4. Piezometry

4.1 Data logger time series

The piezometric data shows different levels of periodical fluctuation. The measured amplitude between the lowest measurements of the groundwater table during the end of the dry season and the highest groundwater table at the end of the rainy season varies from 4 to 6 m.

The lowest groundwater tables are observed with 8 to 10 m below ground. The piezometer HVO-1 shows the deepest groundwater level with around 14 m (Fig. 4.1 and Fig. 4.2). HVO-9 at Bétérou shows the shallowest water level with around 6 m. Recharge in the rainy season lifts the groundwater table to a depth of 4 to 6 m. This means a general rise of 4 to 6 m for each observation well. The only exception is HVO-9 with a rise of only 2 m. This limited recovery at HVO-9 may indicate the fact that water in shallower depths is controlled by the surface morphology and flows horizontally away to any effluent, for example as interflow.

The rise of the groundwater table in 2005 due to the rainfalls is for some loggers less accentuated as for others. The groundwater level at HVO-1, HVO-3 and HVO-11 recovers less than for the other piezometers. The piezometers of concern are situated in the North and the Northwest of the study area. The difference is caused by the retarded and lower precipitation in the year 2005. This is already an indication for a regionally different recharge pattern and is underlined by the trend analyses made later in this chapter.

Theoretically the morphological position of the observation borehole should have a certain influence on the groundwater level and its reaction towards recharge. Most of the observation boreholes are situated at crest positions or in the upper slope.

HVO-9, instead, is placed on the foothills of a slope. HVO-9 is the only borehole with a relatively high groundwater table during the whole year. Water flows from the crests towards its position and thus the groundwater table stays generally higher. HVO-8 and HVO-10 are situated on top of crests. The two hydrographs are quite equilibrated and do not show sharp rise or fall. Recharge at crests would rather runoff towards the slopes than to prevail at the crests. At slopes the recharge of the groundwater table would fast accumulate the water coming from the crests and let it run through towards the lower valleys.

The longer time series (HVO-1, HVO-3, HVO-6, HVO-9 and HVO-11) show very well a decrease in the annual recharge for the years 2005 and 2006 due to reduced rainfalls. For the same period the base groundwater level instead is not decreasing. This means that a certain part of the aquifer stays saturated all the year long and is only slightly affected by dynamic changes of the water level.

Furthermore there is a daily fluctuation of the groundwater table in the observation wells which is thus not caused by pumping or other human intervention. Its amplitude ranges from 8 to 10 cm (see Fig. 4.3). This fluctuation is not only caused by evapotranspiration as it can be seen that the groundwater table descends as well in the night with almost the same amplitude. The observed fluctuations are supposed to be caused by diurnal tidal movements. This effect was also seen by A. Kolodziew (GeoConsult) during
Piezometry

geophysical field experiments in the Collines department in 2004 and 2005. During the rainy season these signals are less clear because of strong recharge.

Fig. 4.1: Groundwater head time series of the HVO data loggers. The data sets are filtered. Only the measurements at 5 am are shown.
Fig. 4.2: Groundwater head time series of the HVO data loggers (continued). The data sets are filtered. Only the measurements at 5 am are shown.
The installation of the data loggers depended on already existing boreholes. The financial means for drilling new observation boreholes close to existing installations like river gaugers and pluviometers were not disposed. However, some divers are next to these stations and their data series (as far as available) can be used for comparison.

The groundwater hydrograph of HVO-9 reacts to rainfalls with fast and sharp rises of the water level (Fig. 4.4). As the observation wells are drilled into the bedrock this reaction cannot be caused by real recharge towards the aquifer. It is rather the hydrostatic equilibrium which transfers the increasing pressure on the diver’s transducers. This fast exchange needs preferential pathways. Those can be fracture structures running out close to the surface. Therefore this location was chosen for an investigation with geoelectric profiling. BOHNENKAMPER (2006) measured thus two structures which can be interpreted as coarse grained clastic materials like quartz bands. Such bands might be remnants from fracture fillings (CHILTON and FOSTER 1993).
The river runoff reacts almost simultaneously with the groundwater levels. This means that the retention through the aquifer is of less importance. Most of the precipitation runs off at the surface or by interflow towards the river. Only when the rainfalls stop it can be seen that there is still base flow from the aquifer towards the Ouémé. But it is as well seen that the groundwater level still falls after the river already ran dry (Fig. 4.5). It is assumed that the withdrawal of the groundwater table during the dry season is due to deep rooting plants. Similar behaviour of an aquifer had been watched for example in studies in Botswana (BAUER et al. 2003).

The data series in Fig. 4.6 shows a gap during the important rainy season. Nevertheless, the groundwater table depletes even after the end of the river runoff. It would be interesting to compare the losses from the groundwater table with a regional estimation of plant’s water consumption.

In all cases the water table drops after the end of the river runoff still for 1.50 to 2 m. Any long range transport in riverbed sediments is not probable as for example the riverbeds of the Ouémé and the Térou River consist of bedrock only along long distances. Flow in fractures depends on their connectivity.
Fig. 4.5: Comparison of rainfall, runoff and groundwater levels at Tchétou (HVO-9 for the year 2004). The data set is filtered for 5:00 am measurements only.

Fig. 4.6: Comparison of rainfall, runoff and groundwater levels at Dogué (HVO-12) for the year 2004. The data set is filtered for 5:00 am measurements only.

The distance groundwater may cover in the bedrock depends on the fracture connectivity. The degree of connectivity within the HVO remains unknown, but is estimated to a maximum of some hundred meters only. The most probable reason for the groundwater depletion during the rest of the year is as said above outtake by plants and evaporation.
For long term interpretation of the time series a trend analysis was done. For each fully available data time series (daily 5:00 am measurements only) the recorded number of days and the measurements are summed up and multiplied. These values are used to calculate the sums of the squared deviation (Q) of the groundwater level below ground (x). For each time series the individual number of time steps (n) is used. The results are found in Tab. 4.1.

\[ Q_x = \sum x^2 - \frac{(\sum x)^2}{n} \]  
\[ (Eq. 4.1) \]

\[ Q_t = \sum t^2 - \frac{(\sum t)^2}{n} \]  
\[ (Eq. 4.2) \]

\[ Q_{xt} = \sum xt - \frac{\sum x \cdot \sum t}{n} \]  
\[ (Eq. 4.3) \]

with
\[ xt = \text{share of the trend} \]
\[ t = \text{time step of the measurement} \]

The squared sums of deviation are used to compute the coefficients \( b_0 \) and \( b_1 \) of the trend equation.

\[ b_0 = \frac{Q_{xt}}{Q_t} \]  
\[ (Eq. 4.4) \]

\[ b_1 = \frac{\sum x}{n} \]  
\[ (Eq. 4.5) \]

The quality of the trend can be evaluated by the linear correlation coefficient \( r \) (Eq. 4.6) (see LANGGUTH 1980):

\[ r = \frac{Q_{xt}}{\sqrt{Q_x \cdot Q_t}} \]  
\[ (Eq. 4.6) \]

The trend equation is formulated as follows:

\[ xt = \left( b_0 \cdot t \right) + b_1 \]  
\[ (Eq. 4.7) \]
Tab. 4.1: Trend analyses of piezometric time series data.

<table>
<thead>
<tr>
<th>Diver</th>
<th>Sum ($x^a$)</th>
<th>Sum ($t^b$)</th>
<th>Sum ($xt$)</th>
<th>Sum ($x^2$)</th>
<th>Sum ($t^2$)</th>
<th>$Q_x^c$</th>
<th>$Q_t^c$</th>
<th>$Q_{xt}^c$</th>
<th>$b_0^d$</th>
<th>$b_1^d$</th>
<th>$r^o$</th>
</tr>
</thead>
<tbody>
<tr>
<td>HVO-1</td>
<td>1.3E+04</td>
<td>5.3E+05</td>
<td>6.7E+06</td>
<td>1.6E+05</td>
<td>3.6E+08</td>
<td>1.5E+03</td>
<td>3.6E+08</td>
<td>1.3E+05</td>
<td>3.6E-04</td>
<td>12.64</td>
<td>0.17772</td>
</tr>
<tr>
<td>HVO-2</td>
<td>4.9E+03</td>
<td>2.2E+05</td>
<td>1.6E+05</td>
<td>3.8E+04</td>
<td>9.6E+07</td>
<td>1.5E+03</td>
<td>9.6E+07</td>
<td>-3.2E+04</td>
<td>-3.4E-04</td>
<td>7.30</td>
<td>-0.085</td>
</tr>
<tr>
<td>HVO-3</td>
<td>8.5E+03</td>
<td>5.3E+05</td>
<td>4.6E+06</td>
<td>7.2E+04</td>
<td>3.6E+08</td>
<td>1.9E+03</td>
<td>3.6E+08</td>
<td>2.3E+05</td>
<td>6.3E-04</td>
<td>8.56</td>
<td>0.27834</td>
</tr>
<tr>
<td>HVO-4</td>
<td>6.3E+03</td>
<td>3.7E+05</td>
<td>2.8E+06</td>
<td>4.7E+04</td>
<td>2.1E+08</td>
<td>1.0E+03</td>
<td>2.1E+08</td>
<td>1.3E+05</td>
<td>6.1E-04</td>
<td>7.60</td>
<td>0.27102</td>
</tr>
<tr>
<td>HVO-6</td>
<td>7.7E+03</td>
<td>5.3E+05</td>
<td>4.1E+06</td>
<td>5.9E+04</td>
<td>3.6E+08</td>
<td>8.8E+02</td>
<td>3.6E+08</td>
<td>7.5E+04</td>
<td>2.1E-04</td>
<td>7.64</td>
<td>0.13276</td>
</tr>
<tr>
<td>HVO-7</td>
<td>4.4E+03</td>
<td>2.2E+05</td>
<td>1.4E+06</td>
<td>3.0E+04</td>
<td>9.7E+07</td>
<td>1.0E+03</td>
<td>9.7E+07</td>
<td>-3.5E+04</td>
<td>-3.6E-04</td>
<td>6.52</td>
<td>-0.111</td>
</tr>
<tr>
<td>HVO-8</td>
<td>1.1E+04</td>
<td>5.6E+05</td>
<td>6.0E+06</td>
<td>1.2E+05</td>
<td>3.9E+08</td>
<td>9.3E+02</td>
<td>3.9E+08</td>
<td>1.7E+05</td>
<td>4.4E-04</td>
<td>10.69</td>
<td>0.28274</td>
</tr>
<tr>
<td>HVO-9</td>
<td>4.9E+03</td>
<td>5.3E+05</td>
<td>2.5E+06</td>
<td>2.4E+04</td>
<td>3.6E+08</td>
<td>5.6E+02</td>
<td>3.6E+08</td>
<td>-1.1E+04</td>
<td>-3.0E-05</td>
<td>4.77</td>
<td>-0.02435</td>
</tr>
<tr>
<td>HVO-10</td>
<td>2.1E+03</td>
<td>6.3E+04</td>
<td>3.6E+05</td>
<td>1.2E+04</td>
<td>1.5E+07</td>
<td>1.2E+02</td>
<td>1.5E+07</td>
<td>-8.6E+03</td>
<td>-5.8E-04</td>
<td>5.75</td>
<td>-0.199</td>
</tr>
<tr>
<td>HVO-11</td>
<td>8.6E+03</td>
<td>5.3E+05</td>
<td>4.5E+06</td>
<td>7.4E+04</td>
<td>3.6E+08</td>
<td>2.2E+03</td>
<td>3.6E+08</td>
<td>4.0E+04</td>
<td>1.1E-04</td>
<td>8.43</td>
<td>0.04457</td>
</tr>
</tbody>
</table>

$^a x =$ groundwater level below ground [m].

$^b t =$ day of the measurement $x$; starting with the first measurement on day zero (0).

$^c Q =$ squared deviation.

$^d b =$ regression coefficients of the trend equation.

$^o r =$ correlation coefficient.
4 Piezometry

SACHS (1997) presents a way how to prove the correlation coefficient directly. He proposed critical values for comparison with the values of $r$ in order to identify any statistical significance. The critical values are related to the number of measurements and their grade of freedom respectively and can be read from tables presented in SACHS (1997).

$$GF = n - 2$$  
(Eq. 4.8)

with

$GF = \text{degree of freedom}$

$n = \text{number of measurements}$

If the amount of $r$ is bigger as the critical values than the 5% significance is statistically assured. The comparison is given in Tab. 4.2.

**Tab. 4.2: Critical values of $r$ for the HVO divers.**

<table>
<thead>
<tr>
<th>Diver</th>
<th>$n$</th>
<th>$GF$</th>
<th>$r$</th>
<th>critical value 5% significance $^1$</th>
</tr>
</thead>
<tbody>
<tr>
<td>HVO-1</td>
<td>1028</td>
<td>1026</td>
<td>0.17772</td>
<td>0.0505</td>
</tr>
<tr>
<td>HVO-2</td>
<td>661</td>
<td>659</td>
<td>-0.08500</td>
<td>0.0740</td>
</tr>
<tr>
<td>HVO-3</td>
<td>1029</td>
<td>1027</td>
<td>0.27834</td>
<td>0.0505</td>
</tr>
<tr>
<td>HVO-4</td>
<td>856</td>
<td>854</td>
<td>0.27102</td>
<td>0.0619</td>
</tr>
<tr>
<td>HVO-6</td>
<td>1028</td>
<td>1026</td>
<td>0.13276</td>
<td>0.0505</td>
</tr>
<tr>
<td>HVO-7</td>
<td>663</td>
<td>661</td>
<td>-0.11100</td>
<td>0.0740</td>
</tr>
<tr>
<td>HVO-8</td>
<td>1057</td>
<td>1055</td>
<td>0.28274</td>
<td>0.0505</td>
</tr>
<tr>
<td>HVO-9</td>
<td>1030</td>
<td>1028</td>
<td>-0.02435</td>
<td>0.0505</td>
</tr>
<tr>
<td>HVO-10</td>
<td>355</td>
<td>353</td>
<td>-0.19900</td>
<td>0.105</td>
</tr>
<tr>
<td>HVO-11</td>
<td>1028</td>
<td>1026</td>
<td>0.04457</td>
<td>0.0505</td>
</tr>
</tbody>
</table>

$^1$ from SACHS (1997).

Out from 10 data series 5 seem to have a significant statistic time-measurement-relationship (HVO-1, HVO-3, HVO-4, HVO-6 and HVO-8). HVO-11 fails this test only close but is included into the above group. All approved series show a positive trend.

At the actual state of observation the properties of the time series are changing too strong that neither stationary nor non-stationary behaviour in time could be determined (MIDDLETON 2000).

The regarded data represents groundwater level below the ground surface. A sufficiently high value of $r$ assures the trend equation. A positive trend would mean a further depletion of the groundwater table while a negative trend represents the opposite. Thus the 6 diver data series with the assured trend would rather represent a general trend to a depletion of the groundwater table (see average trend curve in Fig. 4.7). The remaining 4 series who failed the significance test show all a negative trend. It should be mentioned that their data series are much shorter (with the exception of HVO-9). Thus any trend calculation is much more uncertain.

However, $r$ is only a statistical value. A close view to the data of the 6 approved data series reveals that the measurements started in the end of the dry season at a very low level and ended in most of the series at relatively higher levels. As the available time
series is still very short the influence of the last measured season on the overall series is considerably strong.

Fig. 4.7: Average trend equation \( XT \) for the 6 data logger time series with an approved statistical relevance.

The comparison of 5:00 am measurements at observation wells and at pumped boreholes showed no relevant difference in the depth of the groundwater table. It seems that the groundwater recovers almost fully during the night after being pumped at daytime. In rural areas with low water extraction rates the approach to install data loggers even under non-stationary conditions is reasonable when measurements are taken after sufficient recovery time. More frequently used pumps, e.g. in urban areas, might very well not recover at all.

Tab. 4.3: Coordinates of the IRD piezometers in the Donga catchment.

<table>
<thead>
<tr>
<th>ID</th>
<th>Locality</th>
<th>Type</th>
<th>X (UTM)</th>
<th>Y (UTM)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ANANPZ</td>
<td>Ananinga</td>
<td>dug well</td>
<td>380252</td>
<td>1074308</td>
</tr>
<tr>
<td>BABAPZ</td>
<td>Babayaka</td>
<td>dug well</td>
<td>342425</td>
<td>1077927</td>
</tr>
<tr>
<td>BELEPZ</td>
<td>Belefoungou</td>
<td>dug well</td>
<td>359868</td>
<td>1085200</td>
</tr>
<tr>
<td>BORTOKOPZ</td>
<td>Bortoko</td>
<td>dug well</td>
<td>379276</td>
<td>1083944</td>
</tr>
<tr>
<td>CPR-SOSSOPZ</td>
<td>CPR-SOSSO</td>
<td>piezometer</td>
<td>362792</td>
<td>1087601</td>
</tr>
<tr>
<td>DEND1PZ</td>
<td>Dendougou</td>
<td>dug well</td>
<td>360612</td>
<td>1076431</td>
</tr>
<tr>
<td>DJAKPINGPZ</td>
<td>Djakpingou</td>
<td>dug well</td>
<td>354966</td>
<td>1082777</td>
</tr>
<tr>
<td>DJOUGOUZP</td>
<td>Djougou</td>
<td>piezometer</td>
<td>353994</td>
<td>1073376</td>
</tr>
<tr>
<td>FOUNGapZ</td>
<td>Founga</td>
<td>dug well</td>
<td>345498</td>
<td>1071177</td>
</tr>
<tr>
<td>FOYOPZ</td>
<td>Foyo</td>
<td>dug well</td>
<td>383357</td>
<td>1073726</td>
</tr>
<tr>
<td>GANPZ</td>
<td>Gangamou</td>
<td>dug well</td>
<td>374354</td>
<td>1088880</td>
</tr>
<tr>
<td>GAOUNGAPZ</td>
<td>Gaounga</td>
<td>piezometer</td>
<td>384600</td>
<td>1076758</td>
</tr>
<tr>
<td>KOKOPZ</td>
<td>Koko Sika</td>
<td>dug well</td>
<td>383108</td>
<td>1079976</td>
</tr>
<tr>
<td>KOLOPZ</td>
<td>Kolokonde</td>
<td>dug well</td>
<td>366117</td>
<td>1093400</td>
</tr>
<tr>
<td>KOUAPZ</td>
<td>Koua</td>
<td>dug well</td>
<td>367437</td>
<td>1079381</td>
</tr>
<tr>
<td>MONEMOSPZ</td>
<td>Mone mosquée</td>
<td>dug well</td>
<td>373542</td>
<td>1075300</td>
</tr>
<tr>
<td>PAMIPZ</td>
<td>Pamido</td>
<td>dug well</td>
<td>350265</td>
<td>1074369</td>
</tr>
<tr>
<td>SANKPZ</td>
<td>Sankoro</td>
<td>dug well</td>
<td>369989</td>
<td>1091610</td>
</tr>
<tr>
<td>SERIVERIPZ</td>
<td>Sérivéri</td>
<td>dug well</td>
<td>361901</td>
<td>1074174</td>
</tr>
<tr>
<td>TCHAPZ</td>
<td>Tchakpaissa</td>
<td>dug well</td>
<td>359596</td>
<td>1073986</td>
</tr>
<tr>
<td>TEWAMOUPZ</td>
<td>Téwamou</td>
<td>dug well</td>
<td>377571</td>
<td>1085723</td>
</tr>
</tbody>
</table>
The same ambivalent information is given by trend analyses from different piezometers in the Donga catchment. Since the beginning of the year 2000 the IRD measures the groundwater table with 21 data loggers installed in open dug wells and observation wells distributed in this catchment (see Tab. 4.3). The data was available for analysis by the kind permission of Luc Séguis (IRD Cotonou). The time series were controlled through and equally analysed for trends and periodicity. The very same phenomena are seen in this data set. The correlation coefficient $r$ is sometimes fitting well sometimes not. Trends are in some cases slightly positive in others slightly negative. However, these data series comprises only the years 2000 to 2003. This is hardly more than the time series from the IMPETUS divers and thus is still not sufficient for a clear interpretation.

