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Investigation of Solute Transport in the Porous Aquifer of the Test Site Wilerwald (Switzerland)

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Content

1. Introduction (Ch. LEIBUNDGUT)

The problem of pollutant transport has become increasingly important in recent years, adding a new dimension to groundwater investigations. The interest is focused on how contaminated water moves in various media. Representative parameters are necessary to reproduce and simulate groundwater flow and solute transport. Tracer technology, in combination with solute transport models, is a valuable tool in this area of research.

Previous studies showed that it is not sufficient to solve the problem of solute transport in natural aquifers by laboratory experiment (I. STOBER, 1988). Due to different scales, the transfer of results from the laboratory to the site is difficult and often impossible. Therefore, a natural experimental study area in a porous aquifer was created. After evaluating nine different possibilities, the "Test Site Wilerwald" was found to be most suitable and was established by Ch. LEIBUNDGUT et al. (1988). Due to the aquifer properties known at that time, the assumption of a reliable homogenity of the aquifer was legitimated. Under the condition of a well known "test site" it seams to be possible to investigate the tracer properties in porous aquifers on the one hand and to calibrate the models of solute transport on the other hand.

In order to get the necessary information concerning the distribution of the hydraulic conductivity of the aquifer in the test site initial tracer tests and geophysical investigation were carried out. The tracer tests indicated, that the assumption of a "quite homogenous" structure of the aquifer in the test site was too optimistic. As well the horizontal and vertical distribution of the hydraulic conductivity is very high. In order to get a more detailed knowledge of the test site and the flow mechanism and to improve the interpretation of the tracer tests results, additional tracer tests and geophysical investigations were carried out. Therefore, this paper deals mainly with this prospective investigations of the test site and the modelling by evaluation the tracer tests. Whereas, the paper is not dealing with the tracer methodological aspects and the behaviour of tracers in a porous aquifer.

The investigations have been carried out in collaboration with several institutions of ATH.

2. Test Site (Ch. LEIBUNDGUT)

Establishing a suitable test site in a porous aquifer is difficult. In order to get an optimal test field for the purposes of tracer experiments including the modelling some important boundary conditions must be considered for the evaluation such as non confined groundwater, relation of width to length > 1 : 3, depth of aquifer < 15 m, stable flow conditions, depth of groundwater table < 5 m (because of technical problems with pumping in case of > 5 m) and good accessibility. The test site in the Wilerwald close to Berne in the Swiss Central Plateau fullfilled the above stated criteria best (Fig. 2.1).

Twenty-one observation wells and an injection well were installed in three main half circulated galleries. The groundwater surface shows a general flow direction towards NNW. According to the isohypses a convergent flow may be assumed (Fig. 2.1). The wells of a size of 1 1/2" diameter were conceived in order to enable water sampling at three levels in 5 m, 8 m and 11 m depth (Fig. 2.2). The reason for

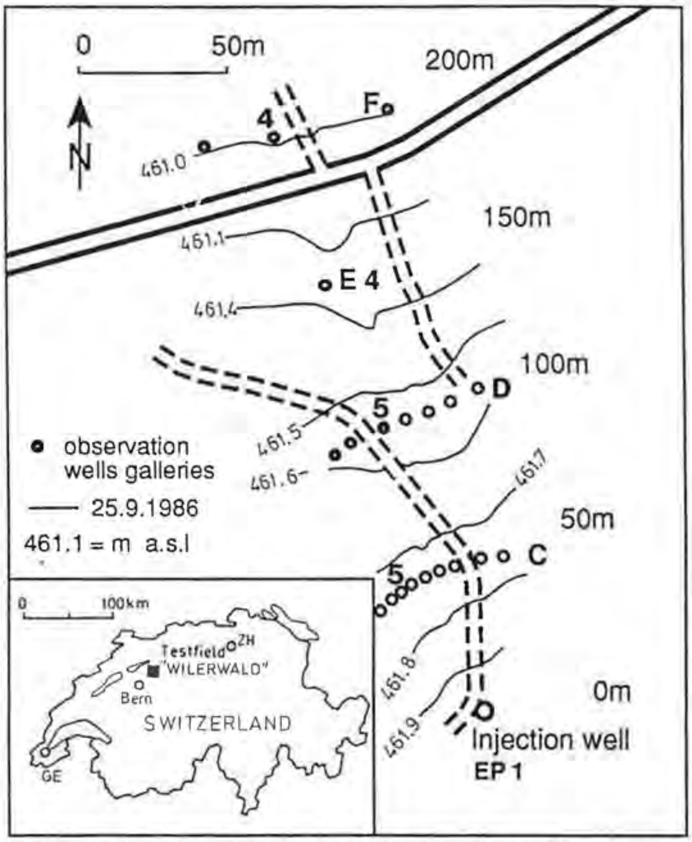

Fig. 2.1: The "Test Field Wilerwald" in Switzerland with the location of wells and the groundwater isohypses indicating the main direction of flow in the aquifer.

that arrangement was the variance of the hydraulic conductivity in the three levels known from probings during installation of observation wells.

The aquifer of the large Emme valley near Wilerwald consists of postglacial Holocene sandy gravels with a considerable variance of thickness from a few meters to 88 meters. The geological composition of the aquifer varies between 40 and 60% quarzites, 30 and 50% Flysch and calcareous sandstones and 5 and 15% crystalline rocks. The depths of groundwater table varies between 0,5 and 3 m. This formation is intercalated between alluvial sediments of low permeability and lacustrine sediments which contitutes the aquitard and which overlay a very irregular Miocaen's surface of molassic marls (H. LEDERMANN, 1978). The sedimentological analyses of the drill core (M within fig. 2.1) revealed the local existence of intercalations of lentiformed sandy and silty levels which lead to a strongly heterogenous distribution of the permeabilities. Although these Quaternary sediments are known to be quite heterogenous, it could be expected a relatively homogenous aquifer in the test field a quite regular thickness of approximatily 15 m and a hydraulic gradient uniformly decreasing towards the N due to its small dimensions.

