al., 2006; Reis et al., 2007, Salvador et al. 2012). This is the case of Ribeiro Meirinho, which undergoes a sharp reduction of the flow rate when it reaches the calcareous formations, having several sinks in its bed. It is estimated that besides direct recharge, an extra amount of 62x 10$^6$ m$^3$/year, originating from surface flow produced on the drainage area outside the aquifer, infiltrates when the rivers reach the aquifer system (Oliveira and Oliveira, 2012).

Regarding aquifer abstractions, Nunes et al. (2006) estimated a mean annual withdrawal for irrigation of 31 hm$^3$, and Stigter et al. (2009) estimated a 10% of mean aquifer recharge is abstracted for urban water supply.

3.3.2 Water available for MAR and sources of water

Water available for artificial recharge uses data computed in previous section linked to each possible source of water.

In the initial phase of testing/demonstrating the MAR facilities, alternative sources of water are used such as water provided by fireman trucks tanks or groundwater pumped from different aquifers to
the infiltration facilities. After these initial tests, the following available sources of water are considered:

- For the Rio Seco case study possible sources of water are river flow and greenhouse rain harvesting.
- For the large wells to be used as infiltration wells the source of water is greenhouse rain harvesting.
- For the Cerro do Bardo case study possible sources of water are river flow and rejected water from Water treatment plants or from surface storage dams.

### 3.3.2.1 Greenhouse rain harvesting

One possible important source of water for MAR in the Campina de Faro aquifer system (CFAS) consists in harvested rainwater from greenhouses (Figure 24) due to the large surface area occupied by these infrastructures. This potential source of water could in some cases be redirected to large diameter wells (Figure 25), which, at CFAS, present a high potential for well water recharge (as determined during large well injection tests described in Deliverable 4.2).

![Figure 24 - Intensive greenhouse agricultural activity in the Campina de Faro aquifer system and traditional system of drains for collection of rainwater](image)

![Figure 25 – Traditional large diameter wells potentially suited for injection of harvested water](image)

In order to estimate the potential rainwater that can be harvested from greenhouses the average distribution of annual and monthly value of rainwater was calculated and overlapped with the location of greenhouses at CFAS.
Average annual rainfall estimates were based on a 32 year average rainfall distribution model (from 1959/60 to 1990/91) consisting in a 1 km² resolution matrix developed by Nicolau (2002). Based on this distribution model, average annual rainfall on the CFAS was estimated as 570 mm with the spatial distribution shown in Figure 26.

Figure 26 - Top: 32 year average annual rainfall spatial distribution on Campina de Faro aquifer system (CFAS) (based on Nicolau, 2002) and greenhouse’s locations (based on APA-ARH Algarve, unpublished). Bottom left: 32 year average monthly and annual rainfall distribution values for the CFAS (based on Nicolau, 2002).

The location of greenhouses and their surface area estimation was based on the survey of the land use, using year 2007 aerial photos, developed by the Algarve Water Basin Regional Administration of the Portuguese Environment Agency (APA-ARH Algarve, unpublished). Based on this survey, the total surface area occupied by greenhouses which are totally within or intercept the CF aquifer is estimated as 2.74 km² with the spatial distribution identified in Figure 26.

Based on the estimates of the average annual and monthly rainfall distribution, the greenhouse’s locations and their surface area (Figure 26), the average annual rainfall that could potentially be
intercepted by greenhouses in the CFAS is calculated as 1.63 hm$^3$/yr. The 32 year monthly and annual averages of rainfall intercepted by the greenhouses are presented in Table 2. It is unlikely that the totality of these amounts can be harvested and used for MAR due to the lack of appropriate greenhouse infrastructures, conduits or close location to large diameter wells. Nonetheless they should be seen as the average maximum potential water available for future MAR solutions.

Table 2 - Estimated average monthly and annual potential rainfall harvested from greenhouses.

<table>
<thead>
<tr>
<th>Month</th>
<th>Intercepted rainfall (hm$^3$)</th>
<th>Month</th>
<th>Intercepted rainfall (hm$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JAN</td>
<td>0.252</td>
<td>JUL</td>
<td>0.00273</td>
</tr>
<tr>
<td>FEB</td>
<td>0.235</td>
<td>AUG</td>
<td>0.00823</td>
</tr>
<tr>
<td>MAR</td>
<td>0.139</td>
<td>SEP</td>
<td>0.0371</td>
</tr>
<tr>
<td>APR</td>
<td>0.112</td>
<td>OCT</td>
<td>0.191</td>
</tr>
<tr>
<td>MAY</td>
<td>0.0625</td>
<td>NOV</td>
<td>0.263</td>
</tr>
<tr>
<td>JUN</td>
<td>0.0275</td>
<td>DEC</td>
<td>0.292</td>
</tr>
<tr>
<td>Annual</td>
<td></td>
<td></td>
<td>1.63</td>
</tr>
</tbody>
</table>

3.4 CLIMATE CHANGE

An analysis of the impact of climate change in the general water budget of the areas under study is performed using projections of the precipitation and temperature series.

3.4.1 Emission scenarios and Predicted climate change

Two different studies are presented.

For the PT2 – Querença-Silves aquifer system, Stigter et al. (2009, 2014) present a study regarding groundwater flow simulation of future scenarios using the ENSEMBLES projections (A1b scenarios) for recharge.

Oliveira et al. (2012) in the framework of the ProWaterMan project studied the PT3 – Melides and also the PT2 – Querença-Silves sites, considering the scenarios of the SIAM II project in Portugal (Santos and Miranda, 2006). The three scenarios used to project changes for year 2100 were the following:

- **Scenario IS92a** – “business as usual” as in 1992.
- **Scenario SRES A2** – heterogeneous world, fragmented socio-economic development, population grows.
- **Scenario SRES B2** – search of local solutions for social, economic and environmental problems, population grows but less than in scenario A2.
The average variation of precipitation and temperature projected by these scenarios are presented in Table 3. This Table also includes PT1 – Rio Seco demo site area projected changes for the same three scenarios. With these projected changes, recent series of precipitation and reference evapotranspiration are modified and used to run the BALSEQ_MOD model.

Table 3 – Projected precipitation and temperature variations in different climate change scenarios

<table>
<thead>
<tr>
<th>Demo site</th>
<th>Process</th>
<th>Scenario SIAM: HadRM2, IS92a</th>
<th>Scenario SIAM HadRM3, SRES A2</th>
<th>Scenario SIAM HadRM3, SRES B2</th>
</tr>
</thead>
<tbody>
<tr>
<td>PT1 – Rio Seco</td>
<td>Monthly Precipitation</td>
<td>Winter: +45%</td>
<td>Winter: -35%</td>
<td>Winter: -20%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Spring: -25%</td>
<td>Spring: -55%</td>
<td>Spring: -25%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Summer: -77%</td>
<td>Summer: -50%</td>
<td>Summer: -15%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Autumn: -55%</td>
<td>Autumn: -45%</td>
<td>Autumn: -35%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Summer: +6 ºC</td>
<td>Summer: +3.75 ºC</td>
<td>Summer: +3 ºC</td>
</tr>
<tr>
<td></td>
<td>Monthly minimum Temperature</td>
<td>Winter: +5.25 ºC</td>
<td>Winter: +2.75 ºC</td>
<td>Winter: +1.75 ºC</td>
</tr>
<tr>
<td>PT2 - Querença-Silves</td>
<td>Monthly Precipitation</td>
<td>Winter: +40%</td>
<td>Winter: [-30%;-40%]</td>
<td>Winter: [-20%;-30%]</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Spring: [-20%;]-30%</td>
<td>Spring: -50%</td>
<td>Spring: [-20%;-30%]</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Summer: [-70%;-85%]</td>
<td>Summer: -65%</td>
<td>Summer: [-30%;-40%]</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Autumn: [-50%;-60%]</td>
<td>Autumn: -40%</td>
<td>Autumn: [-20%;-30%]</td>
</tr>
<tr>
<td></td>
<td>Monthly maximum Temperature</td>
<td>Winter: +4.25 ºC; Spring: +4.75 ºC; Summer: +5.75 ºC; Autumn: +5.25 ºC</td>
<td>Winter: +3 ºC; Spring: +3.5 ºC; Summer: +3.75 ºC; Autumn: +4 ºC</td>
<td>Winter: +2 ºC; Spring: +2 ºC; Summer: +2.5 ºC; Autumn: +2 ºC</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Winter: +4.25 ºC; Spring: +4.75 ºC; Summer: +5.75 ºC; Autumn: +5.25 ºC</td>
<td>Winter: +3.5 ºC; Spring: +3 ºC; Summer: +3 ºC; Autumn: +3 ºC</td>
<td>Winter: +2 ºC; Spring: +2 ºC; Summer: +2.5 ºC; Autumn: +2 ºC</td>
</tr>
<tr>
<td></td>
<td>Monthly minimum Temperature</td>
<td>Winter: +4.75 ºC; Spring: +4.5 ºC; Summer: +5.25 ºC; Autumn: +5.25 ºC</td>
<td>Winter: +3.5 ºC; Spring: +3 ºC; Summer: +3 ºC; Autumn: +3 ºC</td>
<td>Winter: +2 ºC; Spring: +2 ºC; Summer: +2.5 ºC; Autumn: +2 ºC</td>
</tr>
<tr>
<td>PT3 – Melides site</td>
<td>Monthly Precipitation</td>
<td>Winter: [+40%;+50%]</td>
<td>Winter: [-20%;-30%]</td>
<td>Winter: [-20%;-30%]</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Spring: -20%</td>
<td>Spring: -40%</td>
<td>Spring: [-20%;-30%]</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Summer: [-70%;-85%]</td>
<td>Summer: -65%</td>
<td>Summer: [-30%;-40%]</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Autumn: [-50%;-60%]</td>
<td>Autumn: -40%</td>
<td>Autumn: [-10%;-20%]</td>
</tr>
<tr>
<td></td>
<td>Monthly maximum Temperature</td>
<td>Winter: +4.25 ºC; Spring: +5.25 ºC; Summer: +7 ºC; Autumn: +6.75 ºC</td>
<td>Winter: +3 ºC; Spring: +3.5 ºC; Summer: +3.75 ºC; Autumn: +4 ºC</td>
<td>Winter: +2 ºC; Spring: +2.5 ºC; Summer: +3 ºC; Autumn: +3 ºC</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Winter: +4.25 ºC; Spring: +5.25 ºC; Summer: +7 ºC; Autumn: +6.75 ºC</td>
<td>Winter: +3.5 ºC; Spring: +3.5 ºC; Summer: +3.75 ºC; Autumn: +4 ºC</td>
<td>Winter: +2 ºC; Spring: +2.5 ºC; Summer: +3 ºC; Autumn: +3 ºC</td>
</tr>
<tr>
<td></td>
<td>Monthly minimum Temperature</td>
<td>Winter: +5 ºC; Spring: +4.75 ºC; Summer: +6 ºC; Autumn: +5.75 ºC</td>
<td>Winter: +3.25 ºC; Spring: +3 ºC; Summer: +3 ºC; Autumn: +3 ºC</td>
<td>Winter: +1.75 ºC; Spring: +2 ºC; Summer: +2.5 ºC; Autumn: +2 ºC</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Winter: +5 ºC; Spring: +4.75 ºC; Summer: +6 ºC; Autumn: +5.75 ºC</td>
<td>Winter: +3.25 ºC; Spring: +3 ºC; Summer: +3 ºC; Autumn: +3 ºC</td>
<td>Winter: +1.75 ºC; Spring: +2 ºC; Summer: +2.5 ºC; Autumn: +2 ºC</td>
</tr>
</tbody>
</table>