It was shown in this chapter that the available time rows of piezometric data are still not sufficient to determine any trend of the groundwater table development. Actually the impact of increasing groundwater consumption by the people or of the declining precipitation in 2005 and 2006 on the general groundwater table cannot be determined. Longer time series would give the opportunity to investigate in detail the periodicity and autocorrelation of the data.

### 4.2 Regionalisation of piezometric data

It is a common procedure to generate contour lines of the groundwater table for regional research. Groundwater contours allow the determination of the groundwater flow direction. The regional distribution of local groundwater levels is hereby achieved by kriging.

These measurements are interpolated applying the kriging algorithm on the depth of the groundwater table below the ground. The depth of the groundwater level is equal to the thickness of the unsaturated zone. The thickness of the unsaturated zone can be subtracted from the DEM (see Fig. 4.11).

Many parts of the HVO are not accessible or do not show any settlements with wells or other groundwater points. Furthermore the DEM showed that the surface of the Area is slightly undulated and shows a number of smaller subcatchments. Each subcatchment has an individual distribution of crests, slopes and valleys. And at each of these positions the groundwater level would differ. No representative data for groundwater levels for all these subcatchments and their morphological characteristics is available. Only a general picture of the major groundwater flow direction in the HVO can be given. Groundwater flow would be on a local scale directed towards the locally draining river. On the regional scale flow is roughly directed towards the Ouémé valley and from there towards the South. But it has to be understood that the groundwater contours generally follows just the morphological features as the water table is quite close to the surface.

The distance between the data logger positions makes it impossible to integrate the actual morphology in interpolation algorithms unless an extremely simplified terrain model is used. However, during the field campaigns the depth of the groundwater table was regularly measured at numerous open dug wells (Fig. 4.8 and Tab. 4.4, for details see Annex 3). In respect to the size of the HVO, the completion of each measurement campaign required several weeks. The open dug wells are always in use by the villagers.

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8 A subcatchment in the Northwest of the HVO. The Donga catchment is object of intensive research of the CATCH project of the IRD.
Thus, the water table is not stationary, except for measurements in the early morning hours.

Tab. 4.4: Number of manual piezometric measurements on wells realised in and around the study area during different seasons from 2004 to 2006.

<table>
<thead>
<tr>
<th>Season</th>
<th>N° of measurements</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dry season 2004</td>
<td>113</td>
</tr>
<tr>
<td>Rainy season 2004</td>
<td>110</td>
</tr>
<tr>
<td>Dry season 2005</td>
<td>187</td>
</tr>
<tr>
<td>Rainy season 2005</td>
<td>83</td>
</tr>
</tbody>
</table>

Fig. 4.8: Regional distribution of all manual and automatic piezometric measurements realised in the vicinity of the study area (Projection: UTM, Zone 31P, WGS 84).

The software package SURFER© has been used to produce contour maps for the dry seasons 2004/2005 and for the rainy seasons 2004/2005. Each data set was kriged. The resulting grid data was arranged in the same dimensions as the DEM. The same grid size facilitates mathematical grid operations.

The difference between the groundwater surfaces of the dry season 2004 and the rainy season 2004 (Fig. 4.9) and respectively the same for the year 2005 (Fig. 4.10) was calculated. The remaining volume was then multiplied with the storage coefficient for the regolith aquifer (4% as proposed by ENGALENC 1978) in order to obtain the volume of effectively gained groundwater at the end of the rainy season considered as recharge towards the aquifers. The effective recharge between the dry season 2004 and the dry season 2005 was then calculated by the same procedure (Fig. 4.12).
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Fig. 4.9: The groundwater levels during the rainy season are generally higher as in the dry season 2004 (Projection: UTM, Zone 31P, WGS 84).

Fig. 4.10: Distribution of the groundwater differences in 2005 (Projection: UTM, Zone 31P, WGS 84).
Fig. 4.11: Exemplary interpolation of manually made groundwater measurements. The data was interpolated and then subtracted from the DEM.
By this procedure a recharge of 240 mm in the year 2004 and 158 mm for the year 2005 was calculated for the HVO area. Both values are too high in comparison to values of 30 to 100 mm/a proposed by GIERTZ (2005), ENGALENC (1978), ENGALENC (1985), SANDWIDI et al. (2006) and SOGREAH and SCET (1997).

Tab. 4.5: Results from the grid based recharge calculation. The volume is calculated by subtracting the grids of the interpolated manual field measurements.

<table>
<thead>
<tr>
<th>Grid</th>
<th>Storage Volume [m³]</th>
<th>Recharge [mm]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rainy – Dry 2004</td>
<td>3408108893.79</td>
<td>240</td>
</tr>
<tr>
<td>Rainy – Dry 2005</td>
<td>2243401570.47</td>
<td>158</td>
</tr>
<tr>
<td>Dry 2005 – Dry 2004</td>
<td>455483045.27</td>
<td>32</td>
</tr>
</tbody>
</table>

The recharged water between the end of the dry season 2004 and the end of the dry season 2005 was calculated as 32 mm. This means that the groundwater table rose in total from one year to the other. The amount of 32 mm fits very well with the general assumptions for recharge made by the above citations.

The elevated groundwater table differences in the figures above are due to the regional distribution of precipitation but as well by the great time intervals between the measurements. The groundwater contours of the dry season 2004 are used as initial head conditions of the numerical model (Chapter 8).
5. Hydrochemistry

5.1 Physico-chemical characteristics

5.1.1 Temperature

Air and water temperature (in °C) have been measured on-site for each sample. In general, water from the regolith aquifer was sampled from open dug wells where it is in direct contact with the atmosphere. The borewells instead are fully cased. The temperature in borewells is slightly elevated (~1-2°C) compared to the dug wells. The solarised metallic casings (in general Ø=8”) conduct the heat from the surface towards the relatively small water volume stocked inside.

Tab. 5.1 gives an overview about the population statistics of measured water temperatures for all sampling campaigns. Correlation of water against air temperature (Fig. 5.1) showed only a small but positive correspondence ($r^2=0.3$). The water temperature principally depends on the general climatic conditions during the sampling period and not from any inflow from another hydrogeological background as it might be suspected in a fractured environment.

Tab. 5.1: Population statistics of temperature measurements.

<table>
<thead>
<tr>
<th>water temperature (°C)</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Number</td>
<td>246</td>
</tr>
<tr>
<td>Mean</td>
<td>29.49</td>
</tr>
<tr>
<td>Median</td>
<td>29.5</td>
</tr>
<tr>
<td>St. dev.</td>
<td>1.48</td>
</tr>
<tr>
<td>Minimum</td>
<td>25</td>
</tr>
<tr>
<td>Maximum</td>
<td>35.7</td>
</tr>
</tbody>
</table>

Equation $Y = 0.205 \times X + 22.5$
Number of data points used = 184
Coef. of determination, $r^2 = 0.3$

Fig. 5.1: Determination of correlation between air and water temperature.

5.1.2 pH

The pH besides redox potential and water temperature controls the solubility of many agents in groundwater. It further influences the ion exchange and the sorption capacity of
the surrounding rock. The value of pH ranges for the borewells from 5.7 to 7.47 and for the dug wells from 5.43 to 8.03 (see Fig. 5.2).

![Fig. 5.2: pH data of boreholes (a) and dug wells (b) presented respectively in histograms.](image)

A pH around 7 is generally typical for rainwater, and may indicate recently infiltrated recharge or just perched shallow aquifers. Changes of pH values in natural waters are often caused by different contents of carbonic acid. Bicarbonate in water buffers the acid and equilibrates the pH within the range of 5.5 to 8.0 (HÖLTING 1996). When pH<5.5 silicates, like feldspar and clay minerals, may work as buffers. The buffering consumes H+. The content of H2CO3 declines and HCO3− increases. The reaction follows the simplified equilibrium reaction:

\[ H_2CO_3 \leftrightarrow HCO_3^- + H^+ \]  

(Eq. 5.1)

Precipitation is in equilibrium with atmospheric CO2 which has a partial pressure of 10^{-3.5} atm. Due to microbial reactions the soil zone holds more CO2. Typical partial pressures of CO2 achieved in the soil zone are 10^{-2.0} to 10^{-1.5} atm (DREVER 1997).

A higher content of CO2 in the groundwater increases the solubility of CaCO3. Thus more HCO3− dissolves. For each partial pressure pCO2 a certain relationship of pH/HCO3− is achieved. Fig. 5.3 demonstrates the relationships for 3 different pCO2. The curve for atmospheric pCO2 shows a gradual increase of bicarbonate only for higher pH values. This scatter plot gives a general footprint for recharge mechanisms in the study area. Most of the samples are from different sources and taken during different seasons and plot around the curves of the typical pCO2 in soil zones. This means that all sampled water passed through the soil zone.

If groundwater is directly recharged from the surface, for example by open fractures, the samples would plot closer to the curve for pCO2=10^{-3.5} atm. If groundwater travelled a rather long passage under higher pCO2 conditions the scatter of points would be below the soil zone pCO2 relationships.
5.1.3 Electrical conductivity

During sampling it was recognised that the electric conductivity (EC) at most locations did not change during the seasons (see Chapter 5.3). At some dug wells relatively high values are measured (800 to >1500 μS/cm). This behaviour is observed at these points for all seasons. The same is seen for nearby borewells. EC ranges for borewells from 92 to 1780 μS/cm and for dug wells from 10 to 1725 μS/cm. The histograms a) and b) in Fig. 5.4 do not show a clear Gaussian normal distribution for all samples. In each of the histograms a second population appears with typically higher charges. The possible extend of the two additional population is marked by circles.

This picture demonstrates that the source for the higher charge must be in the regolith aquifer wherefrom groundwater infiltrates the bedrock.
The data sets for the dug wells and the boreholes were regionalised by kriging (Fig. 5.5a and Fig. 5.5b). From multiple measurements at the same point only the median EC value was included. The variogram includes a nugget effect of 10 and an exponential model. The comparison of the data shows that elevated EC appears more often in wells. But it is seen as well that the higher charged boreholes are situated where already higher electric conductivities in wells were measured.

Higher EC values appear especially in the South of the study area. This area includes the villages of Dogué, Wari-Maro, Ouannou and Kikélé. While in Dogué, Wari-Maro and Ouannou the high conductivity is observed in borewells too, in Kikélé and Igboromakoro (a village next to Dogué with a handpump) instead no increase of EC in the pumps is observed. It seems that not in all villages higher charged groundwater from the regolith aquifer infiltrated towards the bedrock. These findings will be discussed in Chapter 5.2.

The case of Sonoumoun is discussed in Chapter 5.4.3. At Kori the regolith layer is very thin (<5 m). Contamination from the surface might pass to the bedrock rather fast. Increased concentration at the pump of Sérou might have geogenic origins in regard to its hydrochemical composition (see Annex 1 and Fig. 5.24).

A map generated by kriging of EC measurements taken from the BDI shows that southward of the HVO the EC is generally elevated (see Fig. 5.6). Especially around Dassa and in its South the EC values are higher. The reason for it is a thinner regolith and thus a direct contamination of groundwater due to human activities. The South of the HVO instead is characterised by very thick regolith. At Dogué the regolith is thicker than 20 m. The elevated EC might be caused by a special hydrochemical environment (Chapter 5.2.3).

Fig. 5.5: Regionalised EC data for the (a) bedrock aquifer and for the (b) regolith aquifer (Projection: UTM, Zone 31P, WGS 84).
5.1.4 Redox potential

Redox levels in groundwater are mostly determined by the quantity of oxygen and nitrate, by pH and by the relation of the redox couples (Fe$^{2+}$/Fe$^{3+}$, Mn$^{2+}$/Mn$^{3+}$, S$^{2-}$/SO$_4^{2-}$). The most important variables (after DREVER 1997) in natural systems are therefore:

- Oxygen content of recharge water.
- Distribution of potential redox buffers in the aquifer. The redox levels in groundwater often corresponds to buffering by the redox pairs Mn$^{2+}$/MnO$_2$, Fe$^{3+}$/FeOH$_3$, or Fe$^{2+}$/Fe$_2$O$_3$.
- Circulation rate of the groundwater. The pe of groundwater depends very much on its residence time in the aquifer. Longer residence time causes a lower pe.

The electron activity can be expressed in units of volts (Eh) or in units of electron activity (pe). Eh [V] and pe [-] are related by the equation:

$$pe = \frac{F}{2.303 \cdot R \cdot T} \cdot Eh$$

(Eq. 5.2)

$F =$ Faraday's constant (96.484 KJ)
$T =$ Temperature [K]
$R =$ gas constant \([8.314E-03 \text{ kJ/K*mol}]\)
At 25°C $p_e$ can be approximated by:

$$p_e = 16.9 \cdot Eh$$  
(Eq. 5.3)

The scatter (Fig. 5.7) of oxygen measurements against the redox potential as electron activity ($p_e$) reveals a rather good correlation ($r^2 = 0.84$).

![Fig. 5.7: The oxygen content plotted against the redox potential.](image)

Almost all samples from ground and surface water show Eh values of generally oxidising conditions (compare with Fig. 5.8). 4 samples showed negative values. All of them are borewells (D04-H-ANM-1, R05-H-OUB-P, D06-H-WEWE-P, D06-H-BET-P). The borehole at Ouberou (OUB-P) is defect since a long time. The water is already ochre because of corrosion. Its water must be stagnant and not very well aerated. This sample is excluded from further interpretation.

![Fig. 5.8: Redox potential and $p_e$ range encountered in natural systems at near-neutral pH (modified from Sigg 1999).](image)

The other three borewells are regularly used. They show negative values only during the dry season. But Tab. 5.2 a) - d) demonstrates that there is no typical trend towards low redox potentials during the dry seasons. Borewells and dug wells may even show the opposite behaviour. Possibly the three borewells were still fed by water stored in extended fractures with less oxygen contents before being recharged by fresher water from the regolith aquifer.

The redox electrode needs a long time to stabilise during the measurement. Sigg (2000) showed that only the use of a comparing electrode, e.g. hydrogen, in the laboratory deliver reasonable redox values. The influence of atmospheric oxygen, which can modify
the redox potential, cannot be excluded during field measurements (Käss and Seeburger 1989).

5.1.5 Oxygen

The mean and the median values of oxygen of well samples for all sampling campaigns are higher than those for the borehole samples (see Tab. 5.3). In wells and boreholes groundwater remains rather stagnant under constant exchange with the atmosphere. Thus oxygen contents at these places are barely representing conditions in the aquifer. Whirling and ventilation occur during pumping and drawing of the water by the villagers. This might be the reason why no further correlations between oxygen and other measurements or chemical constituents were found. Oxygen was not regarded for further investigations.

Tab. 5.2: Population statistics of redox measurements in the study area for boreholes (a+c) and wells (b+d) for the dry seasons and for the rainy seasons.

<table>
<thead>
<tr>
<th></th>
<th>a) Dry season</th>
<th>b) Dry season</th>
</tr>
</thead>
<tbody>
<tr>
<td>boreholes</td>
<td></td>
<td>wells</td>
</tr>
<tr>
<td>Number</td>
<td>66</td>
<td>85</td>
</tr>
<tr>
<td>Mean</td>
<td>140.8</td>
<td>197.6</td>
</tr>
<tr>
<td>Median</td>
<td>139</td>
<td>201</td>
</tr>
<tr>
<td>Modus</td>
<td>224</td>
<td>225</td>
</tr>
<tr>
<td>st. Dev.</td>
<td>92.8</td>
<td>73.6</td>
</tr>
<tr>
<td>Minimum</td>
<td>-65</td>
<td>28</td>
</tr>
<tr>
<td>Maximum</td>
<td>320</td>
<td>377</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>c) Rainy season</th>
<th>d) Rainy season</th>
</tr>
</thead>
<tbody>
<tr>
<td>boreholes</td>
<td></td>
<td>wells</td>
</tr>
<tr>
<td>Number</td>
<td>63</td>
<td>30</td>
</tr>
<tr>
<td>Mean</td>
<td>138.1</td>
<td>154.2</td>
</tr>
<tr>
<td>Median</td>
<td>148</td>
<td>155</td>
</tr>
<tr>
<td>Modus</td>
<td>180</td>
<td>182</td>
</tr>
<tr>
<td>st. Dev.</td>
<td>66.9</td>
<td>53.7</td>
</tr>
<tr>
<td>Minimum</td>
<td>15</td>
<td>51</td>
</tr>
<tr>
<td>Maximum</td>
<td>264</td>
<td>255</td>
</tr>
</tbody>
</table>

Tab. 5.3: Population statistics of oxygen measurements of boreholes (left) and wells (right).

<table>
<thead>
<tr>
<th>boreholes</th>
<th>content [mg/l]</th>
<th>saturation [%]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number</td>
<td>67</td>
<td>67</td>
</tr>
<tr>
<td>Mean</td>
<td>2.33</td>
<td>31.39</td>
</tr>
<tr>
<td>Median</td>
<td>2.43</td>
<td>33</td>
</tr>
<tr>
<td>St. dev.</td>
<td>1.05</td>
<td>14.34</td>
</tr>
<tr>
<td>Minimum</td>
<td>0.62</td>
<td>8.4</td>
</tr>
<tr>
<td>Maximum</td>
<td>4.62</td>
<td>65.7</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>wells</th>
<th>content [mg/l]</th>
<th>saturation [%]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number</td>
<td>73</td>
<td>73</td>
</tr>
<tr>
<td>Mean</td>
<td>4.58</td>
<td>63.89</td>
</tr>
<tr>
<td>Median</td>
<td>4.61</td>
<td>66</td>
</tr>
<tr>
<td>St. dev.</td>
<td>0.69</td>
<td>10.64</td>
</tr>
<tr>
<td>Minimum</td>
<td>2.01</td>
<td>26.4</td>
</tr>
<tr>
<td>Maximum</td>
<td>5.93</td>
<td>82.3</td>
</tr>
</tbody>
</table>
5.2 Hydrochemical parameters

5.2.1 Distinction of hydrochemical groups

Due to the field conditions and the lack of technical data it was concluded that dug wells penetrate only the regolith aquifer while the borewells tap water from fractures in the bedrock. The measurements of the electrical conductivity (Chapter 5.1.3) revealed that the groundwater from each aquifer is not normal distributed. A number of samples from both aquifers showed a relatively high conductivity which might be caused by a different hydrochemical environment. In order to describe the different chemical characteristics from each aquifer it was necessary to distinguish between samples with a common lower EC and those with elevated EC.

The results of rang analysis (Fig. 5.9) for dug well samples and for borewell samples plot in the beginning as smooth curves. Both curves have a steep slope after the turning point at 450 µS/cm.