The chemical components indicate a quite hard water. However, no problems must be expected concerning an interaction between water and tracer used.

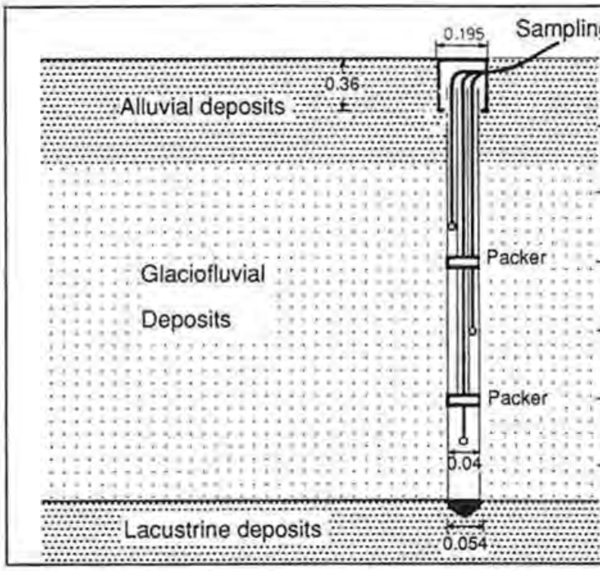

Fig. 2.2: Schematic cross section of aquifer and observation well showing three-level sampling arrangement. The lacustrine deposits act as aquitard.

3. Tracer Tests (Ch. Leibundgut, A. De Carvalho Dill)

A preliminary test was done in May 1986 in order to prove assumptions made regarding ground water flow direction and flow velocity (s. chap. 2.). 0.3 kg uranine and 6 kg naphtionate diluted in 70 l of water have been injected simultaneously in the well EP1. The main transport direction was considered to cross between C4 and the bulk tracer transport at C5. Considerable differences in the concentration distribution of each well and at different depths indicated a horizontal and vertical heterogenity of the aquifer.

In the main tracer experiment, (September 1986) 1 kg uranine and 20 kg naphtionate dissolved in 90 l of water have been injected into each of the assumed layer of the aquifer. The mixed tracer solution was injected simultaneously in the sampling levels of 5 m, 8 m and 11 m (assumption of layers) over a period of one hour (single-shot injection).

The observation wells were sampled for a period of nearly 4,000 hours. Out of the wells included in this test the following levels were sampled with automatic sampling devices: C5 (5 m), C5 (11 m), D5 (5 m), D5 (11 m), and E4 (8 m) every four hours, C5 (8 m) and D5 (8 m) every hour. E4 (5 m), E4 (11 m) and all layers of well F4 were sampled manually at irregular intervals.

Figure 3.1 shows the naphtionate concentration-time curves for each of the three layers in the four wells selected for the reinterpretation study (s. chap. 5.). The uranine curves are not presented here because they display the same characteristics. As can be seen, the hypothesis of constant thickness and permeability for the three layers postulated causing a layer-specific flow and transport behaviour cannot be maintained. In the well C5 the concentration-time curves of the 8 m- and 11 m-levels are almost identical only the 5 m-level shows a distinct difference. In the well D5 those of the 5 m- and 8 m- levels are similar and in the well E4 all layers produced even the same concentration-time curve. However, in the well F4 can be made a distinction among all three levels.

The maximal concentrations occured in the direction of C4–C5–C6–D5–D6 and further towards E4 (Fig. 2.1). No significant maximum were measured in the F-gallery. Probably the piezometers are not situated in the principal direction of flow which was situated between C4 and C5 and between D4 and D5. It is obvious that the expectation of the main flow direction derived from the water table (s. fig. 2.1) is not equal with the results found in the tracer tests. A remarkable delay of tracer transport in the longitudinal direction showed the D-gallerie with the occurence of relatively high concentrations even after 30 days. This variable pattern does not indicate a strict division of the aquifer in three layers of constant thickness but rather the existence of pockets and channels of varying characteristics.

In consequence of the results of the first serie of tracer tests the injections of the second serie were performed only in the 8 m-level in the middle of the aquifer between 6 and 10 m. Diluted in 40 l of water and 20 l of ethylenglycol the tracers uranine (0.5 kg) sulforhodamine B (9.9 kg) and lithium chloride (25 kg) were injected in November 1988 with an injection rate of 1.2 l per minute. 60 l of water were used for rinsing (J. MÄGDEFESSEL, 1990). All wells were sampled at the 8 m-level. The

