Groundwater Field Course in Kappelen

Alessandro Agazzi, René Kaufmann, Karin Treyer

July 1, 2010

This work is made within the scope of the course
"‘Fach- & Computerlabor"’ at the ETH Zürich

Supervisor:
Daniel Braun

Assistance:
Matthias Willmann
Thomy Keller
Li Ning

Authors:
Alessandro Agazzi - aagazzi@student.ethz.ch
René Kaufmann - krene@student.ethz.ch
Karin Treyer - treyerk@student.ethz.ch

Cover Picture:
Pumping Well, Foto: René Kaufmann

Abstract

Between the 14th and the 18th of June 2010, groundwater field work was done at the Kappelen site. Eight experiments provided an overview on the aquifer properties for large and small scale. The most important are presented shortly in the following: In order to determine an overall hydraulic conductivity, a pumping test with and without packer was conducted. A tracer test with naphthenate gave information on field porosity and dispersion. Small-scale variability of important parameters could be analysed with a flow meter test and a dilution test. Other tests helped to round out the picture of the aquifer conditions. Flow direction is parallel to the river "Alte Aare", the hydraulic conductivity was calculated to be between about 0.00285 m/s and 0.0207 m/s. These values fit well into the range of sandy gravel conductivities. With a flow meter test, small-scale variabilities could be investigated. All experiments worked well but data were sometimes difficult to analyze. Like this, handling of real data could be learned.

Contents

List of Figures

List of Tables

1 Introduction

1.1 Research area

The research area is situated between Kappelen and Lyss (see fig. 1.1) close to the motorway A27 in a hardwood forest with mainly beeches (Fagus sylvatica). The ground consists of Molasse delta that was deposited by the Aare river after Würm glacial. Around Kappelen this delta consists of large gravel deposits with a thickness of around 450 m. The aquifer consists mainly of sandy gravel and has a thickness of roughly 15 m. It is supplied by precipitation (900 mm/yr in Payerne with an average evapotranspiration of 450 mm/yr) and by the "Alte Aare" river. However, it is almost impossible to obtain this amount. Overall, the groundwater recharge is approximately 5–10 l/s/km^2 which depends strongly on the depth of the groundwater table and the texture.

The groundwater flow is parallel to the flow direction of the "Alte Aare" river (north).

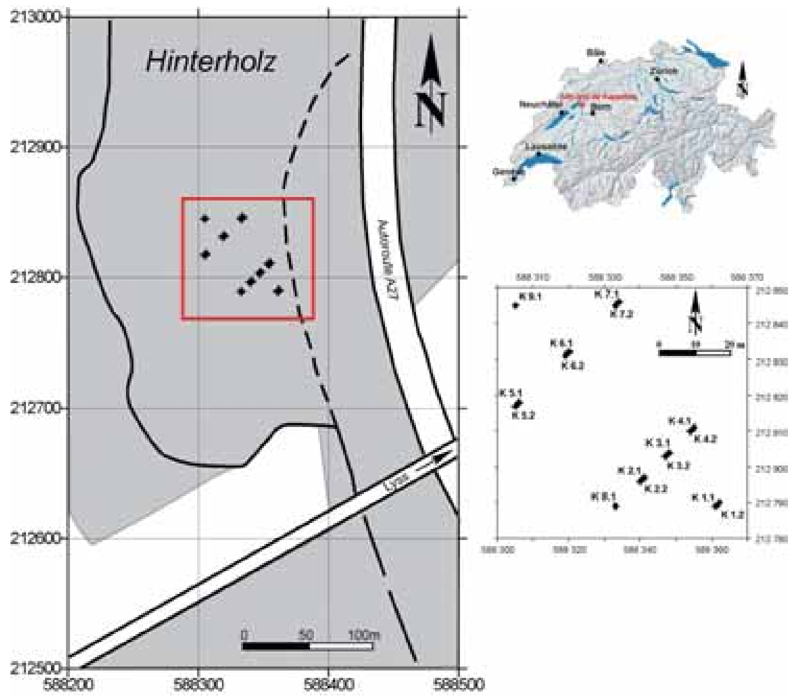

Figure 1.1: Research area near Kappelen

1.2 Motivation

The motivation for studies done at Kappelen originates in applying methods that were introduced in the lectures as well as plan and perform field experiments using field equipment. New methods to determine aquifer paramters are studied. Measurements coming from the field are analyzed, which is not the same as managing well-prepared data for assignments. Working in groups and subgroups has to be organized and can be trained during the field and report work.

1.3 Experiments

The first experiment consists of measuring the groundwater level with a water level conduct meter (Kabellichtlot) at all 16 boreholes in order to be able to draw a flowfield/groundwater map. This data and the knowledge of the flow direction is useful to plan further investigations. At the same time, vertical profiles of temperature and electrical conductivity are registered.

Recharge rate is calculated in the second experiment with the Penman-Monteith FAO method using data provided by the meteorological service in nearby Payerne. The results are discussed with respect to groundwater level fluctuations and the temporal delay between recharge and increase of the water table.

Experiment 3 is about spatio-temporal variability in soil moisture. Measurements of volumetric water content by hand at different places at the research area and automatical measurements for three days are compared. Soil moisture content plays an important role in surface runoff and therefore groundwater recharge.

By doing a pumping test (experiment 4), aquifer parameters are determined (e.g. hydraulic conductivity, storage coefficient). Those parameters are needed for the 2D modelling of the aquifer. The drawdown is analysed with and without an installed packer at the bottom of the well.

With a flowmeter test (experiment 5), a vertical distribution of flow velocities and therefore hydraulic conductivities is found.

Experiment 6 (dual pumping technique) is done to get a vertical profile of the chloride concentration.

Experiment 7 is used to determine the local Darcy velocity measuring the electrical conductivity. This dilution test is done by injecting salted water that is slowly diluted due to diffusion, dispersion and advection.

The last experiment 8 (tracer test) is the longest and the most expensive experiment. 1000 g of napthenate are injected into a pumping well located upstream of the monitoring well. The measured data of the tracer test is compared to data of a 2D model (modelled in Processing Modflow). Out of that comparison, information about porosity and dispersivity over the whole research area are gained and mass transport investigated.

1.4 Weather

The field week took place from 14th to the 18th of June 2010. May has been a quite rainy month with total precipitation of 92.1 mm. Just the day before the field work started, 11.5 mm of rain were measured. Values for the following days amounted to 1.3 mm for Monday, 0.7 mm for Tuesday and 6.8 mm for Wednesday. On Thursday there was no rain until departure. The sun did nearly never shine during the four days.

2 Experiment 1 – Vertical Profile

2.1 Introduction and aims

With information on the groundwater level it is possible to get information on the hydraulic head of the groundwater flow. Three measurements allow predicting the local horizontal direction of the groundwater flow. With time series of hydraulic heads the hydraulic conductivity k (m/s) and the storage coefficient S (-) can be estimated by model calibration techniques.

The vertical profiles of electrical conductivity and temperature give information on processes in the aquifer, e.g. it can indicate the infiltration of river water or give hints on the geological structures. Temperature can show a yearly cycle; it will be shifted and dampened as a function of the depth.

The aim of this experiment is to learn how to measure the groundwater level and the vertical profile of electrical conductivity and groundwater temperature. With the results of the groundwater level out of different boreholes a groundwater map with information about flow direction is created.

2.2 Description of test

In each borehole the surface of the water level is measured with a water level contact meter (dt.: Kabellichtlot). When the electrode contacts the water, a lamp on the cable roller will switch on due to the electrical conductivity of water. On the tape measure the depth of the water level can be read at the upper side of the boreholes (first reference points). To create the groundwater map all boreholes have to be adjusted to one single reference height. For this case, borehole 1.1 is the reference borehole. In table A.1 the correction of the heights is shown.

With the help of the program "Surfer", nearby groundwater depth measurements could then be interpolated with Kriging method and a groundwater field could be created. If making a first assumption of an isotrope homogeneous aquifer, filter velocity as well as interspace velocity can be calculated with the help of the Darcy law. The electrical conductivity and the temperature are measured with KLL-Q, a water level contact meter with a special sensor to detect this and other parameters. The electrical conductivity gives information on the amount of dissolved ions in the water. Generally, a higher electrical conductivity is generated by longer residence time in the aquifer and therefore correlates with the age of the measured groundwater.

2.3 Results

Table A.1 in the appendix gives information on the measured water levels and other information on the boreholes.

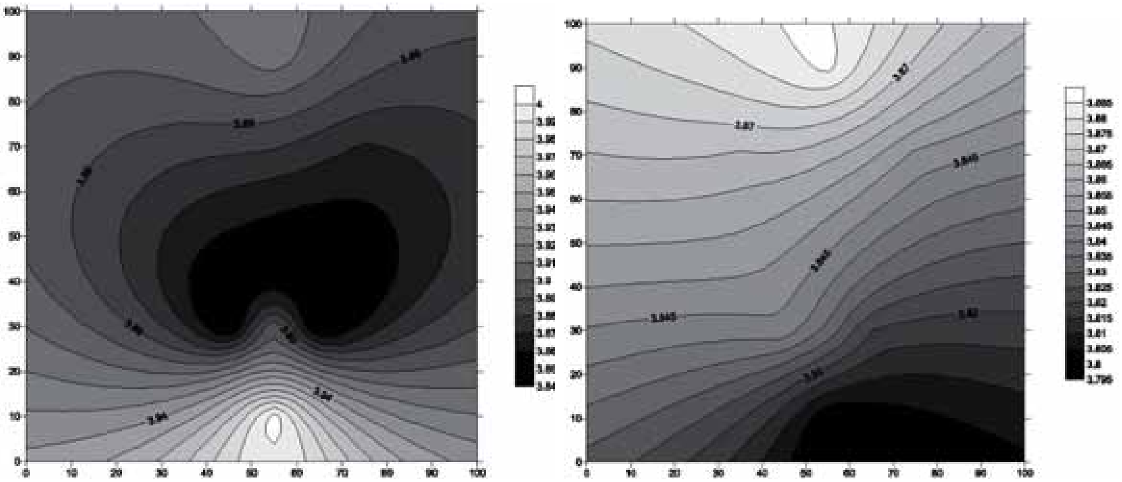

Figure 2.1: Groundwater flow field of the deeper boreholes. Number 1.1 to 9.1. *Left:* Undisturbed groundwater flow field in Kappelen, measured on 7th June 2010. *Right:* Groundwater flow field measured on 14th June 2010. The groundwater flow field is influenced by pumping.

Figure 2.1 shows the groundwater flow field for the boreholes where the filter line is deeper than 8 m from the ground surface. Keep in mind that only the draw down is measured and therefore higher values are lower points.

In the undistributed case the flow field has two high points in the middle of the field, the groundwater level is in a depth of 3.84 m. From these two points the groundwater flows in all direction. The deepest point is located at coordinate 55/10 (borehole 1.2), the draw down lies 0.16 m deeper than the highest point.

While the pump is running the flow field has a gradient. The highest point is near coordinate 60/0 and the lowest point near 60/100. The pump, located at 55/30 (borehole 3.2) influences the flow field. Near this point there is a kink.

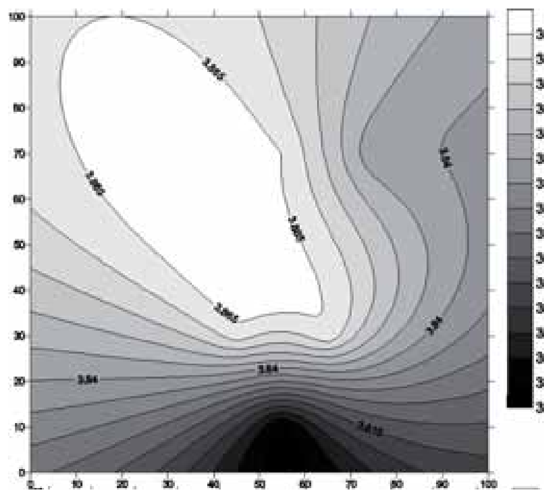

Figure 2.2: Undisturbed groundwater flow field upper boreholes. Number 1.2 to 7.2, measured on 7th June 2010.

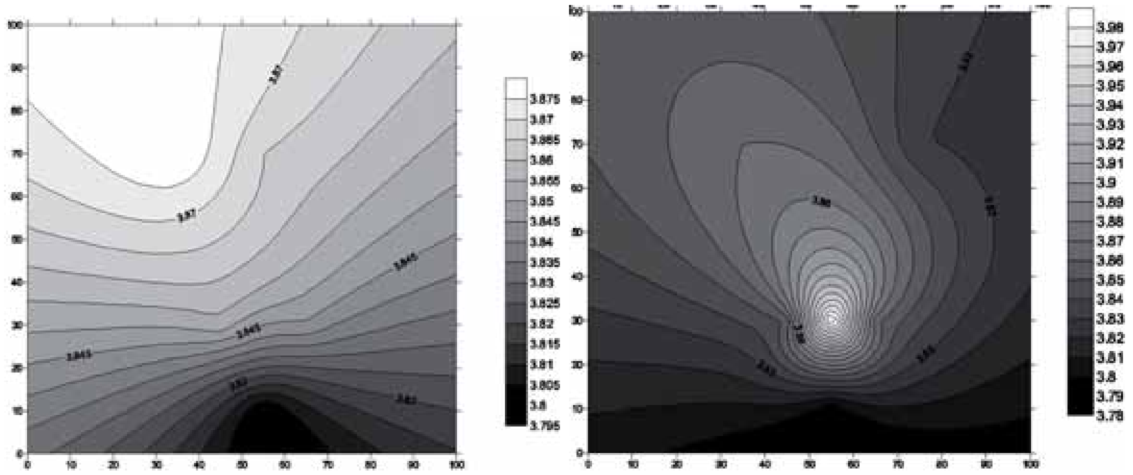

Figure 2.3: Disturbed groundwater flow field upper boreholes. Number 1.2 to 7.2, measured on 14th June 2010. *Left:* Flow field when the draw down in the pumping well is ignored. *Right:* Flow field with draw down in the pumping well.