![Rang distribution in percent of all electric conductivity measurements for dug wells and borewells made in the field during the period of 2004 to 2006.](image)

The dug well samples plot generally lower than the bedrock samples. But above the inflexion point at ~75% (dug wells) and at ~80% (borewells) respectively, it is seen that the higher charge is principally found in the regolith aquifer. However the maximum EC values are for both groups similar.

The samples above the inflexion points are assigned to a so-called group 3. Comparatively, the revised samples from dug wells are called group 1 and the revised borewell samples are called group 2 (see Tab. 5.4). In Chapter 5.3 it is explained that some samples from the regolith aquifer (Group 1) show a strong precipitation signature. They are discarded from the group 1 samples.

As described in Chapter 5.1.3, anthropogenic contamination is the reason for high charges at some locations (Kori, Sonoumoun). Typically, human influence on
groundwater quality is marked by elevated nitrate and chloride values. Samples with excessive nitrate (NO$_3^-$ > 100 mg/l) are excluded from group 3 (see Tab. 5.5).

Additionally it was necessary to exclude 5 samples (Tab. 5.6) from group 2. They are borewells which were already suspicious during sampling because of either low redox values or very high iron and manganese concentrations. All of these borewells were visibly old and people complained about its smell and taste. It appears that corroding casings or clogged filters have changed sufficiently the quality of the water tapped in these boreholes.

Tab. 5.4: Assignment of aquifers to sample groups.

<table>
<thead>
<tr>
<th>Group</th>
<th>Groundwater</th>
<th>Source</th>
<th>Seasons</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>regolith*</td>
<td>dug wells</td>
<td>all</td>
</tr>
<tr>
<td>2</td>
<td>bedrock</td>
<td>borewells</td>
<td>all</td>
</tr>
<tr>
<td>3</td>
<td>regolith +</td>
<td>dug wells +</td>
<td>all</td>
</tr>
<tr>
<td></td>
<td>bedrock</td>
<td>borewells</td>
<td></td>
</tr>
</tbody>
</table>

*Samples with strong precipitation signature were discarded.

The map in Fig. 5.10 shows the distribution of the group samples. Most samples from group 3 are taken in the southern part of the HVO. In group 3 dug wells plot generally higher than the borewells (Fig. 5.9). This may indicate that the origin for this higher charge is to be found in the regolith itself, either by contamination from the surface or by special hydrogeochemical conditions.

Fig. 5.10: The distribution of all hydrochemical groups in the study area (Projection: UTM, Zone 31P, WGS 84).
### 5 Hydrochemistry

#### Tab. 5.5: Samples excluded from group 3 because of too high grades of contamination.

<table>
<thead>
<tr>
<th>Sample*</th>
<th>Location</th>
<th>X UTM</th>
<th>Y UTM</th>
</tr>
</thead>
<tbody>
<tr>
<td>D04-W-BANI-1</td>
<td>Banigri</td>
<td>424768</td>
<td>1012407</td>
</tr>
<tr>
<td>R04-W-BANI-1</td>
<td>Banigri</td>
<td>424768</td>
<td>1012407</td>
</tr>
<tr>
<td>D06-W-BDOG-1</td>
<td>Dogué</td>
<td>383221</td>
<td>1006210</td>
</tr>
<tr>
<td>D04-W-DAR-1</td>
<td>Daringa</td>
<td>393970</td>
<td>1045257</td>
</tr>
<tr>
<td>D04-H-KAR-1</td>
<td>Kari</td>
<td>444818</td>
<td>1100288</td>
</tr>
<tr>
<td>D05-W-KAR-2</td>
<td>Kari</td>
<td>445010</td>
<td>1100335</td>
</tr>
<tr>
<td>D05-H-OUN-P</td>
<td>Ouannou</td>
<td>395894</td>
<td>997902</td>
</tr>
<tr>
<td>R05-H-OUN-P</td>
<td>Ouannou</td>
<td>395894</td>
<td>997902</td>
</tr>
<tr>
<td>D05-F-PRE-1</td>
<td>Péréré</td>
<td>499279</td>
<td>1083351</td>
</tr>
<tr>
<td>D04-W-PELE-1</td>
<td>Pélélina</td>
<td>350353</td>
<td>1047339</td>
</tr>
<tr>
<td>D05-W-PELE-1</td>
<td>Pélélina</td>
<td>350353</td>
<td>1047339</td>
</tr>
<tr>
<td>R05-H-SEB-P</td>
<td>Sébou</td>
<td>443940</td>
<td>1029716</td>
</tr>
<tr>
<td>D06-H-SEB-P</td>
<td>Sébou</td>
<td>443940</td>
<td>1029716</td>
</tr>
<tr>
<td>D04-W-SON-1</td>
<td>Sonoumoun</td>
<td>420434</td>
<td>1079934</td>
</tr>
<tr>
<td>D05-W-SON-3</td>
<td>Sonoumoun</td>
<td>420535</td>
<td>1080105</td>
</tr>
<tr>
<td>R04-W-TAR-2</td>
<td>Tamarou</td>
<td>466791</td>
<td>1076529</td>
</tr>
</tbody>
</table>

#### Tab. 5.6: Samples excluded from group 2.

<table>
<thead>
<tr>
<th>Sample*</th>
<th>Location</th>
<th>X UTM</th>
<th>Y UTM</th>
</tr>
</thead>
<tbody>
<tr>
<td>D04-H-ANM-1</td>
<td>Anoum</td>
<td>334784</td>
<td>1072403</td>
</tr>
<tr>
<td>D05-F-GNG-1</td>
<td>Guingamou</td>
<td>494107</td>
<td>1057124</td>
</tr>
<tr>
<td>R05-H-IGER-P</td>
<td>Igbéré</td>
<td>385405</td>
<td>994399</td>
</tr>
<tr>
<td>R05-H-KPS-P</td>
<td>Kpéssou</td>
<td>409973</td>
<td>1028012</td>
</tr>
<tr>
<td>R05-H-OUTH-P</td>
<td>Ouberou</td>
<td>415657</td>
<td>1018023</td>
</tr>
</tbody>
</table>
5 Hydrochemistry

5.2.2 Hydrochemical characteristics of the groundwater in the HVO

Recharge for all hydrochemical groups in the HVO area originates in precipitation (see also Chapter 6). Percolating downward towards the bedrock, the recharged water alternates progressively with increasing depth. The change is generally so smooth that clear hydrochemical boundaries cannot be drawn. In order to describe the general composition of each hydrochemical group population statistics were taken (Tab. 5.7).

Tab. 5.7: Population statistics of the hydrochemical parameters for each group (see Annex 1).

<table>
<thead>
<tr>
<th>Group 1</th>
<th>pH</th>
<th>EC [µS/cm]</th>
<th>HCO₃⁻ [mg/l]</th>
<th>Cl⁻ [mg/l]</th>
<th>SO₄²⁻ [mg/l]</th>
<th>NO₃⁻ [mg/l]</th>
<th>PO₄³⁻ [mg/l]</th>
<th>SiO₂ [mg/l]</th>
<th>Na⁺ [mg/l]</th>
<th>K⁺ [mg/l]</th>
<th>Ca²⁺ [mg/l]</th>
<th>Mg²⁺ [mg/l]</th>
<th>Fe₅₇ [mg/l]</th>
<th>Mn²⁺ [mg/l]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Min</td>
<td>5.43</td>
<td>85</td>
<td>9.15</td>
<td>0.38</td>
<td>0</td>
<td>0</td>
<td>0.074</td>
<td>4.79</td>
<td>3.08</td>
<td>1.19</td>
<td>6.6</td>
<td>0.55</td>
<td>0</td>
<td>0.006</td>
</tr>
<tr>
<td>Max</td>
<td>7.76</td>
<td>409</td>
<td>244</td>
<td>56.45</td>
<td>30.71</td>
<td>105.47</td>
<td>2.175</td>
<td>54.44</td>
<td>50.5</td>
<td>39.14</td>
<td>50.96</td>
<td>21.06</td>
<td>0.745</td>
<td>1.75</td>
</tr>
<tr>
<td>Mean</td>
<td>6.76</td>
<td>206</td>
<td>94.55</td>
<td>5.43</td>
<td>2.3</td>
<td>6.66</td>
<td>0.715</td>
<td>30.3</td>
<td>14.71</td>
<td>4.12</td>
<td>22</td>
<td>4.98</td>
<td>0.06</td>
<td>0.034</td>
</tr>
<tr>
<td>Median</td>
<td>6.8</td>
<td>215.71</td>
<td>97.25</td>
<td>10.08</td>
<td>3.48</td>
<td>18.39</td>
<td>0.94</td>
<td>29.93</td>
<td>14.84</td>
<td>5.46</td>
<td>21.49</td>
<td>6.05</td>
<td>0.093</td>
<td>0.081</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
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<tbody>
<tr>
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<td>92</td>
<td>61</td>
<td>0.4</td>
<td>0.275</td>
<td>0.275</td>
<td>0.015</td>
<td>7.83</td>
<td>9.9</td>
<td>2.4</td>
<td>6.1</td>
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<td>108.3</td>
<td>5.2</td>
<td>49.13</td>
<td>31</td>
<td>18</td>
<td>71</td>
<td>36</td>
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<tr>
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<td>161.65</td>
<td>3.3</td>
<td>1.7</td>
<td>0.39</td>
<td>0.67</td>
<td>32.77</td>
<td>16.3</td>
<td>5.6</td>
<td>30</td>
<td>10</td>
<td>0.55</td>
<td>0.033</td>
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<tr>
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<td>174.35</td>
<td>8.2</td>
<td>2.94</td>
<td>12.5</td>
<td>0.77</td>
<td>32.55</td>
<td>17.3</td>
<td>5.7</td>
<td>31.57</td>
<td>11.95</td>
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<tbody>
<tr>
<td>Min</td>
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<td>446</td>
<td>228.75</td>
<td>9.11</td>
<td>3.87</td>
<td>3.7</td>
<td>0.043</td>
<td>9.88</td>
<td>20.48</td>
<td>3.19</td>
<td>35</td>
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<td>110.2</td>
<td>160.4</td>
<td>95</td>
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<td>38.5</td>
<td>149.9</td>
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<td>347.7</td>
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<td>27.57</td>
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<td>46.53</td>
<td>0.69</td>
<td>26.23</td>
<td>61.33</td>
<td>20.06</td>
<td>74.97</td>
<td>38.8</td>
<td>0.07</td>
<td>0.13</td>
</tr>
</tbody>
</table>

In Fig. 5.11 the box plots for EC measurements from group 1 (dug wells in the regolith) and group 2 (borewells in the bedrock) are compared. The 25- and the 75-percentile of each plot cover mostly the same range of EC. This picture is conclusive as the groundwater of both aquifers has the same source. The median of group 1 is lower than for the group 2. This is explained by the dilution of regolith groundwater by fresh rainwater, while Bedrock groundwater prevails in the fractures and gets more concentrated.

Fig. 5.11: Whisker-Box-Plot for the electric conductivity measured from dug wells and borewells.
The plot of all three groups into a Piper diagram shows very well their overlapping hydrochemical characteristics (Fig. 5.12). Groundwater in the HVO is generally of a Na-Ca-(Mg)-HCO₃ type and is rich in silicon. This type of water is typical for shallow groundwater systems in crystalline areas (Singhal and Gupta 1999). With increasing depth an increase of magnesium is observed. Any other differentiation is at least hidden.

![Piper diagram of hydrochemical groups in the HVO.](image)

*Fig. 5.12: Piper plot of all three hydrochemical groups in the HVO.*

The Schoeller diagram (Fig. 5.13) shows the original rainfall composition in relation to the designed hydrochemical groups. It is striking that the sulphate concentration for group 1 and group 2 equals that of the precipitation. The general hydrochemical facies of precipitation, group 1 and group 2 are comparable and represent the dissolution of mineral phases during the downward percolation of groundwater.

The ionic make-up of groundwater is driven by weathering reactions and is controlled by the pCO₂ of the recharge water (Hutchings and Petrich 2002). Plagioclase feldspars, biotite, and other dark minerals weather more rapidly than K-feldspars or quartz (Appelo and Postma 1999; Hutchings and Petrich 2002).

Potassium and magnesium are derived primarily from biotite (Hutchings and Petrich 2002). Chloride and sulphate are not dissolved by silicate weathering but generally indicate mixing of water from other sources (see Chapter 5.2.3). The concentration of dissolved silicon depends almost entirely on the weathering of plagioclase and related minerals. Kaolinite is the first alteration product formed by silicate weathering.
Analysis results show that the dominating anion is bicarbonate whose amount is directly dependant from the availability of organic substances and the pH of the precipitation water.

Fresh precipitations are generally slightly acidic. A decrease in pH will weather albite to gibbsite and then to kaolinite. When the water remains stagnant then Na-montmorillonite is formed. The leaching effect depends on the intensity of the rainfalls. Stronger leaching links to kaolinite while less rainfall plots towards the montmorillonite zone (APPELO and POSTMA 1999). Leaching forms kaolinite in the surface layers. This is well demonstrated by the milky coloured water of surface runoff, rivers and shallow dug wells.

Hydrochemical analyses made at the Nalohou test site (UTM 347124/1077311) shows that under a slightly acid pH more Al\(^{3+}\) is dissolved (Fig. 5.14). Under neutral pH conditions in the deeper regolith Al\(^{3+}\) gets less soluble. It is assumed that the remaining aluminium remains in the weathering product. The charge of the dissolved cations must be balanced with H\(^+\).

Influences of rock and soil types are hardly to be found in seepage water and phreatic groundwater regardless of the differences in climate or vegetation (LOEHNERT 1988). Maturation occurs within the saturated zone in which groundwater composition might be affected by the rock nature.
A high CO₂ content related to a decrease of pH causes an accelerated silicate weathering with an increasing HCO₃⁻-value. The hydrolysis of silicates liberates the cations Na⁺, K⁺, Ca²⁺, Mg²⁺ and Fe²⁺/³⁺. Additionally, a higher partial pressure of CO₂ increases the solution of Ca- and Mg-carbonates (STÖBER 1995). Sodium and silicate dioxide are less absorbed by plant roots and are hence relatively enriched in groundwater against other cations. The modification of the Ca²⁺/Mg²⁺-ratio at the expense of the latter is not observed in the unsaturated zone and is definitely established in the saturated zone (ROOSE and LELONG 1981).

The deeper the fractures the deeper groundwater can enter the massif. Groundwater flow in hydrogeological massifs is exclusively connected to fractures and faults. Especially the silicate dioxide content can be used to describe the deep circulation of waters. The plots of Fig. 5.15a) and b) show the relatively similar silicon concentration in all three groups. Group 1 shows the widest range of scatter which is caused by dilution in shallower zones and as well by the very heterogeneous build up of the dug wells. Fig. 5.15a) and b) shows that group 3 has the highest chloride and calcium concentration but does not differ in its silicon concentration.

Fig. 5.15: a) Plot of silicon against chloride and b) plot of silicon against calcium.

Group 2 is the most homogeneous. Its samples scatter mostly in a well limited area for most of the presented diagrams. Fig. 5.16a) shows that the concentration of sodium in group 2 is rather constant, while chloride gets relatively concentrated. Both groups 1 and 3, instead, show a relative increase of the two parameters. On the other hand it is seen as well in Fig. 5.16b) that sodium stays in constant concentration against calcium. The relationship Na/Ca is roughly interpreted as 1:2 and can be explained by ion exchange on clay minerals as represented by Eq. 5.4:

\[
2Na^+ + Ca - X_2 \rightleftharpoons 2Na^- X + Ca^{2+}
\]

Montmorillonite is a typical agent for ion exchange. The observed changes of the concentration of calcium and sodium are probably caused by the following ion exchange reaction (Eq. 5.5) from Ca-montmorillonite to Na-montmorillonite.
$$3[Ca_{0.33}Al_4(\text{Si}_{3.67}\text{Al}_{0.33}O_{10})_2(\text{OH})_4] + Na^+$$

$$\leftrightarrow Ca^{2+} + 6[Na_{0.33}Al_2(\text{Si}_{3.67}\text{Al}_{0.33}O_{10})_2(\text{OH})_4]$$

(Eq. 5.5)

a)

b)

Fig. 5.16: Scatter diagrams of sodium against chloride (a) and against calcium (b).

5.2.3 Discussion of group 3 – the southern province

As the median concentration for all major ions of group 3 is elevated (Fig. 5.13), it is suggested that besides anthropogenic contamination (high concentrations of anions) a special hydrochemical environment exist. The increase of magnesium and calcium is considerable and suggests long residence times of the groundwater. The highest concentrations were found in dug wells, while borewells close to the same were either not affected or mostly to a lesser degree. It is therefore concluded that the hydrochemistry of group 3 is originally under influence from the regolith aquifer.

Most of the samples from group 3 belong to the southern part of the HVO (Fig. 5.10). This regional concentration and the often observed thick regolith layer (>20 m thickness) point at geogenic origins for the elevated EC.

The general thickness of the regolith was interpolated from BDI borehole logs all over the HVO. The resulting map in Fig. 5.17 demonstrates roughly the increased thickness of the
regolith in the southwestern part of the HVO. The map indicates as well thicker regolith layer to the East of the HVO. The groundwater samples from the townships of Nikki and Péréré in the East and outside of the HVO are part of group 3. The direct correlation between regolith thickness and ionic charge, however, is not conclusive as the sampled dug wells of group 3 have different depths and do not always penetrate the whole of the regolith. The eventual siting of boreholes in fracture fillings cannot be determined from the available data.

Group 3 is observed on different rock types and geological series (Fig. 3.19) and though any geogenic influence of bedrock geochemistry is for a start widely excluded.

Stability diagrams computed from ion activities with the software PhreeqC (PARKHURST and APPELO 1999) showed that group 3 plots mostly in the field for Na-montmorillonite. But once more the strong overlapping between all groups is seen (Fig. 5.18) and, partly, samples from the other groups are placed in the same field too. The Na-montmorillonite stability field is an indicator for stagnant or slowly moving water. The kaolinite field is typical for fast flow and groundwater table variations causing a rather fast weathering and transport of dissolved components (APPELO and POSTMA 1999).

TAYLOR and EGGLETON (2001) subdivide the saprolite zone into a succession of layers of different weathering characteristics. At the top, the zone of groundwater table fluctuations is described as kaolinitic. Below this kaolinite zone the saprolite remains saturated without any change. Weathering is slower but very intense and forms as dominant clay mineral montmorillonite.

Some of the group 3 samples belong to dug wells whose depth is definitely below the range of groundwater fluctuations (depth >15 m) they belong certainly to the Na-montmorillonite zone.
The kaolinite-montmorillonite reaction quotient for the hydrochemical data exhibit a sharp increase from about –10 (precipitation) to more than +12 (dug wells) (Fig. 5.19). This means that most of the dug well samples achieved an equilibrium with a single weathering product (kaolinite) but tend already to a horizontal asymptote where a equilibrium exists between kaolinite and montmorillonite (GARRELS 1967). The picture is the same for Ca- and Na-montmorillonite. The first conclusion drawn is that recharge only contributes to a lesser extent to the regolith hydrochemistry. Most of the infiltrating precipitation does, finally, not meet the groundwater body. A second conclusion is that group 3 is not influenced by ion exchange as mentioned for group 2.