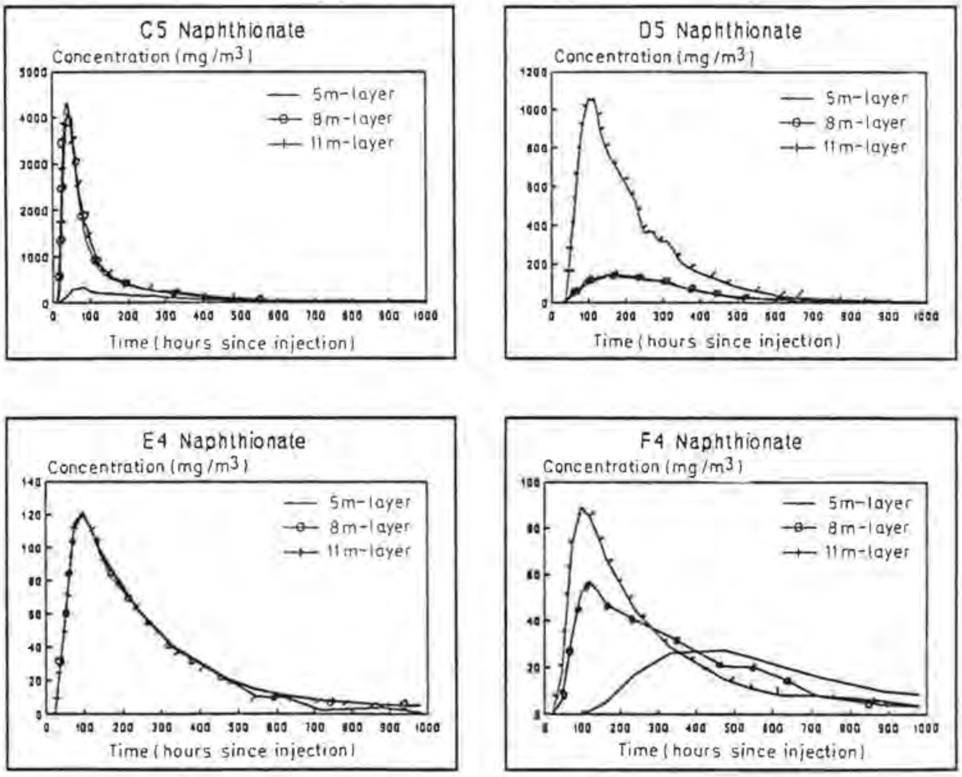

Fig. 3.1: Napthionate concentration-time curves at observation wells C5, D5, E4 and F4 in the different levels of sampling (assumed layers).

wells C6, D2, D4 and E4 were additionally sampled in order to control at 4 and 11 m.

The results were quite different from those ones of the first serie. There was also a vertical difference in the concentrations measured at different dephts but without any tendency to one of the levels. The quickest arrival of all tracers happened in C7, by which all tracer concentrations peaks were obtained still on the day of the injection. The highest concentrations appeared in C4, C6, D6, and D7. In the F-gallery F6 received ten times more tracer than F4.

Comparing the concentration peaks of all tracers in the different wells one can say that the difference in their occurence are on average small in the C-gallery. The appeareance of the peaks oscillate between one hour and two hours in C7, till 14 days and 26 days in C8. The differences increased with distance, being extreme in the eastern part of the field. In D2 e.g. it shows a delay of 24 days for sulforhodamine B but for lithium chloride at least 117 days. In contrary this difference is minimal in D7 and D6 which also presented the highest tracer concentration. Lithium chloride was not always present in the samples. It did not appear in F4, D1 and D5 at all. However, in areas with higher permeability (e.g. C6, D6, D7) the concentration-time curves of lithium chloride were even shorter than those ones of sulphorhodamine B. It must be assumed that the relatively small amount of tracer and sorptions effects are responsible for this astonishing feature. But this methodological aspect is not treated in the contribution on hand.

4. Geophysical Prospecting (A. DE CARVALHO DILL, I. MÜLLER)

In order to get a better understanding of tracer experiments it was necessary to obtain a realistic representation of the permeability distribution at the Wilerwald test field. Very Low Frequency-Resistivity (VLF-R) and refraction seismic surveys were performed for detailed resistivity and transmissivity mapping and for the evaluation of strata thickness. Locations of the electro-magnetic multifrequency VLF-R soundings and the seismic lines are presented in fig. 4.1. The electro-magnetic VLF-R device allows to measure the apparent resistivity Rho_a at different depths between two electrodes spaced 5 m from each other on the surface, as well as that of the dephasing of the electric component of the signal in relation to the magnetic component (Φ) (G. FISCHER et al., 1987).

The apparent resistivity (Rho_a) of the strata between the two points serving to measure Ex can be obtained, according to the following relation:

$$Rho_a = \left(\frac{Ex}{Hy}\right)^2 \cdot \frac{1}{2\pi F \mu_0}, \tag{4.1}$$

where

Rho_a is given in ohm meters,
Ex is given in volt/meters,
Hy is given in ampere/meters,
F being the emitter's frequency is given in Herz (Hz) (emitters available between 13 and 250 Kilohertz),
μ_0 being the electro-magnetic permeability of the vacuum is given in henry/meters.