Adapted from Santos and Miranda (2006, PT3 and PT2 in Oliveira et al., 2012)
3.4.2 Climate change impacts

For the study area DEMO Site 2 – PT2 Querença-Silves (QS), regarding groundwater flow simulation of future scenarios using the ENSEMBLES projections (A1b scenarios) for recharge, the summarized achieved conclusions presented in the study of Stigter et al. (2009, 2014) were:

- Period 2020–2050: Changes in recharge, particularly due to a reduction in autumn rainfall resulting in a longer dry period. More frequent droughts are predicted at the QS aquifer.
- Toward the end of the century (2069–2099): results indicate a significant decrease (mean 25 %) in recharge at QS aquifer, with an high decrease in absolute terms (mean 134 mm/year).
- Scenario modelling of groundwater flow shows its response to the predicted decreases in recharge and increases in pumping rates, with strongly reduced outflow into the coastal wetlands, whereas changes due to sea level rise are negligible.

For the case of the study developed by Oliveira et al. (2012) for PT2 – Querença-Silves and PT3 – Melides sites, and now developed for the PT1 – Rio Seco site, the future precipitation series, temperature series and reference evapotranspiration series were estimated using the methodologies presented in Annex 2.

The results obtained for the PT2 – Querença-Silves aquifer system are shown in Figure 27 for the three emission scenarios. Figure 28 shows the equivalent results for the PT3 – Melides hydrographic basin.

Figure 27 – Climate change impact in the water cycle in PT2 – Querença-Silves aquifer system (source: Oliveira et al., 2012)
Table 4 – Average yearly values of natural groundwater recharge and surface runoff obtained by the actual 1979-2009 time series and estimated values under the different climate change scenarios

Table 4 summarises the results obtained for the three case study areas. It is possible to see that natural groundwater recharge may be as low as 40% of the actual recharge and that surface runoff may be as low as 51% of the actual surface runoff under scenario A2. Despite the scenario B2 is not so unfavourable it still accounts for large reductions on the water available in the water cycle. These results are a useful starting point regarding the role of MAR in this aquifer system, taking into account the future trends of climate change patterns expected in this area.

![Figure 28 – Climate change impact in the water cycle in PT3 – Melide hydrographic basin (source: Oliveira et al., 2012)](image)

Table 4 – Average yearly values of natural groundwater recharge and surface runoff obtained by the actual 1979-2009 time series and estimated values under the different climate change scenarios

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>PT1 - Rio Seco</td>
<td>GW Recharge</td>
<td>294</td>
<td>245 (84%)</td>
<td>136 (46%)</td>
<td>186 (63%)</td>
</tr>
<tr>
<td></td>
<td>Surface runoff</td>
<td>115</td>
<td>100 (88%)</td>
<td>59 (51%)</td>
<td>79 (69%)</td>
</tr>
<tr>
<td>PT2 - Querença-Silves</td>
<td>GW Recharge</td>
<td>119</td>
<td>88 (74%)</td>
<td>47 (40%)</td>
<td>64 (54%)</td>
</tr>
<tr>
<td></td>
<td>Surface runoff</td>
<td>199</td>
<td>179 (90%)</td>
<td>114 (58%)</td>
<td>142 (72%)</td>
</tr>
</tbody>
</table>
4. DEMO SITE 3: LOS ARENALES AQUIFER, CASTILE AND LEÓN, SPAIN

4.1 WATER BALANCE (AS THE AVAILABLE MODEL WAS PERFORMED)

4.1.1 Santiuste basin

Table 5 – Datos disponibles para la resolución de la ecuación general del balance hídrico. Cubeta de Santiuste. Año hidrológico 2002/03.

<table>
<thead>
<tr>
<th>ENTRADAS</th>
<th>Datos disponibles para la resolución de la ecuación general del balance hídrico. Cubeta de Santiuste. Año hidrológico 2002/03.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Infiltración directa del agua de lluvia</td>
<td>De 2,825 a 3,287 hm$^3$/año</td>
</tr>
<tr>
<td>Infiltración a través de la escorrentía superficial</td>
<td></td>
</tr>
<tr>
<td>Importaciones</td>
<td>0,18</td>
</tr>
<tr>
<td>Entradas por arroyos</td>
<td>0,056</td>
</tr>
<tr>
<td>Retornos de riegos</td>
<td>0,26 a 0,30 (0,28)</td>
</tr>
</tbody>
</table>

| RECARGAS | Recarga artificial del acuífero | 0,933 a 1,344 hm$^3$/año |
| SALIDAS | Fluro subterráneo y manantiales entrantes | 0,061 hm$^3$/año |
| Regadíos | Destino del agua de las captaciones: |
| Abastecimiento urbano | (Importaciones) |
| Abastecimiento industrial | 0,165 hm$^3$/año |
| Abastecimiento ganadero | 0,154 hm$^3$/año |
| Salidas por cauces superficiales. | 0* |
| Percolación hacia el acuífero profundo. | 0,762 a 1,107 (1,0) hm$^3$/año |

| DESCARGAS | Descargas Manantiales y rezumes | De 0,241 (0,3) a 0,677 hm$^3$/año |
| Drenaje subterráneo y subterráneo hacia el norte de la Cubeta | |
| TOTAL | E (de 3,382 a 3,817) + (AR) - S (De 4,010 a 5,108) = ΔV /AR = volumen de entradas por recarga artificial. |
| De - 0,628 a -1,291 hm$^3$/año (promedio = -0,959 hm$^3$/año) | |
| 3,422- 4,307= 0,885 hm$^3$ |

4.1.2 Carracillo district

<table>
<thead>
<tr>
<th>Muestra</th>
<th>Zona Acuífero</th>
<th>Muestra</th>
<th>Zona Acuífero</th>
<th>Muestra</th>
<th>Zona Acuífero</th>
</tr>
</thead>
<tbody>
<tr>
<td>Muestra</td>
<td>Profundidad (metros)</td>
<td>Limites Breding (metros/día)</td>
<td>Permeabilidad ensayos laboratorio (metros/día)</td>
<td>Permeabilidad ensayos de bombeo (metros/día)</td>
<td></td>
</tr>
<tr>
<td>----------</td>
<td>--------------</td>
<td>----------</td>
<td>--------------</td>
<td>----------</td>
<td>--------------</td>
</tr>
<tr>
<td>limacín, superficial.</td>
<td>168/197</td>
<td>8 – 43</td>
<td>31.1</td>
<td>25.35 (Bombeo n° 2 y 12)</td>
<td></td>
</tr>
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<td>4.3 – 86</td>
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<td>197 A</td>
<td>4.3 – 8</td>
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<tr>
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<td>60/65/35A</td>
<td>8 – 43</td>
<td>13.8</td>
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<td>8.86</td>
<td>28.5</td>
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</table>
**Volumen Máximo Gestionable**: 8,24 hm³.

**Volumen Mínimo Gestionable**: 0,00 hm³ a 4,86 hm³.

**Volumen Medio Gestionable**: 2,36 hm³ - 3,64 hm³.

NB: Information has been submitted by Tragsa
5. DEMO SITE 7: MENASHE INFILTRATION BASIN, HADERA, ISRAEL

5.1 INTRODUCTION

The Menashe facility is being used over 40 years for groundwater enrichment of the coastal aquifer using collected surface water. During the winter, as the major streams draining the southern Mt. Carmel slopes and the Menashe hills start to flow, part of the water is diverted by a system of dams and canals into a main conveyance canal. Driven further 16 Km west by gravitation the water is introduced into infiltration ponds for coastal aquifer enrichment, and later pumped by production wells (Figure 29). In addition to runoff infiltration, desalinated water originating from the nearby Hadera desalination plant is periodically introduced into the aquifer according to maintenance requirements. The physical and chemical processes involved in this type of aquifer enrichment are the essence of the demo site activity and research.

The mid-scale model deals with saturated flow within the coastal aquifer, underlying the Menashe infiltration facility area. It will correspond to both the larger watershed scale and the smaller, infiltration pond scale future models.

Figure 29 - Menashe infiltration site and catchment area – Location map and setting
5.2 CONCEPTUAL MODEL

The Pleistocene coastal aquifer stretches along most of the coast of Israel. It reaches some 100 m depth along the coast just west of the Menashe site, and gradually wedges out towards east till it vanishes about 12 Km from the coast. The aquifer consists of Pleistocene aged, coastal environment rocks series, dominated by calcareous sandstone and interbedded by conglomerates, silt and clay layers. At places thicker marine clay layers divide the aquifer into sub horizontal sections. This division is more dominant towards the coast and affects the western parts of the Menashe site. The entire aquifer complex is underlain in most areas by a thick, practically impermeable Neogene rock unit (Saqiye Group) and is separated from the deeper Cretaceous aquifer of the Judea Group (Figure 30). The aquifer is open towards west to the sea and is bounded by the fresh - seawater interface, allowing some freshwater seepage offshore.
Figure 30 - Geological setting of the Menashe site and its catchment area

The geological data processed from well logs, geological and structural maps serve as the basis for the conceptual model, constructed via the GMS software package. The variety of rock types is represented by four types of materials. Following the marine clay structural maps, low permeability cells divide in places the western part of the model into sections. These layers are imbedded into the otherwise geostatistical based model computed by the T-PROGS software (Figure 31). The output of the T-PROGS is conditioned to the borehole data, the computed materials proportions and the material transition trends observed in the borehole data. The final horizontal model borders are still to be set according to further water level data.
5.3 WATER BUDGET

The modelled area water balance is influenced by artificial enrichment, production wells activity, direct rainfall, subsurface inflow (mainly from the east) and seepage to the sea. While the average annual artificial recharge and pumping amounts are controlled (these are very roughly similar and equal ~ 12 MCM), the other parameters are yet to be computed by the model, following its final border definition and calibration.
5.4 WATERSHED MODEL WATER BUDGET AND CLIMATE CHANGE

5.4.1 Watershed Scale Model

For the quantification of the water budget of all the components of the hydrologic cycle, coupled surface water – groundwater model of the Menashe streams watershed is being developed. The model includes a surface water model of the four major streams of the Menashe hills, which will be iteratively coupled to an unsaturated/saturated zone model of a chalk aquitard connected to the coastal sandstone aquifer (Figure 32).