Figure 2.2 shows the undisturbed groundwater flow field. The direction of the flow field points to the upper left corner.

In figure 2.3 the disturbed groundwater flow field is shown. The left figure shows the groundwater flow field when the draw down in the pumping well is ignored. The highest draw down is in the upper left corner. The right figure shows the groundwater flow field with the draw down in the well. It shows the highest draw down in the well.

As next step, the vertical profiles of electrical conductivity and temperature were measured. Figure 2.4 shows the temperature profile in boreholes 5.1 and 5.2.

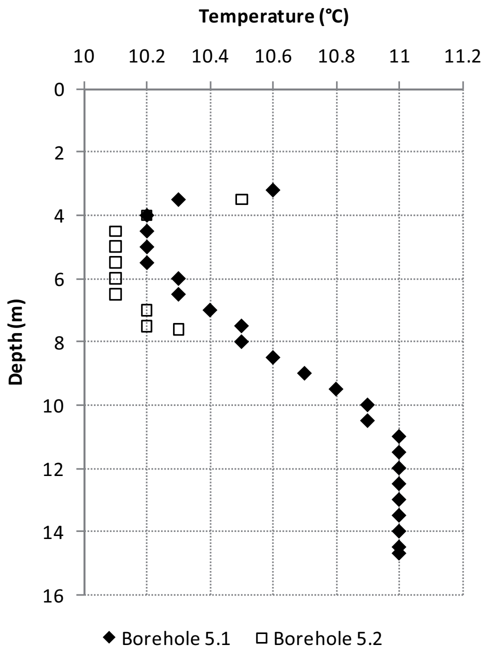

Figure 2.4: Temperature measurements in boreholes 5.1 (deeper) and 5.2 in °C on 14th of June 2010.

In the first few meters, temperature decreases around half a degree. For the depths between 4 m and 6 m the temperature is stable. Between 6 m and 11 m the water temperature increases linearly to 11°C, where it stays stable for lower depth.

In the same two boreholes, electrical conductivity was measured (see fig. 2.5). In contrast to the temperature, the values remain about the same with depth, especially in borehole 5.1. Borehole 5.2 has a extreme value at its bottom (7.6 m). The depth of the borehole is only just under 8 m.

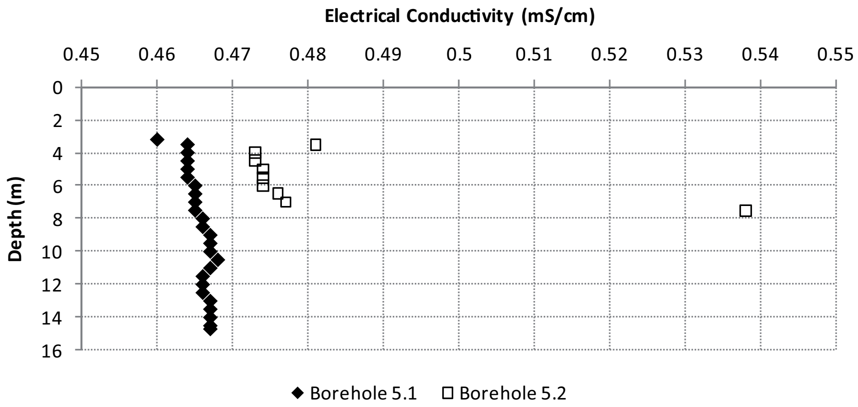

Figure 2.5: Measurements of electrical conductivity in boreholes 5.1 and 5.2 in mS/cm of 14. June 2010.

2.4 Discussion and conclusions

The position of the filter line has a large influence on the groundwater flow field map. In the undisturbed case the general flow direction is completely different. The flow field in the upper part shows that the flow field is on the lower boundary, in the case of the lower filer line the highest point is near the wells 2.1/ 2.2 and 3.1/ 3.2. The data were measured by another group a week before the measurement of our group was made. There are no other data for comparison. A source of error can be a measurement error.

For the pumping case there are two important statements to do. First, the draw down in the pumping well has to be measured. In this borehole the draw down is the deepest one and so it influences the flow field. This is shown in figure 2.3. Second, the flow field, in the boreholes where the filter line is in the upper part, is much more influenced by the pump than in the boreholes where the filter line is in the lower part.

Also when ignoring the pumping well, the data gives information about the flow field. It shows clearer the path way of the general flow field. The kink in figure 2.3 shows where the pump is installed.

The phenomenon of higher temperature near the water surface than a few centimeters below occurs because of gas exchange between the air and the water in the borehole. The air temperature was higher than the water temperature the day of the measurement. Over a year the water temperature has a typical cycle (see fig. A.1 in the appendix.)

For the high electrical conductivity values in borehole 5.2 two explanations are probable:

1. The measurement device touched the ground and slug influenced the measurement
2. It could be that in the week before, a dilution test has taken place in this borehole. It is possible that a certain amount of salt remained at the bottom of the borehole.

More measurements should have been taken in borehole 5.2 in order to make sure there have not been made any measurement mistakes. If those measurements would have supported the values above, all 0.1 m should have been taken a measurement to get a finer resolution.

The vertical profiles indicate a quite homogeneous aquifer without big peculiarities. It should be paid attention on taking the vertical profiles in as much undisturbed boreholes as possible. Previous mixing of the water should be avoided. It can be assumed that these conditions have been accepted in the experiment at hand because the temperature profile does not show homogeneity. Data for the two boreholes do not show any inflow of water coming from another source. The values for electrical conductivity are rather low, so that a long hydraulic retention time is not probable. This indicates an aquifer with a high hydraulic conductivity.

3 Experiment 2 – Recharge rates calculation

3.1 Introduction and aims

Besides factors like interactions, pumping or other boundary conditions, the groundwater level depends a lot on the recharge rate of the area. This rate can be calculated starting from the combined water balance for the vadose zone and the saturated zone:

$$P + L_{in} + SR_{in} + SSR_{in} \pm \Delta S = ET + Q + L_{out} + SR_{out} + SSR_{out} \tag{3.1}$$

Where P precipitation $[\mathrm{L}^3/\mathrm{T}]$, ET is evapotranspiration $[\mathrm{L}^3/\mathrm{T}]$, L_{in} infiltrating water from rivers $[\mathrm{L}^3/\mathrm{T}]$, L_{out} exfiltrating water to rivers $[\mathrm{L}^3/\mathrm{T}]$, Q the amount of extracted (pumped) water $[\mathrm{L}^3/\mathrm{T}]$, SR is surface runoff (the subscript in indicates incoming runoff, the subscript out outgoing runoff) $[\mathrm{L}^3/\mathrm{T}]$, SSR is subsurface runoff $[\mathrm{L}^3/\mathrm{T}]$ and ΔS is the change in subsurface water storage in the area including soil moisture and groundwater $[\mathrm{L}^3/\mathrm{T}]$. If ΔS is positive, the storage increases.

In the case of the Kappelen site, surface runoff as well as pumping activities can be neglected. Precipitation and evapotranspiration are probably the main actors for the change in subsurface water storage. Infiltration or exfiltration could take place at the river "Alte Aare", but the vertical profiles of temperature and electrical conductivity don't give hints for such an interaction. Furthermore, subsurface runoff can be neglected because of the geological situation. Resulting from all of these considerations, equation 3.2 takes the form of

$$P \pm \Delta S = ET \tag{3.2}$$

Evapotranspiration is calculated with the FAO Penman-Monteith equation [1]

$$\mathrm{ET}_0 = \frac{0.408 \cdot \Delta \cdot (R_n - G) + \gamma \cdot \frac{900}{T+273} \cdot u_2 \cdot (e_s - e_a)}{\Delta + \gamma \cdot (1 + 0.34 \cdot u_2)} \tag{3.3}$$

ET_0	Reference Evapotranspiration $(\mathrm{mm/d})$
Δ	Slope of the water vapor pressure curve $(\mathrm{kPa/^\circ C})$
R_n	Net radiation $(\mathrm{MJ/m^2/d})$
G	Soil heat flux $(\mathrm{MJ/m^2/d})$
γ	psychrometric constant $(\mathrm{kPa/^\circ C})$
T	mean daily air temperature at 2 m height $(^\circ\mathrm{C})$
u_2	Windspeed at 2 m height $(\mathrm{m/s})$
e_s	Saturation vapour pressure (kPa)
e_a	Actual vapour pressure (kPa)

[1] Source Allen 1998

Equation 3.3 is a close, simple representation of the physical and physiological factors governing the evapotranspiration process. All of the parameters are available for Payerne in form of an excel sheet. The FAO manual provides a large overview on how to calculate the potential as well as the actual evapotranspiration rate. For this experiment, many data were provided over a time frame of one year, measured in Payerne. This is one of the official stations for meteorological data measurement. Payerne is 45 km away from Kappelen and about at the same height. However, climate must not necessarily be the same! At Kappelen site itself, pressure as well as temperature were measured over a whole year in defined boreholes. Because of vandalism, data from the last month are lost and had to be replaced with data from June 2008 to June 2009. The given data were the following:

- For borehole 4.1: Pressure p (bar) and temperature T (C), measured between 18th of June 2008 and 3rd of June 2009 two times a day (at 6:00 and 18:00).
- An excel sheet with many data on e.g. precipitation, air temperature, global radiation, wind, vapour pressure etc. Additionally, potential and actual evapotranspiration are calculated in this sheet according to the FAO Penman-Monteith method.

Pressure data of the boreholes have to be corrected with air pressure and are then converted into mWs.

3.2 Results

Figure 3.1 shows the calculated recharge rate and the measured groundwater level. Recharge rate was calculated by subtracting the evapotranspiration from precipitation.

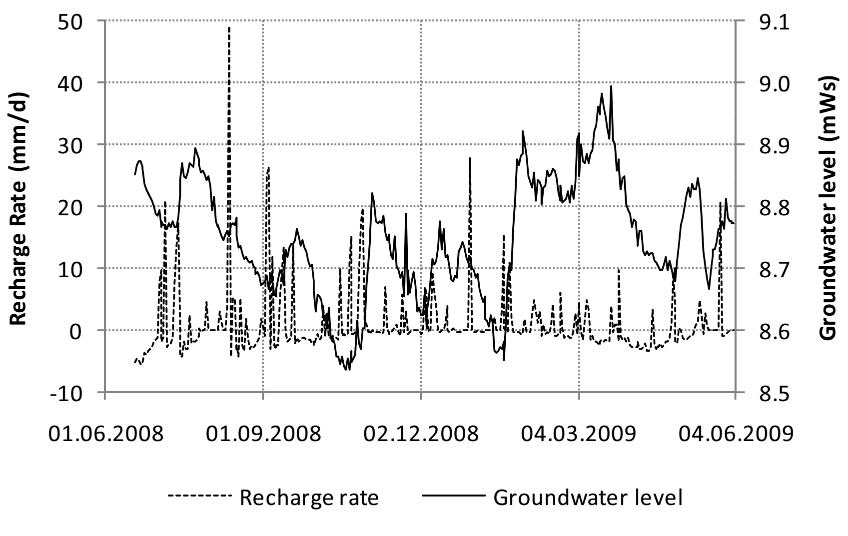

Figure 3.1: Recharge rate versus the groundwater level on the Kappelen site

3.3 Discussion

It could the case be that the recharge rate has a direct influence on the groundwater level, but especially in the second half of the graph, this assumption is not supported. Recharge rate is rather low in winter and spring time (January to May), whereas the groundwater level is highest during these months. The variance of the groundwater level fluctuations can be influenced by

- temporal shift between the precipitation event and the final groundwater recharge
- critical parameters for the recharge rate: Precipitation, wind, radiation, unsaturated zone soil properties, vegetation properties etc.
- discharge
- river-aquifer interactions

Temporal shift can rather be excluded when taking into account the above figure, assuming that it cannot go over several months. In case this could be an influencing factor, the delayed response of groundwater levels to precipitation can e.g. be estimated with a kinematic wave formulation. During winter and spring, a renewal of the groundwater can be observed, even if the recharge rate is not high or even negative at that time. Precipitation measurements include snow – in winter 08/09, temperatures were often below zero degrees. The question is if the soil was frozen or if there could take place a recharge of water in form of melting snow. Astonishingly, data show a rising groundwater table with sinking temperatures. Therefore the high groundwater recharge e.g. in January cannot be explained. Maybe there takes place an interaction between the river "Alte Aare" and the aquifer.

Another reason for the low correlation between recharge rate and groundwater level could be the uncertainty on provided data. These are valid for Payerne which is about 45 km away from Kappelen and 490 m above sea level. However, Switzerland shows a high diversity of small-scale climates.

4 Experiment 3 – Soil Moisture

4.1 Introduction and aims

Soil moisture plays a crucial role in hydrology. On the one hand in the vadose zone hydrology for crop in agricultural production, on the other hand in hydrometeorology for runoff generation or groundwater recharge respectively. For instance, it gets more and more difficult for plants to transpire if the soil gets drier. Over initially wet or very dry soils, precipitation leads much faster to surface runoff.

In this experiment, the temporal variation in soil moisture content at different places during four days is investigated as well as the spatial variability on a 1-by-1 m area. The results of measurements in the wood and on the meadow are discussed.

4.2 Description of test

The measurements are all done with the 10HS Soil Moisture Sensor (see fig. 4.1). This sensor is based on the time domain reflectometry (TDR). It generates an electromagnetic field which emits electromagnetic waves. These waves travel faster through soil than through water due to the fact that the dielectric constant of water is much higher than that of soil. The resulting differences in travel time are then correlated to the volumetric water content (VWC).