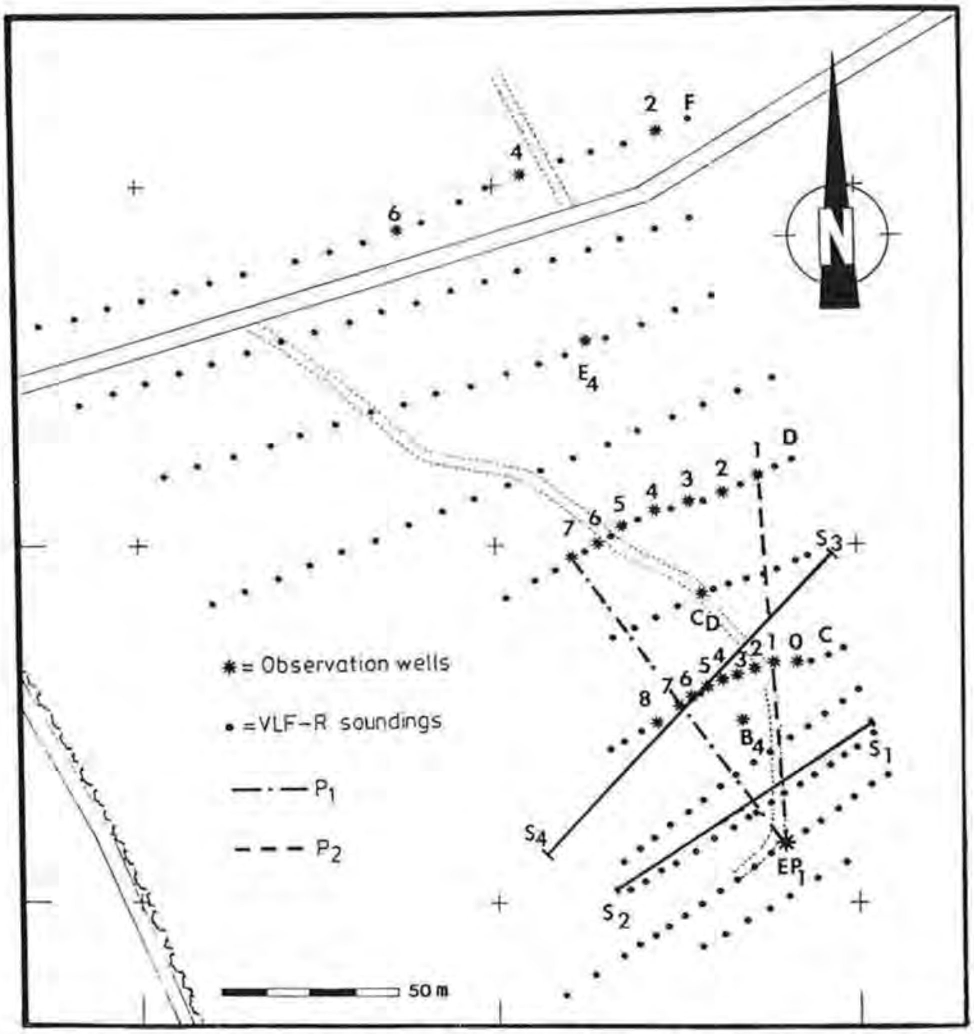

Fig. 4.1: Location of the geophysical vertical VLF-R soundings and the seismic spreads within the frame of observation wells. The profiles P1 and P2 related to VLF-R measuring points connect three of the piezometers with the chosen breakthrough curves (comp. fig. 4.3).

The depth investigation is given within the 12 to 240 KHz frequency by the relation stated below:

$$D = 503 \sqrt{Rho_a/F}\,, \tag{4.2}$$

where Rho_a is the apparent resistivity in Ω m (measured) and D the investigation depth in meters. Thanks to the multifrequency device, we can "diagnose" the rock at different depths and compute layer thickness.

Therefore the VLF technique is a reliable tool in order to investigate the degree of variability of the quarternary formation. To be able to map it, the Wilerwald test field area was divided into 11 profiles (Fig. 4.1). In each one the measuring points were separated by 5–10 m. Three frequencies were used: 183 kHz, 60 kHz and

19 kHz. This simultaneous use enables the calculation of actual resistivities and of the lokal thickness of the aquifer with the help of a computer "best fit" program.

The use of the 183 kHz apparent resistivity data allowed the construction of a Countour Map (Fig. 4.2). Due to the almost linear function of resistivity and permeability the map gives already a good picture of the structure of the aquifer. As the greater resistivity values are the more permeable ones (Tab. 4.1), one can visualize the existence of a true "channel" in a "meander"-like structure. A quick view on the map permits to localize the several piezometers regarding the structure. The C7 one e.g. is situated in the channel zone, D1 and F2 are in a less permeable region, which circles the NE-area of the test field. The injection point lies in a less permeable zone.