Aim of the simulation is to reproduce the quantities of the hydrologic components on the basis of climate and stream discharge data of the past 50 years and then further use the...
developed model to simulate possible future climate scenarios and to evaluate the impact of a climatic change on the water budget.

5.4.2 Natural Water Budget
The current long-term water budget of the watershed is an average precipitation of about 600 mm per year, which occur between September to May, with a precipitation maximum in January and February. The potential evapotranspiration is about 1600 mm year, actual evapotranspiration is estimated to be about half of the precipitation amount and the average stream discharge is about 100 mm year which occurs from December to June. Precipitation as well as stream discharge are highly variable from year to year. The annual precipitation varies between below 400 mm and above 1000 mm, the annual discharge varies between 0 to 30 million m³ (Figure 33).

![Figure 33. Annual precipitation sum in mm of the Regavim station (1965/66 - 2012/13) of the Israel Meteorological Service and monthly stream discharge at the infiltration basin inlet in Mio. m³ (1966/67 – 2012/13) from MEKOROT Company.](image)

5.4.3 Climate Change
Possible future climate scenarios will be based on downscaled global climate models and regional climate models based on IPCC scenarios.

For example, Alpert et al. (2008) evaluated a regional climate model of the Eastern Mediterranean, which supports the large-scale predictions for the entire Mediterranean to
have rainfall reductions up to 35% and an average temperature rise of 3–5 °C by 2071–2100. A regional study in northern Israel supports the tendency to a more extreme climate, of both wetter and drier years. The results further suggest a significant factor of increase in the number of the heavy rain days over the Jordan River basin in Israel, which in this region would lead to an average rainfall increase of about 10%.

Black (2009) calculated a reduction in winter rain in the area of Jordan and Israel by the end of the 21st century, which will be reflected in reductions of the frequency and duration of rainy events and hence the number of rainy days.
6. CONCLUSIONS

As mentioned in the DoW, regarding WP12, “For selected sites, strategies for selecting positions for MAR facilities will be presented and water budgets on a watershed scale elaborated to calculate water availability for MAR, and predictions on the influence of future climatic changes will be made. These sites will serve as reference for modelling strategies where the MAR installations are included into a more generalized water budget approach to allow long term predictions on MAR efficiency and economic feasibility. ... The main tasks of this work package are: Task 12.1: Methods evaluation (Task Leader: LNEC): Literature review on potential and currently applied modelling approaches for MAR sites and the evaluation of weaknesses and strengths (concluded) and Task 12.2: Water budget and conceptual modelling (Task Leader: LNEC): For selected sites, GIS layers of information for conceptual modelling will be prepared. Achievements for selected sites reported in this Deliverable 12.1.

As mentioned in the Introduction “Complementary information on the water budget and on modelling is being presented in the specific case-study and/or Thematic Deliverables, e.g. for Portugal in Deliverable 4.2 on “South Portugal MARSOL demonstration sites characterisation” and Deliverable 8.1 on “DSS with integrated modeling capabilities”, with modelling examples from Italy, e.g. “first application of the FREEWAT preliminary scripts on the S. Alessio site. This Deliverable will be further extended and reviewed during the progress of the project as, e.g., Task 12.3: Climate change impact (Task Leader: as in D12.1 also LNEC) is due by month 17, to include a summary of the Water budget and climate change impact assessment of Demo Site 4: Llobregat River Infiltration Basins, Sant Vicenç Dels Horts, Catalonia, Spain; Demo Site 5: River Brenta Catchment, Vicenza, Italy; And, Demo Site 6: Serchio River Well Field, Tuscany, Italy.”

Three Workshops have been organized to facilitate the work reported: (1) a preliminary Workshop during the Kick-off meeting, held in TUDarmstadt (Germany), January 2014; (2) a Workshop on Modelling organized by LNEC in Lisbon (Portugal), July 14th and 15th 2014, with the following objectives (a) To present the current state of the art approaches for the modelling of MAR sites; (b) Definition of appropriate modelling approaches for MAR sites, and, (c) Preparations of guidelines for the selection of appropriate MAR modelling schemes; and, (3) a “Progress Report” Modelling Workshop during the 1st Year MARSOL Meeting held in Tel-Aviv (Israel), December 2014.

7. REFERENCES


8. ANNEX 1 - SOIL DAILY SEQUENTIAL WATER BALANCE

8.1 INTRODUCTION TO ANNEX 1

For the conceptual case of an area where there is no artificial recharge, no surface flow entering the area, and the groundwater level is always below the soil zone, the water balance equation for the soil of that area can be expressed by (Fig. 1):

\[ P - RET - \Delta A_l - Sr - Dp = \varepsilon \]  

where \( P \) is the precipitation, \( RET \) is the effective evapotranspiration, \( \Delta A_l \) is the variation (final - initial) of the water stored in the soil, \( Sr \) is surface runoff, \( Dp \) is deep percolation and \( \varepsilon \) is the calculation error of the balance.

\[ P = \text{precipitation} \]
\[ Sr = \text{surface runoff} \]
\[ Is = \text{surface infiltration} \]
\[ \Delta A_l = \text{difference of water stored in the soil in the end of the day and in the beginning of that day} \]
\[ RET = \text{effective evapotranspiration} \]
\[ Dp = \text{deep percolation} \]
\[ R = \text{recharge} \]
\[ D = \text{groundwater discharge} \]

Fig. 1 – Soil water balance of an area with no discharge of groundwater and no surface flow entering in the system.

The sequential mass balance approach intends to measure or estimate and compute \( P \), \( RET \), \( Sr \) and \( \Delta A_l \) parameters, computing \( Dp \) by solving Eq. 1 considering \( \varepsilon = 0 \). The sequential water balance is carried out in a determined time step, for instance the daily time step.

Recharge (\( R \)) is then assumed to be equal to \( Dp \):

\[ R = Dp = P - RET - \Delta A_l - Sr \]  

Eq. 2

The soil daily sequential water balance method is a good method to forecast differences on total recharge in response to changing daily precipitation pattern. Moreover as a general characteristic of the method it allows for the determination of seasonal recharge. However it must be taken into account that the presented method provides a value of the water available for deep percolation, and that this deep percolation will take some time to reach the aquifer.
8.2 THE BALSEQ NUMERICAL MODEL

A soil daily sequential water balance methodology was implemented in BALSEQ numerical model (Lobo Ferreira, 1981; Lobo Ferreira & Delgado Rodrigues, 1988). Fig. 2 shows the flowchart of the BALSEQ model. In this model the runoff curve number ($NC$) that depends on soil permeability and on land use, is used in the process of estimating surface runoff. $NC$ values vary between 0 (corresponds to the area with very high permeability, where all water infiltrates into the soil), and 100 (corresponds to a completely impermeable zone).

The effective evapotranspiration is calculated using the potential evapotranspiration (the evapotranspiration that would occur if the water available in the soil was not a limiting factor) and the amount of water available in the soil. This water available in the soil is calculated by a sequential water balance that daily updates the water stored in the soil.

The computation of deep percolation depends on the maximum amount of water available in the soil for evapotranspiration ($AGUT$):

$$AGUT = (sr - wp) \cdot rd$$  \hspace{1cm} Eq. 3

in which $sr$ is the specific retention (or field capacity), $wp$ is the wilting point and $rd$ is the depth of the plant roots. If after the process of evapotranspiration the water stored in the soil is above the $AGUT$ value, the water in excess to $AGUT$ becomes deep percolation.
\( P = \text{precipitation} \)

\( \text{PET} = \text{potential evapotranspiration} \)

\( \text{NC} = \text{runoff curve number} \)

\( \text{Sr} = \text{surface runoff} \)

\( \text{Is} = \text{surface infiltration} \)

\( \text{Al} = \text{water stored in the soil in the end of the day} \)

\( \text{HI} = \text{water stored in the soil along the day} \)

\( \text{RET} = \text{evapotranspiration} \)

\( \text{AGUT} = \text{maximum amount of water available for evapotranspiration} \)

\( \text{Dp} = \text{deep percolation} \)

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**Fig. 2 – Chart flow of BALSEQ model for daily sequential water balance in the soil**

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**8.3 THE BALSEQ_MOD NUMERICAL MODEL**

BALSEQ numerical model has been subject to changes and new methods have been implemented to calculate surface infiltration, effective evapotranspiration and deep percolation. These methods, developed in Oliveira (2004) have all been included in the BALSEQ_MOD numerical model and are presented in the next sections.

**8.3.1 Computation of surface infiltration**

This procedure was developed and presented in Oliveira (2004, 2007), using the results of the application of the Philip surface infiltration model, to a set of situations that could be representative of the different infiltration conditions (namely soil texture, daily precipitation, precipitation distribution and initial soil moisture).
The procedure computes surface infiltration (Is) using the following formula:

\[
Is = \begin{cases} 
   P & \text{if } P \leq P_{\text{lim}} \\
   a.P + b & \text{if } P > P_{\text{lim}} 
\end{cases}
\]  \hspace{1cm} \text{Eq. 4}

where \( P \) is precipitation; \( P_{\text{lim}} \) is precipitation limit or threshold computed by the intersection of two straight lines of equations \( Is = P \) and \( Is = a.P + b \), that is \( P_{\text{lim}} = b / (1-a) \); \( a \) and \( b \) are the parameters of the straight line and are presented on Table 6 as a function of the textural soil class (check Fig. 3 for the definition of the textural class) and of the initial soil moisture (\( \theta \)). If \( \theta \) is not the one presented on Table 6 then the parameters of the straight line are calculated assuming a linear variation between \( a \) and \( b \) parameters of the nearest (above and below) straight line parameters:

\[
\begin{align*}
   a &= a_1 + \frac{(a_2 - a_1)}{(\theta_2 - \theta_1)}(\theta_1 - \theta_1) \\
   b &= b_1 + \frac{(b_2 - b_1)}{(\theta_2 - \theta_1)}(\theta_1 - \theta_1)
\end{align*}
\]  \hspace{1cm} \text{Eq. 5}

where \( \theta_1 \) is the known initial soil moisture below \( \theta \), \( a_1 \) and \( b_1 \) are the corresponding known straight line parameters, and \( \theta_2 \) is the known initial soil moisture above \( \theta \), and \( a_2 \) and \( b_2 \) are the corresponding known straight line parameters.