Figure 4.1: 10HS Soil Moisture Sensor based on time domain reflectometry (TDR)

Handheld measurements At different places in the forest, along the edge of the forest and on the nearby meadow, measurements are done with the TDR once a day at four days. In addition to these point measurements describing roughly the temporal variability, two 1-by-1 m areas (forest and meadow) are chosen to investigate the spatial variability at a certain point in time. That square is divided into smaller squares (0.2-by-0.2 m) where at each grid point the VWC is measured. Those results are compared and used for a variogram.

Long-term measurements In order to get a more detailed distribution of the VWC in time, several TDR are put into a hole at both forest and meadow at different soil depths (30 cm and 50 cm). During a four days run, the measuring interval is 30 min. The hole is covered by a wooden board. Apart from the VWC, temperature is registered.

4.3 Results

Handheld measurements Table 4.1 shows the temporal variability (mean and standard deviation) of the VWC at four different places starting on 14th June until 17th June 2010. The VWC decreases until 16th of June and rises afterwards again.

Table 4.1: Temporal variability in VWC (m^3/m^3) at different places (cut and uncut meadow, bareland and forest) from 14th to 17th of June 2010

		Meadow, cut		Meadow, uncut		Bareland		Forest	
Date	Time	Mean	Stddev.	Mean	Stddev.	Mean	Stddev.	Mean	Stddev.
14/06/10	afternoon	0.20	0.02	0.25	0.04	0.30	0.02	0.22	0.05
15/06/10	afternoon	0.15	0.02	0.17	0.05	0.19	0.09	0.17	0.06
16/06/10	afternoon	0.11	0.00	0.12	0.00	0.16	0.00	0.09	0.00
17/06/10	morning	0.16	0.01	0.17	0.01	0.18	0.02	0.14	0.04

Table 4.2 lists the mean, standard deviation and minimum and maximum of VWC of a 1-by-1 m square in the forest. The equivalent measurements on the meadow are neglected (explanation see variogram).

Table 4.2: Mean, standard deviation, minimum and maximum of VWC (m^3/m^3) measured within a 1-by-1 m area in the forest

Date	Time	Mean	Standarddeviation
14/06/2010	afternoon	0.22	0.05
15/06/2010	afternoon	0.17	0.06
16/06/2010	afternoon	0.09	0.00
17/06/2010	morning	0.14	0.04

The following figure 4.2 shows the variogram of the VWC measurements conducted at the 1-by-1 m square in the forest and on the meadow. The data set consists of 36 measurements. Due to a quite compact soil on the meadow, the TDR-sensor couldn't be pushed completely into the soil. Therefore, the data used for the variogram of the meadow represents only that VWC which was measured tucking the sensor by half of its length into the soil.

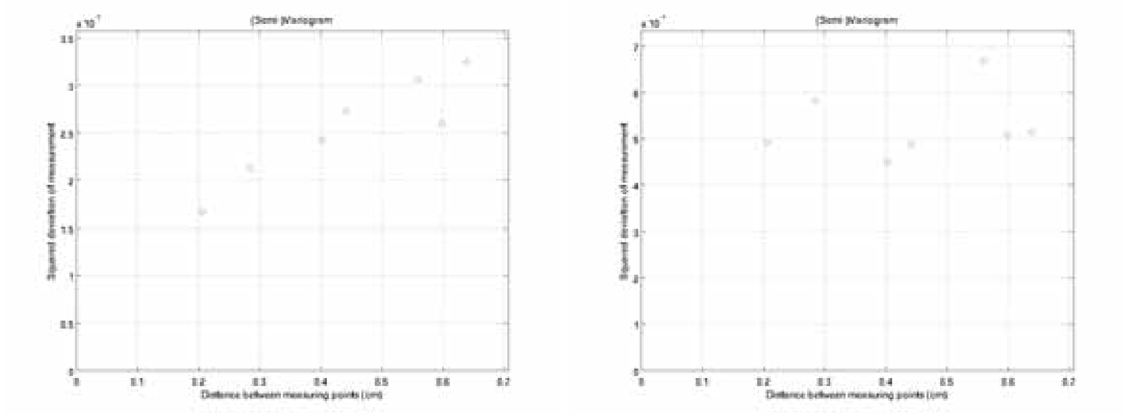

Figure 4.2: Variogram for the results of the 1-by-1 m area in the forest (left) and on the meadow (right)

Longterm measurements Figure 4.3 and 4.4 show the temporal variability of the VWC registered by the TDR placed in a hole on the meadow and in the forest. The VWC stays – independent of meadow or forest – more or less constant over the whole period. The deeper the sensor is located, the less variability it measures. The temperature profile shows a nice daily variability which is smoothed by the (wooden) cover of the hole.

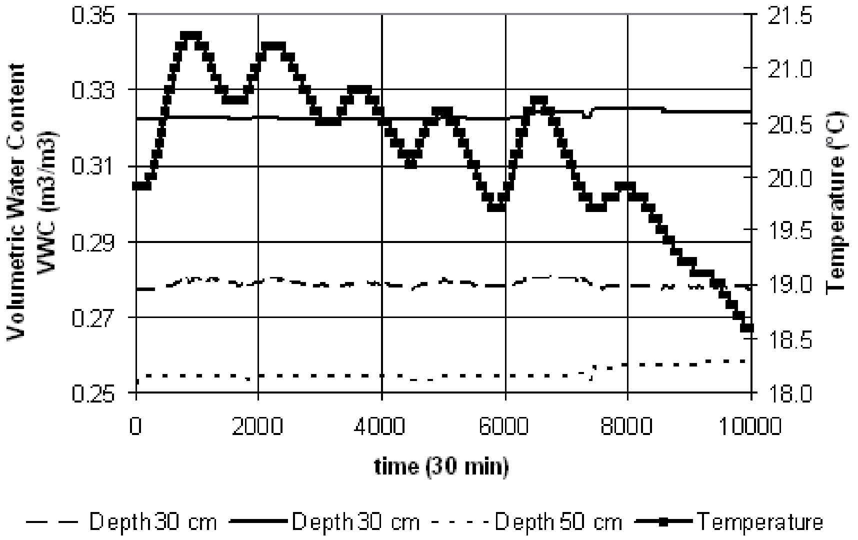

Figure 4.3: Temporal variability of the VWC and temperature on the meadow. Duration 10th–17th June 2010

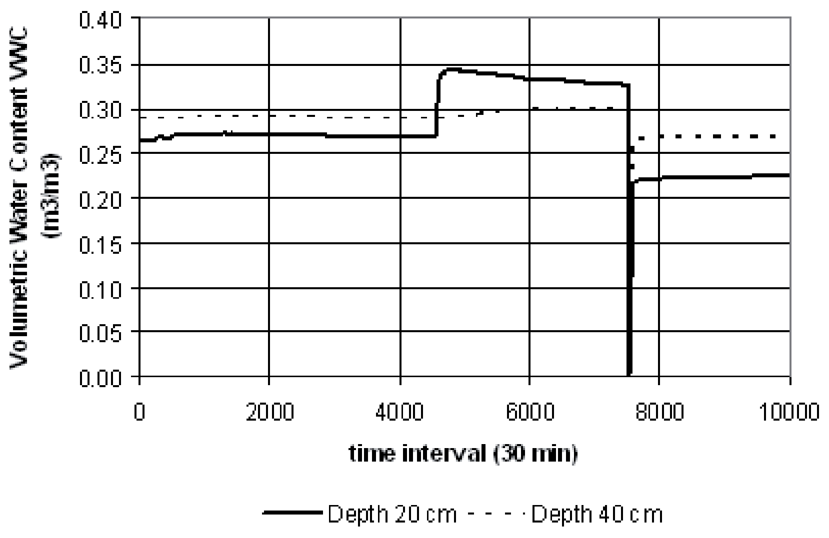

Figure 4.4: Temporal variability of the VWC in the forest. Duration 10th–17th June 2010

4.4 Discussion

As mentioned in the introduction of this report, precipitation values during the field days amounted to 1.3 mm for Monday, 0.7 mm for Tuesday and 6.8 mm for Wednesday. On Thursday there was no rain until departure. The sun did nearly never shine during the four days.

Handheld measurements The values shown in table 4.1 represent the average of measurements taken by three groups with the identical device at slightly different places on the meadow and in the forest. However, there is a negative trend in VWC which is strongest on the bareland. This isn't astonishing due to the fact that the bareland isn't protected of the sun by any leaves (e.g. in the forest) or grass cover (e.g. on the meadow). The total decrease of VWC at all locations might be due to transpiration by plants and trees. Evapotranspiration probably doesn't take much influence because of the fairly bad weather.

The rise in VWC after 16th June 2010 stems from a long ongoing rainfall event with 6.8 mm of rain. It is slower in the forest due to higher interception than on the open meadow. Nevertheless, it is odd that the VWC on the bareland increases just by 0.02 whereas it is more than doubled this value at all the other places. In general, the VWC is higher on the uncut than on the cut meadow. At the investigated site, the VWC of the forest is similar to that on the meadow which isn't what one would expect.

The standard deviations are mainly small meaning that the overall differences in VWC aren't that big and the device works reliably. The standard deviation on 16th June is equal to zero because there is just one measurement.

When looking at the measurements taken in the 1-by-1 square (see tab. 4.2), they show a high heterogeneity across a wide range (factor 2–4). Even within the three different areas, the mean differs by a factor of 1.9. However, the minimum and maximum values are similar.

The variogram for the measurements taken in the forest shows a positive trend meaning that the further the measured points lie apart, the less correlated they are. The variogram of another group for measurements in the forest looks similar eventhough the differences in squared deviations are smaller by a factor 2. The variogram of the meadow has a much smaller trend with even smaller differences in the squared deviation. However, it is important to note that the TDR-sensor returns very different results if the sensor is tuck into the soil completely or – as it is the case in figure 3 – just by half its length. Therefore it is difficult to compare these two.

Longterm measurements The lower the sensor is, the lower the VWC is (see fig. 4.3). The variability is very low (standard deviation of around 1E-03). The two topmost sensors should at least show some variability due to the experienced weather, for instance due to the long ongoing rain on 16th/17th June. In figure 4.4 the upper sensor shows more variability than the lower one. Both top most sensors differ by around 4 % VWC eventhough they are located at the same depth of 30 cm. This difference might stem from calibration.

Figure 4.4 rejects the statement made above. Those soil layers with a higher VWC lie beneath those with a lower VWC. Around time 7500. there are some outliers which might stem from removing or readjusting the sensor. There is no sensible explanation for the turnaround in VWC between time 4500 and 7500.

4.5 Conclusions

In general, the soil moisture shows a high spatial heterogeneity and smaller temporal variability which gets even the smaller, the deeper the sensor is located in the soil. In order to describe temporal variability in VWC, few measurements are enough (e.g. two per day). There are differences in VWC for the forest and the meadow with a trend to a more stable and slightly higher soil moisture content in the forest.

5 Experiment 4 – Pumping test

5.1 Introduction and aims

With the help of a pumping test, the hydraulic conductivity as well as the storage capacity of an aquifer can be determined. A pump is installed in the borehole; through manual and electronic measurements, data of the groundwater depth are collected and then interpreted.

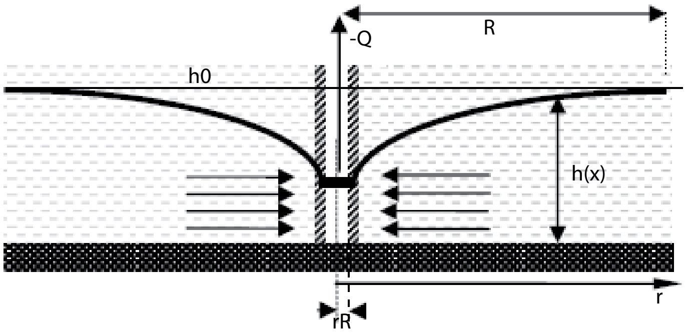

Figure 5.1: Aquifer during a pumping test

In figure 5.1, a stationary confined well in a finite aquifer is shown Kinzelbach and Stauffer 2005. Q is the pumping rate, R the distance between the pump and the measurement, $s(r)$ is the subsidence or recovery, m is the thickness of groundwater body, r_R is the radius of the borehole.

There are several types of pumping tests that can be conducted:

- Instationary, confined
- Quasi-stationary, unconfined
- Instationary, unconfined

The most general approach to analyse pumping tests is the inverse numerical calibration. The method with instationary and confined conditions is widely approved and applied in practice and serves as an approximation for unconfined aquifers. Instationary unconfined pumping tests can be evaluated as to Theiss (logarithmical) or Jacob (semi-logarithmical $s(\log t)$), whereas the latter is used in this experiment. The measured groundwater depth values s are plotted against time t on a semi-logarithmic scale. Then, a line is fitted to the points. After that, transmissivity, storage coefficient and hydraulic conductivity can be calculated according to the following equations (valid for $u < 0.01$):

The transmissivity is T is calculated by equation 5.1

$$T = \frac{-2.3 \cdot Q}{4 \cdot \pi \cdot \Delta s} \qquad [T] = \mathrm{m}^2/\mathrm{s} \tag{5.1}$$

where Q is the pumping rate $(\mathrm{m}^3/\mathrm{s})$ and Δs is the logarithmic cycle[1]

[1] Difference of s between to orders of magnitude.

The storage coefficient S calculated by equation 5.2

$$S = \frac{2.25 \cdot T \cdot t_0}{r^2} \qquad [S] = - \qquad (5.2)$$

where t_0 is the time when the fitted line is at 0 on the y-axis (s), r is the distance between the pumping borehole and the measured borehole (m).

The horizontal hydraulic conductivity k_f is calculated by equation 5.3

$$k_f = \frac{T}{M} \qquad [k_f] = \mathrm{m/s} \qquad (5.3)$$

where M is the thickness of groundwater body (m).