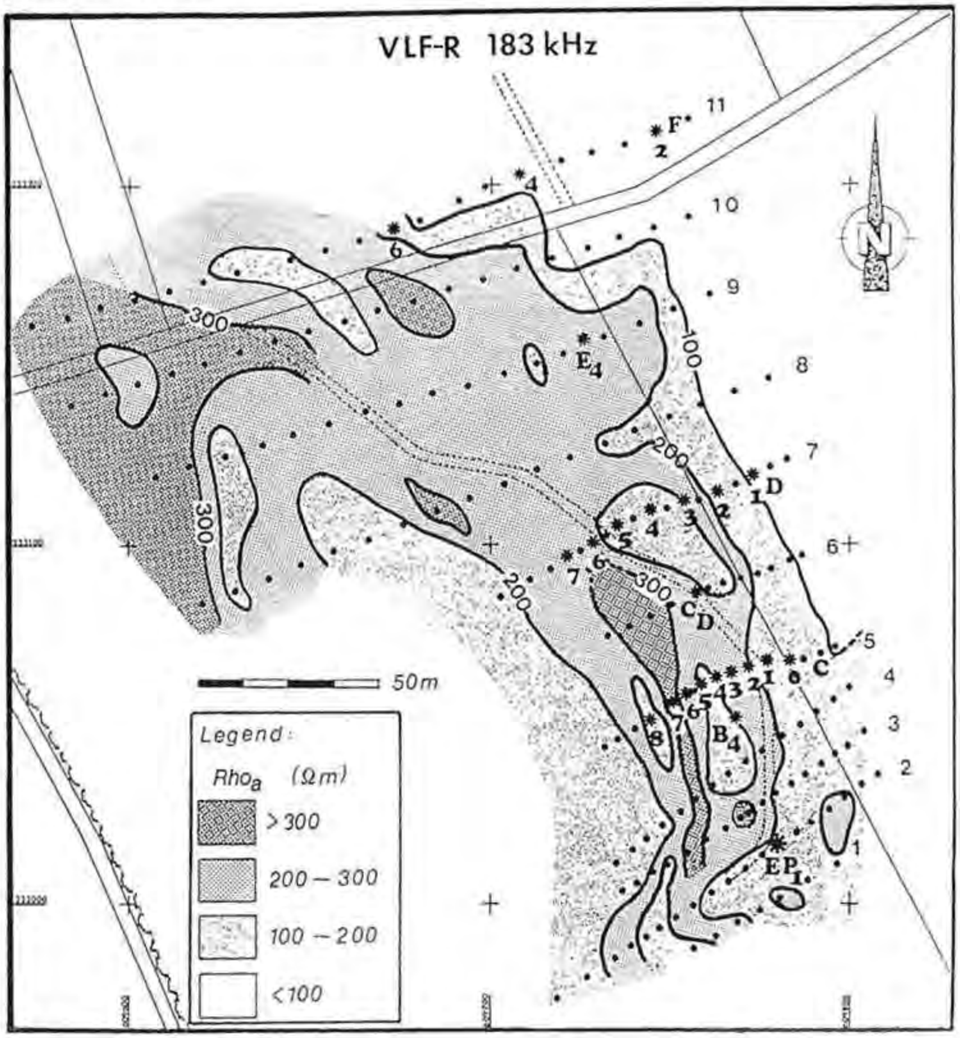

Fig. 4.2: Apparent Resistivity Contour Map obtained with 183 kHz, clearly indicating the higher permeable channel (Rho_a > 300 ohm), and its disposition relative to the net of piezometer (A. DE CARVALHO DILL & I. MÜLLER). For signatures s. fig. 4.1.

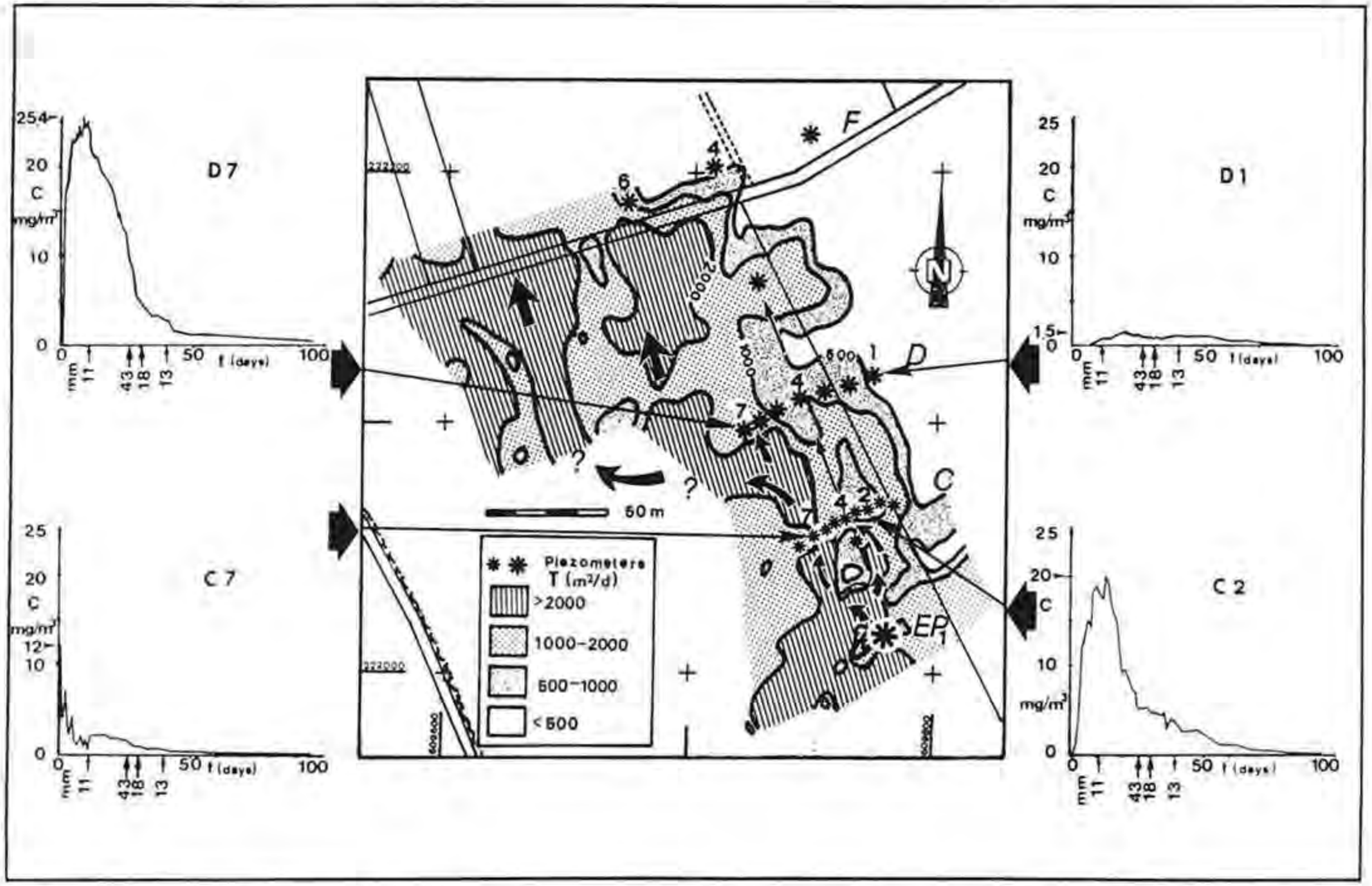

Fig. 4.3: Transmissivity Map showing the main direction of tracer transport (fat arrows) in relation with four chosen breakthrough curves (uranine tracer test 1988/89) and the main direction of water flow (A. DE CARVALHO DILL & I. MÜLLER).