If we take it as a typical pattern that the majority of group 3 is placed in the Na-montmorillonite field still some questions remain open. Stagnant water should be relatively older. Tritium analysis (Chapter 6.3) has not shown an important increase in water age with rising depth within the regolith. Therefore it is difficult to say that groundwater of group 3 would be older than those of the other groups. The high share of nitrate and chloride due to contamination demands supply of contaminated water in order to maintain the nitrate input. Nitrate is not stable under the current groundwater conditions (see Chapter 5.4.3).

The high contents of sulphate in group 3 derive either from human input or by the pyrite oxidation. Pyrite (FeS₂) is found in traces in many types of crystalline bedrock facies and dykes and is possibly oxidised to sulphate. Sulphate concentrations vary widely in group 3. Only 2 (H-PTA-1 and H-SER-P) wells show constantly elevated concentrations twice as high as for the rest of this group. The correlation between chloride and sulphate in Fig. 5.20 shows a clear linear trend for all groups. Thus, the sulphate concentration may have been caused by contamination.

On the other hand it is principally among the samples from group 3 that elevate fluoride occurs (see Chapter 5.4.2). Weathering of basement rocks under stagnant groundwater conditions and ion exchange releases fluoride what is affirmed by the explicative approach made in this chapter.
Further considerations about the hydrochemistry of group 3 will rest assumptive as the level of contamination obscures the natural signals.

Fig. 5.19: Changes in the kaolinite-montmorillonite reaction quotient for the hydrochemical groundwater groups in the HVO (modified after GARRELS 1967).

Fig. 5.20: Correlation diagram of sulphate against chloride. The dashed line represents a 1:1 relationship for both constituents.
5.3 Seasonal variations

The hydrochemical analyses from all campaigns and independent from any group affiliation were screened for seasonal changes caused by precipitation and/or evaporation. The plot of all seasonal data in Gibb’s diagrams (see Fig. 5.21) revealed that the influence of precipitation and evaporation on groundwater chemistry is generally low for most of the samples.

![Gibb’s diagram](image)

*Fig. 5.21: Plot of all groundwater samples (grey and black crosses) in Gibb’s diagram. Samples discussed in the text are marked with black crosses. Arrows show the relative position of different samples from the same location. Circles embrace samples of the same sampling point or equal habits. The wells with strong precipitation signature are found in Tab. 5.8.*

In general it is seen that those samples taken from the same location plot as well in the vicinity of one to another. This means that most of the wells are not affected by surface processes, although the open dug wells are directly exposed to the atmosphere.

On the other side it is shown that wells from the same village or region do not plot necessarily nearby in the diagram. Note for example that the samples from W-BDOG-1 and W-BDOG-2 for all seasons form separated clusters although the wells are only a few hundred meter distant from each other. Furthermore, the sample from D04-P-GWB-3, which lies in the vicinity of Dogué, plots even more distant from the others. This borehole...
is only half as deep as the dug wells in Dogué and outside of the village. Dilution or contamination may play a more important role in these shallow depths. In the village of Sonoumoun different characteristics for closely situated wells are observed too. D04-W-SON-1 is a well with high mineral charge in the centre of the village; D05-W-SON-3 is a well at the village periphery, which is only moderately charged.

Samples from borewells (fractured aquifer) and from dug wells (regolith aquifer) differ in different ways. In the village of Kikélé the well KIK-6 plots closer to the field of evaporation than the handpump KIK-P. In Ouannou it is the handpump (H-OUN-P) sample which plots in the evaporation field while the dug well W-OUN-1 plots below. In contrast to these examples the village of Wari-Maro shows that the sampled well and the handpump plot equally in the very same cluster (W-WARI-1 and H-WARI-P). The 3 samples from the footpump in Birni (F-BRI-3) instead plot in every season differently.

These observations show that even in a comparatively small area like a village (generally a few km² only) the hydrogeological and hydrochemical conditions may change profoundly horizontally from one well to another and vertically from wells in the regolith aquifer towards pumps in the bedrock aquifer. Therefore local particularities cannot be explained by a generalised regional sampling. The detection of local contamination sources demands intensified local research (see Chapter 5.4).

The local existence of perched aquifers in the weathered crystalline basement areas has been described in literature before (WRIGHT 1992 and ENGALENC 1978) and was proven for the HVO by FASS (2004) and BOHENKÄMPER (2006). The hydrochemistry of wells which exploits this aquifer is mainly influenced by precipitation. The well of Guéssou (D04/R04/D05-W-GUE-1) is situated in a perched aquifer (BOHENKÄMPER 2006). Eventually, other wells are situated in perched aquifers too. The wells with an assumed strong precipitation signature are represented in Tab. 5.8.

Tab. 5.8: Names of samples under strong precipitation influence.

<table>
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<th>Sample*</th>
<th>Location</th>
<th>X UTM</th>
<th>Y UTM</th>
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<tr>
<td>D04/R04/D05-W-BODI</td>
<td>Bodi</td>
<td>342672</td>
<td>1036773</td>
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<tr>
<td>D04/D05-W-TCH-1</td>
<td>Tchatchou</td>
<td>451119</td>
<td>1009028</td>
</tr>
<tr>
<td>D04-W-PAP-2</td>
<td>Partago</td>
<td>379998</td>
<td>1054149</td>
</tr>
<tr>
<td>D04-W-BRI-1</td>
<td>Birni</td>
<td>338520</td>
<td>1104741</td>
</tr>
<tr>
<td>D04/R04/D05-W-GUE-1</td>
<td>Guéssou</td>
<td>461405</td>
<td>1109699</td>
</tr>
<tr>
<td>R04-W-WEWE-2</td>
<td>Wé-Wé</td>
<td>402344</td>
<td>1038032</td>
</tr>
<tr>
<td>R04-W-KBG-1</td>
<td>Kpabégou</td>
<td>343561</td>
<td>1082842</td>
</tr>
<tr>
<td>D04-W-BAK-1</td>
<td>Bakou</td>
<td>397280</td>
<td>1042949</td>
</tr>
<tr>
<td>D04-W-BUV-1</td>
<td>Djougou</td>
<td>354286</td>
<td>1072507</td>
</tr>
<tr>
<td>D04/R04-W-GOS-1</td>
<td>Gosso</td>
<td>396533</td>
<td>1077975</td>
</tr>
<tr>
<td>R04-W-KAKI-S</td>
<td>Kakioka</td>
<td>426212</td>
<td>1018523</td>
</tr>
<tr>
<td>R04-W-OSA-1</td>
<td>Wassa</td>
<td>349599</td>
<td>1063068</td>
</tr>
</tbody>
</table>

* Samples from the same location but from different seasons are put together by just repeating the season’s code.

Two wells and a handpump (W-BDOG-2, W-SON-1 and H-OUN-P) show an influence of evaporation. But under close regard it is not very reasonable to take just evaporation as the reason for the higher charge of these water points. OUN-P is a closed handpump, SON-1 is a well which is clearly contaminated by human residues and W-BDOG-2 shows
such a general increase in mineralisation that any increase of chloride cannot be clearly caused by evapotranspiration.

Spots of evaporation would be depressions with ponds of water, principally during the dry season. In these locations generally no wells or borewells are found. However, the observations made by isotope analysis (Chapter 6) reveal no evaporation signature on groundwater.

In order to observe the seasonal variation in each village Schoeller diagrams were made to visualise differences between each sampling campaign (see Fig. 5.22 to Fig. 5.25).

In general it can be said that the hydrochemical facies for each site (wells and boreholes) does not change much during the year. Especially the wells of Dogué show that fluctuations due to dilution or concentration by evaporation are very low. The same is seen for a couple of other wells (Kikélé, Pénéssoulou and Kakikoka). At some places it is observed that the well water gets diluted during the rainy season (e. g. R04-W-WEWE-2 in the village of Wé-Wé; R04-W-PEB-1 in Pélébina; R04-W-OSA-1 in Wass). The opposite behaviour may occur as well (e. g. D05-W-DDU-1 in Dendougou, R04-W-OUB-1 in Oubérou).

A reason for this may be the extraction of water by the local people. Already during the early morning hours (around 6 am) people starts looking for water. At Dogué and at Dendougou it was observed that the women are able to extract enough water to lower the groundwater table for a couple of meters in a few hours only. Therefore the sampling for this thesis is regularly done on groundwater freshly seeped in from the aquifer.

It is a common habit in Benin to disinfect their wells regularly with concentrated disinfectants or chlorine. Other contamination sources are possible as well, e.g. dead animals fallen into the wells or the simple water clogging devices used by the women which often stored on the ground without cleaning before the next use.

For many pumps no change in hydrochemistry is observed during the year (e. g. Bétérou, Bio-Sikka, Kikélé, Parakou, Sérou in Fig. 5.22 to Fig. 5.25). Water in fractures is not directly recharged from the surface but by a passage via the regolith first.

During the cooperation with other IMPETUS Workgroups possible pathways for bacteria and virus contaminations of well water were discussed. This here presented study concentrates on a regional scale. Answers to the posed questions demand more detailed local observations. From the hydrochemical observation no typical hint can be deduced to say whether or not rainfall may flush surface deposits and faeces directly into a well or into a badly cased borehole. One point of interest was the pump of Vanhoui. The analyses showed that sample R04-H-VAN-1 has a higher anion charge, principally in chloride but nitrate as indicator for faecal contamination was not detected. The bacteriologically critical wells of Kakikoka show no seasonal fluctuations.
Fig. 5.22: Hydrochemical facies of different points in the HVO represented in Schoeller diagrams. The diagrams combine the seasonal samples from all points within the vicinity of a village (with exception of the village of Dogué, because of visibility). The villages are in alphabetical order.
Fig. 5.23: (continued) Hydrochemical facies of different points in the HVO represented in Schoeller diagrams. The diagrams combine the seasonal samples from all points within the vicinity of a village (with exception of the village of Dogué, because of visibility). The villages are in alphabetical order.
Fig. 5.24: (continued) Hydrochemical facies of different points in the HVO represented in Schoeller diagrams. The diagrams combine the seasonal samples from all points within the vicinity of a village (with exception of the village of Dogué, because of visibility). The villages are in alphabetical order.
Fig. 5.25: (continued) Hydrochemical facies of different points in the HVO represented in Schoeller diagrams. The diagrams combine the seasonal samples from all points within the vicinity of a village (with exception of the village of Dogué, because of visibility). The villages are in alphabetical order.
5 Hydrochemistry

5.4 Groundwater quality

5.4.1 Physico-chemical quality

The general physico-chemical quality of groundwater in the HVO is good, with the exception of elevated nitrate concentrations (see Chapter 5.4.3) in water points close to settlements.

Contents of major ions are all in acceptable limits (WHO 2004). Elevated values for potassium, magnesium, iron and manganese are not rigorously limited but can be objected by consumers because of aesthetic of physiological considerations. Water temperature is often elevated but generally water is stored in cooler places before consumption. In total 94% of the samples in the HVO satisfied the WHO drinking water guideline values.

When asked, the rural people underline that they rather prefer water from the regolith wells than from the borewells because of its taste. It was observed that most of the people do not like to drink the water with to high charge, e.g. the well B-DOG-2 in Dogué. In some cases elevated fluoride values are related to samples with a higher electric conductivity. Its appearance is discussed in Chapter 5.4.2 Fluoride.

The disposal of water points to contamination due to their proximity to locations of diverse daily use, e.g. market places, latrines, showers, in combination to a shallow water table is of major concern for groundwater quality. Nitrate as a good indicator for this source of contamination. The results from the analyses are explained in Chapter 5.4.3.

A greater thickness of the unsaturated zone provides more time for the downward percolation of contaminated water and, consequently, enhanced protection of groundwater (HOWARD et al. 2005). But during the rainy season, the water table rises often to very small distance towards the surface (3-5 m). The groundwater table is thus on he same level as the latrine holes and their deposits.

5.4.2 Fluoride

5.4.2.1 Geological sources

In crystalline areas groundwater is mostly sensitive to high fluoride concentrations where granites and alkaline rocks occur (TRAIVI 1993, FLEISCHER and ROBINSON 1963). It is present in primary minerals (fluorite, biotite, hornblende and apatite). Fluoride can be leached out during weathering and circulation of water in rocks and soils.

Metamorphic rocks have a fluoride concentration from 100 ppm (regional metamorphism) up to > 5000 ppm (contact metamorphism). In these rocks the original minerals are enriched with fluoride by metasomatic processes (FRENCKEN 1992).

Fluoride contents in most freshwaters are low (< 1 mg/l). Fluoride can replace hydroxide ions on mineral surfaces under low and neutral pH. In presence of $\text{Ca}^{2+}$ the fluoride content is controlled by the solubility product of fluorite ($\text{CaF}_2$).

High fluoride concentrations can be built up in groundwater which has a long residence time (FRENCKEN 1992). Increased fluoride concentrations are possible when the $\text{Ca}^{2+}$-
activity decreases, for example due to ion exchange in Na-SO₄-HCO₃-water (MATTHESS 1990).

Dykes and pegmatites bear high concentrations of fluoride. It is likely that these rocks could be providing higher fluoride to groundwater during weathering (SINGHAL and GUPTA 2005).

The WHO (2004) determined the upper limit of Fluoride in drinking-water as 1.5 mg/l for daily consumption. Repeated ingestion of high fluoride concentrations over a long period may cause fluorosis of bones and teeth. A minimum dose of fluoride is essential for dental health (Tab. 5.9).

\textit{Tab. 5.9: Impact of fluoride in drinking water on health (DISSANAYAKE 1991).}

<table>
<thead>
<tr>
<th>Concentration of fluoride [mg/l]</th>
<th>Impact on health</th>
</tr>
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<tbody>
<tr>
<td>0</td>
<td>limited growth and fertility</td>
</tr>
<tr>
<td>0.0 - 0.5</td>
<td>dental caries</td>
</tr>
<tr>
<td>0.5 - 1.5</td>
<td>promotes dental health</td>
</tr>
<tr>
<td>1.5 - 4.0</td>
<td>dental fluorosis (mottling of teeth)</td>
</tr>
<tr>
<td>4.0 - 10.0</td>
<td>dental and skeletal fluorosis (pain in back and neck bones)</td>
</tr>
<tr>
<td>&gt; 10.0</td>
<td>crippling fluorosis</td>
</tr>
</tbody>
</table>

For livestock breeding fluorosis symptoms must be compared to the economical usefulness of an animal. The acceptable level of fluoride in the feedstuff without economic loss of the animal is around 30-40 ppm (LEEMANN 1970).

5.4.2.2 Fluoride in the study area

Around Benin fluorosis cases are reported from Niger (TRAVI 1993) and Ghana (BGS 2006). The DGEau is therefore worried about its occurrence in Benin. Until 2005 fluoride values from 5 to 80 mg/l were measured in 56 boreholes in Benin (Tab. 5.10, oral. comm. L. DOVONAN, DGEAU 2005). Until recently, none of these boreholes is restricted to public use. Information about their exact position and geological environment was not available at time. The highest values had been observed in the village of Bouloum next to Sérou in the Department Donga.

\textit{Tab. 5.10: Borewells contaminated by fluoride in Benin (oral comm. L. DOVONAN, DGEAU, March 2005).}

<table>
<thead>
<tr>
<th>Department</th>
<th>contaminated pumps</th>
</tr>
</thead>
<tbody>
<tr>
<td>Collines</td>
<td>42</td>
</tr>
<tr>
<td>Borgou</td>
<td>13</td>
</tr>
<tr>
<td>Donga</td>
<td>1</td>
</tr>
</tbody>
</table>

The groundwater samples in the study area show only low or not detectable values in most cases. Tab. 5.11 shows the most affected wells (all are open dug wells). The results of the two field campaigns in 2004 showed constant increased values for the well W-BDOG-2 in Dogué. For all later field campaigns no fluoride content above the detection limits was measured. This effect might be caused by a changed sensitivity of IC analysis in the laboratory or as well by dilution, but rainfall occurred as well during the
early field campaigns. The general level of fluoride in the sampled wells is low and there is no object for concern despite the well W-BDOG-2 in Dogué who shows a generally high mineralization.

The markedly local occurrence of high fluoride values is due to the eventual appearance of fluoride bearing rocks such as pegmatite or crystallised drop outs in fractures.

Because of the spatially distribution of such rocks and fracture fillings it is not to be excluded that unsampled wells and pumps in the study area show higher fluoride values.

**Tab. 5.11: Fluoride concentrations observed at three well during the seasons from 2004 to 2006.**

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>W-BDOG-2</td>
<td>1.26</td>
<td>1.8</td>
<td>&lt; 1</td>
<td>&lt; 1</td>
<td>&lt; 1</td>
</tr>
<tr>
<td>W-KAKI</td>
<td>1.2</td>
<td>&lt; 1</td>
<td>&lt; 1</td>
<td>&lt; 1</td>
<td>&lt; 1</td>
</tr>
<tr>
<td>W-KIK-6</td>
<td>0.83</td>
<td>1.3</td>
<td>&lt; 1</td>
<td>&lt; 1</td>
<td>&lt; 1</td>
</tr>
</tbody>
</table>

### 5.4.3 Nitrate and Nitrite

#### 5.4.3.1 Nitrogen compounds in the environment

Almost all nitrogen (N) found in the soil and subsoil originates from the atmosphere which is made of 78.1% nitrogen. In water nitrogen may oxidise and occur as nitrate (NO$_3^-$) or nitrite (NO$_2^-$), in reduced form nitrogen may appear as ammonium (NH$_4^+$). They circulate in the natural environment in the so-called nitrogen cycle (MATTHESS 1994). Natural background nitrate concentrations are evaluated in the report of ECETOC (1988) to be 0 to 10 mg/l in groundwater and up to 5 mg/l in surface water. Excessive concentrations of nitrate in drinking water may cause methamoglobinemia to small children (blue-baby syndrome). It hinders the oxygen transport in the blood and leads to suffocation (HEM 1985). The internationally mostly respected limits for nitrate in potable water are given in **Tab. 5.12**.

**Tab. 5.12: Limits for nitrate in drinking water in international use. It should be noted that some countries may chose other limits following their own policies.**

<table>
<thead>
<tr>
<th>Limit [mg/l]</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>50</td>
<td>WHO (2004)</td>
</tr>
<tr>
<td>44.3</td>
<td>US EPA (2002)</td>
</tr>
</tbody>
</table>

In this thesis nitrate analyses are reported as mg/l NO$_3^-$. Some laboratories may refer to the nitrogen content of the analysed nitrate (NO$_3^-$N). 1 mg/l of NO$_3^-$N equals 4.43 mg/l of NO$_3^-$.  

APPELO and POSTMA (1999) consider the extensive application of fertilizers and manure in agriculture as the main cause of high nitrate concentrations in shallow groundwater. Artificial fertilisers are generally used on cotton fields which are generally distant from villages. Cotton planters use around 200 kg/ha of nitrogen bearing fertilisers (oral comm. T. MARTIN, IRD Cotonou 2007).
The study area is mainly of rural character. Industrial influx of nitrate can therefore be excluded. Besides agricultural and industrial nitrate sources it is human excreta which contains considerable amounts of nitrate and N-compounds.

Within the study area, the population either uses simple pit latrines or has no form of sanitation at all. Nitrate contaminations under equal conditions are found all over in Africa, as in Botswana (JACKS et al. 1999), Tanzania (NKOTAGU 1996) and The Republic of Guinea (GÉLINAS et al. 1996). Up to 95% of the nitrate derived from the human excreta originally belongs to urine (JACKS et al. 1999).