5.2 Description of test

This experiment was carried out on Tuesday, 15th of June 2010 in the morning. Light rain was falling.

The pump is placed in borehole 3.2, the drawdowns at all other shallow wells (X.2) are monitored with hand measurements (water level contact meter). One electrical device is placed in the pumping borehole itself. Immediately after starting or switching off the pump, very frequent measurements should be taken. Borehole 3.2 has a total perforated tube of 9 m depth out of which approximately 1 m is silted. The groundwater level is at about 4 m depth. Two tests are carried out:

1. Subsidence and recovery when pumping without a packer. The pump is put in a depth of 6 m.
2. Subsidence and recovery when pumping with a packer of about 1 m placed on the bottom of the borehole.

In both cases, the pumping rate of $18\,\mathrm{m^3/hr}$ or 5 l/s is based on information on transmissivity tests carried out already before. It is important to choose an appropriate pumping rate because of the danger of falling dry. Furthermore, care should be taken that the pump does not overheat. In this case, a cooling jacket is not needed. No air should get in order to avoid Venturi-effect. Just when the pumping starts, measurements of the groundwater depth level should be taken in a fine time resolution. The recovery of the aquifer can be measured when the pump stopps.

As a second test, a packer is introduced into the borehole. Generally, two packers can isolate specific sections of a borehole. As a result, e.g. the vertical distribution of water quality and hydraulic conductivity in an aquifer can be determined.

The first test without packer took half an hour: 15 minutes with the pump being on and then 15 minutes without pumping to measure recovery. After that, the packer was installed in the borehole. About 25 minutes after the end of the first test, the pump was put on again and the packer test started.

5.3 Results

Hand measurements as well as electronic measurements showed a subsidence with adjacent recovery in both experiments and all boreholes. However, especially the measurements for the test without packer seem to be very irregular. For example, time for subsidence varies between 5 and 515 s without algorithm for e.g. distance. Table 5.1 shows selected hand measurements with total subsidence. Recovery was very difficult to define and is neglected in that table (see fig. 5.2 and fig. 5.3).

Table 5.1: Hand measurements for pumping test in different boreholes, with and without packer.

		Subsidence (cm)		Time needed (s)	
Borehole	Distance to pumping well (m)	Without packer	With packer	Without packer	With packer
2.1	10	2	–	5	–
3.1	0	1.7	–	515	–
4.2	10	6	3	180	30
2.2	10	3	1	30	210

Subsidences are very different in the boreholes, the minimum is 1.7 cm, maximum was measured in borehole 4.2 with 6 cm. This could be explained with geological properties of the aquifer. Normally, subsidence should be higher when using a packer which is not the case here. In contrast, the electrical measurements in the pumping borehole show a higher subsidence for the case with packer (see fig. 5.4).

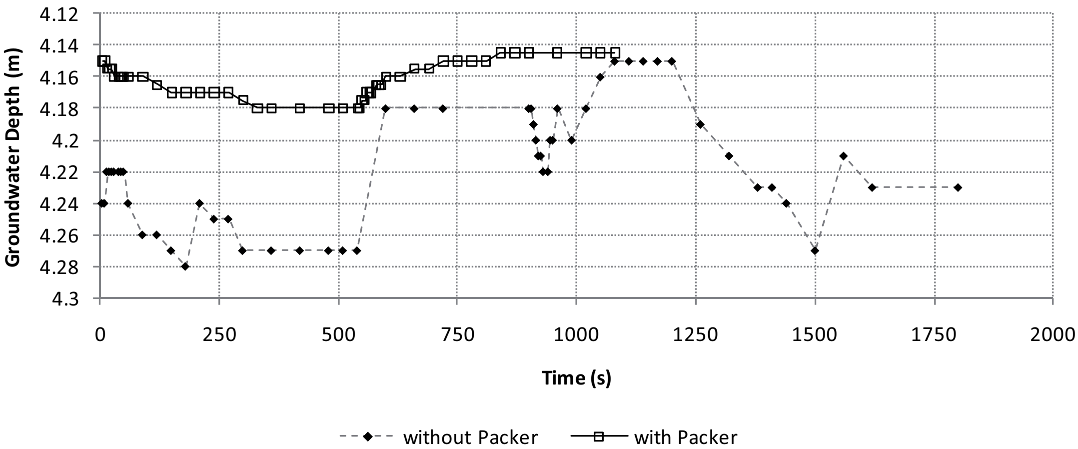

Figure 5.2: Hand measurements during pumping test in borehole 4.2; with and without packer.

It can be seen that recovery even goes higher than the initial value, but decreases again. These irregularities could for example be explained with a changing pumping rate. However, this should then be seen in the other measurements too, which is not the case for example in borehole 2.2. Another interesting thing is that the initial groundwater level of the test with packer is much higher than it was at the end of the test before (without packer).

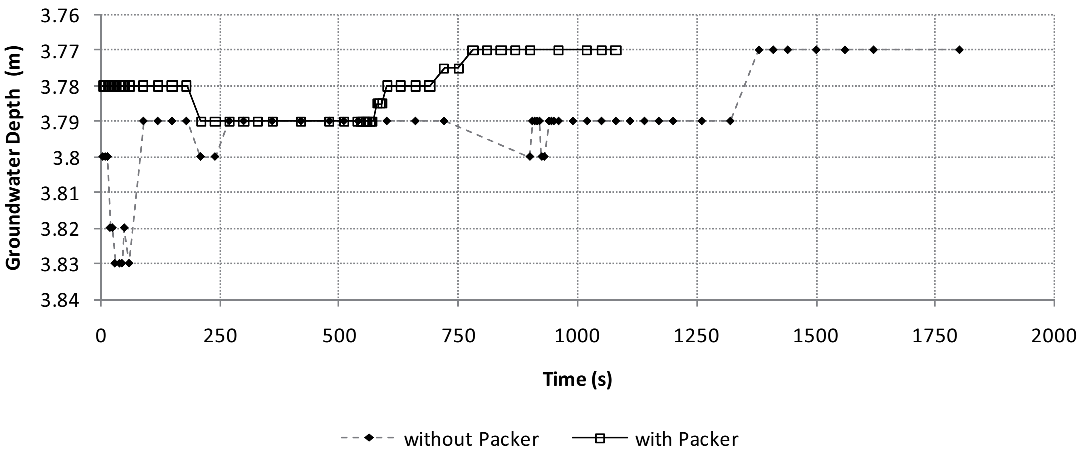

Figure 5.3: Hand measurements during pumping test in borehole 2.2; with and without packer.

In one of the measurements, a rebound effect was observed when switching off the pump. This fact was prevented by simultaneously closing a piston valve in the tube going away from the pump.

The electronic measurements generally support the hand measurements (see fig. 5.4). The groundwater level is higher at the beginning of the second test. Subsidence takes places quite fast. However, these measurements show a bigger influence of the packer: pressure decreases over 0.06 bar in contrast to 0.02 bar in the case without packer.

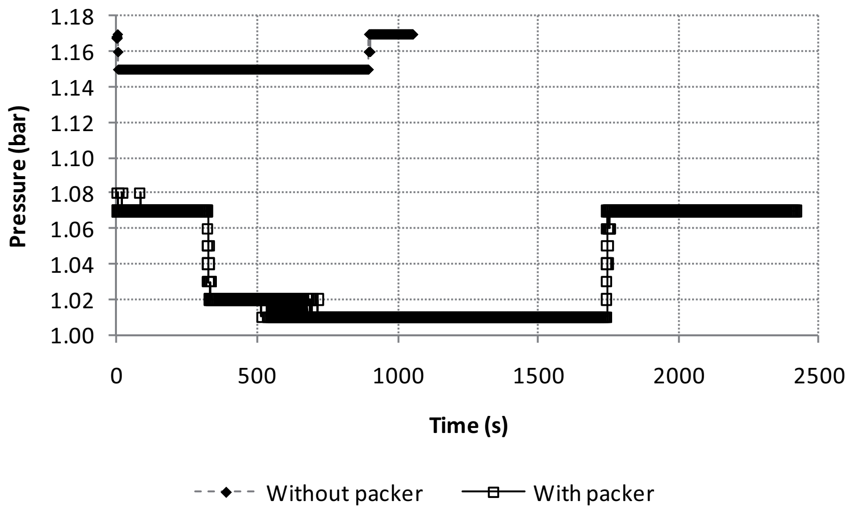

Figure 5.4: Electronic measurements of pumping test in the pumping well (borehole 3.2)

As mentioned before, pumping tests serve for determining hydraulic conductivity and storage capacity of an aquifer. Table 5.2 shows the calculation values and results for the hydraulic conductivity k_f and storage coefficient S.

Table 5.2: Calculation values and results for the pumping test in borehole 4.2

	Without Packer	With Packer
Pumping rate Q $(\mathrm{m}^3/\mathrm{s})$	0.005	0.005
Logarithmic cycle Δs (m)	-0.0325	-0.0166
t_0 (s)	12.84	9.89
Transmissivity T $(\mathrm{m}^2/\mathrm{s})$	0.0229	0.0552
Thickness of groundwater body (m)	10	9
Storage coefficient S (-)	0.00815	0.01229
Horizontal hydraulic conductivity k_f $(\mathrm{m/s})$	2.85E-03	6.13E-03

5.4 Discussion

The values for transmissivity, storage coefficient and hydraulic conductivity are about twice as high for the case with packer than without. The thickness of the groundwater body had to be estimated and cannot be taken as fix value. Hydraulic conductivity corresponds well with literature data for a gravel-sandy aquifer which is geologically the case [2].

With the help of a pumping test, subsidence and recovery of the aquifer as well as the horizontal hydraulic conductivity are explored. Results for the latter correspond well with the local geological situation. However, assumptions had to be made for the thickness of the groundwater body. It is not sure how much the packer reduces this value. For modelling, an instationary, homogeneous and confined aquifer was assumed. These are simplifications, what should be taken into account when looking at the calculated values. As the results show a plausible value for the hydraulic conductivity, the models seem appropriate for this case.

Measurements seem sometimes to be confused and cannot be compared well. Possible sources of errors can be wrong measurement technique, a changing pumping rate or insufficient time management during the tests. A leaking packer would influence all boreholes the same way and should not lead to such different values. A further point is that the boreholes are not perforated at their whole length, except for the pumping boreholes 3.2 itself. Therefore, not the whole theoretically possible flow area can be used by the water.

[2]Source: Kinzelbach and Stauffer 2005

6 Experiment 5 – Flow meter test

6.1 Introduction and aims

The aquifer of a groundwater body is in general inhomogeneous and measurements of aquifer properties are limited. Due to high spatial and vertical variability of hydraulic conductivity K, models of groundwater flow and solute transport show large uncertainties. The pumping test gives information on K for a large aquifer volume, but not for small-scale variability. However, knowledge on the small-scale variability is important for the spread of a contaminant plume.

With the flow meter experiment the small-scale vertical variability of hydraulic conductivity can be characterized in a borehole. To do this two conditions are needed: the natural flow condition and the condition while pumping.

If the mean of the hydraulic conductivity $\bar{K}$ is known, the hydraulic conductivity K of a section i can be determined by the following equation (6.1). The relative hydraulic conductivity η_i of section i is the ratio of the hydraulic conductivity of section i and the mean of the hydraulic conductivity. $\bar{K}$ can be estimated with the help of a pumping test.

$$\eta_i = \frac{K_i}{\bar{K}} = \frac{\Delta Q_i \cdot b}{Q_p \Delta z_i} \tag{6.1}$$

ΔQ_i is the groundwater inflow in the section for pumping condition minus the groundwater flow in the section for natural flow condition (m^3/s), b the screened well length (m), Q_p the total pumping rate and Δz_i the length of the section (m).

The aim of the flow meter test is to determine the vertical profile of the hydraulic conductivity in a certain region of the groundwater aquifer.

6.2 Description of test

A flow through a well can be measured with the Haferland flow meter (see fig. 6.1). It works like a propeller gauge for estimating the flow in a river. Water flow creates rotation of the propeller that sends three impulses per turn to the measurement device. With the numbers of impulses in a specific time and a conversion factor, the flow can be estimated in a given cross section, e.g. in a well.

Figure 6.1: Nose of the Haferland flow meter where the propeller is installed. Foto: René Kaufmann

The measurements are done in borehole 9.1 with a total depth of 16 m but about 0.5 m of slump at the bottom. The diameter of the well is 11.5 cm. The groundwater level lies in a depth of around 4 m; the filter line starts at a depth of 10 m. The pump is installed in a depth of 8 m, 2 m above the filter line.

Three measurements were done in two different configurations. For the first configuration no pumping takes place while the pump is running for the second configuration.

The first measurement is done with a constant vertical velocity downwards from 10 m to 15.5 m depth. The second measurement is the same but upwards from 15.5 m to 10 m. Every 5 cm an automatic measurement is taken. The third measurement is a time measurement where every 3 s a measurement is taken. The flow meter is stopped every 0.5 m to take measurements.

While the flow meter is moving up- or downwards the program of the flow meter control knows the vertical velocity. Due to the vertical speed the flow through the flow meter is higher or lower than in reality. The program automatically corrects the flow with the known speed. This flow as well as other data is saved on the hard disk.

6.3 Results

Figure 6.2 shows the measured flow in borehole 9.1 for depths between 10 m and 15.5 m. For the case the flow meters moves downwards (see fig. 6.2, right) the flow decreases from 0 l/s to around -0.1 l/s. For the upward moving (see fig. 6.2, middle) the flow fluctuates around 0 l/s with larger differences in the bottom. When taking measurements at every meter the flow is 0 l/s for each depth.