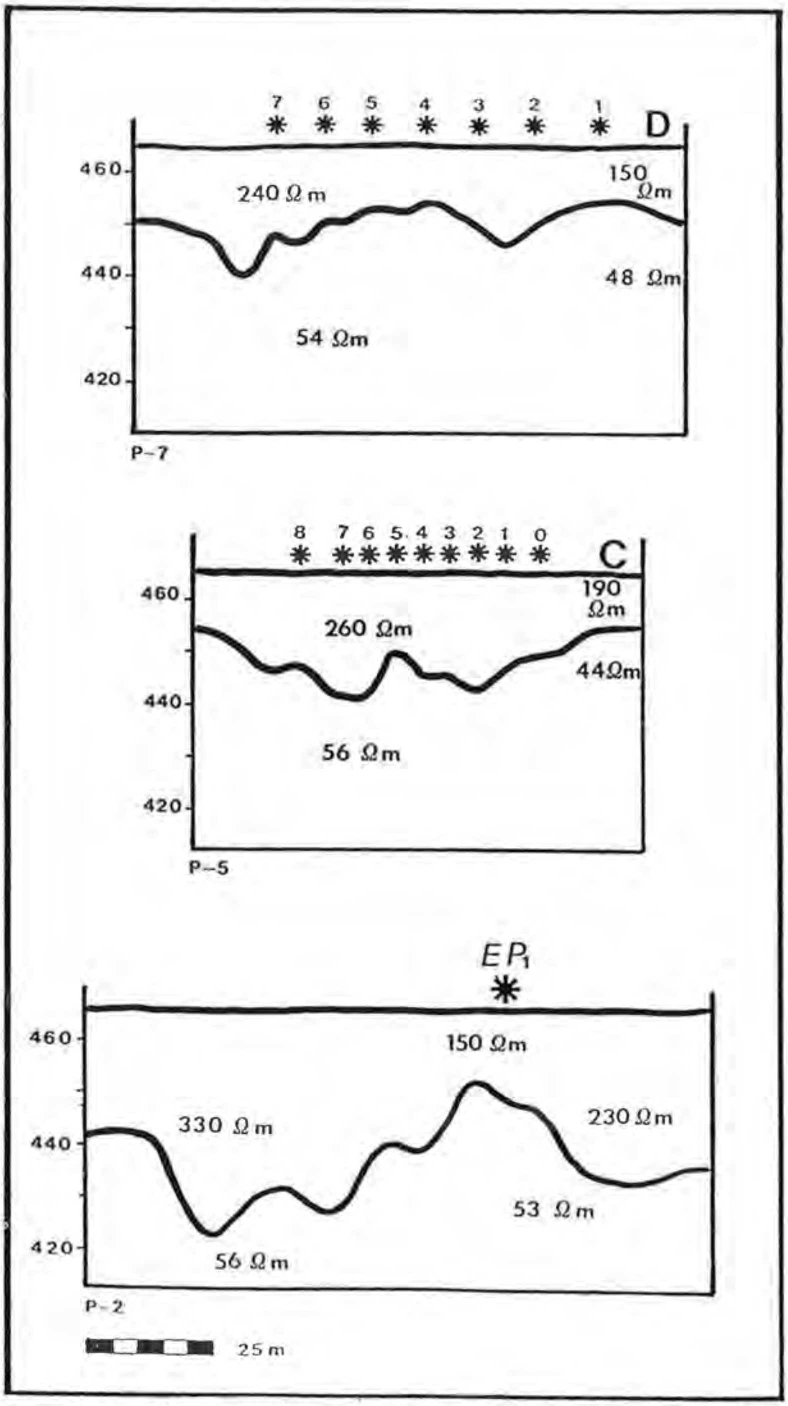

Fig. 4.4: Resistivity profiles along the well galleries EP1, C and D obtained from VLF-R soundings showing the thickness of the permeable gravels and the bounding layer of the aquifer with their true resistivities.

Tab. 4.1: Relation between Rho_a (ohm) and permeability (k) values.

Rho_a (ohm)	k (m/s)
> 300 200–300 100–200 < 100	$\sim > 2 \times 10^{-3}$ $\sim 5 \times 10^{-4} – 2 \times 10^{-3}$ $\sim 5 \times 10^{-5} – 5 \times 10^{-4}$ $\sim < 5 \times 10^{-5}$

Two seismic surveys (s. fig. 4.1) were carried out to verify the VLF-R results and the existence of the loose-gravel-composed-channel. The seismic profiles were oriented perpendicular to the channel. The seismic energy had to travel across the loose gravel channel to reach the geophones. The effect was a reduced seismic velocity in that direction. The existence of the channel was proven by the refraction survey S_3–S_4 (Fig. 4.1). The channel effect can be seen by means of the loss of velocity up to the 12^{th} geophone, because the looser material, which functions as a vertical dyke, attenuates the vibrations. The true resistivity and thickness values were calculated with the help of an empirical relationship between resistivity of saturated loose deposits (log R) and permeability (log k) for the saturated zone.

Figure 4.3 shows the Transmissivity Map and four chosen uranine breakthrough curves (uranine tracer test 1988/89) of the wells D7, C7, D1 and C2. The corresponding resistivity profiles are plotted in fig. 4.4. It shows clearly the leading role which permeability plays in its shape, especially by the time taken to reach the maximum on the one hand (D7, C7, C2) and the spreading of the breakthrough curve by D1 ($T < 500$ m²/d) on the other hand.