As an example, surface infiltration on a silty clay soil (assuming \( sr = 0.387 \) and \( n = 0.479 \)), with \( \theta_1 = 0.35sr + 0.65n = 44.7 \% \) is given by:

\[
Is = \begin{cases} 
   P & \text{if } P \leq 1.00 \\
   0.203P + 0.797 & \text{if } P > 1.00
\end{cases}
\]  \hspace{1cm} \text{Eq. 6}

This equation was calculated using the straight line equations (Table 6) for \( \theta_2 = 0.25sr+0.75n = 45.6\% \) (\( Is = 0.182P + 0.775; a_2 = 0.182 \) and \( b_2 = 0.775 \)) and for \( \theta_1 = 0.5sr+0.5n = 43.3\% \) (\( Is = 0.236P + 0.832; a_1 = 0.236 \) and \( b_1 = 0.832 \)) and Eq. 5 and Eq. 4 with \( P_{\text{lim}} = b / (1 - a) \).
<table>
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<th>$P_{\text{lim}}$ (cm/d)</th>
<th>$a$ in</th>
<th>$\theta$</th>
<th>$P_{\text{lim}}$ (cm/d)</th>
<th>$a$ in</th>
<th>$\theta$</th>
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<td>Loamy sand wp = 5.5%, sr = 12.5%, n = 43.7%</td>
<td>5.72 0.924 0.838</td>
<td>$\theta = wp$</td>
<td>1.54 1.000 0.351</td>
<td>5.64 0.967 0.828</td>
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<td>5.44 1.005 0.819</td>
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<td>3.34 1.064 0.682</td>
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<td>2.17 1.460 0.327</td>
<td>1.01 0.770 0.235</td>
<td>$\theta = 0.5wp + 0.5sr$</td>
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<td>0.81 0.677 0.169</td>
<td>$\theta = 0.5sr + 0.5n$</td>
<td>1.12 0.809 0.275</td>
<td>0.74 0.637 0.139</td>
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<td>0.54 0.491 0.093</td>
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<td>1.09 0.832 0.236</td>
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<td>2.67 1.254 0.531</td>
<td>0.87 0.772 0.115</td>
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<td>0.38 0.373 0.007</td>
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<td>$\theta = n$</td>
<td>For sand soil $Is = P$</td>
<td>3.72 1.077 0.710</td>
<td>3.63 1.158 0.681</td>
<td></td>
</tr>
</tbody>
</table>

$Is$ = surface infiltration; $P$ = precipitation; $P_{\text{lim}}$ = precipitation threshold; $sr$ = specific retention (or field capacity); $n$ = porosity; $wp$ = wilting point; $\theta$ = initial soil moisture. The $wp$, $sr$ and $n$ values are average values taken or calculated (in the case of the silt texture) from Rawls and Brakensiek (1989).
In the BALSEQ_MOD numerical model, surface infiltration can be calculated either by the methodology already implemented in the BALSEQ numerical model (Lobo Ferreira, 1981) by using the runoff curve number (see section 8.1), or it can be computed, assuming that there is no surface medium water storage (and hence no evaporation from the surface water medium), by the difference between precipitation and surface infiltration:

\[ S_r = P - I_s \]  

**Eq. 7**

### 8.3.2 Computation of evapotranspiration

#### 8.3.2.1 Introduction

The effective evapotranspiration (RET) is estimated by (Allen *et al.*, 1998):

\[ RET = ( K_o \cdot K_{cb} + K_e ) \cdot ET_o \]  

**Eq. 8**

where \( ET_o \) is the reference evapotranspiration, \( K_o \) is the basal crop coefficient, \( K_e \) is the soil water evaporation coefficient and \( K_o \) is the water stress coefficient.

The \( ET_o \) refers to the evaporation from a hypothetical reference crop.
The $K_{cb}$ and $K_e$ terms of the equation integrate the physical and physiological differences between the specific field crop and the reference crop; hence their values vary along the time (depending on the vegetative stage). The use of the two different coefficients, $K_{cb}$ and $K_e$, constitutes the dual crop coefficient approach.

The $K_a$ term is related to the stress conditions in which the crop develops and is mainly dependent on the water available in the soil during the crop growth.

The determination of the effective evapotranspiration is not straightforward. Hence, methods based on climatic information, vegetation type and water availability in the soil, are used.

### 8.3.2.2 Estimation of the reference evapotranspiration

The reference evapotranspiration ($ET_o$), refers to the evapotranspiration of a surface that Allen et al. (1998) define as a hypothetical reference crop with an assumed crop height of 0.12 m, a fixed surface resistance of 70 s.m$^{-1}$ and an albedo of 0.23. In $ET_o$ estimation, the water available in the soil is not limiting the transpiration. To its estimation Allen et al. (1998) selected the FAO Penman-Monteith method, and presented a description of the method and of the process to quantify its parameters.

The FAO Penman-Monteith method was derived from the original method of Penman-Monteith that uses an aerodynamic resistance and a crop resistance defined for the grass reference surface. The FAO Penman-Monteith method is described by (Allen et al., 1998):

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

Eq. 9

where:

- $ET_{o}$ reference evapotranspiration (mm.d$^{-1}$);
- $R_n$ net radiation at the crop surface (MJ.m$^{-2}$.d$^{-1}$);
- $G$ soil heat flux density (MJ.m$^{-2}$.d$^{-1}$);
- $T$ mean daily air temperature at 2 m height (ºC);
- $u_2$ wind speed at 2 m height (m.s$^{-1}$);
- $e_s$ saturation vapour pressure (kPa);
- $e_a$ actual vapour pressure (kPa);
- $e_s - e_a$ saturation vapour pressure deficit (kPa);
- $\Delta$ slope vapour pressure curve (kPa.(ºC)$^{-1}$);
- $\gamma$ psychrometric constant (kPa.(ºC)$^{-1}$).

The equations to compute the several variables of Eq. 9 are presented accordingly to Allen et al. (1998). To understand how these formulas were derived Allen et al. (1998) work may be consulted.

- $\Delta$ - slope vapour pressure curve (kPa/ºC)
\[ \Delta = \frac{4098 \times e'(T)}{(T + 237.3)^2} \quad \text{Eq. 10} \]

where \( T \) is mean air temperature (°C); \( e'(T) \) is the saturation vapour pressure at temperature \( T \) (kPa);

- \( T \) – mean air temperature (°C)

\[ T = \frac{T_{\text{max}} + T_{\text{min}}}{2} \quad \text{Eq. 11} \]

where \( T_{\text{max}} \) is the maximum air temperature (°C); \( T_{\text{min}} \) is the minimum air temperature (°C);

- \( e'(T) \) – saturation vapour pressure at temperature \( T \) (kPa)

\[ e'(T) = 0.6108 \times \exp \left( \frac{17.27 \times T}{T + 237.3} \right) \quad \text{Eq. 12} \]

- \( e_s \) – mean saturation vapour pressure (kPa)

\[ e_s = \frac{e'(T_{\text{max}}) + e'(T_{\text{min}})}{2} \quad \text{Eq. 13} \]

- \( e_a \) – actual vapour pressure (kPa)

\[ e_a = \frac{e'(T_{\text{min}}) \times RH_{\text{max}} + e'(T_{\text{max}}) \times RH_{\text{min}}}{100} \quad \text{Eq. 14} \]

where \( T_{\text{min}} \) stands for minimum temperature (°C); \( RH_{\text{max}} \) stands for maximum relative humidity (%).

- \( \gamma \) – psychrometric constant (kPa/°C)

\[ \gamma = 0.000665 \times P \quad \text{Eq. 15} \]

where \( P \) is atmospheric pressure (kPa).

- \( u_2 \) – wind speed at 2 m height (m.s\(^{-1}\))

\[ u_2 = u_z \times \frac{4.87}{\ln(67.8 \times z - 5.42)} \quad \text{Eq. 16} \]

where \( u_z \) is the measured wind speed at \( z \) m above ground surface (m/s).

- \( R_n \) – net radiation at the crop surface (MJ.m\(^{-2}\).d\(^{-1}\))

\[ R_n = R_{ns} - R_{nl} \quad \text{Eq. 17} \]

where \( R_{ns} \) is net solar radiation (MJ/m\(^2\)/d); \( R_{nl} \) net long wave radiation (MJ/m\(^2\)/d).
• $R_{nl}$ – net long wave radiation (MJ.m\(^2\).d\(^{-1}\))

$$R_{nl} = \sigma \times \left[ \left( T_{\max} + 273.16 \right)^4 + \left( T_{\min} + 273.16 \right)^4 \right] \times \left( 0.34 - 0.14 \times \sqrt{e_u} \right) \times \left( 1.35 \times \frac{R_s}{R_{so}} - 0.35 \right)$$  \text{Eq. 18}

where $\sigma$ is the Stefan-Boltzmann constant = 4.903 x 10\(^{-9}\) MJ.K\(^{-4}\).m\(^{-2}\).d\(^{-1}\); $R_s$ solar radiation (MJ.m\(^2\).d\(^{-1}\)); $R_{so}$ solar radiation that would occur if there was clear-sky conditions (MJ.m\(^2\).d\(^{-1}\)).

• $R_s$ – solar radiation (MJ.m\(^2\).d\(^{-1}\))

$$R_s = \left( a_s + b_s \times \frac{n}{N} \right) R_a$$  \text{Eq. 19}

where $n$ is the actual duration of sunshine (hour); $N$ is the maximum possible duration of sunshine or daylight hours (hour); $R_a$ is the extra-terrestrial radiation (MJ.m\(^2\).d\(^{-1}\)); $a_s$ and $b_s$ are two parameters (dimensionless), that should be calibrated for local conditions. If this calibration has not been carried out the values of $a_s = 0.25$ and $b_s = 0.50$ are recommended. If there is no data available on $n$, the relation $n/N$ may be approximated using the cloudiness observations as represented in Table 7 for oktas and in Table 8 for tenths.