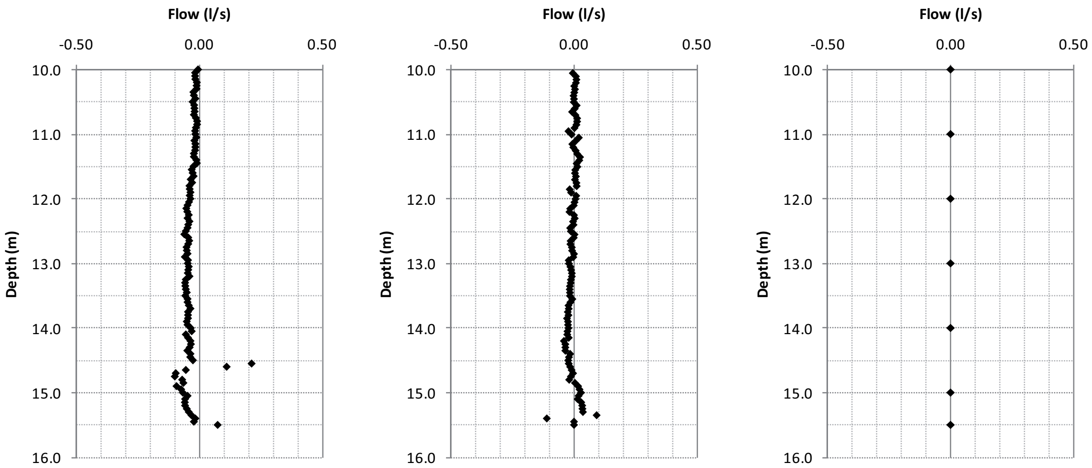

Figure 6.2: Vertical profile of the flow in borehole 9.1 while the pump is not running. *Left:* Flow meter moving downwards. *Middle:* Flow meter moving upwards. *Right:* Flow meter stops for a while every 1.0 m.

Figure 6.3 shows the measured flow in the well while the pump is running. It pumps 0.475 l/s (see tab. B.1 in the appendix). While moving downwards between 10.0 m and 11.5 m depth the flow is maximum and constant with a value of 0.446 l/s. Below 11.5 m the flow decreases continuously. While moving the flow meter upwards the flow increases continuously until a depth of 11.25 m. At this depth a jump in the flow from 0.343 l/s to 0.565 l/s is observed.

For the case the flow meter is stopped every 0.5 m the flow has a constant value between 10.0 m and 11.5 m depth. Below this depth, the flow decreases linearly until a depth of around 14.5 m. At a depth of 15 m the flow has a value of 0 l/s.

Figure 6.4 shows the hydraulic conductivity calculated with equation 6.1. The highest values are at a depth between 11.5 m and 12.0 m and 14.5 m and 15.0 m. Between the depth of 12.0 m and 14.5 m the hydraulic conductivity decreases near to 0 m/s. In the upper part, between 10 m and 11 m the conductivity is very low.

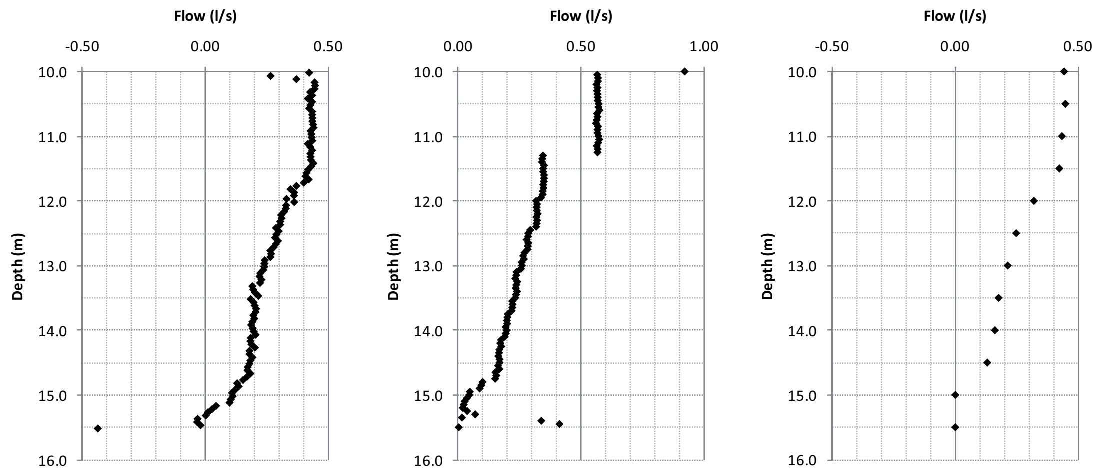

Figure 6.3: Vertical profile of the flow in borehole 9.1 while the pump is running. *Left:* Flow meter moving downwards. *Middle:* Flow meter moving upwards. *Right:* Flow meter stands for a while every 0.5 m.

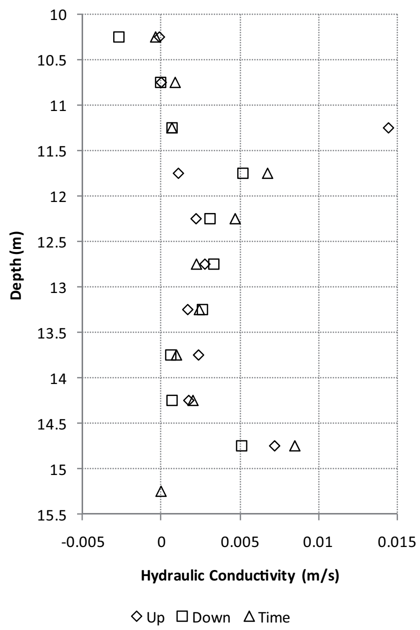

Figure 6.4: Hydraulic conductivity at different depths, calculated for moving up and down the flow meter and measuring every half meter.

6.4 Discussion and conclusions

The flow meter corrects the flow depending on the vertical velocity (see fig. 6.5). While being in the air, the flow meter provides negative values for flow. In water the flow would move the propeller while the flow meter is moving in the vertical direction. The flow meter program corrects this automatically and instantanious. While slowing down or accelerating the movement, the device gives out some errors which can be seen around the depth of 4.5 m. For interpretation these results can't be used.

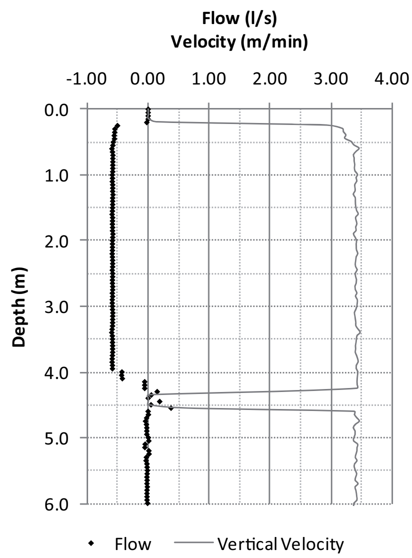

Figure 6.5: Vertical velocity and calculated flow through the device. The water level is at a depth of around 4.5 m. The device was slowed down while going into the water. Below a depth of 4.5 m the flow has a value of 0 l/s.

Logger measurements In the configuration where the pump is not running, the flow should be 0 l/s for all measurements . While moving up or down the flow meter, the fluctuations lie between -0.1 l/s and +0.1 l/s. For the case where the flow meter is moving to the ground, the errors are larger than while moving upwards. Especially in the lowest region between 14.5 m and 15.5 m, the offset is large. While moving the flow meter down, it is stopped at 14.5 m (two very high flow points). This offset is due to starting and stopping the flow meter for the last one meter.

When moving upwards, the measurements only fluctuate around ± 0.05 l/s with expectation in the lowest half meter. Here, the starting process is disturbing the correction of the measurements.

The most accurate measurements can be achieved when the flow meter is stopped on a given interval, in this case every meter.

In the pumping configuration there can be found several interpretations. While the flow meter is moving downward, the flow shows a closed line. There are three break-points where the slope of the line changes. In the upwards movement, these breaking points are visible too. However, there is a large gap at a depth of 11.25 m. This effect is due to the large change of the flow in this part. The propeller in the flow meter has less friction and starts to run with low force. In the case when the flow meter is moving down, the propeller has the possibility to run out. During the upward movement the flow increases very rapidly. The flow in the depth between 10.0 m and 11.25 m is larger than the downward moving of the flow meter. Probably this is due the effect that the flow and the upward moving let to increase the propeller.

When the flow is measured at a specific point, the measurements are more accurate. Out of the data the maximal standard deviation is 0.015 l/s. One can see three break-points, two of them can be seen very good.

The slope of the flow is an indicator for the hydraulic conductivity of a layer. If the change of the flow between two points is small and therefore the slope is steep, the hydraulic conductivity is small. The main part of water will come from the part below this layer. If the slope is flat between two points, the hydraulic conductivity will be high.

In the upper part of the filter line, the borehole has a high permeable layer. It reaches from a depth of 10.0 m to around 11.5 m. The part from 11.5 m to 12.5 m is the highest permeable part in this borehole. The layer below this part is more permeable than the supreme layer, but less permeable than the second layer. A conclusion for the hydraulic conductivity below 14.5 m is difficult. It's possible that the pump does not reach this region.

For further research the part between the depth of 11.0 m and 12.0 m would be interesting because there are two layers of different hydraulic conductivities. In this part high variability of measurements was observed.

Hand measurements The hand measurement of the pumping rate and the comparison with the data shows that the flow meter provides good results. With the hand measurements the pumping rate was determined as 0.475 l/s, for the measurements where the flow meter is stopped every half meter and the flow meter is moving downwards, the results comes to an agreement. Only for the case where the flow meter is moving upward, the result of the flow meter is too high.

For all three measurement cases, the hydraulic conductivity shows similar results. Only for the section between 11.0 m and 11.5 m, the result of the upward moving flow meter has a very large value. This has the same reason as discussed for the flow.

The high conductivity in the lowest part of the well is due the high increase of the flow. At 15 m depth no flow can be found.

Conclusions Over all, the flow meter brings good results. The most accurate results can be reached when the measurements are taken at specific heights for a while. Disadvantage of this method is that it takes a lot of time and money. For a coarse overview of the groundwater layers, the flow measurements can be done with the moving flow meter. In that case it's recommended to take measurements in both directions. For high quality analysis this method has too much uncertainty.

7 Experiment 6 – Dual Pumping Test

7.1 Introduction and aims

By pumping water out of an aquifer, the concentration profile of substances is mixed up. Therefore it isn't possible to distinguish different sections of higher or lower concentrations within the borehole or the aquifer respectively. In order to get a vertical concentration profile, the dual-pumping test can be applied.

In this experiment, the dual-pumping test is applied to get a vertical concentration profile of chloride.

7.2 Description of test

At borehole 6.1 with a total depth of around 14.9 m, two pumps are installed at roughly 14.3 m and 8.5 m deepness. Both pumps are regulated via the frequency of the current. In a first step, just one pump operates at the time in order to derive its pumping rate according to the chosen frequency. The mean of two samples is taken to minimise the measuring error.

Afterwards, both pumps operate at the same time but with different pumping or frequency ratios (see 7.1). This technique induces a stagnation point which makes it possible to get two separate samples at different heights of the well (see fig. 7.1).

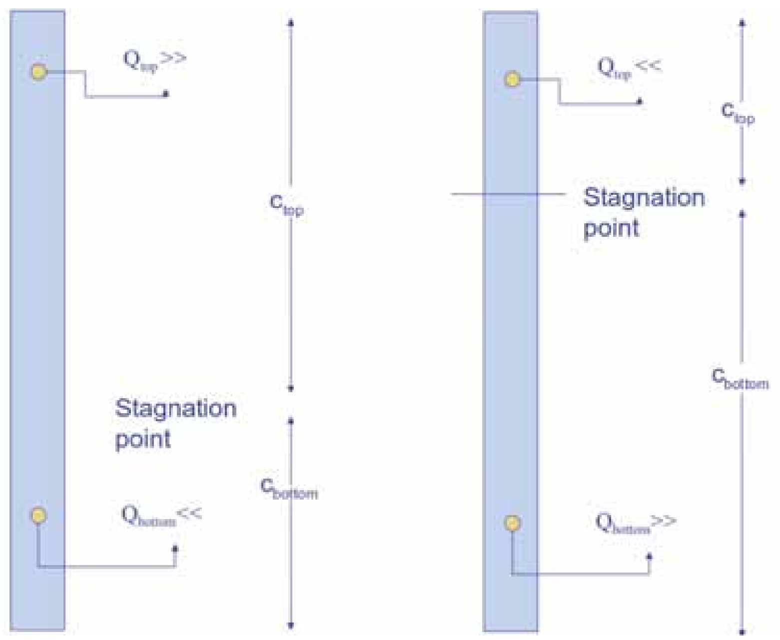

Figure 7.1: Setting for Dual-Pumping Test with two pumps operating at different pumping rates

Table 7.1: Frequencies of bottom pump 1 (fP1) (Hz) and top pump 2 (fP2) (Hz) and corresponding proportion of pumping rate Q1/Q2. Q1 and Q2 are derived experimentally by operating just on pump.

fP1:fP2	Q1/Q2
350:100	0.48/0.09
260:180	0.37:0.26
180:260	0.25:0.41
100:350	0.07/0.53

The vertical concentration profiles are calculated by applying the method illustrated in Thullner et al. 2000 and Rapp et al. 1998.

First, the concentrations derived from the samples are flux weighted as

$$C_{ave}(\gamma_t) = \frac{c_t(\gamma_t) \cdot Q_t + c_b(\gamma_b) \cdot Q_b}{Q_t + Q_b} \tag{7.1}$$

where t being top and b the bottom pump and γ the ratio of Q_i. In order to minimise the influence of scattering of the data, the values are normalised using the mean of the flux weighted concentrations calculated above.

The normalised concentration values are plotted and fitted with an interpolation function using a polynomial 3rd grade (see fig. 7.2).