Considering the tracer flow it seems that the tracer has a tendency to deflect to the left side of the field (Fig. 4.3). It is therefore a consequence of permeability distribution corresponding to the quickest flow path, which influence the direction of water flow at the local scale. Nevertheless, though C7 has the quickest tracer arrival, the main transport occured in C4 and C6 piezometer. The depression between C7 and EP1 can act as a capture for rather dense solutions. Remarkable is the localisation of the piezometer C4, since it is situated next to a good permeable area, followed by a narrow path, located between two less permeable areas. High tracer concentrations are detected in this well.

5. Mathematical Modelling of the Tracer Experiment Performed in the Test Site Wilerwald (P. Maloszewski, Ch. Leibundgut, J. Schneider)

As described in the former chapters the porous aquifer of the test field Wilerwald is strongly heterogeneous vertically and horizontally (s. fig. 4.1 and 4.2). The exact description of the area under investigation is given in chap. 2. and 4. For the modelling the following mean values of aquifer parameters have been assumed: mean thickness of 10 m, the mean hydraulic conductivity of approximately 2.4×10^{-3} m/s, the average saturated porosity of 12–17% and a mean natural gradient equal to 0.4%).

In this field several dye tracer experiments were performed to determine the aquifer and transport parameters (comp. chap. 3.). The experiment carried out in the year

1985/86 for the distance up to 200 m (M. SANSONI et al., 1988) is reinterpreted in this study to obtain improved information about aquifer properties.

M. SANSONI et al. have assumed that the aquifer consists of three distinct layers (flow domains) with constant thickness and different hydraulic conductivities. Water samples were collected from three depths. The sampling points were fixed in the same depths in all observation wells (Fig. 2.2). In that way in each well were obtained three different tracer concentration curves. Each tracer curve was interpreted separately assuming that the tracer flows through single homogeneous layer (flow domain) having constant transport parameters. Since the aquifer had already been divided into expected three zones having the constant thickness and different hydraulic conductivities by three-level wells (three flow domains), the authors excluded possible vertical heterogenity of each zone. However, it was hydrogeologically not sound to assume that the layers with different permeabilities have the constant thicknesses in all places of the aquifer. As an result the dispersivities found surpassed by far the orders of magnitude values given by other authors for similar formation what indicated to check the interpretation.

Due to this fact in the present reinterpretation study the average weighted concentrations were calculated for each observation well. Similary to the former study it is assumed that the tracer spreads through the multilayered aquifer according to the aquifer's structure and texture. However in contrary to it, without defining the number of layers and their thicknesses. Should the tracer curve exhibit several peaks, this can be interpreted as a result of flow through approximately parallel layers of different permeabilities (J. SCHNEIDER, 1991).

5.1. Transport Model

The general three dimensional transport equation including tensor form of dispersion and vector form of water flow velocity can be found in J. BEAR (1972). In the case of nonreactive tracer injected through the whole thickness of the homogeneous, saturated groundwater aquifer and when the (x,y,z)-coordinate system is chosen with the x-axis always parallel to the direction of flow ($v_x = v$, $v_y = 0$, $v_z = 0$), the 3-D transport equation can be simplified to the following two dimensional one:

$$D_L \frac{\partial^2 C}{\partial x^2} + D_T \frac{\partial^2 C}{\partial y^2} - v \frac{\partial C}{\partial x} = \frac{\partial C}{\partial t}, \tag{5.1}$$

where C is the solute concentration in the water, v is the mean water velocity, x and y are the space coordinates and t is the time variable. D_L and D_T are longitudinal and transversal (lateral) dispersion coefficients, respectively, equal for neglectable molecular diffusion to:

$$D_L = \alpha_L \cdot v, \tag{5.2}$$

$$D_T = \alpha_T \cdot v, \tag{5.3}$$

where α_L and α_T are longitudinal and transversal dispersivities, respectively.

The solution to the eq. (5.1) for instantaneous injection (Dirac impulse in $x = y = 0$) was given by A. LENDA & A. ZUBER (1970) and has the following form:

$$C(x,y,t) = \frac{M}{nH} \frac{x}{4\pi v t^2 \sqrt{D_L D_T}} \exp\left[-\frac{(x - vt)^2}{4D_L t} - \frac{y^2}{4D_T t}\right], \tag{5.4}$$

where

x = the distance between the injection well and an observation well measured parallelly to the streamline,
y = the distance between the streamline exiting from the injection well and the observation well measured perpendiculary to that streamline,
M = the mass of tracer injected,
n = the porosity of the water layer,
H = the thickness of the aquifer.

In the present study the mathematical modelling was limited to the four wells situated directly on the streamline exiting the injection well. In such a case $y = 0$ and eq. (5.4) is simplified to:

$$C(x,0,t) = \frac{M}{nH} \frac{x}{4\pi v t^2 \sqrt{D_L D_T}} \exp\left[-\frac{(x-vt)^2}{4D_L t}\right]. \qquad (5.5)$$

The three unknown flow parameters v, D_L and D_T cannot be determined simultaneously from eq. (5.5). To overcome this difficulty eq. (5.5) can be normalized to the maximal concentration C_{max} measured at time moment $t = t_{max}$ what gives:

$$C(x,0,t) = C_{max} \cdot (t_{max}/t)^2 \exp\left[-\frac{(x-vt)^2}{4D_L t} + \frac{(x-vt_{max})^2}{4D_L t_{max}}\right]. \qquad (5.6)$$

Two transport (model) parameters (v and D_L) can be easly found by fitting the theoretical solution (eq. 5.6) to the concentrations measured by using for example the method of least squares as it was shown by P. MALOSZEWSKI (1981). In some cases, when the exact lenght of the streamline between injection and detection wells (X) is not known, it is better to use instead of v and D_L the following two parameters: the mean transit time of water

$$t_0 = X/v \qquad (5.7)$$

and the Peclet number

$$Pe = vX/D_L = X/\alpha_L. \qquad (5.8)$$

After introducing (5.7) and (5.8) eq. (5.6) is reduced to the following form:

$$C(t) = C_{max} \cdot (t_{max}/t)^2 \exp\left[-\frac{Pe(1-t/t_0)^2}{4t/t_0} + \frac{Pe(1-t_{max}/t_0)^2}{4t_{max}/t_0}\right]. \qquad (5.9)$$

It describes the tracer concentration as a function of time measured in the observation well situated directly on the same streamline as the injection well.