Nitrate is fast transported in flowing groundwater. Disposal of human excreta for example in latrines are therefore commonly known as origin of nitrate contamination of the groundwater. Further on the traditional village wells are often in a poor condition. Cracks and lack of sealing may cause the infiltration of contaminated water.

### 5.4.3.2 Nitrate in the study area

Analysis of groundwater samples from all field campaigns revealed high amounts of nitrates. The highest nitrate content, found in the study area, is 648.77 mg/l at a well in the village Sonoumoun (D04-W-SON-1) during the dry season 2004. All analysis results are represented in the Annex 1.

Of the 164 groundwater samples taken within the study area, 93 samples (56.7%) were found to have nitrate concentrations above 10 mg/l, which indicates anthropogenic influence (ECETOC 1988). 42 among these samples have nitrate concentrations exceeding 50 mg/l (45.2% of the 93 samples and 25.6% of the total population).

The plot of all samples in a pe-pH diagram (see Fig. 5.26) reveals that the ruling conditions favour a complete decay of all nitrogen compounds to nitrogen. Therefore the encountered nitrate concentrations are either conserved in some part of the aquifer with other redox conditions or there is a continuous enrichment.

**Fig. 5.26:** Stability phases of nitrogen compounds in a pe/pH diagram (N-O-H system). The samples are plotted for each season respectively.

BARRETT et al. (2000) suggest that a good correlation between NO₃⁻ and Cl⁻ is given when they are of the same source. Furthermore, a low Cl⁻/NO₃⁻-ratio is indicative of a
faecal origin when \( \text{NO}_3^- > 10 \text{ mg/l} \) (MORRIS et al. 1994). This is shown for the study area in Fig. 5.27. Human activity can therefore be interpreted as origin of contamination.

Isotope investigations in the groundwater of the Collines department by CRANE (2006) proofed the very same. Human waste and manure are the principal origin of nitrate (Fig. 5.28).

A borehole drilled in 2002 in the village of Dogué (UTM 383200/1006349) was abandoned because of bad water quality (EC > 4000 µS/cm²). In 2002 nitrate concentration was determined as 16.83 mg/l (Source: BDI). In 2007 at the same location a much higher nitrate concentration was found (>400 mg/l). This value was proven by the hydrochemical laboratory of the DGEau at Cotonou as well as by the IMPETUS laboratory at Parakou. The occurrence of exposed excrements at the surface around the borewell was reported by A. UESBECK (IMPETUS subproject A5). The high nitrate concentration in this borehole is therefore caused anthropogenic contamination.
Such a high nitrate value is only reported from the open dug well in the centre of the village Sonoumoun (D04-W-SON-1, UTM 420434/1079934) with a nitrate concentration of 648.77 mg/l. The origin of this very high value is clear. The regolith layer in Sonoumoun is very thin (< 1 m). Mica schist is outcropping directly in the dug well which is almost enclosed by near standing houses and latrines. This well has only a depth of around 14 m. Contaminated water may infiltrate almost directly from latrine pits into the well. The borewell at Dogué, instead, is deeper than 40 m and is installed in a regolith with a thickness of around 25 m. Preferential flow paths (macropores, fissures) or a bad boreholes casing would be the cause of the observed contamination.

Alternatively, other possible sources of high nitrate levels in groundwater exist. SCHWIEDE et al. (2005) describe high nitrate contents in soils and sandstone aquifers in Botswana due to cattle rising (up to 600 mg/l). Cattle breeders are the half nomadic Fulbe people who come by casually and cannot be seen as relevant source for nitrate contamination. BOLGER et al. (1999) found in arid Australia soil nitrate concentrations around termite mounds of up to 2,000 mg/l. In Benin termite mounds can be found all over the country therefore it is not convincing that only the village of Dogué would be affected by a termite induced nitrate contamination. Elevated nitrate contents occur as well in spinifex and other grasses, leaf litter in mulga, in surface crusts and in bare sandy soil covered with bushfire ash. When rain falls, this nitrate is leached from the ash and percolates toward the groundwater table. In Benin it is a common practice to burn wild grasses to clear area and fertilise agricultural soils.

In Dogué two dug wells (W-BDOG-1 at the central market place and W-BDOG-2 at the village edge) were investigated from 2001 until 2006. Both show elevated nitrate values but a different behaviour during the last years (see Fig. 5.29).

![Graph](image)

*Fig. 5.29: Evolution of nitrate concentrations in groundwater from W-BDOG-1 and W-BDOG-2 in Dogué from 2001 – 2006. Data from 2001 to 2002 was collected by FASS (2004). No field campaign in 2003.*

The well at the village edge is around 500 m distant from the market place. Its nitrate level is rather constant during the time with a level of around 20 to 30 mg/l. The other
5 Hydrochemistry

well instead shows an increasing level of nitrate with already 90.2 mg/l in the dry season 2006.

The differing nitrate levels of both wells can be clearly explained by the difference in use and location. The market place is regularly frequented by many people and shows many buildings while the other well is only achievable by a small tray and is encircled by some trees. Nevertheless both wells are regularly frequented but the well W-BDOG-2 is not used for drinking water. This seems paradox as it shows at least less nitrate. The villagers refer to it as having a bad taste which they refuse. The reason might be the generally higher level of solved minerals in this well (see Tab. 5.13).

Tab. 5.13: Average mineralisation of the two regularly sampled dug wells in Dogué.

<table>
<thead>
<tr>
<th>Location</th>
<th>Depth [m]</th>
<th>Average mineralisation [mg/l]</th>
</tr>
</thead>
<tbody>
<tr>
<td>W-BDOG-1</td>
<td>22</td>
<td>694.8</td>
</tr>
<tr>
<td>W-BDOG-2</td>
<td>24</td>
<td>1343.9</td>
</tr>
</tbody>
</table>

The distribution of nitrate within the regolith might be either advantaged by microfissures in the saprolite clay or by lateral flow in the interface zone between sandy soils and argillaceous saprolite.

5.4.4 Heavy metals

The elements with a density >4.5 g/cm³ are called heavy metals. Although groundwater contamination of heavy metals is often related to the infiltration of industrial water and slugs. A geologic origin is as well possible. Crystalline rocks contain heavy metals in variable concentrations. Mn²⁺ and Fe²⁺ are separately described together with the major cations.

The HVO is a rural area without any industries except at Parakou. But all over the area especially in the vicinity of settlements disposals like car batteries, energy cells and many others are found. Sometimes the disposals can even be found in wells. It is as well often seen that people wash their cars, trucks and motorcycles directly next to water points or even parking inside lakes, rivers or bas-fonds. In order to determine any impact of these disposals on the shallow groundwater in the HVO, samples from the rainy season 2004 and from the dry season 2005 were additionally analysed for a limited variety of heavy metals, among them arsenic (As), cadmium (Cd), copper (Cu), lead (Pb) and nickel (Ni), was analysed. For the results and the drinking water limits (WHO 2004) refer to Annex 1.

All of them are very poisonous (APPELO and POSTMA 1999). Cu is less toxic but it is good indicator for contamination by simple industrial forms as steel and metal work. Cd and Pb are typically found in energy cells and are therefore investigated (WHO 2004). Contamination by geogenic arsenic is reported from many places in the world, for example in Bangladesh, India or Mexico (AGGARWAL et al. 2000). Arsenic can be released from the rock matrix to groundwater under anaerobic conditions but also in acidic groundwater under oxidising conditions. The later case tends to occur in arid and semiarid settings resulting from extensive mineral reaction and evaporation (WSP 2005). Cases of arsenic in rural groundwater are reported from Ghana by SMEDLEY et al. (1996). In the rainy season 2004 53 samples were analysed, in the dry season 2005 66
samples. The samples include groundwater from the regolith and from the bedrock aquifer. Surface water from lakes and rivers were tested as well. The reservoirs of Djougou (DJO-BR) and Parakou (OKP-1) are almost exclusively in use for the public water supply of these cities. No traces of any contamination by the here presented heavy metals were found (see Annex 1).

5.4.5 Sodium adsorption ratio (SAR)

Artificial irrigation is not popular in Benin. Longer drought periods and new agricultural policies may change this in future. In the HVO groundwater is available throughout the whole year and may serve as an important source for decentralised irrigation.

In order to prove the suitability of groundwater for this purpose the sodium adsorption ration (SAR) was calculated. The SAR (Eq. 5.6) was proposed by the US DEPARTMENT OF AGRICULTURE (USDA 1954) as classification scheme for groundwater.

\[
SAR = \frac{Na^+}{\sqrt{\frac{1}{2} (Ca^{2+} + Mg^{2+})}} \tag{Eq. 5.6}
\]

with

\begin{align*}
SAR &= \text{Sodium adsorption ratio} \\
Na^+, Ca^{2+}, Mg^{2+} &\text{ in meq/l}
\end{align*}

The SAR of the groundwater samples is plotted against their electric conductivity [µS/cm] respectively into a classification diagram (Fig. 5.30). The salinity hazard is described by 3 intervals of electrical conductivity (C1, C2 and C3). The higher the grade the more likely it is that the irrigated soil gets saline. The sodium hazard (S1, S2 and S3) indicates how far harmful levels of exchangeable Na⁺ are produced in the soil water. The majority of the samples represent a low (EC < 250 µS/cm) or medium salinity hazard (250 – 750 µS/cm) when used for irrigation. There is couple of samples with a relatively high salinity hazard (EC > 750 µS/cm). In general those are always the same samples for each season. They belong mainly to group 3.

High salinity water cannot be used on soils with restricted drainage (USDA 1954). The locations of concern (Tab. 5.14) are mostly situated in the southern part of the HVO. But other areas as well outside the HVO are indicated (for example F-NIK-2 and F-PRE-1). The SAR is an appropriate method to check long term influence of groundwater hydrochemistry on the soil properties. Groundwater within the study area can generally be used for irrigation with a low risk of sodium hazard.
Fig. 5.30: Groundwater classification for the HVO for all sampled seasons. Grey triangles = dug wells; black crosses = borewells.

Tab. 5.14: Sample locations with high salinity hazard.

<table>
<thead>
<tr>
<th>Village</th>
<th>Sample</th>
<th>N° of samples from March 2004 to March 2006</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alfakpara</td>
<td>W-ALF-1</td>
<td>1</td>
</tr>
<tr>
<td>Dogué</td>
<td>W-BDOG-1</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>W-BDOG-2</td>
<td>5</td>
</tr>
<tr>
<td>Kari</td>
<td>H-KAR-1</td>
<td>1</td>
</tr>
<tr>
<td>Kikélé</td>
<td>W-KIK-6</td>
<td>3</td>
</tr>
<tr>
<td>Nikki</td>
<td>F-NIK-2</td>
<td>1</td>
</tr>
<tr>
<td>Ouannou</td>
<td>H-OUN-P</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>W-OUN-1</td>
<td>1</td>
</tr>
<tr>
<td>Péréré</td>
<td>F-PRE-1</td>
<td>1</td>
</tr>
<tr>
<td>Sérou</td>
<td>H-SER-P</td>
<td>4</td>
</tr>
<tr>
<td>Sonoumoun</td>
<td>W-SON-1</td>
<td>1</td>
</tr>
<tr>
<td>Tamarou</td>
<td>W-TAR-2</td>
<td>1</td>
</tr>
<tr>
<td>Wari-Maro</td>
<td>W-WARI-1</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>H-WARI-P</td>
<td>3</td>
</tr>
</tbody>
</table>
6. Environmental isotopes

6.1 Stable isotopes in precipitation and surface water

Since the rainy season 2004 precipitation was regularly sampled at three sites in the HVO. They are at Sérou (UTM 357472/1068612), Dogué (UTM 383034/1006505) and Parakou (UTM 469862.5/1035435.8). More than 210 precipitation samples were taken for analysis until the dry season 2006. By the time of writing this thesis the results of these samples were still not available. Thus, only 8 rainfall samples taken during the field campaigns 2004 were used (see Annex 2). 13 samples from earlier campaigns (FASS 2004) were added.

![Fig. 6.1: a) Comparison of GNIP data (Kano-Nigeria = light grey x / Niamey-Niger = grey +) with the HVO rainfall data (black rhombus). The global meteoric water line (GMWL as grey dashed line) is calculated by Craig's notation (Eq. 6.1). b) Groundwater analyses (grey +) are shown in relation to the GMWL. Analyses from surface waters (grey x) are grouped around an evaporation line (light gray dashed line).](image)

From the GNIP database (IAEA/WMO 2004) the measurements from Niamey and Kano were additionally chosen. All data sets were compared with the GMWL (Fig. 6.1 a). The isotopic range of HVO precipitation in 2002 to 2004 lies between –6.24‰ and 2.845‰ for δ18O, and between –35.75‰ and -24.95‰ for δ2H. The stable isotope composition of the HVO samples plots around the GMWL with a median d-excess of approximately 9. The relative plot of the groundwater samples shows that they are much less scattered as the precipitation samples (see Fig. 6.1 b).

The surface water samples are plotted into the same diagram. They are influenced by evaporation and plot around an evaporation line with the equation:

\[ \delta^2H = 4.44 \cdot \delta^{18}O - 1.22 \]  

(Eq. 6.1)

The evaporation line crosses the GMWL in the scatter of the groundwater samples in the encircled area. This means that surface water and groundwater types have the same
origin. The groundwater reservoir is a mainly semi-confined aquifer. It is the mixing pot for all isotopic rainwater signatures. Another origin of groundwater in the HVO is not distinguished. Groundwater samples are not affected by evaporation in contrast to the sampled surface waters. This implies that the principal source of groundwater recharge in the HVO is the direct infiltration of rainfall. The remaining surface water runs off via rivers or gets evaporated in residual lakes or bas-fonds.

### 6.2 Stable isotopes in Groundwater

Stable isotopes were regionally sampled from wells and pumps (see Annex 2). It was observed that no clear distinction can be made between the regolith and the bedrock samples (Fig. 6.2). The relative position of the samples marked by letters (a to h) shows that the isotope composition even in the deeper bedrock aquifer is not stable. As shown in Fig. 6.2 groundwater plots around the GMWL. Bedrock and regolith groundwater has infiltration of rain water as common source. The isotope composition may change during the time as a result of mixing either with fresh water or with downward percolating water due to extraction.

![Fig. 6.2](image)

**Fig. 6.2:** (left) Dry season 2004: Isotope relationships for groundwater samples from wells (circles=regolith) and pumps (triangles=bedrock). (right) Rainy season 2004: Isotope relationships for groundwater samples from wells (circles=regolith) and pumps (triangles=bedrock). The letters (a) to (h) signify a choice of samples from the same locations.

While all samples plot during the dry seasons around the GMWL (Fig. 6.3 left) they show a different behaviour in the rainy seasons (Fig. 6.3 right). Mainly the samples from the rainy season 2002 (from FASS 2004) show a higher D-Excess. The D-Excess for these samples has been re-calculated by a strict linear fitting to be rather +12 (the value of +12.5 was originally determined by FASS 2004).

At the village of Dogué water from precipitation was regularly sampled and analysed for its stable isotope content (samples for the period 2001 – 2002 from FASS 2004). The time row from 2001 to 2004 shows a rise in D-excess in when the rainy season in 2002, and respectively in 2004, starts (Fig. 6.4).
A D-excess $> +10$ can be explained when air masses are enriched by condensed water which rises up from evapotranspiration surfaces. These may occur when the air passes areas of lakes or dense forests (among others: oral comm. by Y. TRAVI in 2005, KENDALL and MCDONNELL 1998, CLARK and FRITZ 1997).

The continuation of stable isotope sampling from groundwater and precipitation helped to prove that the observed difference in the D-Excess is not a result of differing origins of the groundwater but rather of the different isotope composition of the precipitation during the year as shown by Fig. 6.4. In correspondence to the pH observations (Fig. 5.3) and the interpretation of the seasonally sampled hydrochemical analyses (in Chapter 5.2 and 5.3) the local character of recharge by precipitation can be assumed.

Groundwater flow over distances of several kilometres from distant areas, e.g. outside the HVO, through a rather thin regolith aquifer (average thickness = 5-15 m) remains unimaginable. Chapter 4 showed that subsurface flow in the regolith of the slightly
undulated HVO area heads towards the local depressions. The conclusion can only be that the groundwater available in local subcatchments is first of all a product of local recharge.

No clear end members can be defined in this hydraulic system. As consequence stable isotopes serve not to determine mixing ratios in order to separate groundwater base flow to rivers from surface runoff. Other tracers might be more promising for this task.

### 6.3 Tritium data from the HVO

In total 34 samples for tritium analyses were chosen on hydrogeological base and in order to receive a regional distribution across the whole HVO (see Annex 2). The samples were taken from groundwater in different depths. Samples from open dug wells normally shall describe the TU contents of groundwater in the regolith while samples from pumps reflect conditions in the fractured bedrock.

Additionally 33 results from tritium sampling during the rainy season 2002 from FASS (2004) are available (including samples from 2 rainfalls, 5 surface water sources and 26 shallow groundwater sources).

Tritium concentrations in precipitation and surface waters in the study area have been compared to measurements from the GNIP stations at Bamako (Mali), Kano (Nigeria) and N’Djamena (Chad) and for reference at São Tomé as published by the IAEA/WMO (2004). The first three stations are more of less all at the same latitudes. Kano is the closest station to the HVO in Benin. São Tomé has been chosen to show the increase of tritium concentrations moving from the seaside to the inner continent (see Fig. 6.5). The Fig. 6.7 shows that the course of the curves is quite similar for Kano, Bamako and N’Djamena and tritium units fluctuate around the same amounts.

![Fig. 6.5: Seasonal fluctuations of TU in precipitation measured at different stations of the GNIP database (IAEA/WMO 2001) with an enlarged view for the years 1970 to 2000.](image-url)
It is seen that for N’Djamena and Bamako the development of the curves is quite comparable. Both start in the early 60ies with values exceeding 1000 TU when monthly variations are still the strongest. In the midst of the 1970s values decreased to around 50 TU. Bamako shows consequently in 1998 as last and lowest measurement 4.7 TU. Analyses of precipitation and surface waters in the HVO show a range of 2.8 to 3.8 ± 0.7 TU. This is slightly less than at Bamako but reasonable as the whole study area is situated more meridional. The decrease of tritium towards the South can be seen as well by the course of the curve for São Tomé which shows the same trend but lower values.

Tritium analyses for precipitation and surface water from the rainy season 2002 (FASS 2004) show an equal range of 2.7 to 3.9 ± 0.7 TU. By comparison of these few samples from the dry and the rainy season no seasonal variations in tritium contents are obvious. For further considerations the median of the available results 3.5 ± 0.7 TU is assumed as the average content of fresh recharge water. It cannot be excluded that precipitation with > 4 TU might occur as for example the oral communication with TRAVI (2005) revealed an actually expected tritium concentrations in precipitations in the Sahel (including the North of Benin) vary between 6 – 10 TU.