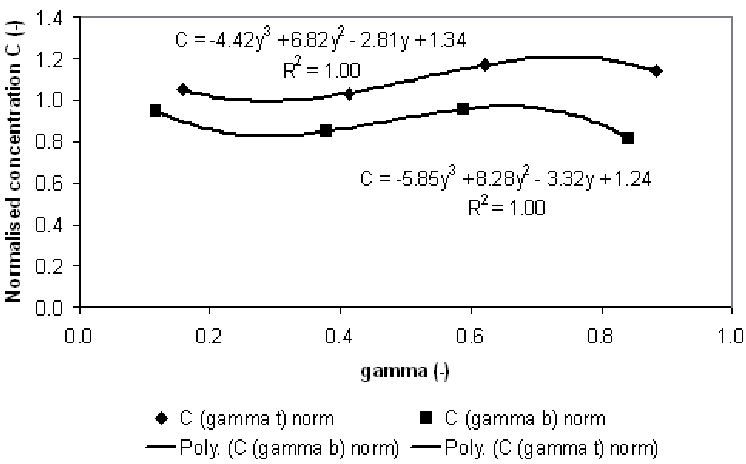

Figure 7.2: Interpolation with a polynomial (3rd grade) of the normalised, flux weighted concentrations

Using the equations below Rapp et al. 1998, the concentrations profiles can be calculated.

$$c_t(z(\gamma_t)) = C(\gamma_t) + \frac{dC(\gamma_t)}{d\gamma_t} \tag{7.2}$$

$$c_b(z(\gamma_b)) = C(\gamma_b) + \frac{dC(\gamma_b)}{d\gamma_b} \tag{7.3}$$

where γ_t and $\gamma_b = \gamma_t$ - 1 being the relative pumping ratios, $c_t(z(\gamma_t))$ and $c_b(z(\gamma_b))$ the top/bottom calculated concentration (mg/l) in depth z. $C(\gamma_t)$ and $C(\gamma_b)$ are the interpolation functions and the derivatives stem from the 1st derivative of the interpolations functions (see fig. 7.2). Depth z equals the depth of the water divide which was calculated for all four ratios of γ as (assumption homogeneity)

$$\begin{array}{ll} \mathrm{top-pump} & z_{\mathrm{top-pump}} + \gamma_t \cdot (z_{\mathrm{top-pump}} - z_{\mathrm{bottom-pump}}) \\ \mathrm{bottom-pump} & z_{\mathrm{bottom-pump}} - \gamma_b \cdot (z_{\mathrm{top-pump}} - z_{\mathrm{bottom-pump}}) \end{array} \tag{7.4}$$

Equation (7.2) returns two profiles showing the same profile but once derived from the bottom and once from the top pump. These two vertical profiles are interpolated as

$$c^{int}(z(\gamma_t)) = c_t(z(\gamma_t)) \cdot (1 - \gamma_t) + c_b(z(\gamma_t)) \cdot \gamma_t \tag{7.5}$$

7.3 Results

The following figure 7.3 shows the vertical concentration profiles of the chloride derived from the bottom pump (left), the top pump (right) and their interpolation (see eq. (7.2)).

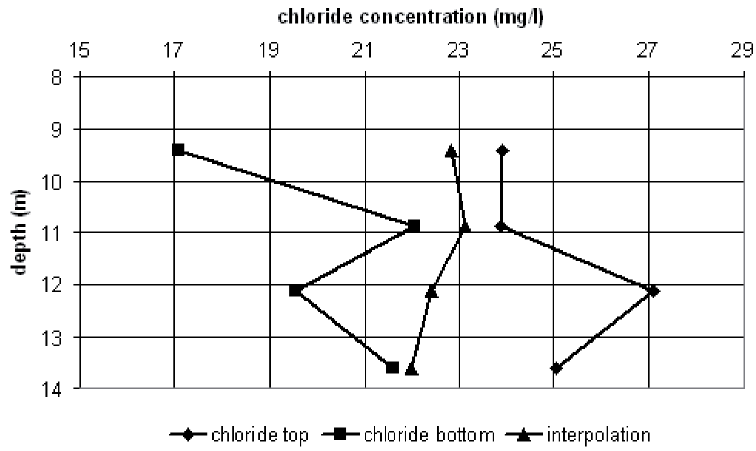

Figure 7.3: Vertical concentration profile of chloride

7.4 Discussion and conclusions

According to Thullner et al. 2000 and Rapp et al. 1998, the resulting vertical concentration profiles should coincide and almost overlap. However, the retrieved ones show major differences and they don't look equally at all. As a result, it is difficult to say much about the vertical distribution of the chloride concentration in borehole 6.1. The interpolated values are more or less constant over the depth and they lie mostly between 22–23 mg/l which is similar to the measured concentrations. Rapp gives as an explanation for large deviations between the two profiles that measurment errors or unsuitable interpolation schemes might be a reason. Nevertheless, the applied interpolation function (see fig. 7.2) fits the data very well. A polynom of 2nd grade didn't fit the data at all and was therefore not used for further calculations. In addition, the field work is thought to be done as accurately as possible.

The dual pumping technique is an easy way to obtain a vertical concentration profile. However, measurement errors and difficulties in interpolation and analysis might lead to odd provfiles with large deviations resulting in a difficult interpretation.

8 Experiment 7 – Dilution test

8.1 Introduction and aims

With this experiment, the Darcy velocity and thus the hydraulic conductivity can easily be determined. A tracer, in this case salt, is inserted into the borehole while pumping the water and re-injecting again, creating a water circle. The tracer is diluted, what can be measured with a KLL-Q device with the electrical conductivity.

The following equation 8.1 can then be put into a graph in order to get the Darcy velocity v_d:

$$\ln \frac{C}{C_0} = v_d \cdot \frac{-2 \cdot \alpha \cdot t}{\pi \cdot r_i} \tag{8.1}$$

In equation 8.1 is C the concentration, C_0 is the highest concentration, v_D the Darcy velocity (m/s), α a factor accounting for change in the flow field (-), t is the time since injection (s), and r is the radius of the borehole (m).

8.2 Description of test

The experiment was conducted on the 16th of June, 9:30-11am. No rain had occurred in the hours before.

Salted water is injected in a borehole while pumping at the bottom of the well. The water is re-injected into the borehole which makes the net extraction zero and the borehole content well mixed. The salted water will be diluted due to diffusion, dispersion and advection. Salt concentration is measured by hand as well as with a data logger. At the beginning, a fine time resolution of measurements should be chosen. The data logger "Diver" is programmed for measuring a time interval of 5 s. The Measurement of the "Diver" is limited for concentrations between $0\,\mu\mathrm{S}$ and $5\,\mu\mathrm{S}$.

The experiment is made in borehole 5.2 with a groundwater level of 3.5 m and a depth of 7.6 m. The pump is installed at 7 m depth, the data logger at 5.5 m. The pumped water is filled in at a depth of 4 m. Pumping frequency is 350 Hz.

100.1 g NaCl are mixed with 1001.5 g water pumped from a borehole. This mixture is then injected into the borehole while already pumping. Care should be taken that no salt stays at the borders of the borehole. Measurements are taken for half an hour, then the experiment is stopped because recovery to the initial state will need much more time.

8.3 Results

In figure 8.1 the measurements shows very high values in the first two minutes, going up to 17.54 mS/cm. From time 01:50 to 02:00, a very sharp decline takes place (from 17.54 mS/cm down to an electrical conductivity of 4.76 mS/cm). It is probable that the measurement device got a direct inflow of the test solution. For the remaining measurement time, values oscillate around 4.5 mS/cm.

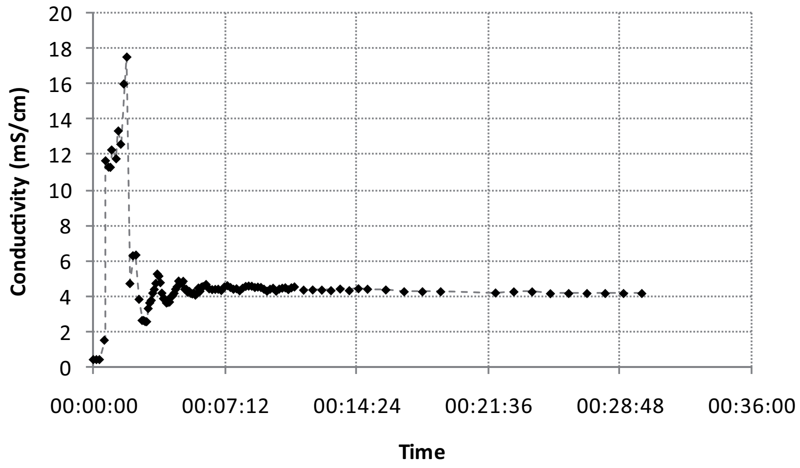

Figure 8.1: Dilution test in borehole 5.2 on 16th of June 2010; Hand measurements

For programming default reasons we did not get any measurement data from the electrical device, leading us to take data from another group. There, 100 g NaCl in 1000.3 g pumped water were injected in borehole 7.2. Measurements were taken first all 5 s, later in a 10 s interval during 43 min (see fig. 8.2).

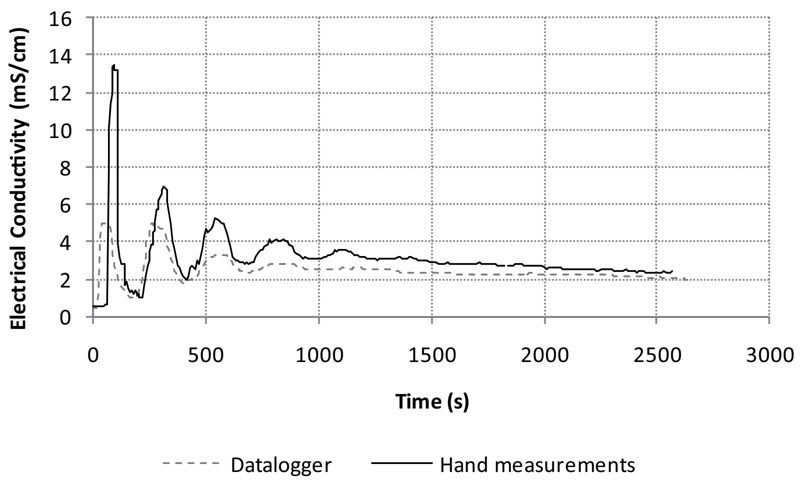

Figure 8.2: Dilution test at borehole 7.2 on 16th of June 2010; Hand measurements and data logger measurements.

In this case again, a very quick peak in electrical conductivity is reached after 01:30 with 13.46 mS/cm, remaining there for 20 s. Then, a very sharp decrease goes down to about 3 mS/cm. For the first 1000 s, oscillations take place. Probably there was not reached a homogeneous dispersion of the salt at the beginning. After that, a more constant degradation occurs. Measurements were stopped after 43 min at an electrical conductivity of about 2.4 mS/cm. The electronic data logger device cannot measure values above 5 mS/cm which makes the peaks much more flat than with the control measurements by hand. Behalf of that, the values match together very well.

Electrical conductivity and concentration act linearly in certain ranges. For NaCl, this is the case in the range of about 0 mol/l to 2 mol/l [1].

The initial concentration of the salt is 1.7 mol/l (100 g NaCl in 1000 g water), so that electrical conductivity can be taken as substituent to concentration. The highest measured value is taken as C_0 in equation (8.1). With α being 1 and the radius of the well 0.115 m, figure 8.3 can be drawn and the Darcy velocity read out

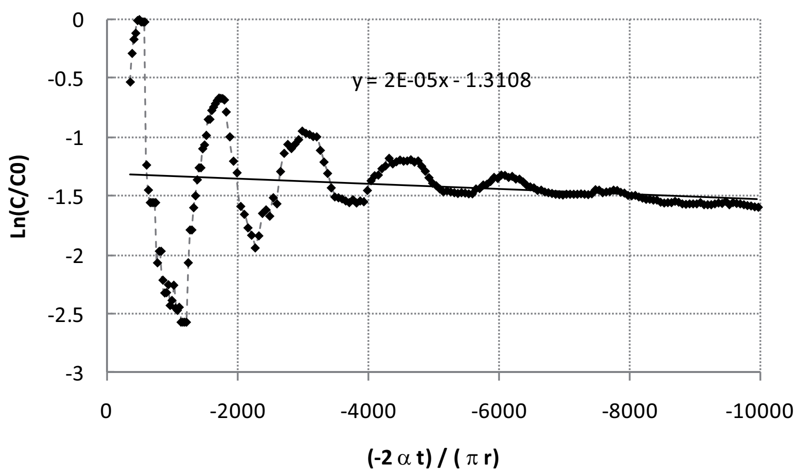

Figure 8.3: Semi-log plot of the measured electrical conductivities and a second term in order to determine v_D.

8.4 Discussion and conclusions

The needed hydraulic gradient is calculated between borehole 1.2 and 7.2 and has an amount of 0.000964 m/s. Therefore, the hydraulic conductivity k_f amounts to 0.0207 m/s. This value is seven times higher than the value determined in the pumping test without packer ($k_f = 0.002848\,\mathrm{m/s}$) and 3.4 times higher than in the test with packer ($k_f = 0.0061\,\mathrm{m/s}$). From a geological point of view, this can be possible and would indicate a gravel layer near borehole 7.2. Geological investigations showed that there is gravel in the test field. It should be kept in mind that the tube is perforated only in 4 out of 8 m depth. Therefore, the data are not valid for the whole borehole depth.

The salt should be distributed very homogeneously in the borehole; nothing should remain at the frame of the tube. This possible error source can be avoided by a careful handling of the salty solution.

Because of a programming error, the data logger could not be used for interpretation. It would be important to check each time if all devices are working well before starting a test.

This experiment is very easy and can help to determine local hydraulic conductivities.