<table>
<thead>
<tr>
<th>Oktas</th>
<th>0</th>
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<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
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</thead>
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<tr>
<td>n/N</td>
<td>0.95</td>
<td>0.85</td>
<td>0.75</td>
<td>0.65</td>
<td>0.55</td>
<td>0.45</td>
<td>0.35</td>
<td>0.15</td>
<td>-</td>
</tr>
</tbody>
</table>

Source: Doorenbos and Pruitt (1977)

<table>
<thead>
<tr>
<th>Tenths</th>
<th>0</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
</tr>
</thead>
<tbody>
<tr>
<td>n/N</td>
<td>0.95</td>
<td>0.85</td>
<td>0.8</td>
<td>0.75</td>
<td>0.65</td>
<td>0.55</td>
<td>0.5</td>
<td>0.4</td>
<td>0.3</td>
<td>0.15</td>
<td>-</td>
</tr>
</tbody>
</table>

Source: Doorenbos and Pruitt (1977)

• $R_{so}$ – solar radiation that would occur if there was clear-sky conditions (MJ.m\(^2\).d\(^{-1}\))

$$R_{so} = \left( 0.75 + 2 \times 10^{-5} \times a \right) \times R_a$$  \text{Eq. 20}

where $a$ is altitude above sea level (m); $R_a$ extraterrestrial radiation (MJ.m\(^2\).d\(^{-1}\)).

• $R_a$ – extraterrestrial radiation (MJ.m\(^2\).d\(^{-1}\))

$$R_a = \frac{1440}{\pi} \times G_{sc} \times d_s \times \left[ \omega_s \times \sin(\varphi) \times \sin(\delta) + \cos(\varphi) \times \cos(\delta) \times \sin(\omega_s) \right]$$  \text{Eq. 21}

where $G_{sc}$ is the solar constant = 0.0820 MJ.m\(^2\).min\(^{-1}\); $d_s$ is the inverse relative distance Earth-Sun (rad); $\omega_s$ is the sunset hour angle (rad), $\varphi$ is latitude (rad); $\delta$ is solar declination (rad).

• $d_s$ – inverse relative distance Earth-Sun (rad)
\[ d_r = 1 + 0.033 \times \cos\left(2 \times \pi \times \frac{365}{365} \times J\right) \tag{Eq. 22} \]

where \( J \) is the number of the day in the year, starting at 1 January.

- \( \omega_s \) – sunset hour angle (rad)

\[ \omega_s = \arccos(-\tan(\varphi) \cdot \tan(\delta)) \tag{Eq. 23} \]

- \( \delta \) – solar declination (rad)

\[ \delta = 0.409 \times \sin\left(2 \times \pi \times \frac{365}{365} \times J - 1.39\right) \tag{Eq. 24} \]

- \( R_{ns} \) – net solar radiation (MJ.m\(^{-2}\).d\(^{-1}\))

\[ R_{ns} = 0.77 \times R_s \tag{Eq. 25} \]

- \( G \) – soil heat flux density (MJ.m\(^{-2}\).d\(^{-1}\))

For the case of daily or 10-day reference evapotranspiration computations, the value of \( G \) is low when compared with \( R_n \). For this reason Allen et al. (1998) suggest that for these time periods \( G \) may be ignored.

For the computation of monthly reference evapotranspiration, using monthly mean values of the parameters, \( G \) may be approximated by the following equation:

\[ G = 0.14 \left( T_{\text{month}} - T_{\text{month before}} \right) \tag{Eq. 26} \]

where \( T_{\text{month}} \) is mean air temperature at the current month (ºC) and \( T_{\text{month before}} \) is mean air temperature of the previous month (ºC).

### 8.3.2.3 Determination of the basal crop coefficient (\( K_{cb} \))

Four distinct vegetative stages may be identified during the crop growth: (1) initial (up to 10% ground cover), (2) crop development (until total ground cover), (3) mid-season (until the start of maturity); and (4) late-season (until harvest or full senescence).

The definition of a crop coefficient curve as a function of the vegetative growth is made by defining the crop coefficient values for initial (\( K_{cb \ ini} \)), middle (\( K_{cb \ mid} \)) and late (\( K_{cb \ end} \)) seasons, and the length of each crop growth stage (Fig. 4).
Indicative values for $K_{cb}$ may be consulted in Table 17 of Allen et al. (1998). In the case of minimum relative humidity conditions and wind speed different from the “standard” ones (45 % and 2 m.s$^{-1}$, respectively), the $K_{cb\_mid}$ and $K_{cb\_end}$ values should be adjusted using the formula:

$$K_{cb\_i} = K_{cb\_{(table)}} + [0.04.(u_2 - 2) - 0.004.(HR_m - 45)].(a_i/3)^{0.3}$$  \hspace{1cm} \text{Eq. 27}$$

where $K_{cb\_{(table)}}$ is the $K_{cb}$ value determined for “standard” conditions for the stage $i$ ($i = \text{mid}$ or $i = \text{end}$), $a$ (m) is the average crop height, $u_2$ (m.s$^{-1}$) is average wind speed during stage $i$, determined or corrected at a 2 m distance above the ground and $HR_m$ (%) is the average, for the stage $i$, of the minimum daily relative humidity.

Fig. 5 shows an example of the basal crop coefficients distribution for two vegetation types that co-exist in a cartographic unit (non-irrigated wheat and orchard), along four hydrological years that start in October, the 1$^{st}$, 1995. For the non-irrigated wheat $K_{cb\_ini} = 0.15$, $K_{cb\_mid} = 1.10$, $K_{cb\_end} = 0.15$. For the orchard $K_{cb\_ini} = 0.35$, $K_{cb\_mid} = 0.9$, $K_{cb\_end} = 0.65$. For the non-irrigated wheat the vegetation cycle periods are the following: 1$^{st}$ day of the crop after the 1$^{st}$ of October ($\text{day\_ini}$) = day 32, initial stage length ($\text{L\_ini}$) = 30 day, crop development length ($\text{L\_dev}$) = 140 day, mid-season length ($\text{L\_mid}$) = 40 day, late-season length ($\text{L\_end}$) = 30 day. For the orchard, $\text{day\_ini} = \text{day 152}$, $\text{L\_ini} = 30$ day, $\text{L\_dev} = 50$ day, $\text{L\_mid} = 130$ day, and $\text{L\_end} = 30$ day.
Fig. 5 – Basal crop coefficient ($K_{cb}$) for each vegetation cover, not considering the interdependency between the crop coefficients

For the $K_{cb}$ values that are not tabulated, Allen et al. (1998) propose the use of the following formulas:

- Natural vegetation and crops not listed in Table 17 of Allen et al. (1998):

For areas above a few hectares $K_{cb\ full}$ may be approximated by:

$$K_{cb\ full} = \min(1.20; 1.0 + 0.1. a) + [0.04.(u - 2) - 0.004.(HR_m - 45)].(a / 3)^{0.3}$$  

Eq. 28

- Sparse vegetation:

$$K_{cb\ mid\ adj} = K_{c\ min} + (K_{cb\ full} - K_{c\ min}) \cdot \min\left[1; 2. f_c; \left(f_{c\ eff}\right)^{\frac{1}{1.5a}}\right]$$  

Eq. 29

where $K_{cb\ mid\ adj}$ is the estimated $K_{cb}$ during the mid-season period when plant density and/or leaf area are lower than full cover conditions; $K_{cb\ full}$ is the estimated $K_{cb}$ during the mid-season for vegetation having full ground cover or leaf area index > 3; $K_{c\ min}$ is the minimum $K_c$ for bare soil (approx. 0.15-0.20); $f_c$ is the fraction of soil surface that is covered by vegetation as observed from top (0.01 – 1); $f_{c\ eff}$ is the effective fraction of soil surface covered or shaded by vegetation (0.01 – 1) – in BALSEQ.MOD it is assumed equal to $f_c$; $a$ is the plant height (m). $K_{cb\ full}$ is given by $K_{cb\ mid}$ for the conditions of full-cover crop given in table 17 of Allen et al. (1998), corrected for climate conditions (Eq. 27). Fig. 6 shows the sparse basal crop coefficients distribution for the previous example.
Fig. 6 – Sparse basal crop coefficient ($K_{cb}$) for each vegetation cover, not considering the interdependency between the crop coefficients

For the late-season period $K_{cb\text{ end adj}}$ may also be computed using Eq. 29 with $K_{cb\text{ end}}$ instead of $K_{cb\text{ mid}}$.

In the case of the presence of two vegetation-types in the same area, Allen et al. (1998) suggest some procedures that take into account the existence of an energy upper limit for the evapotranspiration, given by $K_{c\text{ max}}$, with $a$ given by the tallest vegetation and $K_{cb}$ by the largest value:

$$K_{c\text{ max}} = \max\left\{\left\{1.2 + \left[0.04(u_2 - 2) - 0.004(HR_m - 45)\right](a/3)^{0.3}\right\}; \{K_{cb} + 0.05\}\right\}$$  \hspace{1cm} \text{Eq. 30}

$K_{c\text{ max}}$ presents values between 1.05 and 1.30.

In the BALSEQ_MOD model (Oliveira, 2004), for the cases where two crops exist, the following procedure was used:

$$K_{cb\text{ adj (crop 1, corrected)}} = K_{cb\text{ adj (crop 1)}} \cdot \frac{K_{c\text{ max}}}{(K_{cb\text{ adj (crop 1)}} + K_{cb\text{ adj (crop 2)}})}$$

$$K_{cb\text{ adj (crop 2, corrected)}} = K_{cb\text{ adj (crop 2)}} \cdot \frac{K_{c\text{ max}}}{(K_{cb\text{ adj (crop 1)}} + K_{cb\text{ adj (crop 2)}})}$$  \hspace{1cm} \text{Eq. 31}

where $K_{cb\text{ adj (crop 1)}}$ refers to the $K_{cb\text{ adj}}$ of crop 1, estimated by Eq. 27 as if this crop existed alone. The same applies for crop 2. Fig. 7 shows the adjusted sparse basal crop coefficients for the considered example.
8.3.2.4 **Determination of the soil water evaporation coefficient** ($K_e$)

In terms of the evaporation component of RET (related to $K_e$), when the soil is wet, evaporation from the soil occurs at the maximum rate. However, the sum $K_e + K_{cb}$ cannot exceed a limit value ($K_{c\text{ max}} - K_{cb}$, Eq. 30), which is determined by the energy available in the soil for evapotranspiration ($K_{c\text{ max}} - K_{cb}$).

On another hand $K_e$ may not exceed the available energy in the wet exposed fraction of the soil ($f_{ew}$):

$$K_e \leq f_{ew} \cdot K_{c\text{ max}}.$$  

$K_e$ is given by:

$$K_e = \min(K_r \cdot (K_{c\text{ max}} - K_{cb}) ; f_{ew} \cdot K_{c\text{ max}})$$  

**Eq. 32**

where $K_r$ is a evaporation reduction coefficient that depends on the amount of water stored in the upper part of the soil subject to evaporation (topsoil).