### 6.4 Tritium age determination of groundwater in the HVO

Most of the samples were taken from dug wells and boreholes. As dug wells normally penetrate the regolith zone in depths from 5 to 20 m, boreholes catch water most of times in the fractured bedrock in depths > 30 m. Fig. 6.6a) show that tritium contents of the samples vary with the depth. Most of the analyses of the samples from shallow groundwater tables show TU in the range of the precipitation reported above. Most of the samples from groundwater belonging to the bedrock aquifer show tritium concentrations < 2 TU instead. Still a couple of samples from dug wells as from boreholes show concentrations of > 5 TU although actual rainwater has only 2.2 – 4.6 TU. Rainwater with originally higher tritium contents is still stored in some layers. These layers are most probable within the montmorillonitic weathering zone where less water is exchanged or in the bedrock where water from above infiltrates. Pumping lowers the water level within the fractures quite fast as they have usually only a small storage capacity (see Fig. 6.6). The older pre-bomb test water (age > 50 years) is then mixed with younger water producing elevated values of TU.

<table>
<thead>
<tr>
<th>Sample*</th>
<th>Wet season TU</th>
<th>Dry season TU</th>
<th>st. Dev. ±2σ</th>
<th>Depth [m]</th>
</tr>
</thead>
<tbody>
<tr>
<td>PELE - 1</td>
<td>3.8</td>
<td>7.3</td>
<td>0.7</td>
<td>12.68</td>
</tr>
<tr>
<td>WEW - 2</td>
<td>3.0</td>
<td>4.3</td>
<td>0.7</td>
<td>11.28</td>
</tr>
<tr>
<td>OUB - 1</td>
<td>3.1</td>
<td>3.7</td>
<td>0.7</td>
<td>20.2</td>
</tr>
<tr>
<td>GWB-3</td>
<td>3.0 (GWB-1)</td>
<td>2.1</td>
<td>0.7</td>
<td>11.60</td>
</tr>
<tr>
<td>SER - 2</td>
<td>3.0</td>
<td>2.7</td>
<td>0.7</td>
<td>12.56</td>
</tr>
<tr>
<td>DRG</td>
<td>3.0</td>
<td>2.9</td>
<td>0.7</td>
<td>12.62</td>
</tr>
</tbody>
</table>

* All locations are open dug wells with the exception of GWB 1 and 3. These are observation wells within the regolith which were installed by FASS (2004).
When the TU exceeds the value of the precipitation samples it can clearly be said that recharge belongs to the bomb-linked water. The question remains how water from that period still prevails in open dug wells when fresh water regularly infiltrates. As said in Chapter 4 most of the dug wells do not dry up completely during the dry season. They are still connected to a water table within the regolith aquifer. It might be possible that layers or lenses, e.g. perched aquifers, of older bomb-water prevail and are still infiltrating into these wells too. A higher amount of TU can be admixing this older groundwater.

Comparison of samples from the dry seasons 2004 and 2005 with data from the rainy season 2002 shows a sharp increase of the tritium amount at PELE-1, WEW-2 and OUB-1. In the case of SER-2 the value for TU stays the same or is even decreasing but keeps including the standard deviation still in the range of the precipitation tritium (see Tab. 6.1).

Age signals of tritium may be hidden in mixed water. The age of recent water correlates positively with tritium concentrations in precipitation. Sampling of fracture water is only possible at pumped boreholes where it gets mixed (see above). Some samples from the bedrock aquifer show TU < 0.5. This water must have been recharged before 1952 (MAZOR 1997). Groundwater recharged since that time got mixed already during infiltration into the aquifer and continued percolating downward. Therefore it could still exist a layering of different water ages in the aquifer. The highest observed value for TU is 7.3.
Some assumptions were made for calculating mixing ratios and determine the percentage of actual precipitation in groundwater (Eq. 6.2). For the precipitation the mean value of 3.5 TU is just considered as non-changing and as well the mean value of 0.7 for bedrock samples is not modified as well. A further assumption is that there is no decay of tritium during precipitation and infiltration to the groundwater.

\[ x = \frac{T_R - T_F}{T_R - T_P} \times 100 \]  
(Eq. 6.2)

\[ x = \text{share of recharge in the mixing water in } \% \]
\[ T_R = \text{TU of the regolith mixing water (from 4.0 to supposed 10 TU)} \]
\[ T_F = \text{TU of the fracture water (mean: 0.7 TU)} \]
\[ T_P = \text{TU of the precipitation (mean: 3.5 TU)} \]

Under these assumptions the maximum mixing ratio between recent recharged water to sampled older water can be calculated to be somewhere between 30.1 to 85.0% (Tab. 6.2). This shows that recharge water delivers for samples with > 6 TU only around 50% of the mixing water. This value emphasises the presence of “older” water. The calculation above is only a first approach. It must be kept in mind that the sampled groundwater from fractures could as well be mixed. More data is needed to base conclusions on these findings.

**Tab. 6.2: Mixing ratio of recently recharged water to groundwater from the regolith and the bedrock aquifer. Results from Eq. 6.2.**

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>4 0.7 3.5</td>
<td>85</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5 0.7 3.5</td>
<td>65.1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>6 0.7 3.5</td>
<td>52.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>7 0.7 3.5</td>
<td>44.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>8 0.7 3.5</td>
<td>38.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>9 0.7 3.5</td>
<td>33.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>10 0.7 3.5</td>
<td>30.1</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Nevertheless, attempts were made to calculate some eventual water ages by black-box-models using the Boxmodel V2-3 sheets of ZOELLMANN and AESCHBACH-HERTIG (2001). As reference tritium measurements of the GNIP database at Bamako, Mali, were used. Restriction is given by the lack of yearly data. Mean values were calculated for each year and the interpolated linearly over the years without measurements. The program calculates reasonable water ages only until the year 2000 but this is sufficient taking in account that values in the rainy season 2002 are not that different from values measured in the following years. It has to be kept in mind that regional recharge is only around 10 to 15% from the total amount of precipitation. The dispersion model (DM) was chosen as it serves well to simulate partly covered aquifers with samples from different depth levels (DVWK 1995).

The choice of input parameters is presented in Tab. 6.3. The year of observation is technically limited to the year 2000. In order to obtain a possible age for the groundwater
Environmental isotopes

with > 6 TU a value of 10 TU is entered. Residence times between 30 to 50 years are assumed. BAUER (2004) determined dispersion parameters by forced gradient tracer tests in the Aguima catchment. Values for $P_D$ range from 0.003 to 0.11 with an average of 0.05 which is used in the DM. During modelling values had been adjusted and changed several times. As a result the DM results show that older groundwater should be recharged in the 70ies (1968 – 1980).

The mean atmospheric concentration of tritium measured at Kano from 1971 to 1973 was calculated as 69 TU. All mixing processes excluded, radioactive decay in groundwater would lead to 12.25 TU in the year 2000. The more southwards situated HVO should show lesser concentrations in the precipitation in 2000 (longitudinal effect).

Considering the above outlined processes it seems possible that originally recharged groundwater in deep or bad connected fractures could be very well much older than 50 a.

*Tab. 6.3: Entry parameters into the user interface of the Boxmodel V2-3© (by IHW, ETH-Zurich, ZOELLMANN et al. 2001).*

<table>
<thead>
<tr>
<th>Model Code (pm,em,dm)</th>
<th>dm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tau [a] (for transfer and output graphs)</td>
<td>30</td>
</tr>
<tr>
<td>Tau Step [a] (for tau graph)</td>
<td>1</td>
</tr>
<tr>
<td>Delta (dispersion parameter, $P_D$)</td>
<td>0.5</td>
</tr>
<tr>
<td>Tritium Factor (tritium input scaling)</td>
<td>1</td>
</tr>
<tr>
<td>Tracer Code (tr, cfc, kr, he)</td>
<td>tr</td>
</tr>
<tr>
<td>Year of Observation</td>
<td>2000</td>
</tr>
<tr>
<td>C_obs (observed concentration)</td>
<td></td>
</tr>
</tbody>
</table>

The study of CEFIGRE (1990) presented results from a tritium research (BRGM-AQUATER 1986) on borehole groundwater in the Leo-Ranch area in Burkina Faso during 1984 to 1985. It showed that in the crystalline massif of Burkina Faso water ages ranges around 30 years – younger towards the North and older towards the South. DRAY et al. (1988) used isotopes ($^3$H and $^{14}$C) to determine groundwater age in the coastal basin of Benin. They showed that groundwater in the Mesozoic aquifers has an age of around 2000 a in the North and is getting older towards the South (>6000 a). The coastal basin aquifers are recharged by mainly by surface waters in the northern unconfined areas while the South of the basin is mainly confined. In relation to the crystalline basement it can be stated that there is a certain amount of water which is transferred from the basement towards the coastal basin too.

Age determination by the $^{14}$C-method is reasonable when the minimum age of the sample is >1000 a (oral communication by W. STICHLER, GSF 2005). Although a high percentage of bedrock water should be much higher age than 50 a, mixture occurs especially around places with pumping where samples can be taken. Therefore it is not reasonable to take such water to $^{14}$C analysis. Within the scope of this study it was not priority to know the exact groundwater age. But the tritium analysis revealed the relatively slow regional vertical exchange of groundwater and a very homogeneous horizontal distribution with the exception of locally pumped areas.
7. Conceptual hydrogeological model

The observations described in the foregoing chapters are fitting to the regional model for regolith areas (Fig. 7.1) as proposed by DANIEL et al. (1997). Recharge occurs areally at the ground’s surface but principally at the crests (Chapter 3.7.2). The valleys are preferential for groundwater discharge, either by transpiration or by fracture flow. Groundwater flow heads towards local depressions. The groundwater table moves mostly within the regolith and is the lowest at valleys.

![Conceptual model of the regional hydrogeology](image)

Fig. 7.1: Conceptual model of the regional hydrogeology (modified from DANIEL et al. 1997; not scaled; vertically exaggerated). Effective recharge takes place at tophill – discharge downhill. The groundwater table (blue) is set in the regolith. Groundwater in the saturated zone flows towards local morphological depressions. Flow in the bedrock is limited to the fracture zones.

The results from the piezometric observations (Chapter 4) and the investigation in hydrochemistry (Chapter 5) and environmental isotopes (Chapter 6) revealed that groundwater recharge and flow highly depend on local fractures systems.

Regional flow in the bedrock can be excluded due to its lacking fracture connectivity. A regional flow pattern in the regolith is unlikely too, due to the low overall gradient of the terrain (from Northwest to Southeast around 1.5 m/km). Additionally the terrain is undulated and the average median regolith thickness only around 20 m. Groundwater flow would therefore head towards the drains of the local subbasins.
As consequence the groundwater table reflects the surface elevation (Fig. 4.11). Groundwater flow is therefore limited to the local subcatchments. Local conceptual models from the literature represent mostly x-sections of these subcatchments (e.g. MARTIN 2006, FASS 2004, CEFIGRE 1990 and ENGALENC 1978).

The case example from a subcatchment of the Ara River in the North of Djougou shows that groundwater levels in valleys are very diverging (Fig. 7.2). In the boreholes with 2 m depths at all three positions the water table is generally shallow, but in the valley it is the shallowest. The deeper boreholes (10 m and 20 m) show instead that the water levels measured in the valleys are the deepest. Those measured at the crest are still deeper than along the slope. This means for the valleys that perched aquifers on top of clayish valley fillings exist. Below these fillings the groundwater table is lowered either by the extraction from deep rooting plants in the riverine forests (compare with Chapter 3.5) or either by a system of well connected fractures draining off the water towards any distant feature of discharge.

The groundwater table at the crest is relatively lower than at the slope. This behaviour is caused by the equilibration of the piezometric surface in relation to the morphologically higher situated borehole position at the crest.

The comparison of EC measurements at Nalohou revealed an increase of concentration downhill. The highest values were observed in the deeper groundwater levels of the valleys. It is supposed that outwash and dissolved content from the crest and the slope is
accumulated within the valley filling. Assuming that the valley is formed on a well drained fracture zone, groundwater should experience a relatively rapid exchange and thus a decrease in EC. This is not the case; instead it has to be assumed from the observation described for Fig. 7.2 that the lower groundwater level in the river valleys is caused by transpiration through the riverine forests. The roots of trees in these forests but as well of the alone standing trees and in the tree Savannah can easily achieve depths from 10 to 20 m (L. SÉGUIS, IRD, oral comm. 2007; C. ZUNINO, oral comm. 2006; Y. TRAVIS, LHA, oral comm. 2005). Thus, they are able to extract groundwater throughout the whole year. This assumption was also made before by L. SÉGUIS (IRD, oral communication 2007) and by ENGALENC (1978) for comparable sites in Benin and in West Africa.

As said before in Chapter 1.3 only a few regional groundwater models in a sub-Saharan setting are described in literature. In two case studies from southern Africa groundwater models were developed including discharge of groundwater by trees. KLOCK et al. (2001) assumed a discharge of 20-30 mm/a from Kalahari aquifers by Acacia trees.

Bauer (2004) determined the groundwater discharge by riverine forests in the Okavango delta as varying between 21.9 and 1569.5 mm/a. He also observed that the downstream groundwater velocity is very low compared to the lateral flow velocities feeding the water demand by the riverine forests.

The interaction of groundwater with surface water is discussed in Chapter 4. The probably very small contribution of groundwater from deeper aquifer levels in the regolith and the bedrock is additionally demonstrated by the electric conductivity. The general electric conductivity of running river water is < 100 µS/cm. Groundwater is progressively higher charged with increasing depth. Thus any infiltration would only be caused by the shallowest groundwater levels and interflow. FASS (2004) calculated the possible share of bedrock groundwater on rivers as 2%. But his references for bedrock water were the wells and observation boreholes of Dogué which, although relatively deep, are not dug in the basement.

It is concluded that bedrock groundwater has generally no influence on surface water. Wherever surface water bodies (lakes and rivers, bas-fonds) are connected to the fractured aquifer below it would certainly influence its hydrochemistry. This area of influence is controlled by the fracture connectivity. Elsewhere the influence of surface water on groundwater chemistry was excluded by isotopic evidence (Chapter 6).
8. Groundwater flow model

8.1 Objectives of the model

In the previous chapters it was already described that the hydrogeological conditions of the crystalline basement aquifers in Central Benin as well as the available data base limit the application of a regional groundwater modelling approach. The encountered groundwater hydraulics does not fit the conditions found in other catchment areas of comparable size with thick sedimentary aquifers and a considerable groundwater flow velocity. Instead, the HVO catchment consists of many subcatchments that are not necessarily well connected among another. Regional flow is limited through the fracture connectivity and the very low general permeability of the regolith. The total aquifer thickness (regolith + bedrock) achieves a maximum of 80 m. The relationship between this value and the horizontal extension HVO area is roughly equal to 1:100.000. This means the aquifer only represents a very thin layer parallel to the surface. Under such measures described before it seems not reasonable to build a detailed regional numerical model. Further on there exist not enough piezometric time rows at sufficient observation points in the HVO. The uncertainty of model results could not be controlled by any other means than by the comparison with the generally observed position of the groundwater level at different seasons. Accepting that the lack of input would cause uncertainties within a several meters range they would almost always fit the originally observed groundwater levels.

As consequence only a general approach was applied for simulating groundwater flow in the HVO area as described by the conceptual model in Chapter 7. The resulting model simplifies the HVO area to a great measure by its geometric dimensions and by the distribution of its hydraulic parameters. It was tried to use parameter values within the order of observed field data, values from literature and from other models.

The purpose of the model is to prove the assumption made for the conceptual model that the average groundwater level observed in the field can be maintained by recharge/discharge within the model area.

The constraints given to this modelling approach are discussed in the following chapters respectively. The impact of groundwater exploitation on the groundwater resources was considered under regard of the IMPETUS-scenario information.

8.2 Model geometry

The numerical groundwater flow model is a three-dimensional finite element model with triangular elements of different size. The model area covers almost the whole HVO (Fig. 8.2). The natural boundary of the catchment is the hydrological watershed. The course of the natural watershed is very complex and shows many turns and edges. Discretisation close to the course of this boundary demands an increased number of mesh nodes on the boundary itself. Thus the numerical weight of the boundary elements would greatly influence the inner model solution. To avoid this problem the outline of the model area is chosen within the HVO area and is straighter than the watershed’s course. The model area is about 13,985 km² and thus 3.5% smaller than the total HVO area. The average element size is, under regard of their number per layer, approximately 1 km² (Tab. 8.1).
Local features like inselbergs, duricrusts and bas-fonds were not modelled. The regolith aquifer is generally unconfined.

Horizontally the finite element mesh was automatically distributed by the T-mesh Delaunay procedure (DIERSCH 2005). The refinement along the rivers was done manually. Initial model tests showed that a refinement around the villages was not necessary. Only in the case of Djougou the model mesh around the village’s coordinates was refined to reduce the sphere of influence of drawdown caused by groundwater exploitation in this area.

Tab. 8.1: Number of nodes and elements in the model after the final refinement.

<table>
<thead>
<tr>
<th>Layers</th>
<th>Nº Elements</th>
<th>Nº Nodes</th>
</tr>
</thead>
<tbody>
<tr>
<td>single layer</td>
<td>13,635</td>
<td>9,284</td>
</tr>
<tr>
<td>all layers</td>
<td>40,905</td>
<td>27,852</td>
</tr>
</tbody>
</table>

Vertically the model is structured in three layers. The bottom layer represents the fractured bedrock aquifer while the two top layers simulate the regolith layer. The total thickness of the regolith aquifer was assumed to be 20 m (see Fig. 8.1). The regolith aquifer was then divided by equal parts into an upper and a lower layer for numerical reasons given by DIERSCH (2005). DIERSCH (2005) advises to divide thin layers in two or more layers to produce more elements for water exchange. The bottom depth of the bedrock aquifer was determined as 80 m below ground. This is the maximum depth where groundwater can be exploited from boreholes in Benin under economical conditions (SOGREAH and SCET 1997).

The model slices are determined as plane (Fig. 8.1). Note that the top elevation of the first slice is zero. Groundwater heads are measured as meters below ground level.

Fig. 8.1: Presentation of the model area in a 3D sketch. Three layers, with a plane surface but a dip from the North to the South, represent the regional geology.

8.3 Boundary Conditions

The schematic Fig. 8.2 shows the distribution of the boundary conditions (BC) of the model area.

It is assumed that the hydrogeological watershed of the HVO is at the same position as the hydrological. This is likely because of the shallow water table and the observation of a generally morphology driven groundwater flow (Fig. 8.3 a). The connectivity of
fractures especially in the bedrock aquifer may result in different flow ways (see Fig. 8.3 b). In regard to the size of the HVO it is assumed that inflow and outflow of groundwater beyond the model boundary are in general equal.

The whole HVO groundwater divide is considered as a BC of the 2nd type – constant flow (also known as Neumann condition). In this special case it is a no-flow boundary (flow = 0). Flux in and out along the HVO groundwater divide may occur but is because of the slow groundwater movement which is based on local subbasins only of importance for the subbasins very close to this divide (refer to Chapters 7 and 8.1).

In FEFLOW® special constraints can be given to each boundary condition. It was chosen to supply the Neummann condition with a head constraint of minimum 8 m bgl to a maximum of 5 m bgl. This procedure was chosen to assure that boundary nodes would not fall dry and is in accordance to the head observations made in the HVO area as described in Chapter 4.

![Diagram](image.png)

Fig. 8.2: Distribution of the boundary conditions in the model area for the 3rd layer (bedrock). For the regolith layer the distribution is the same, but without the well boundaries (Projection: UTM, Zone 31P, WGS 84).