[1] Source: *Die spezifische Leitfähigkeit* 2010

9 Experiment 8 – Tracer test

9.1 Introduction

The aim of the tracer test is to evaluate the mobile porosity and the dispersivity in the aquifer. Tracer tests are ideally performed within natural conditions because they show the natural gradient. The disadvantage of this method is that it takes a lot of time and money, especially in low conductive aquifers. To speed up tracer dispersion one can force the gradient with a pump. The advantage of pumping is that more tracer substance is recovered but the gradient is changed. Under natural gradient condition, the 1D solution of the concentration curve has the bell shape as it is in equation (9.1)

$$c(x,t) = \frac{M/(n \cdot A)}{\sigma \cdot \sqrt{2 \cdot \pi}} \exp\left[\frac{-(x-u\cdot t)^2}{2\cdot\sigma^2}\right] \tag{9.1}$$

where $\sigma = \sqrt{2 \cdot D \cdot L \cdot t}$, M is the mass injected and A is the cross-sectional area.

To check the quality of a tracer test, the recovered mass of the tracer is an indicator. At the pumping well the recovered mass is always smaller than the injected mass due to adsorption, losses at the injection well, incomplete tracer recovery at the pumping well and other losses. Out of this, the initial mass of the tracer has to be modified for interpretation.

In a 2D transport model the results can be better interpreted. The desired parameter like porosity and dispersivity can be estimated directly.

Table 9.1 shows the parameters for the PMWIN flow model.

Table 9.1: Parameters for the PMWIN flow model

Parameter	Dimension	Value
Mesh Size	L	100 × 100 Grid
Cell Size	L	1 × 1
Layer Type	-	Unconfined
Transmissivity	L^2/T	Calculated
Boundary Condition -		Fixed Head
	L	upper head: -3.789
	L	lower head: -3.897
Top of layer	L	0
Bottom of layer	L	-8
Horizontal Hydraulic Condutivity	L/T	2.81873E-03
Effective Porosity	L^3/L^3	0.4
Initial Concentration	M/L^3	1.15E-5
Dispersivity	L	17

9.2 Description of test

1000 g of naphthenate are given into a tank with 25 l of water from ETH Hönggerberg in Zurich. This solution is given into a tank with a volume of 1000 l. This solution is the tracer which will be recovered at borehole 3.2 where the pump is installed with a pumping rate of about $18\,\mathrm{m}^3/\mathrm{hr}$. The tracer was injected in borehole 1.2.

After injecting the tracer in the borehole, a sample is taken every hour out of the pumping well. These samples are analyzed in the laboratory with a fluorescence meter.

9.3 Results

A sample of 10 ml out of the 1000 l tank is taken and diluted with 1 l of fresh local water. With this sample and the known concentration, the online sensor can be calibrated. The online sensor is logging only the measurements in mV, so the results have to be converted from mV to ppb. The tracer sample out of the 1000 l tank gives a value of 638.15 mV which is equal to 1107.25 ppb.

Due to a tracer test conducted one week before, a contamination with uranine can be detected in the samples. For this the calibration line for uranine is also needed. In figure 9.1 the calibration line of uranine and naphthenate is shown. With this line the conversion from counts of the fluorescence meter to concentration can be made.

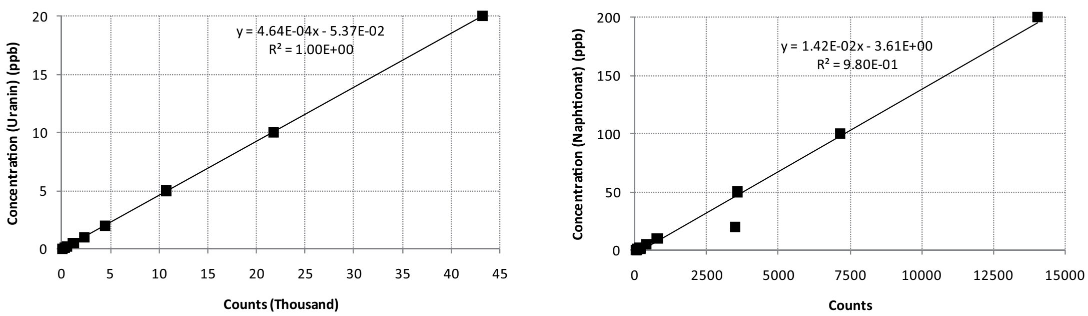

Figure 9.1: Calibration line for uranine (left) and naphthenate (right) for conversion from counts of the fluorescence meter to concentration.

Figure 9.2 shows the measured and corrected concentration of naphthenate. The correction has to be made because the detector of naphthenate detect uranine, too. The aim of this correction is to bring the non zero values of naphthenate to zero before the tracer test was started.

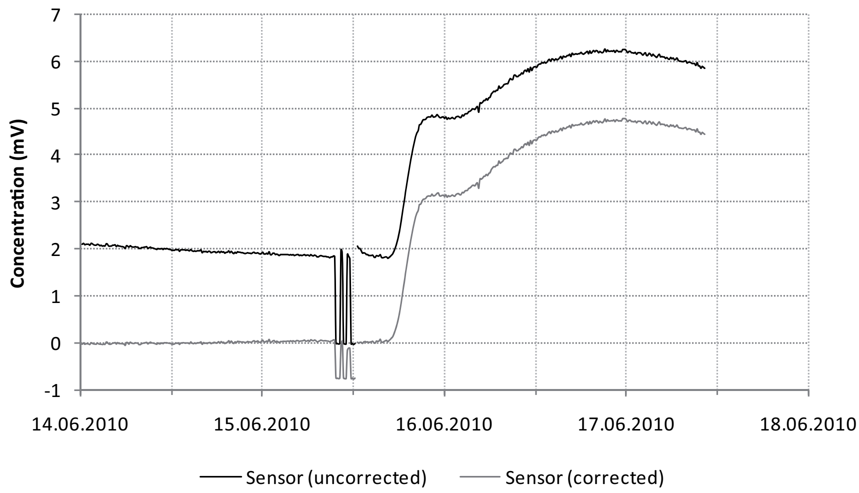

Figure 9.2: Measured (black line) and corrected (gray line) concentration of naphthenate in mV

Figure 9.3 shows the concentration of naphthenate measured with the sensor in the pumping well every 10 minutes as well as samples taken by hand every hour and analyzed in the laboratory. The sensor data has been calibrated with the initial value where 638.25 mV is equal to 1107.3 ppb and multiplied with 56.21. This correction factor is needed because the sensor has no linear dependency for very high concentration.

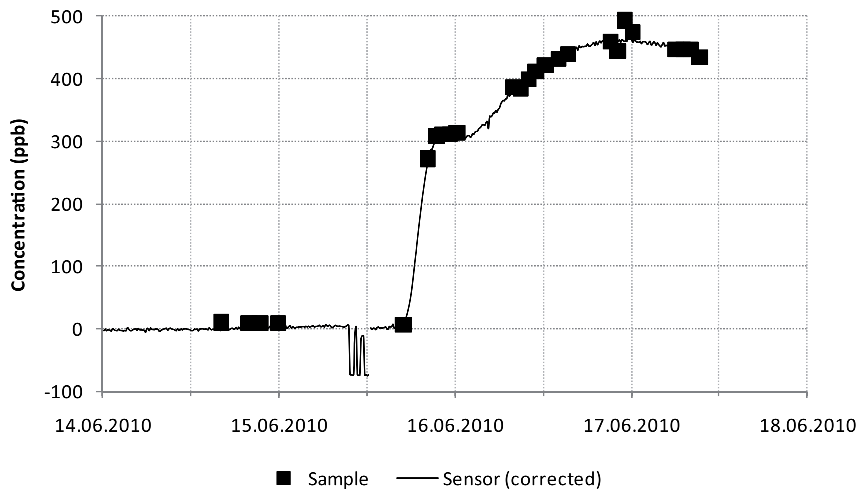

Figure 9.3: Comparison of concentration of naphthenate measured with sensor and samples analyzed in the laboratory.

Figure 9.4 shows the flow field while the pump is running. A large area is influenced by the pump. The contamination plume has its maximum at the injection well. In the direction of the pumping well the plume has its largest expansion. Due to diffusion, the plume has also an expansion to the side.

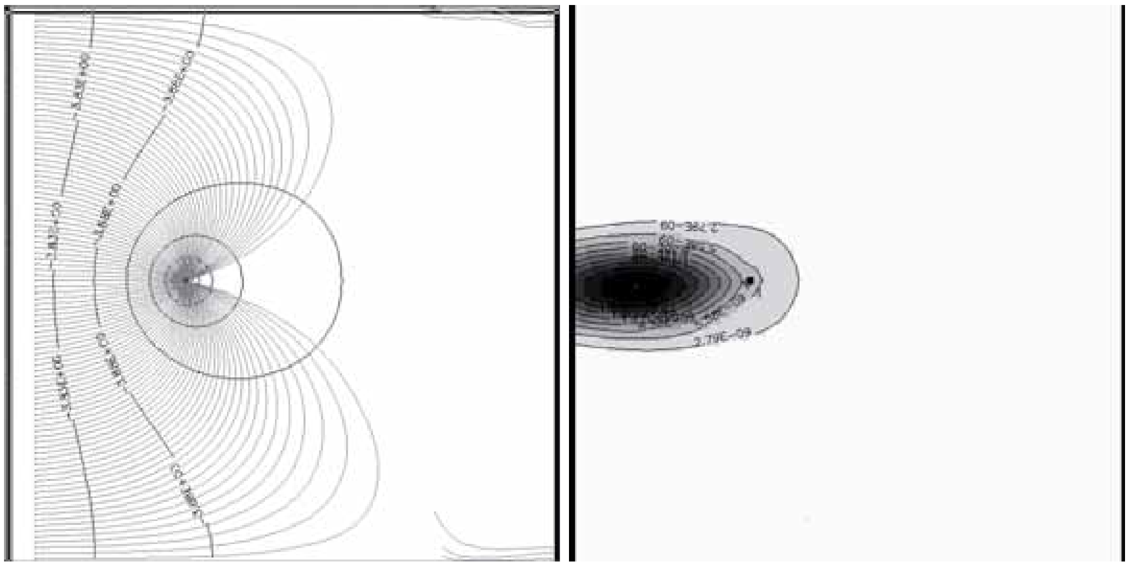

Figure 9.4: *Left:* Flow field while the pump is running. *Right:* Plume of the contamination after 4.5 hours. The plume reached the well.

With the parameters used out of table B.2 (see appendix), the comparison of the measured sensor data and modeled data can be seen in figure 9.5. In the flow model the increasing slope is too flat and the peak is too early. The sensor data shows a high increase, then a stagnation and again an increase of the tracer concentration. The peak of the modeled and measured data has the same height.

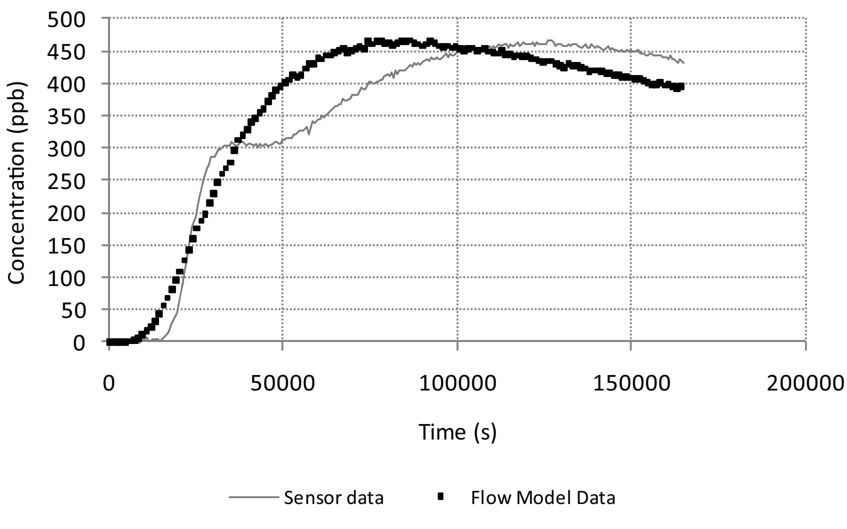

Figure 9.5: *Left:* Flow field while the pump is running. *Right:* Plume of the contamination after 4.5 hours. The plume reached the well.

Figure 9.6 shows the tracer concentration in all boreholes two days after the injection of the tracer. The lowest concentration in the upper aquifer is found near the pump. Along the line from borehole 1.1 to 9.1 the concentration are smaller than in the borehole on the side. In the lower aquifer the highest concentration can be found at the injection borehole. Low concentration can be found on the left side in flow direction. On the right side the concentration are higher.

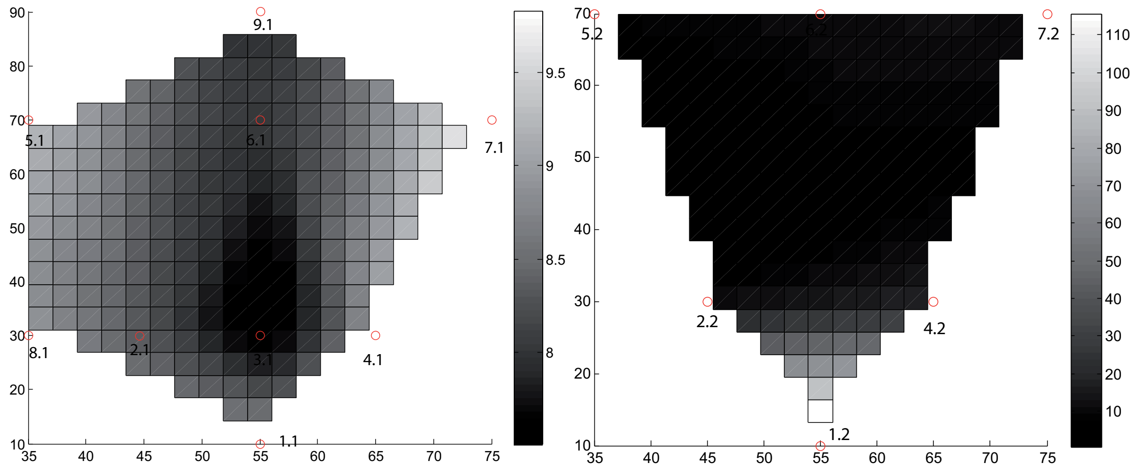

Figure 9.6: Distribution of tracer concentration (ppb) in the boreholes for the higher (right) and lower (left) part of the aquifer.