Quantification of $K_r$ requires a topsoil daily water balance. Its value varies between 1 for a soil with water content equal to field capacity ($sr$) and 0 for a soil at the wilting point ($wp$). While the water content is above a threshold given by $(1 - p) \cdot \text{available water}$, $K_r = 1$. Available water is given by the difference between field capacity and wilting point. Below this threshold $K_r$ is given by:

$$K_r = (\theta - wp) \cdot (1 - p) \cdot (sr - wp) \cdot (1 - \theta)$$  

**Eq. 33**

where $\theta$ is the water content.

As an average, and looking at the values published in Allen *et al.* (1998) for different soil types, the value of $p$ may be assumed as 42 % of the available water.

In the case of a bare soil, while the water content is above the threshold value $(1 - p) \cdot \text{available water}$, a value of $K_e = 1.15$ may be assumed.

8.3.2.5 **Determination of the water stress coefficient** ($K_a$)

$K_a$ is a parameter similar to $K_r$. But, in this case, it depends also on the type of crop. Its value varies between 1 for a soil with water content equal to $sr$ and 0 for a soil at $wp$. While the water content is above a threshold given by $(1 - p) \cdot \text{available water}$, $K_a = 1$. Below this threshold $K_a$ is given by:
\[ K_a = (\theta - wp). \left( (1 - p) . (sr - wp) \right)^{-1} \]  
\text{Eq. 34}

where \( \theta \) is the water content.

The values for \( p \) are published on Table 22 of Allen et al. (1998). These values apply for \( Etc = (K_{cb} + K_{e}) \cdot ET_o = 5 \text{ mm.day}^{-1} \). It can be adjusted for different \( ETc \) using:

\[ p = p_{(ETc=5\text{mm/d})} + 0.04 \cdot (5 - ETc) \]  
\text{Eq. 35}

The adjusted \( p \) is limited to \( 0.1 \leq p \leq 0.8 \).

A value of 0.50 is commonly used for many crops. The value of \( p \) may also be corrected as a function of the soil type. For fine textured soils (clay) \( p_{(ETc=5\text{mm/d})} \) should be reduced by 5 % to 10 %. For more coarse textured soils (sand) it should be increased by the same amount (Allen et al., 1998).

### 8.3.2.6 Determination of the fraction of the soil surface covered by vegetation as observed from the top \((f_c)\) and of the wet exposed fraction of the soil \((f_{ew})\)

It is assumed that the influence zone of the roots is horizontally given by the area occupied by the vegetation. So it is assumed equal to the fraction of soil surface that is covered by vegetation as observed from top \((f_c)\). \( f_c \) varies along the time, depending on the development stage of the vegetation. Up to two vegetation types may be considered. The sum of the \( f_c \) of each vegetation type cannot be larger than 1. The following parameters must be characterised for each vegetation type: maximum area \((f_{c\_max})\), minimum area \((f_{c\_min})\) equal to 10 % or in the case of perennial crops equal to \( f_{c\_max} \) and rest time area \((f_{c\_0})\), equal to 0 in the case of dormancy or plant inexistence, or equal to \( f_{c\_max} \) in the case of evergreen forests). During a year-cycle \( f_c \) assumes the following values:

- before initial stage, or after the end of the late-season stage: \( f_c = f_{c\_0} \);
- in the initial stage: \( f_c = f_{c\_min} \);
- in the crop development stage: \( f_c \) varies linearly between \( f_{c\_min} \) in the first day and \( f_{c\_max} \) in the last day;
- in the mid-season and late-season: \( f_c = f_{c\_max} \).

The wet exposed fraction of the soil \((f_{ew})\) is given by 

\[ f_{ew} = 1 - (f_{c1} + f_{c2}) \],

where \( f_{c1} \) and \( f_{c2} \) stand for the fractions occupied by vegetation 1 and vegetation 2.

Fig. 8 shows an example of the area fractions occupied by two crops (non-irrigated wheat and orchard) and by the bare soil, along a hydrological year that starts in the 1\textsuperscript{st} of October. For the non-irrigated wheat \( f_{c\_min} = 10 \% , f_{c\_max} = 60 \% , f_{c\_0} = 0 \% \). For the orchard \( f_{c\_min} = 20 \% , f_{c\_max} = 80 \% , f_{c\_0} = 0 \% \).
Determination of the soil moisture

The computation of the $K_a$ parameter requires the quantification of the soil water content (expressed in % volume of water / volume of soil) above the wilting point ($\theta - wp$). The water content above the wilting point reflects the water that may be mobilised by the plants for the evapotranspiration process. As the plants can withdraw water along the depth of their roots, instead of the soil water content, a different variable, $A_l$, is used that refers to the amount of water stored in the root depth ($rd$) that may be mobilised by the plants, which is given by:

$$A_l = (\theta - wp) \cdot rd$$

Eq. 36

In the BALSEQ_MOD model the daily amount of water available for evapotranspiration, $A_{l_ETR}(\text{day, cover } i)$, is given by:

$$A_{l_ETR}(\text{day, cover } i) = A_{l_{ini}}(\text{day, cover } i) + I_s(\text{day, cover } i) + A_{l_{inc}}(\text{day, cover } i)$$

Eq. 37

where $cover_i$ refers to the vegetation or crop type 1 or 2 or to the bare soil, $A_{l_{ini}}$ is the amount of water that exists in the soil in the end of the previous day of the sequential water balance, $I_s$ is the surface infiltration computed for the current day, and $A_{l_{inc}}$ represents, for the case of the vegetation cover, the increase, from the previous to the current day, of the amount of water in the soil due to the increase of the area covered by the vegetation or due to the increase of the root depth.

For the case of the bare soil, $A_{l_{inc}}$ is null, except for the day in which one vegetation cover becomes inactive. In that day, the amount of water in the bare soil is increased by the amount of water that existed in the vegetation cover area in the previous day, and $A_{l_{inc}}$ of the bare soil is:

$$A_{l_{inc}}(\text{day, bare soil}) = [A_{l_{ini}}(\text{day, cover}) / rd_{1\text{ last day}}(cover) \cdot \text{thick}(day) \cdot f_{c}(day-1, cover) / f_{c}(day, bare soil)]$$

Eq. 38

where $\text{thick}$ represents the bare soil thickness subject to evapotranspiration, $rd_{1\text{ last day}}$ is the root depth of the cover in the last day that it existed, $f_{c}$ is the fraction of the area occupied by the cover, $day$ the current day and $day-1$ the previous day.
For each vegetation cover, the following terms are related to the increase of the amount of water in the soil, expressed in terms of water column in the area occupied by the vegetation crop (Fig. 9):

**a) Term related to the increase of the root depth** ($A_{l1}$), considering the soil water content that exists in the growth zone of the plant roots:

$$A_{l1} = [rd_{(day,cover)} - rd_{(day-1)}] \cdot (sr - wp - \theta_{l \, def}) \cdot \frac{fc_{(day-1,cover)}}{fc_{(day,cover)}} \quad \text{Eq. 39}$$

where $\theta_{l \, def}$ is given by:

$$\theta_{l \, def} = \frac{A_{l \, def \, (day-1,cover,1)}}{rd_1(cover) - rd_{(day-1,cover)}} \quad \text{Eq. 40}$$

$A_{l \, def \, (day-1,cover,1)}$ represents the deficit of water in the soil thickness between the root depth in the previous day [$rd_{(day-1)}$] and the maximum root depth of the plants ($rd_1$), the soil water content reaches the field capacity ($sr$).

**Fig. 9 – Situations considered during the water balance for the case of the increase or the reduction of the fraction occupied by a specific vegetation cover or of the soil thickness subject to evapotranspiration.**

Due to the increase of the plant root depth, the term $A_{l \, def \, (day,cover,1a)}$ is updated for the new depth that still has to be fulfilled by the plant roots [between $rd_1$ and $rd_{(day)}$]:

$$A_{l \, def \, (day,cover,1a)} = \frac{A_{l \, def \, (day-1,cover,1)}}{[rd_1(cover) - rd_{(day-1,cover)}] \cdot \frac{fc_{(day-1,cover)}}{fc_{(day,cover)}}} \quad \text{Eq. 41}$$
b) Term related to the increase of the area in the zone of the bare soil ($A_{l2}$), considering the soil water content that exists in this zone:

$$A_{l2} = [fc_{(day, cover)} - fc_{(day-1, cover)}] \times A_{l \text{ ini}}_{(day, bare soil)} / fc_{(day, cover)} \quad \text{Eq. 42}$$

c) Term related to the increase of the area below the depth subject to evaporation of the bare soil ($A_{l3}$, applicable if $rd_{(day)} > \text{thickness of the evaporating zone } [thick_{(day)}]$), considering the water content that exists in the soil in the zone of increase of the plant’s root depth:

$$A_{l3} = [rd_{(day,cover)} - \text{thick}_{(day)}] \cdot (sr - wp - \theta_{def}) \cdot [fc_{(day,cover)} - fc_{(day-1,cover)}] / fc_{(day,cover)} \quad \text{Eq. 43}$$

where $\theta_{def}$ is given by:

$$\theta_{def} = A_{l \text{ def}}{(day-1,cover,2)} / [rd_{(cover)} - \text{thick}_{(day-1)}] \quad \text{Eq. 44}$$

$A_{l \text{ def}}{(day-1,cover,2)}$ represents the deficit of water in the soil thickness between the bare soil bottom in the previous day $[thick_{(day-1)}]$ and the maximum root depth of the plants ($rd_{1}$), required to increase the soil water content to the field capacity ($sr$).

In the area previously below the bare soil and that currently is also occupied by the vegetation cover, there is a change in the amount of the water deficit. As this area is now part of the fraction occupied by the vegetation cover, the following applies:

$$A_{l \text{ def}}{(day,cover,1b)} = A_{l \text{ def}}{(day-1,cover,2)} / [rd_{(cover)} - \text{thick}_{(day-1)}] \cdot [rd_{(cover)} - rd_{(day,cover)}] \cdot [fc_{(day,cover)} - fc_{(day-1,cover)}] / fc_{(day,cover)} \quad \text{Eq. 45}$$

The water increase that results from increasing both the thickness and the area of the vegetative cover, expressed in height of the water column in the vegetation area fraction, is:

$$A_{l \text{ inc}}{(day, cover)} = A_{l1} + A_{l2} + A_{l3} \quad \text{Eq. 46}$$

and the water required to fulfil the field capacity between the plant root depth and its maximum depth is:

$$A_{l \text{ def}}{(day,cover,1)} = A_{l \text{ def}}{(day,cover,1a)} + A_{l \text{ def}}{(day,cover,1b)} \quad \text{Eq. 47}$$

If the day in which the vegetation cover becomes inactive, the amount of water related to the cover becomes null:

$$A_{l \text{ RET}}{(day,cover)} = 0 \quad \text{Eq. 48}$$

and the terms related to $A_{l \text{ def}}$ become:

$$A_{l \text{ def}}{(day,cover,1)} = 0 \quad \text{Eq. 49}$$

because the vegetation cover does not exist anymore. In the area below the new bare soil, that in the next plant cycle will be occupied by the plant roots again, $A_{l \text{ def}}$ is:
8.3.2.8 Determination of the root depth

The root depth (\(rd\)) is important to define the amount of water available for evapotranspiration: It depends on the development stage of the vegetation. Up to two vegetation types may be considered. For each one, the following parameters must be characterised: minimum root depth (\(rd_0\)), maximum root depth (\(rd_1\)). During a year cycle \(rd\) assumes the following values:

- before initial stage, or after the end of the late-season stage: \(rd = 0\);
- in the initial stage: \(rd = rd_0\);
- in the crop development stage: \(rd\) varies linearly between \(rd_0\) in the first day and \(rd_1\) in the last day;
- in the mid-season and late-season: \(rd = rd_1\).