The model needs water discharge in its internal area otherwise it would stockpile without limit. As stated before regional groundwater flow is excluded. Downstream flow appears under these conditions (Chapter 4 and 7) not effective enough to deliver an important discharge function. In regard to the assumptions made in Chapter 7 discharge through vegetation was included into the model. Based on the conceptual model a buffer area around the rivers was created to represent the riverine forests (see Chapter 8.4).
These buffers areas had to be secured from running dry as water transfer due to the relatively low overall aquifer permeability (Chapter 8.4) would be too slow for a refilling in time. Therefore the rivers were configured as fixed head conditions of the 1st type (Dirichlet condition) with a general value of 10 m bgl. This is a couple of meters deeper than where the groundwater table’s position is found according to the observation in Fig. 7.2. The reason was to provide generally more space for groundwater fluctuations below the top layer of the model.

![Fig. 8.3: Case A - Accordance of the hydrological and hydrogeological watershed. Case B – Shift of the hydrogeological watershed due to fracture connectivity.](image)

FEFLOW® defines wells and pumps as 4th kind BCs. The position of the well conditions is copied to every model layer but the extraction is assigned to the bottom nodes only (Fig. 8.4). Ideally the extracted water at the bottom layer should be equilibrated by water from the 2nd layer. Exploitation rates for domestic use are rather small and generally negligible for regional models.

![Fig. 8.4: The placement of the well conditions at the bottom of the third model layer is based on the assumed filter position in drill holes. Extracted water leaves the model without any redistribution. Therefore the mesh nodes of concern in the 1st and 2nd layer are described as well conditions with 0 m³/d extraction. The nodes set vertically above one to another will act numerically as a connected tube. Extracted water will be equilibrated by inflow from the other layers.](image)
8 Groundwater flow model

8.4 Hydraulic parameters

The characterisation of the hydraulic properties of the HVO aquifers demanded a high grade of generalisation of the most important parameters for saturated flow: the hydraulic conductivity and the storage coefficient.

It is quite difficult to generalise $k_f$-values in the regolith aquifer due to local differences like macropores, roots, duricrusts or different weathering intensity as presented by FASS (2004) and BAUER (2004). However, most of the $k_f$ values found in the Beninese regolith vary from 1E-04 m/s to 1E-08 m/s. In respect to the regional scale, the fractured bedrock was considered as a porous medium with a lower permeability. The values applied in Tab. 8.2 are found during the process of model calibration. A lower permeability would hinder the exchange of groundwater in this model and it would start to pile up. Higher permeabilities caused the regional equilibration of the groundwater table and the discharge areas became disguised.

The storage coefficients presented in Tab. 8.2 were taken from literature (SOGREAH and SCET 1997; BOUKARI et al. 1985). FEFLOW® requests vertical hydraulic conductivities which were assumed as 1E-01 less than those of the horizontal hydraulic conductivity (LÄNGGUTH 2004).

Tab. 8.2: Hydraulic parameters as applied to the model layers.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Type</th>
<th>Horizontal $k_f$ [m/s]</th>
<th>Vertical $k_f$ [m/s]</th>
<th>Storage coefficient S [-]</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Regolith</td>
<td>5E-06</td>
<td>5E-07</td>
<td>0.03</td>
</tr>
<tr>
<td>2</td>
<td>Regolith</td>
<td>5E-06</td>
<td>5E-07</td>
<td>0.03</td>
</tr>
<tr>
<td>3</td>
<td>Bedrock</td>
<td>1E-08</td>
<td>1E-09</td>
<td>0.000001</td>
</tr>
</tbody>
</table>

Recharge was applied on the top layer. Details about the recharge distribution are given in Chapter 8.5.3. In accordance with the conceptual model (Chapter 7) it was necessary to establish discharge areas close to the river net. A buffer area with its limit in 1 km distance from the river to each direction was created (Fig. 8.2). By this way an area of 2,291 km² got covered (16.4% of the total model area).

It is likely that discharge through transpiration by plants is an important factor in the water budget. Most of the trees in the study area have roots reaching deeper than 10 m. Their water consumption might be the reason for the seasonal draining of the groundwater level. The constant base level of the groundwater hydrographs may thus represent a general balance between recharge and evapotranspiration.

Evapotranspiration was already subtracted from the recharge calculation by the UHP model (chapter 8.5.3). But transpiration still is an important process affecting the groundwater table in greater depths.

The extension of the river buffers is greater than the riverine forest cover in reality. But the buffer area shall represent as well all off the assumed area in the HVO covered by trees with root systems feeding from the groundwater table (compare with the dense vegetation pattern in Fig. 3.13). Fig. 3.26 shows that the preferential zones for effective recharge are found uphill in the area of the interfluvies. Thus the river buffers are representing the areas without recharge but with discharge instead. Additionally, the creation of the distinct buffers serves the model purpose to create local flow patterns towards the local drains. The manifold branches of the real river system in the HVO are
difficult to identify and cannot be included in such detail into a regional numerical model. Thus only the main rivers were included and the buffers chosen can only be coarse approximation to the real tree cover area with impact on the groundwater table. A value of 73 mm/a is consigned to the discharge area of the river buffers (Fig. 8.2) on the top layer where it replaces the net recharge. The value mentioned for discharge is a compromise between values from literature (BAUER 2004, KLOCK et al. 2001 and KLINE et al. 1970) and an iterating approach on both the stationary and the transient models (Chapter 8.6). However, the choice of the size of the river buffers is mainly assumptious and serves primarily the intention to create a simplified model for regional flow.

8.5 Integration of scenario information

8.5.1 Climate scenarios

The Intergovernmental Panel on Climate Change (IPCC) described in its Special Report on Emission Scenarios (SRES published as IPCC 2000) different pathways for the world’s future development until the year 2001. A set of 4 divergent but still plausible storylines (A1, A2, B1 and B2) was created. The storylines are subdivided by emission schemes or other assumptions. The outcome is a whole family of possible scenario descriptions for every storyline. IMPETUS refers to the scenarios A1B and B1 as input of driving forces to their models (BRUCHER et al. 2005). A short description of the scenarios is taken from the SRES (IPCC 2000):

A1 – A future world with rapid economic growth and a global population that peaks in mid-century and declines thereafter and the rapid introduction of new and more efficient technologies. The A1 scenario family develops into three groups that describe alternative directions of technological change in the energy system. The three A1 groups are distinguished by their technological emphasis: fossil intensive (A1FI), non-fossil energy sources (A1T), or a balance across all sources (A1B) (where balanced is defined as not relying too heavily on one particular energy source, on the assumption that similar improvement rates apply to all energy supply and end use technologies).

B1 - A convergent world with the same global population that peaks in mid-century and declines thereafter, as in the A1 storyline but with rapid change in economic structures toward a service and information economy, with reductions in material intensity and the introduction of clean and resource-efficient technologies. The emphasis is on global solutions to economic, social and environmental sustainability, including improved equity, but without additional climate initiatives.

The results from the global climate model ECHAM 5 (ROECKNER et al. 2003) for each IPCC-scenario was downscaled to a regional climate model REMO (JACOB 2001) applied by H. Paeth (IMPETUS A1, PAETH 2006). The REMO integrates as well the land use scenarios modelled by the IMPETUS subproject A3. REMO delivers the climatic input parameters for the hydrological modelling with the UHP model for the A1B and the B1 scenarios.
8.5.2 Socio-economic scenarios

IMPETUS (2006) worked out different scenarios for the development of Benin embedded in a general global context. These scenarios and their general storyline background are in short abbreviated as follows:

BI - Economic growth and stable decentralisation;
BII - Economic stagnation and institutional uncertainty;
BIII - Business as usual.

The scenarios should cover the eventual socioeconomic changes in the Ouémé catchment for the future. However, it was necessary to make a regional subdivision. The catchment was structured in three parts: the Upper Ouémé, Middle Ouémé and Lower Ouémé. Their limits were chosen based on general environmental features and not on hydrological boundaries. The impact of the socio-economic scenarios in regard to HVO groundwater resources concerns mostly the development of the rural population (see Chapter 8.5.4). The scenario data of the Upper Ouémé scenario subregion was adapted to the smaller HVO area.

8.5.3 Recharge

The value for recharge was derived from data computed by S. Giertz (IMPETUS subproject A2, Workpackage Hydrology) with the conceptual, hydrological model UHP-HRU 1.1 (BORMANN and DIEKZRÜGER 2003, GIERTZ 2004). The UHP model subdivided the HVO area in 525 so-called hydrological response units (HRU) with an average size of 30.5 km². The software automatically determines the coverage of the model area by HRUs as function of the prescribed area and the terrain elevation. Flow to the aquifer storage is described as net recharge. The climate input data for all HRUs is exclusively taken from the Parakou climate station. Precipitation data was obtained from several climate stations installed by the IMPETUS project in the HVO area.

The average trend of recharge in Fig. 8.5 was calculated by the unweighted average of recharge at all HRUs at each time step respectively. The UHP model starts already in the year 1993 (GIERTZ et al. 2006). The groundwater model instead operates with an initial recharge amount calculated from the UHP model results for the year 2001.

![Graph](image_url)

Fig. 8.5: Average yearly recharge (unweighted) from all HRUs for each scenario A1B and B1 with trend (dashed lines) and trend equation (original recharge data from GIERTZ 2004).
The recharge in the B1 scenario is generally higher than for the A1B scenario but is fluctuating stronger during the year. However, the average curve for B1 (Fig. 8.5) shows a clearly decreasing trend, although in the case of some HRUs the recharge in 2025 is still higher than in 2001.

This groundwater model was only produced in the aftermath of the scenario calculations by the UHP model. A coupled approach for both models to compute the net recharge towards the aquifers was not achieved.

The actual simplified groundwater model uses the initial average recharge for both scenarios (44 mm/a) in the year 2001 for the whole HVO area as entry parameter. Scenario A1B shows no change of recharge during the transient model period because of the almost linear recharge trend. For scenario B1 the annual change of recharge in the HVO was calculated by the B1 trend equation (see Fig. 8.5).

In the transient model it was distinguished between dry months and months with recharge notable the dry season and the rainy season (Fig. 8.6). During the dry season (October to April) recharge was set to zero. The total yearly recharge was distributed to the months of the rainy season (May to September).

Fig. 8.6: Effective recharge towards the groundwater table; calculated input data for the model scenarios A1B and B1. Recharge is limited to the months of the rainy season. The model starts in the beginning of the year 2002 in the middle of the dry season. Scenario A1B shows constant recharge while B1 is characterised by a decreasing recharge.
8.5.4 Water use

Many hundreds of wells and pumps are found in the HVO area. As mentioned in Chapter 2.2 efforts are made to register the public water points in the BDI. As the BDI remains incomplete in regard to the private dug wells and unofficial water points it was decided to calculate the groundwater extraction from the model area preferentially from the average consumption of the inhabitants. Therefore the numerous water extraction points in a township are combined to a single well condition. The total water consumption by the inhabitants respectively is summed up. The potential demographic development for the scenarios BI, BII and BIII is based on the national census data (INSAE 2003) from 2002 (see Fig. 8.7).

The census counted 119 townships in the HVO area. They are converted to water extraction points. When close to another, they were unified to simplify the finite element net. In total 66 extraction points remained and were implemented into the model as well conditions (see Fig. 8.8). The townships Djougou and Tourou (in the periphery of Parakou) are considered as towns, all the others are classified as villages.

The water consumption analysis of SCHOPP (2004) proved that the improved water supply in urban and peri-urban areas and water abundance during the rainy season cause higher water consumption (Tab. 8.3).

Tab. 8.3: Mean values of the water consumption in l/d per capita in different types of townships in the HVO as measured from 2002 to 2003 (SCHOPP 2004).

<table>
<thead>
<tr>
<th>Month</th>
<th>Town</th>
<th>Periphery</th>
<th>Village</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jul</td>
<td>24.9</td>
<td>22.3</td>
<td>22.0</td>
</tr>
<tr>
<td>Aug</td>
<td>46.4</td>
<td>22.8</td>
<td>17.9</td>
</tr>
<tr>
<td>Sep</td>
<td>39.0</td>
<td>19.6</td>
<td>20.9</td>
</tr>
<tr>
<td>Oct</td>
<td>26.1</td>
<td>19.5</td>
<td>18.6</td>
</tr>
<tr>
<td>Nov</td>
<td>24.6</td>
<td>20.9</td>
<td>15.1</td>
</tr>
<tr>
<td>Dec</td>
<td>10.4</td>
<td>19.9</td>
<td>12.0</td>
</tr>
<tr>
<td>Jan</td>
<td>12.0</td>
<td>14.5</td>
<td>14.9</td>
</tr>
<tr>
<td>Feb</td>
<td>18.3</td>
<td>13.7</td>
<td>14.2</td>
</tr>
<tr>
<td>Mar</td>
<td>31.8</td>
<td>12.8</td>
<td>14.4</td>
</tr>
<tr>
<td>Apr</td>
<td>22.8</td>
<td>15.7</td>
<td>13.2</td>
</tr>
<tr>
<td>May</td>
<td>12.1</td>
<td>17.7</td>
<td>15.4</td>
</tr>
<tr>
<td>Jun</td>
<td>11.8</td>
<td>16.0</td>
<td>15.2</td>
</tr>
<tr>
<td>Mean:</td>
<td>22.8</td>
<td>17.9</td>
<td>16.2</td>
</tr>
</tbody>
</table>
Tapped water from the public supply system originates from surface water resources. Based on the water consumption analysis (SCHOPP 2004) it was tried to identify the share of groundwater on public supply (Tab. 8.4). The observed period comprises not a complete hydrological year but only the period from August 2002 to April 2003.

Fig. 8.8: Position of the remaining 66 villages after aggregation of the census data set from INSAE (2003). Each village presents a well condition in the model mesh (Projection: UTM, Zone 31P, WGS 84).

Tab. 8.4: Distributed use of the different water sources used to satisfy the general water demand based on the observations of SCHOPP (2004).

<table>
<thead>
<tr>
<th>Month</th>
<th>Well</th>
<th>Marigot</th>
<th>Rain</th>
<th>Pump</th>
<th>Groundwater (Well+Pump)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aug</td>
<td>92.19</td>
<td>7.59</td>
<td>0.22</td>
<td></td>
<td>92.19</td>
</tr>
<tr>
<td>Sep</td>
<td>90.48</td>
<td>9.51</td>
<td></td>
<td></td>
<td>90.48</td>
</tr>
<tr>
<td>Oct</td>
<td>87.88</td>
<td>12.12</td>
<td></td>
<td></td>
<td>87.88</td>
</tr>
<tr>
<td>Nov</td>
<td>89.3</td>
<td>10.69</td>
<td></td>
<td></td>
<td>89.3</td>
</tr>
<tr>
<td>Dec</td>
<td>80.82</td>
<td>19.16</td>
<td></td>
<td></td>
<td>80.82</td>
</tr>
<tr>
<td>Jan</td>
<td>67.8</td>
<td>16.25</td>
<td>15.94</td>
<td></td>
<td>83.74</td>
</tr>
<tr>
<td>Feb</td>
<td>10.17</td>
<td>19.45</td>
<td>70.38</td>
<td></td>
<td>80.55</td>
</tr>
<tr>
<td>Mar</td>
<td>17.88</td>
<td>16.17</td>
<td>65.95</td>
<td></td>
<td>83.83</td>
</tr>
<tr>
<td>Apr</td>
<td>30.53</td>
<td>9.14</td>
<td>60.35</td>
<td></td>
<td>90.88</td>
</tr>
</tbody>
</table>

During the months of the rainy season water from wells is preferentially used (approx. 90% of the water needs). Water from marigots is used as well although more for cleaning and washing. The village wells are less productive during the dry season. Hence people start to look for water elsewhere, from marigots in the first place. From the middle of the dry season on, pumps become the principal water source. In general it can be said that
groundwater abstraction covers around 80% of the total water consumption in the rainy season, and respectively 90% in the dry season. It is assumed that these percentages stay constant during the modelled time period. The water consumption for each of the 66 townships is calculated by multiplying the population respectively with the daily consumption [m³/d] and the seasonal percentage of groundwater extraction. The resulting data is imported into FEFLOW® as the 4th boundary condition.

A simplified calculation demonstrates the great potential of groundwater in the HVO to satisfy the public demand (Tab. 8.5). Assuming a worst case scenario the water content for both aquifers is assumed to be very low and recharge is ignored. Maximum water extraction from a population as projected for the year 2025 is added (Scenario B2). Human consumption will thus discharge the total aquifer storage only by 1.4%.

Tab. 8.5: Groundwater volume in the aquifers of the HVO. Minimum water content for both aquifers is assumed (low saturation level for one year). Recharge is ignored. Maximum water extraction from a population as projected for the year 2025 is added (Scenario B2).

<table>
<thead>
<tr>
<th>Aquifer</th>
<th>Saturated thickness [m]</th>
<th>Storage coefficient²</th>
<th>Area [km²]</th>
<th>Water Volume [m³]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Regolith</td>
<td>5</td>
<td>0.03</td>
<td>14,300</td>
<td>21,450,000,000</td>
</tr>
<tr>
<td>Bedrock</td>
<td>20</td>
<td>0.00001</td>
<td>14,300</td>
<td>2,860,000</td>
</tr>
<tr>
<td>Total Volume</td>
<td></td>
<td></td>
<td>21,452,860,000</td>
<td></td>
</tr>
<tr>
<td>Exploitation (2025)</td>
<td></td>
<td></td>
<td>3,056,950.11</td>
<td></td>
</tr>
<tr>
<td>Partition of exploitation on total water volume</td>
<td></td>
<td></td>
<td>0.014 (1.4%)</td>
<td></td>
</tr>
</tbody>
</table>

¹Minimum values for the saturated zone.  
²Minimum values from ENGALENC (1978).

The calculation of discharge by pumping for the HVO area (Tab. 8.6) in the years 2002 and 2025 show that the groundwater exploitation by humans is negligible in comparison to the recharge. However, local exhaustion by extensive pumping, eventually in combination with bad hydraulic conditions, cannot be excluded.

Tab. 8.6: Calculation of the discharge (in mm/a) caused by pumping in the HVO area for the comparison with the regional recharge.

<table>
<thead>
<tr>
<th></th>
<th>Year 2002</th>
<th>Year 2025</th>
</tr>
</thead>
<tbody>
<tr>
<td>Population (HVO)</td>
<td>291,125</td>
<td>564,125</td>
</tr>
<tr>
<td>Consumption [l/d]</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>Total consumption [l/a]</td>
<td>2,125,212,500</td>
<td>4,118,112,500</td>
</tr>
<tr>
<td>Areal discharge [mm/a]</td>
<td>0.15</td>
<td>0.28</td>
</tr>
</tbody>
</table>

It is seen that the demographic development for all IMPETUS scenarios differs only slightly after the year 2015 (Fig. 8.7). The FEFLOW® model is therefore run with the well exploitation data for the scenario BII which shows the strongest increase in population and in water use respectively.
8.6 Stationary model

In regard to the seasonal conditions observed in the HVO it was difficult to decide about the setting of the stationary model. A stationary model for the dry season (no recharge) would produce only the piezometric base line as shown in the Fig. 4.4, Fig. 4.5 and Fig. 4.6. The model would be obsolete. On the other side it is obvious that a model representing recharge conditions will display an overfilling with water which could be managed only by artificial outlets. However, this second option was chosen to quantify the surplus of water and cope with it in the transient models.