9.4 Discussion and conclusion

After some conversion, the curve of the sensor measurement and the points of the analyzed samples seem to look good. One of the problem is the calibration. In the field, the sensor has to be calibrated. This can be done by putting a known tracer with known concentration into the device and taking then a sample. This procedure has the disadvantage that for high tracer concentrations, the linearity of the conversion factor from mV to ppb is not guaranteed. Because of this, two conversions have been made. Figure D.1 in the appendix shows this difference of the sample and the sensor measured data where the sensor is calibrated only with the high tracer concentration. The difference reaches two orders of magnitude.

For the PMWIN flow model three parameters have been changed in order to get a similar curve of the measurements. With the effective porosity the time when the peak of the concentration should reach the well can be adjusted. With the longitudinal dispersivity, the width of the curve can be changed. The third parameter is the initial concentration of the tracer which is set to 0.0065 $[\mathrm{kg/m^3}]$. The real initial concentration is $1\,\mathrm{kg/m^3}$.

The recovery of the tracer after 150'000 s is around 0.06 kg, or 6%. This is a small recovery rate but for two days a good one. Out of the curve it is not clear how long it takes until all of the tracer is pumped out.

The low concentration near the pumping well can be caused due to pumping. Most of the water flow will go through the pumping well. The higher concentration at the outer boreholes can be explained by advection and diffusion of the tracer.

The low concentration at the left side of the flow field could be due to low hydraulic conductivity. The flow field in this lower part of the aquifer seems not to be the same as the flow field in the upper aquifer.

The tracer experiment is a very time-consuming experiment. Every hour a sample has to be taken. In the laboratory further analysis and calibrations take a lot of time. This test has also many error sources like false tracer concentration, dilution errors, wrong handling while injecting the tracer, calibration of the sensor, errors in the laboratory etc. For this experiment it is a huge advantage if enough time and – therefore money is available.

10 Discussion and Conclusion

During four days, groundwater field work was done at the Kappelen site. With the help of several tests, aquifer properties were investigated. Hydraulic conductivity, flow velocities, recharge and interactions as well as vertical profiles were of special interest. As consequence, there were made experiments reaching the whole test field on the one hand; on the other hand, single boreholes were investigated.

In general, the experiments worked well and measurements could be taken. No big problems occurred using measurement devices such as battery problems, obvious wrong measurements or software problems. Evaluations showed that hand measurements correspond well with electronic measurements. Still, hand measurements should never be neglected even if the electronic devices can be trusted - if not, one does not have any data any more in the case of an unforeseen event such as vandalism ...

The aquifer of the investigated test field consists mainly of sandy gravel which could be approved with the pumping test, dilution test and flow meter test. Variability of the hydraulic conductivity is not too high over the field; values remain in the range for sandy gravel (10-3 to [10-1] m). The dilution test gives a k-value being up to seven times higher than pumping test values which is justifiable with local irregularities and measurement accuracy. The flow meter test gave insight into small-scal variability of the hydraulic conductivity, what functioned well. Recharge calculations showed that recharge alone cannot be responsible for groundwater level fluctuations. It is probable that there take place infiltration or exfiltration processes from or to the nearby river "Alte Aare". However, this could not be proven with vertical profiles of electrical conductivity and temperature.

It can be said that a good overview on the test field could be gained with the experiments at hand. Still, it has to be said that it will not be that easy in other cases, mainly due to two reasons

1. The test field is very small, homogeneous and well-prepared for tests. This will not be the case when investigating an aquifer for an engineering office.
2. Measurement devices were provided and are well-maintained. In practice, problems can occur and cause expensive delays when tests have to be redone.

Working on the field and conducting the experiments often takes more time than one thinks. In general, all necessary notes were taken in order to carry out proper evaluation. However, it would sometimes have been easier if data sheets would have been prepared by the groups with a heading to fill out every time. Important information such as weather, time, pumping rates and depth etc. will then be gathered in any case. Especially when doing experiments in big groups, this could help to make sure all information are written down the same way. But even if all necessary information are noted, it is often the case that measurement errors or weaknesses are only detected when trying to analyse the results.

Bibliography

Allen, Richard G. (1998). *Crop evapotranspiration: guidelines for computing crop water requirements*. Rome: FAO.

Die spezifische Leitfähigkeit (2010). Universität Ulm. URL: `http://www.pctheory.uni-ulm.de/didactics/elekleit/html/index042.html`.

Kinzelbach, Wolfgang and Fritz Stauffer (2005). *Vorlesungsskript Grundwasser I*. ETH Zürich.

Rapp, M. C. et al. (1998). "The dual pumping technique (DPT) for level-determined sampling in fully screened groundwater wells". In: *Journal of Hydrology* 207, pp. 220–235.

Thullner, M. et al. (2000). "Validation of the dual pumping technique for level-determined groundwater sampling in a contaminated aquifer". In: *Journal of Hydrology* 235, pp. 104–116.

A Experiment 1 – Vertical Profile

Table A.1: Depth of measured and corrected groundwater level (GWL), measured on 14th June, 2010.

Borehole	x-Coord (m)	y-Coord (m)	Measured GWL (m)	Correction (m)	Adjusted GWL (m)
1.1	55	10	3.80	0.0000	3.800
2.1	45	30	3.83	-0.0130	3.843
3.1	55	30	3.98	0.1465	3.834
4.1	65	30	4.18	0.3600	3.820
5.1	35	70	3.19	-0.6743	3.864
6.1	55	70	3.88	0.0188	3.861
7.1	75	70	3.77	-0.0785	3.849
8.1	35	30	3.95	0.1080	3.842
9.1	55	90	4.00	0.1137	3.886
1.2	55	10	4.05	0.2585	3.792
2.2	45	30	3.53	-0.0325	3.563
3.2	55	30	Pumping station	0.2450	-
4.2	65	30	4.21	0.3460	3.864
5.2	35	70	3.50	-0.3703	3.870
6.2	55	70	3.90	0.0348	3.865
7.2	75	70	3.96	0.1195	3.841

Table A.2: Depth of measured and corrected groundwater level (GWL), measured on 7th June, 2010.

Borehole	Measured GWL (m)	Adjusted GWL (m)
1.1	3.81	3.81
2.1	3.83	3.83
3.1	4.05	3.95
4.1	4.20	4.82
5.1	3.21	3.42
6.1	3.90	3.90
7.1	3.79	3.79
8.1	3.98	3.98
9.1	4.03	4.03
1.2	4.06	4.16
2.2	3.82	3.71
3.2	4.09	5.01
4.2	4.19	4.42
5.2	3.51	3.51
6.2	3.90	4.04
7.2	3.98	3.98

Table A.3: Borehole properties

Borehole	Water lever (m)	Borehole Depth Sump (m)	Filter line (m)	Pressure sensor	Sensor Installation Depth (m)
K1.1	3.65	13.00	8	102	12.3
K2.1	3.68	15.65	10	104	12.3
K3.1	3.82	14.48	10	DIVER	12.3
K4.1	4.02	14.56	10	3	12.3
K5.1	3.2	14.68	10	4/Baro	12.3
K6.1	3.72	14.84	10	106	12.3
K7.1	3.82	14.34	10	5	12.3
K8.1	3.8	14.93	10	2	12.3
K9.1	3.89	16.24	10	108	12.3
K1.2	3.91	7.9	4	101	7.3
K2.2	3.64	8.22	4	103	7.3
K3.2	3.92	8.37	4	105	8.1
K4.2	4.01	4.89	?	107	4.8
K5.2	3.32	7.6	4	109	7.3
K6.2	3.73	8	4	111	7.8
K7.2	3.8	8.32	4	113	7.8

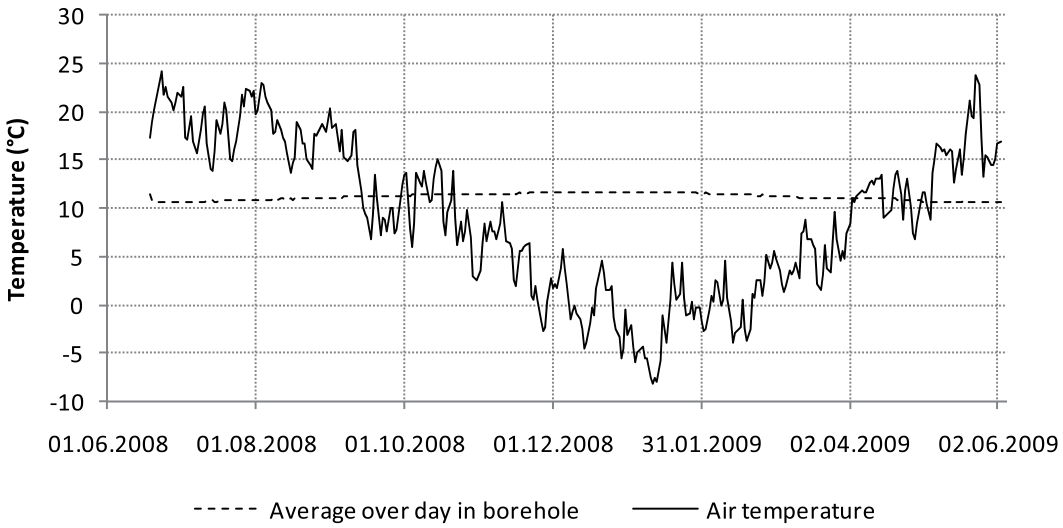

Figure A.1: Comparison of groundwater temperature and air temperature over a year. Source: Electronic measurement device in borehole 4.1 from 18.6.2008 to 3.6.2009.

B Experiment 5 – Flow meter test

Table B.1: Pumping rate in borehole 9.1 for the flow meter test

Frequency (Hz)	Volume (l)	Time (s)	Flow (l/s)
350	5	10.6	0.47
350	5	10.43	0.48
Mean			0.475

Table B.2: Parameter for flow meter test

Parameter	Value	Unit
Mean hydraulic conductivity $\bar{K}$	0.00282	m/s
Screened well length b	5.5	m
Length of section Δz_i	0.5	m
Pumping rate Q_p	0.475	l/s

Table B.3: Flow and hydraulic conductivity at different depths

Depth	Flow (m^3/s)			Hydraulic Conductivity ($cx\mathrm{m/s}$)		
(m)	Up	Down	Time	Up	Down	Time
10.25	-0.0014	-0.0413	-0.0053	-0.0001	-0.0027	-0.0003
10.75	0.0004	-0.0010	0.0138	0.0000	-0.0001	0.0009
11.25	0.2217	0.0108	0.0106	0.0145	0.0007	0.0007
11.75	0.0171	0.0798	0.1040	0.0011	0.0052	0.0068
12.25	0.0343	0.0478	0.0724	0.0022	0.0031	0.0047
12.75	0.0429	0.0512	0.0346	0.0028	0.0033	0.0023
13.25	0.0262	0.0407	0.0374	0.0017	0.0027	0.0024
13.75	0.0367	0.0097	0.0151	0.0024	0.0006	0.0010
14.25	0.0272	0.0110	0.0312	0.0018	0.0007	0.0020
14.75	0.1108	0.0779	0.1306	0.0072	0.0051	0.0085
15.25	-0.2421	0.1883	0.0000			0.0000

C Experiment 6 – Dual Pumping Test

Table C.1: Vertically averaged concentrations of chloride from top and bottom pump analyzed in the lab.

Qb/Qt -	top pump mg/l	bottom pump mg/l
0.07/0.53	24.06	18.64
0.25/0.41	23.65	21.86
0.37/0.26	26.77	19.39
0.48/0.09	26.03	21.78

D Experiment 8 – Tracer Test

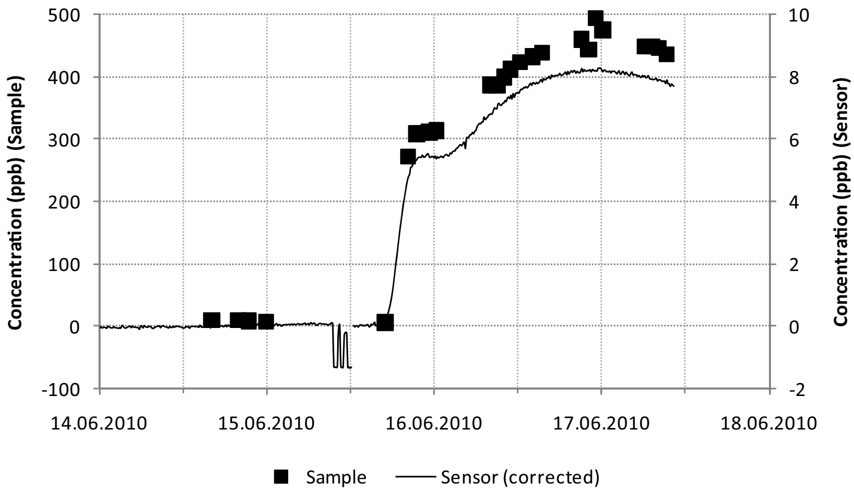

Figure D.1: Comparison of sensor data with sample data. Sensor data is calibrated on initial value of 638.15 mV = 1107.3 ppb