For the bare soil fraction a constant value is assumed along the year – only \(rd_1\) (or \(thick\)) – is defined. According to Allen et al. (1998), the depth of the upper part of the soil that is subject to drying by evaporation is 10-15 cm.

Fig. 10 shows an example of the soil thickness subject to evapotranspiration for two vegetation covers (non-irrigated wheat and orchard) and for the bare soil, along a hydrological year that starts in the 1st of October. For the non-irrigated wheat \(rd_0 = 150\) mm, \(rd_1 = 1,200\) mm. For the orchard \(rd_0 = rd_1 = 1,500\) mm. For the bare soil \(thick = rd_1 = 150\) mm.

\[
A_{i, def(day,cover,2)} = [rd_1(cover) - thick_{(day-1)}] * \left[ sr - \left( wp + A_{i, in(day,cover)} / rd_{(day-1,cover)} \right) \right] \quad \text{Eq. 50}
\]
Computation of effective evapotranspiration (RET)

Using the above described methodology the effective evapotranspiration is computed for each one of the up to 2 land covers and for the bare soil. Fig. 11 shows the RET values for the presented example of the non-irrigated wheat, the orchard and the bare soil, all of them expressed in terms of the water height in relation to the fraction of the soil surface covered by each vegetation type or by the bare soil.

For the complete area, the effective evapotranspiration (RET) – Fig. 12 – is obtained by:

\[
RET = RET_{\text{cover} 1} \cdot f_{\text{cover} 1} + RET_{\text{cover} 2} \cdot f_{\text{cover} 2} + RET_{\text{bare soil}} \cdot f_{\text{bare soil}}
\]

Eq. 51

where \( fc \) is the fraction of soil surface occupied by each cover or by the bare soil [Note that the sum of the \( fc \) is equal to 1].

Fig. 11 – Effective evapotranspiration of each vegetation cover or bare soil (values referred to the soil surface occupied by them).

Fig. 12 – Effective evapotranspiration of the complete area.
8.3.2.10 Summary of the information required to estimate RET in the BALSEQ_MOD model

The following information is required to run the BALSEQ_MOD model, in order to compute the effective evapotranspiration:

- daily surface infiltration \( (I_s) \);
- daily reference evapotranspiration \( (ET_o) \);
- the fraction of the area occupied by each land cover \( (f_c) \) or by the bare soil \( (f ew) \); in the case of vegetation cover it is necessary to know the area fraction occupied by vegetation during mid-season and late-season stages \( (f c_{max}) \), and the area fraction occupied by vegetation during the initial stage of the development \( (f c_{min}) \); for static land covers the same fractions are required but \( f c_{max} = f c_{min} \);
- the soil depth subject to evapotranspiration. For the vegetation cover, two soil depths are defined accordingly to the development stage of the vegetation: the initial stage \( (rd_0) \), and the mid-season and late season crop development stage \( (rd_1) \). For bare soil a depth of 15 cm subject to evaporation is assumed;
- the basal crop coefficients. Applied only for the vegetation covers, three values are required for each vegetation cover for initial \( (K cb_{ini}) \), middle \( (K cb_{mid}) \) and late \( (K cb_{end}) \) seasons. These depend on the vegetation height, the air relative humidity, the wind speed, the fraction of land surface covered by the vegetation;
- the first day of the initial stage \( (day_{ini}) \), and the length of each crop growth stage: initial stage length \( (L_{ini}) \), crop development length \( (L_{dev}) \), mid-season length \( (L_{mid}) \) and late-season length \( (L_{end}) \);
- threshold values for the minimum amount of water stored in the soil that allow the effective evapotranspiration to occur at the maximum rate \( (p) \), both for the vegetation cover and for the bare soil.

8.3.3 Computation of deep percolation

The soil moisture storage variation \( (\Delta A_l) \) and the deep percolation \( (D_p) \) are computed by sequential water balance:

\[
\Delta A_{l_{(day, cover i)}} + D_{p_{(day, cover i)}} = I_{s_{(day, cover i)}} + A_{l inc_{(day, cover i)}} - R_{ET_{(day, cover i)}} \tag{52}
\]

As \( \Delta A_{l_{(day, cover i)}} = A_{l_{end_{(day, cover i)}}} - A_{l_{ini_{(day, cover i)}}} \) and from the sequential water balance \( A_{l_{ini_{(day, cover i)}}} + I_{s_{(day, cover i)}} + R_{ET_{(day, cover i)}} \) are already computed, it is needed to decompose \( A_{l_{end_{(day, cover i)}}} + D_{p_{(day, cover i)}} \) from the following equation:

\[
A_{l_{end_{(day, cover i)}}} + D_{p_{(day, cover i)}} = A_{l_{ini_{(day, cover i)}}} + I_{s_{(day, cover i)}} + A_{l inc_{(day, cover i)}} - R_{ET_{(day, cover i)}} \tag{53}
\]
The most straightforward process to compute deep percolation is used in the BALSEQ model (Lobo Ferreira, 1981), which assumes that all the water that drains freely under the action of gravity will become deep percolation. In this case the amount of water stored in the soil is upper limited by 
\[ AG_{sr(day, cover i)} = rd_{day, cover i} \times sr, \]
where \( rd \) is root depth and \( sr \) is specific retention (or field capacity). When this amount is larger than \( AG_{sr(day, cover i)} \) the water flows in depth, becoming deep percolation:

\[
DP_{(day, cover i)} = \max\{A_i\_{ini(day, cover i)} + Is_{(day, cover i)} + A_i\_{inc(day, cover i)} - RET_{(day, cover i)} - AG_{sr(day, cover i)} ; 0\} \quad \text{Eq. 54}
\]

Knowing \( DP_{(day, cover i)} \) and substituting in Eq. 53, \( A_i\_{end} \) is computed:

\[
A_i\_{end(day, cover i)} = \min\{A_i\_{ini(day, cover i)} + Is_{(day, cover i)} + A_i\_{inc(day, cover i)} - RET_{(day, cover i)} - AG_{sr(day, cover i)} ; 0\} \quad \text{Eq. 55}
\]

These equations are valid when the phreatic level is below the soil bottom. The use of these equations in the daily sequential water balance model imply that in the end of the day all the water present in the soil exceeding the storage corresponding to the field capacity value has been able to drain through all the soil thickness. In many cases that may not happen, as drainage depends on the hydraulic conductivity, the hydraulic head and on the soil thickness. If the water remains in the soil, this will have an amount of stored water larger than \( AG_{sr} \) and this water may be used in the evapotranspiration process in the next day.

For the BALSEQ_MOD program the procedure referred to by Samper et al. (1999) was adopted, where deep percolation is given by the water in the soil exceeding \( AG_{sr} \) however limited by the maximum amount of water that the soil may transmit in the considered time interval (\( K_s \times \Delta t \)), where \( K_s \) is the saturated vertical hydraulic conductivity of the soil and \( \Delta t \) is the time step (1 day):

\[
DP_{(day, cover i)} = \min\{\max\{A_i\_{ini(day, cover i)} + Is_{(day, cover i)} + A_i\_{inc(day, cover i)} - RET_{(day, cover i)} - AG_{sr(day, cover i)} ; 0\} ; (K_s \times \Delta t)\} \quad \text{Eq. 56}
\]

Replacing \( DP_{(day, cover i)} \) in Eq. 55 \( A_i\_{end(day, cover i)} \) is obtained. However with the application of this equation the amount of water in the soil may exceed the maximum amount of water that the soil may contain (\( AG_{(day, cover i)} = rp_{(day, cover i)} \times n \) where \( n \) is porosity). In this case it is assumed that if \( A_i\_{end(day, cover i)} \) given by Eq. 55 exceeds \( AG_{(day, cover i)} \), the difference will be added to the direct runoff or to the water stored in the surface medium (ponding) (that in BALSEQ_MOD is considered null):

\[
Sr_{(day, cover i)} = Sr_{\text{previously computed}(day, cover i)} + (A_i\_{end(day, cover i)} - AG_{(day, cover i)}) \quad \text{Eq. 57}
\]

The amount of water stored in the soil becomes:

\[
A_i\_{end(day, cover i)} = AG_{(day, cover i)} \quad \text{Eq. 58}
\]

Deep percolation thus calculated may still not translate the deep percolation of a day. This is due to the fact that while plant roots are developing and growing in depth, the volume of soil existing between the root depth in that day \( [rp_{(day)}] \) and the maximum depth that will be achieved by the plants \( (rp_{,1}) \) may present a moisture content below the field capacity.
The deep percolation calculated by Eq. 54 or Eq. 56, now on designated as $D_{p, soil}$, is not draining downward by gravity forces, but will fulfill the soil voids until its moisture content attains the field capacity.

Consider the three zones represented on Fig. 13:

1. below the land area occupied by vegetal cover 1 [$f_{c(\text{day}, \text{cover 1})}$];
2. below the land area occupied by vegetal cover 2 [$f_{c(\text{day}, \text{cover 2})}$] – if it exists;
3. below the bare soil, that can also be decomposed in three sub-zones:
   3.1 – the area that during the vegetal cover 1 development will be occupied, that is, the area below [$\text{fraction}_1 - f_{c(\text{day}, \text{cover 1})}$];
   3.2 – the area that during the vegetal cover 2 development (if it exists) will be occupied, that is, the area below [$\text{fraction}_2 - f_{c(\text{day}, \text{cover 2})}$];
   3.3 – the area below the bare soil that will never be occupied by the vegetal cover (1 – [$\text{fraction}_1 + \text{fraction}_2$]).