The stationary model for the HVO represents an average rainy season. It was run with an initial recharge of 44 mm/a and a discharge of 73 mm/a as discussed above. All hydraulic parameters and geometric proportions remain constant.

The resulting groundwater heads are in accordance with the conceptual model. The modelled groundwater contours are represented in Fig. 8.9. The groundwater flow regime is under control by the discharge from the areas around the closest river system.
respectively. Supraregional flow patterns are not observed. Drawdown by pumping does not show impact on the groundwater table under stationary conditions.

In some areas the groundwater table are lower than 10 m bgl. These areas are generally found close to river joints. Some are marked by black circles. The reason for this decrease is simply the areal extent of the discharge zone without a nearby recharge. However, these areas are only a few and were ignored as the aim of this model is showing the regional trends and not the exact analogue to natural conditions.

The total water balance for the stationary model reveals (Tab. 8.7) that the boundary conditions (especially the 1st BC) transfer quite big amounts of water to the outside of the model.

Tab. 8.7: Total water balance of the stationary model. The well flux occurs at the 3rd layer only.

<table>
<thead>
<tr>
<th>Flux type</th>
<th>Flux in (+) Q [m3/d]</th>
<th>Flux out(-) Q [m3/d]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dirichlet, 1st BC</td>
<td>1.55E+04</td>
<td>-9.72E+04</td>
</tr>
<tr>
<td>Neumann, 2nd BC</td>
<td>1.88E+03</td>
<td>-9.33E+03</td>
</tr>
<tr>
<td>Well flux</td>
<td>0</td>
<td>-3.16E+03</td>
</tr>
<tr>
<td>Aereal flux</td>
<td>1.23E+05</td>
<td>-3.08E+04</td>
</tr>
<tr>
<td>Imbalance</td>
<td>0</td>
<td>-5.4</td>
</tr>
</tbody>
</table>

The comparison of discharge and recharge shows that recharge in the stationary model is about three times greater than the discharge (Tab. 8.8). The excess water is removed by the BCs by the order of 28.6 mm/a. The 1st order BC is applied only to the river course itself and not to the total river buffer area. The elements of concern cover an area of around the half of the buffer area. Thus the influence of the 1st BD is objectionable. A more adjusted water balance would be achieved if a lower recharge of 22 to 33 mm/a would be applied. Then the discharge and recharge would compensate related to the area of concern.

Tab. 8.8: Comparison of discharge and recharge in the HVO model related to their share of the HVO surface.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Coverage of the total HVO area</th>
<th>Water input/output related to the area of concern [mm/a]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Recharge (5 months per year)¹</td>
<td>83.62 %</td>
<td>15.3</td>
</tr>
<tr>
<td>Discharge (buffer area)</td>
<td>16.38 %</td>
<td>12</td>
</tr>
<tr>
<td>Imbalance stationary model</td>
<td></td>
<td>24.8</td>
</tr>
<tr>
<td>Imbalance transient model</td>
<td></td>
<td>3.3</td>
</tr>
<tr>
<td>1st BC</td>
<td>50 % (of the total buffer area)</td>
<td>26</td>
</tr>
</tbody>
</table>

¹ Reduction of recharge in scenario B1 leads to 11.5 mm/a in the year 2025 and thus a gradual equilibration of the water balance takes place.

For the transient models instead it has to be kept in mind that the original recharge is only distributed to 5 months of the yearly rainy season. The imbalance of the water
budget is then smaller. A difference of around 3.3 mm/a is computed which is automatically equilibrated by the BCs. In regard to the very simple model conditions this value appears relatively small and adjustable by minor corrections of the chosen parameters. These changes would not essentially change the order of the here chosen values for recharge and discharge. The simplified model approach supports the assumption that a regional groundwater flow is not needed to maintain the hydraulic groundwater conditions as observed in the field but rather a local flow control by drainage due to the riverine vegetation on a subcatchment scale. As consequence of the lack of field data to quantify and finally prove this assumption the model results remain unsatisfying.

The groundwater contours are used as initial head distribution in the transient models afterwards (Chapter 8.7). In FEFLOW® the stationary model has to be run first before the transient model in order to conserve the general flow system.

The stationary model was proved on its sensitivity to recharge. Therefore it was run with values for recharge ranging from zero to 200% changing in 25% steps (Tab. 8.9). The limit for the numerical head error was determined as 1E-03. In most cases the iteration was successful after 4 computing steps.

Tab. 8.9: Stepwise variation of recharge as input for the stationary model. For each case the error on the hydraulic heads was controlled. Minimum limit for successful computing is 1E-03 in less than 12 iteration steps (FEFLOW® default conditions, DIERSCH 2005).

<table>
<thead>
<tr>
<th>Case</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>
</tr>
</thead>
<tbody>
<tr>
<td>Recharge [%]</td>
<td>0</td>
<td>25</td>
<td>50</td>
<td>75</td>
<td>100</td>
<td>125</td>
<td>150</td>
<td>175</td>
<td>200</td>
</tr>
<tr>
<td>Recharge [mm/a]</td>
<td>0</td>
<td>11</td>
<td>22</td>
<td>33</td>
<td>44</td>
<td>55</td>
<td>66</td>
<td>77</td>
<td>88</td>
</tr>
<tr>
<td>Iteration steps</td>
<td>&gt;12</td>
<td>&gt;12</td>
<td>4</td>
<td>4</td>
<td>4</td>
<td>4</td>
<td>4</td>
<td>4</td>
<td>4</td>
</tr>
</tbody>
</table>

*initial case

For each case the average standard deviation of the areal groundwater head distribution was computed and is shown in Fig. 8.10. The model solutions for cases 1 and 2 are numerical insufficient in computing the residual heads. Adjustment in other parameters is needed, e.g. discharge. The options for modification are manifold and the lack of hard data does not justify any special approach. Case 6 still shows groundwater tables below ground surface. As shown above, this is due to the boundary conditions. All other cases with higher recharge produce tables partly above the surface.

The solutions with decreasing recharge show a lower deviation. However, the cases 3 to 6 produce all groundwater table contours within the range of the observations made in nature (Chapter 4). The recharge applied to these cases fit also to the assumptions made before in Chapter 4.2. Recharge may vary about -50% to +25% and is not very sensitive parameter for model calibration. An areal recharge of 22 mm/a could be outbalanced by the discharge introduced and thus represent a probable better start value. However, it preference was given to a higher value closer to the less assumptive UHP model results.
8 Groundwater flow model

Fig. 8.10: Standard deviation of model solutions for different recharge cases. The area between the 22 and 55 mm/a represents equally reasonable model solutions.

8.7 Transient models

The transient models A1B and B1 use the same input parameters and geometric conditions. The initial conditions of both models are described by the stationary model (see Chapter above). Both models start at the 01.01.2002 and run until the 31.12.2025. The models compute daily time steps. The input data about exploitation from wells varies in monthly time steps (Chapter 8.5.4).

The only difference between both models is the data input of recharge. Recharge changes in seasonal steps (Chapter 8.5.3). While model A1B shows constant recharge (44 mm/a), model B1 shows an annual decrease of around 0.48 mm/a.

The calculation in Tab. 8.5 hints already that changes in recharge will have only a minor influence on the model results. But the models show as well the increasing groundwater exploitation by the population.

8.7.1 Scenario model A1B

The groundwater contours at the final time step of model A1B (Fig. 8.11) is mostly similar to the result of the stationary model (Fig. 8.9). Drawdown around villages is principally seen at Tourou. Tourou belongs already to the outskirts of Parakou and is one of the major settlements in the HVO. It has more than 13,000 habitants in 2002 (> 30,000 in 2025). Although Djougou has a higher population (> 50,000 habitants in 2002) drawdown cones are not observed in the same extent. The reason is the relative position of both locations in relation to the discharge zones. At Djougou water is exploited but in the same time it receives additional water from flow towards the rivers. Tourou instead is located closer to a centre of interfluves. Groundwater flow heads towards the two adjacent rivers additionally.
The map of groundwater differences in the HVO shows that the relative drawdown of the groundwater table is only concentrated on areas around wells (Fig. 8.12). Most of the area is only slightly affected by a lowering the groundwater table (< 0.5 m). The relative drawdown around is even more visible by this kind of presentation. In some cases the initial groundwater levels are lower than in the final phase of the scenario model. Those areas are mostly the same which has shown already stronger drawdown (Fig. 8.9). Some areas close to the model boundaries have slightly higher groundwater levels too. This is caused by remaining recharge from the rainy season before and which is still not equilibrated all over the area due to the relatively low hydraulic permeability.

As said before the natural reaction of the aquifers might be very different due to their fractured characteristics.
8.7.2 Scenario model B1

The groundwater contours at the final time step of model B1 (Fig. 8.13) show as well no significant changes in the groundwater flow pattern. Drawdown around Tourou is consequently more evident than in the A1B model. Around other villages as well drawdown cones can be recognised.

The contour map of the table differences (Fig. 8.14) between the final time step of model B1 and the initial head distribution shows for this scenario more extended areas of drawdown. At many interfluves the general groundwater level is lowered around 0.5 to 1 m. The total drawdown at Tourou is still the same as in the A1B model. This means that still sufficient horizontal flow may occur to cover the losses by pumping at this location.
Fig. 8.13: Groundwater contours of the HVO model area from the final time step of the B1 scenario model. Groundwater flow heads generally towards the closely lying river system. Especially around the village of Tourou (see arrow) groundwater drawdown can be observed.

The areas with a higher groundwater table in the final stage of the transient model are decreased. Especially around the rivers they lack but still some are found close to the boundaries. This shows that the lower recharge in the final seasons modelled in the B1 scenario does need less equilibration.
8.7.3 Model comparison

As described above the two scenarios show no important change in regional groundwater flow. It is observed that especially in model B1 local drawdown cones around the villages occur and may locally influence the groundwater direction. The calculation of the difference of the groundwater contours for each scenario is visualised in Fig. 8.15. The general difference between the two contour maps ranges from 0 to 0.35 m. The difference between the transient models is relatively small. The drawdown exemplarily measured at the observation point set into the village of Tourou remains almost the same for both models. The curve observed over the model period of the A1B scenario is only slightly higher (5 to 10 mm) than the one of the B1 scenario. The dynamic is identical. Fig. 8.16 shows the hydrograph of the well at Tourou in the A1B model. The figure shows that the groundwater table is locally mostly under influence of the seasonally adapted water consumption pattern.
Fig. 8.15: Groundwater level differences between the final time steps of model A1B and model B1. Positive values indicate the drawdown of the groundwater table in the A1B model in relation to the initial conditions.

Fig. 8.16: Drawdown at the pumping well of Tourou (black line) and the consumption by its population (blue line). The seasonal fluctuations of water consumption can be traced by the behaviour of the modelled groundwater table.

The groundwater table depletes with the ongoing modelling time and seems to be locally more affected by the increasing local withdrawal from wells and generally less by
decreasing recharge. The amplitude of groundwater fluctuations ranges from 1 to 5 m which is in accordance to the observations described in Chapter 4.

The transient model B1 simulates slightly decreasing groundwater tables due to decrease in recharge. The drawdown is relatively small and thus it is assumed that decreasing recharge is actually no threat to the development of groundwater resources in the HVO area. The aquifers of the HVO will not run dry. But it can be predicted, e.g. by the case of Tourou, that it is possible to exhaust the aquifers locally by continuous pumping. This problem was already considered in the planning of water supply for the bigger settlements in the study area, like Djougou, Bassila or especially Parakou and its periphery. For these settlements lake reservoirs were developed to allow public water supply throughout the year. Decentralised water supply by groundwater pumping for customers at the urban periphery and for rural settlements is in general no problem. Exceptions may exist were the local hydrogeology is not in favour for sufficient groundwater storage.

8.8 Uncertainties and constraints
The major constraints to regional modelling in the HVO are explained below. A quantification of the error of the calculation of the piezometric heads in the model is not possible. The reason is that a fractured environment is modelled as a porous milieu (see section e). However, the modelling results represent an average solution. The following points give a general overview about the general problems limiting the application of a regional numerical groundwater flow model for the HVO area:

a) **Scaling problem:** The model covers an area of around 14,000 km². The smooth changes of the vertical terrain elevation are negligible when compared to the horizontal model extension. An intensive mesh refinement in order to achieve a better representation of the surface demands extensive computational power and a very refined digital elevation model as input source.

b) **Piezometric data:** Piezometer time series are very important for transient models. The piezometric data obtained from the IMPETUS data loggers covers a period of two hydrological years (May 2004 to February 2006). Model prediction for more than twice the time of the gauged period is not recommended (ESSINK 2000; BEAR 1992). Additionally the piezometer groundwater hydrographs may show very individual reactions of the groundwater caused by local features e.g. fractures as discussed in the case of HVO-9 (Chapter 4.1). The regional model approach is not able to integrate these local features. A direct comparison of model data with observed piezometric data seems therefore unreasonable. However, it was possible to compare the modelled data with the overall level of the groundwater table in the studied area.

c) **Conceptual model:** The conceptual model makes difference neither for the differing bedrock types nor for the regolith formed through weathering out of the first. A reason is the lack of consistent data about relevant outcrops and boreholes and the unrefined level of the hydrogeological and geological maps available for the study area. The achieved data gave no clue for differing hydraulic characteristics of the bedrock. As consequence
the conceptual model generalises the main flow components for the whole study area. Single terrain features, e.g. the Kandi fault and the Inselbergs, are not integrated because of the reasons given above (lack of data, scale problem).

d) **Hydrological data**: For the here presented model only average recharge values were applied. The order of magnitude was derived from the regionally differentiated UHP model data (IMPETUS subproject A2, Workpackage Hydrology). As the groundwater model design is simpler as the UHP model an approach to calibrate the net recharge towards the aquifer by coupling the models was not realised.

e) **Fractured bedrock aquifer**: The bedrock layer is modelled like a porous, homogeneous continuum. Thus, the in nature observed sharp peaks and immediate reaction of the groundwater table to precipitation cannot be reproduced in the model. In a porous aquifer the groundwater table outbalances itself and reacts smoother on external influences.

f) **Regolith aquifer**: The model demanded a simplification of the regolith structure. The occurrence of thus semi-confined conditions, due to duricrusts and intensely weathered clay layers, is ignored. Additionally, a constant thickness of the regolith is supposed. Field observations revealed the strongly varying regolith thickness and changes in texture and mineral composition over short distances.

g) **Model limits**: The groundwater divide of the HVO is supposedly equal to the surface water divide. In reality no information about the position and nature of the real groundwater divide is available.
9 Conclusions

9. Conclusions

In the beginning of this study data about the regional hydrogeological and hydrochemical character of the regolith and the bedrock aquifer was only available through geological and hydrogeological maps. It was therefore necessary to undertake several measurement campaigns in order to get an overview about the regional groundwater resources.

The interpretation of results from analyses of groundwater hydrochemistry, stable isotope composition of water resources, tritium and piezometric measurements produced important information about recharge, water age and the evolution of hydrochemical facies.

Based on these results a conceptual regional model was designed. In regard to this conceptual model and some general assumptions about hydraulic parameters in the study area it was tried to create and perform a numerical model for the Upper Ouémé catchment.

The transfer of the hydrogeological features of the HVO into a regional numerical model demanded for strong simplification because of the great extent of the study area. A regional groundwater flow field is assumed for this measure but in the same time its existence remains disputable. As shown in Chapter 4 groundwater tables in the HVO are a reprint of the general surface morphology. Pumping or other types of groundwater extraction would have only very local impact on the available groundwater resources.

It is obvious that the major objective for the numerical model in this environment is not to predict exact piezometric heads but to prove the general assumptions made about the conceptual model. The discussion about the regional groundwater flow model may serve as reference for the development of other regional groundwater models in such areas of complex hydrogeological conditions like the crystalline bedrock with its regolith cover in Central Benin. However, the actual state of available input data reminds that the here model is only a simplified starting point for further modelling.

In relation to the IMPETUS scenario data the groundwater model was able to show that, despite declining recharge, supply of groundwater for public demand until the year 2025 will be sufficient. The experience from the field shows that still some villages are found sparse of groundwater during the dry season. Those villages are found mostly in areas where the regolith is thin or almost not existent and thus lacks important groundwater storage. Additionally important water bearing fractures in the bedrock may lack or where failed to be localised by groundwater exploring activities. But in such places the provision of water supply is rather more a problem of infrastructure and investment than that of a general water shortage.

Reason for concern is the groundwater quality in the vicinity of settlements. Here contamination by human activities is omnipresent. The nitrate concentration in the groundwater already achieved alerting levels in many places. Health risks from fluoride or heavy metals could be excluded for the HVO area.

In respect to the heterogeneous character of the aquifers in the Upper Ouémé catchment it is necessary to underline that all investigations for this study were realised on a regional scale and thus conclusions drawn on them have to be regarded absolutely under this generalising aspect.
Nevertheless, the here presented regional approach produced a good conceptual idea about hydrogeological conditions and hydrochemical processes in the crystalline basement aquifers of Central Benin. Many findings are made in agreement with those of other researchers at comparable sites. Citations are found in the text above.

The regional field work produced a large database about hydrogeology, piezometric measurements, hydrochemistry and environmental isotope concentrations in the HVO area. It completes the already vast data base collected by the IMPETUS fellow researchers and may thus, hopefully, serve well for future research in the Upper Ouémé catchment and in adjacent areas.
10. **Recommendations and outlook**

In rural areas of Benin and elsewhere in Africa groundwater is still the best solution for water supply (WURZEL 2001). The relatively small borehole yields are sufficient for rural water needs (ENGALENC 1985). Groundwater is ubiquitous albeit water bearing fractures have first to be localised. The clue to better implantation strategies are intensive geophysical investigations, the development and exploitation of consistent borehole databases (e.g. BDI), the interpretation of morphological features from a DEM and interpretation of satellite image data. Any improvement in these fields of research is of great concern for the local water supply.

Multi-seasonal observation of the hydrochemical data revealed less importance of seasonal factors for the deeper regolith and the bedrock aquifer. But vertical changes of hydrochemical characteristics especially in the regolith seem to give more information about flow velocity and weathering processes. Detailed hydrochemical research in the regolith in comparison with soil analyses would be of interest. The great number of already available hydrochemical analyses permits as well a profound statistical study and hydrochemical modelling, for example with the PHREEQC modelling software.

Tritium proved the existence of different ages in a vertical sequence in the aquifers. A closer view to the tritium ages in combination with transport models (e.g. with the NETPATH model) could give a better understanding to flow velocities in the regolith (e.g. BORONINA et al. 2005).

Since October 2004 rainwater was collected at three locations in the HVO (Parakou, Dogué and Séröu). The samples were sent to the GSF research centre in order to measure the content of deuterium and oxygen-18. It might be interesting to include these measurements in meteorological models in order to locate the origin of precipitation causing recharge. These samples may serve as calibration tool in meteorological models. An exemplary application is described in STURM et al. (2005).

The groundwater monitoring via data loggers should be continued in order to obtain substantial time series for conclusive statistical analysis of trend components and periodicity.

The here presented FEFLOW® model permits the easy integration of new data. Further refinement can be done in areas of interest and closer investigation. FEFLOW® provides a free programmable interface. This interface can be used for coupling with procedures from other models, e.g. the hydrological UHP model, in order to obtain a conclusive modelling environment. Model coupling was not part of this thesis but would be a great task for further model work in Benin.
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