Fig. 13 – Terms used for the computation of deep percolation when the dual crop coefficient is used for the computation of evapotranspiration

In the case of the first two zones the water amount required to fill the soil voids until its moisture content attains the field capacity is represented by $A_{l, \text{def}(\text{day}, \text{cover 1}, 1)}$ and $A_{l, \text{def}(\text{day}, \text{cover 2}, 1)}$ as calculated in Eq. 47. $D_{p, soil}$ is calculated for the vegetal covers [$D_{p, soil(\text{day}, \text{cover 1})}$ and $D_{p, soil(\text{day}, \text{cover 2})}$]. Deep percolation and the new values of $A_{l, \text{def}}$ are given by:

$$D_{p(\text{day, cover } i)} = \max(D_{p, soil(\text{day, cover } i)} - A_{l, \text{def} \text{ Eq. 47(\text{day, cover } i,1)}}, 0) \quad \text{Eq. 59}$$

$$A_{l, \text{def(\text{day, cover } i,1)}} = \max(A_{l, \text{def Eq. 47(\text{day, cover } i,1)}}, D_{p, soil(\text{day, cover } i)} - 0) \quad \text{Eq. 60}$$

where $i$ assumes the values 1 or 2. In the case that cover 2 does not exist, $i$ only assumes value 1.
To determine deep percolation of the third zone the three sub-zones must be considered. In the case of sub-zones 3.1 and 3.2, the water amount required to fill the soil voids until its moisture content attains the field capacity is given by $A_{l_{def}(day,\,cover\;1,\;2)}$ and $A_{l_{def}(day-1,\;cover\;2,\;2)}$. The terms $A_{l_{def}}$ refer to the calculated values of the previous day, as they have not been updated to the present day of the balance. $Dp$ and $A_{l_{def}}$ are given by:

$$Dp_{3.1}(day) = \max(Dp_{soil}(day,\,bare\;soil) - A_{l_{def}(day-1,\;cover\;1,\;2)}; 0)$$  \hspace{1cm} \text{Eq. 61}

$$A_{l_{def}(day,\,cover\;1,\;2)} = \max(A_{l_{def}(day-1,\;cover\;1,\;2) - Dp_{soil}(day,\,bare\;soil)}; 0)$$  \hspace{1cm} \text{Eq. 62}

where $i$ assumes values 1 or 2. In the case that cover 2 does not exist, $i$ only assumes value 1.

In the case of sub-zone 3.3, water contents below field capacity are not occurring. In this case $A_{l_{def}}$ is not defined and deep percolation is given directly by $Dp_{soil}(day,\,bare\;soil)$:

$$Dp_{3.3}(day) = Dp_{soil}(day,\,bare\;soil)$$  \hspace{1cm} \text{Eq. 63}

For the whole area of the bare soil, $Dp_{(day,\,bare\;soil)}$ is given by:

$$Dp_{(day,\,bare\;soil)} = \{Dp_{3.1}(day) \cdot [\text{fraction}_1 \cdot fc_{(day,\,cover\;1)}] + Dp_{3.2}(day) \cdot [\text{fraction}_2 - fc_{(day,\,cover\;2)}] + Dp_{3.3}(day) \cdot (1 - [\text{fraction}_1 + \text{fraction}_2]) \} / (1 - [fc_{(day,\,cover\;1)} + fc_{(day,\,cover\;2)}])$$  \hspace{1cm} \text{Eq. 64}

Fig. 14 shows the distribution of deep percolation for three different land covers, expressed in water column height in relation to the land fraction they occupy. Fig. 15 shows the global value of $Dp$ for the study area, computed with:

$$Dp = Dp_{\text{non-irrigated}} \cdot fc_{\text{non-irrigated}} + Dp_{\text{orchards}} \cdot fc_{\text{orchards}} + Dp_{\text{bare\;soil}} \cdot fc_{\text{bare\;soil}}$$  \hspace{1cm} \text{Eq. 65}

being $fc_i$ the fraction of the area occupied by each one of the cultures or by the bare soil. [Note: sum of $fc_i = 1$].

As mentioned in section 8.1, the deep percolation is assumed to be equal to the recharge of groundwater.
Fig. 14 – Deep percolation by vegetal cover or bare soil (values referred to the area occupied by each cover or bare soil).

Fig. 15 – Deep percolation for the whole area.
9. ANNEX 2 – METHODOLOGY FOR THE ESTIMATION OF SERIES OF DAILY PRECIPITATION, TEMPERATURE AND REFERENCE EVAPOTRANSPIRATION FOR THE CLIMATE CHANGE SCENARIOS

9.1 PRECIPITATION

The precipitation series were changed by season using the precipitation variation rates projected as a function of the climate change emission scenarios and circulation models. For the case studies the projected changes are presented in Table 3. The months of December, January and February are considered winter, the months of March, April, May are spring, June, July and August are Summer and September, October and November are autumn.

Two distinct methodologies that consider two different precipitation patterns were used:

(1) Constant variation by season: each precipitation event is affected by the variation rate of the corresponding season; for instance a variation rate of -25% means that in the projected series the projected value is equal to the actual value of the series multiplied by (1-25%). Along each year of the actual series, four distinct variation rates are applied in sequence, one for each season.

(2) Variation by removing the lower precipitation events, by season: in this methodology a season variation rate is also used, which is constant for each season of the whole time series, but in this case the lower precipitation events are removed, leaving only the higher precipitation events. The elimination of the lower precipitation events is carried out by sorting in descending way the daily values of precipitation, by accumulating precipitation using the sorted series and by eliminating the precipitation events when the total accumulated at the season reaches the projected precipitation value corresponding to the variation rate of the season.

For instance, if the winter variation rate is -25%, the sum of the projected winter precipitations should be (1-25%) = 75% of the actual total sum of winter precipitation. By sorting all winter values in descendent direction and adding each value to the previous one, an accumulated value is obtained. When this value is equal to 75% of the sum of all the winter values of the actual series, all other precipitation values are set to zero, and the remaining values of precipitation of the projected series are kept equal to the initial actual series.

With this methodology, the higher values of precipitation do not decrease, which means that the higher values of precipitation are preserved over the lower values and thus approaching the projection that with climate change precipitation is more concentrated.

In the case of increasing variation rate of precipitation (generally in winter), the same method is applied but previously all the values of the precipitation series are affected by a factor which is 1.5 times the positive variation rate. Exemplifying, if in winter the variation rate is +4%, all the values in winter are multiplied by 106% (= 1 + 4% x 1.5 = 1 + 6%), and then the series is sorted descending, and when the accumulated projected precipitation is higher than 104% of the actual precipitation series
the projected values are set to zero precipitation. Along the year the four distinct variation rates are applied in sequence, accordingly with the season.

9.2 TEMPERATURE

The required temperature series for the computation of evapotranspiration, are minimum and maximum monthly or daily temperatures. Unlike precipitation that may not occur in several days, temperature is only modified using the constant variation by season method. So temperature is corrected by adding the corrective factor given by the average season variation on temperature. During one year, four distinct constant average variations are applied in sequence, one for the spring, another for the summer, a third one for autumn and the last one to winter. These values are distinct for maximum and minimum temperatures, as referred to in Table 3.

9.3 REFERENCE EVAPOTRANSPIRATION

The changes of the climate variables due to climate change will also provoke a change on the potential and reference evapotranspiration. The most easily available climate variable on climate change scenarios is temperature. Other variables of climate that imply on potential or reference evapotranspiration are wind speed, relative humidity and solar radiation. However these climate variable changes are not so easily available. So the presented methodology for reference evapotranspiration change will only be dependent on temperature change. Also relative humidity which is dependent on vapour pressure is also dependent on temperature, which can be taken into account. It can be stated that reference evapotranspiration is more sensitive to changes on air temperature than on other climate variables. So a projection of reference evapotranspiration based only on air temperature may provide relatively acceptable values.

The FAO Penman-Monteith method (Allen et al, 1998, see section 8.3.2.2) is used to compute reference evapotranspiration.

The estimation of relative humidity on a climate change scenario follows the sequence:

1st – When only having actual values of mean actual relative humidity (HRmed) instead of minimum and maximum relative humidity, estimation of the actual vapour pressure (ea) using actual maximum (Tmax) and minimum (Tmin) temperatures:

\[ e_a = HR_{med} e_s \]  \hspace{1cm} \text{Eq. 66}

Where es is mean saturation vapour pressure, given by

\[ e_s = \frac{e^o(T_{\text{max}}) + e^o(T_{\text{min}})}{2} \]  \hspace{1cm} \text{Eq. 67}

being \( e^o(T) \) the saturation vapour pressure (kPa) at temperature \( T \) (ºC):

\[ e^o(T) = 0.6108 \exp \left( \frac{17.27 T}{T + 237.3} \right) \]  \hspace{1cm} \text{Eq. 68}
2nd – Assuming that vapour pressure remains constant, maximum (HRmax) and minimum (HRmin) relative humidity are estimated using equations:

\[
HR_{max} = \frac{e_a}{e_a^o(T_{min})}; \quad HR_{min} = \frac{e_a}{e_a^o(T_{max})}
\]  

Eq. 69

3rd – By averaging the maximum and minimum relative humidity a new calculation of relative humidity is obtained (HRmed*);

4th – By this last process it is verified that, due to the non-linearity of the variation of relative humidity with the temperature, the value now calculated (HRmed*) is different from the actual value (HRmed). Using these values a corrective factor corrHR is calculated by HRmed* / HRmed.

5th – For the calculation of the reference evapotranspiration for the situation of climate change, it is assumed that the vapour pressure is the same as in the actual series (which is a drawback of the methodology as the amount of water in the atmosphere will also vary).

6th – Thus, the minimum and the maximum relative humidity are calculated using the projected maximum and minimum temperatures for the climate change situation (using Eq. 69) and the vapour pressure (ea) calculated using Eq. 66, and taking into account the comment referred in 5th;

7th – The mean relative humidity is calculated using the minimum and the maximum relative humidity calculated in 6th. This mean relative humidity is corrected by multiplying it by the correction factor corrHR obtained in 4th in order to correct it for the non-linearity of the variation of the relative humidity with the temperature.

8th – The new value of mean relative humidity projected for the climate change scenario is used to estimate minimum relative humidity needed for running the BALSEQ_MOD model, by rearranging ans solving to HRmin Eq. 69, Eq. 67 and Eq. 66:

\[
HR_{min} = \frac{e_a^o(T_{max}) + e_a^o(T_{min})}{2HR_{med}}
\]  

Eq. 70

With the new values of relative humidity and temperature, and keeping constant the remaining variables of the Penman-Monteith equation, the reference evapotranspiration series for climate change scenarios are computed.