5.2. Multilayered Flow

The theoretical possibilities of determining hydrogeological parameters from tracer experiments performed in multilayered systems was presented by A. ZUBER (1974). A. KREFT et al. (1974) applied this approach to interpreting combined pumping and tracer tests (radial convergent flow). The theoretical model developed by A. ZUBER (1974) was based on unidimensional solution of the dispersion-convection equation, whereas in the present study two dimensional solution (eq. 5.9) is used. The basic model assumptions are:

- the tracer is transported through N-parallel layers;
- the possible tracer exchange between layers is neglectably small;
- the tracer concentration measured in the obserwation well is the weighted mean concentration from all layers;
- the sum of volumetric flow rates (q_j) through the layers is equal to the total volumetric flow rate (q) through the aquifer;
- each j-th layer is considered as an separate flow domain with the constant transport parameters (t_{0j} and Pe_j);
- the tracer mass M_j transported through j-th flow domain (layer) is proportional to the flow rate q_j in that layer:

$$M_j/M = q_j/q = n_j h_j v_j/(nHv), \tag{5.10}$$

where n_j, h_j and v_j are the porosity, the thickness and the mean water velocity, respectively. n is the mean average porosity of the aquifer and v is the mean average water velocity through the whole aquifer

$$v = ik/n, \tag{5.11}$$

whereas the water velocity in the j-th layer is equal

$$v_j = ik_j/n_j. \tag{5.12}$$

i is the hydraulic gradient between injection and detection wells, k_j is the hydraulic conductivity of the j-th layer and k is the mean hydraulic conductivity of the whole aquifer equal

$$k = \sum_{j=1}^{N} (h_j k_j)/H, \tag{5.13}$$

where N is the number of layers.

The basis of the interpretation of the tracer concentration curve measured in the observation well is that the contribution of each partial curve to the total concentration curve expressed as an ratio R_j of the surface (A_j) under the j-th curve to the surface (A) under the whole tracer concentration curve is identical to the contribution of the volumetric flow rate in j-th flow domain to the volumetric flow rate through the whole aquifer and consequently equal to the ratio of the masses M_j to M

$$R_j = A_j/A = A_j/\sum_{j=1}^{N} (A_j) = M_j/M. \tag{5.14}$$

Combining eq. (5.10) and (5.13) and assuming that the porosities in all layers are the same, one finaly obtains that

$$R_j = h_j k_j/(Hk). \tag{5.15}$$

According to A. ZUBER (1974), the mean transit time for the system consists of N-layers can be calculated as follows:

$$t_0 = \sum_{j=1}^{N} (R_j t_{0j}), \tag{5.16}$$

where t_{0j} is the mean transit time of water for j-th layer.

5.3. Estimation of Parameters

The modelling of multilayered groundwater systems includes three processes. First, the theoretical solution (eq. 5.9) is fitted to initial part of the experimental curve (including the first peak) to determine two transport parameters t_{01} and Pe_1 for the first layer. Simultaneously the surface (A_1) under the theoretical curve is calculated. Second, the resulting theoretical concentration curve is substracted from the experimental curve. Now the theoretical solution is fitted to the remaining part of the experimental curve. The parameters t_{02}, Pe_2 and A_2 for the second layer are then found. The theoretical curve is substracted again from the experimental curve. Third, this second process will be repeated as many times as necessary to fit all part of the remaining experimental curve. After that the number of partial curves (N) is automatically found and the portions R_j can be calculated from the surfaces (A_j) by using eq. (5.14).

Finally, if the mean hydraulic gradient and aquifer porosity are known or estimated, the mean hydraulic conductivity of the whole aquifer can be calculated by combining eq. (5.7) and (5.11):

$$k = Xn/(it_0). \tag{5.17}$$

The above k-value is the mean aquifer hydraulic conductivity averaged over the N-layers and the flow distance X between injection and observation well (along the streamline exiting injection well).

If k is known or estimated, the hydraulic conductivity of each layer k_j can easily be calculated under the assumption that the porosities in all layers are the same. By combining eq. (5.7), (5.11) and (5.12) one obtains

$$k_j = kt_0/t_{0j}. \tag{5.18}$$

Additionally, the thicknesses of all layers h_j can be found as follows:

$$h_j = R_jHk/k_j. \tag{5.19}$$

The k_j and h_j values are the mean values for each separate layer calculated as an average values for the flow distance X between injection and detection well.

5.4. Results of Modelling

As it was mentioned earlier the modelling of dye tracer experiment performed by M. SANSONI et al. (1988) was done only for four wells situated approximately on the main streamline EP1–C5–D5–E4–F4. These wells are situated in the distance of 50 m, 100 m, 150 m and 200 m from the injection well, respectively (Fig. 2.1).

Similarly to the earlier study (first serie of tracer tests) it was found that the aquifer consits of three distinctly different layers (N = 3).

The model (transport) parameters for each layer determined for both tracers used (uranine and naphthionate) are summarized in tab. 5.1.

The best fits for uranine are shown in fig. 5.1–5.4 and for naphthionate in fig. 5.5–5.8.

According to the model parameter listed in tab. 5.1 the aquifer consists of three distinctly different layers (N = 3).

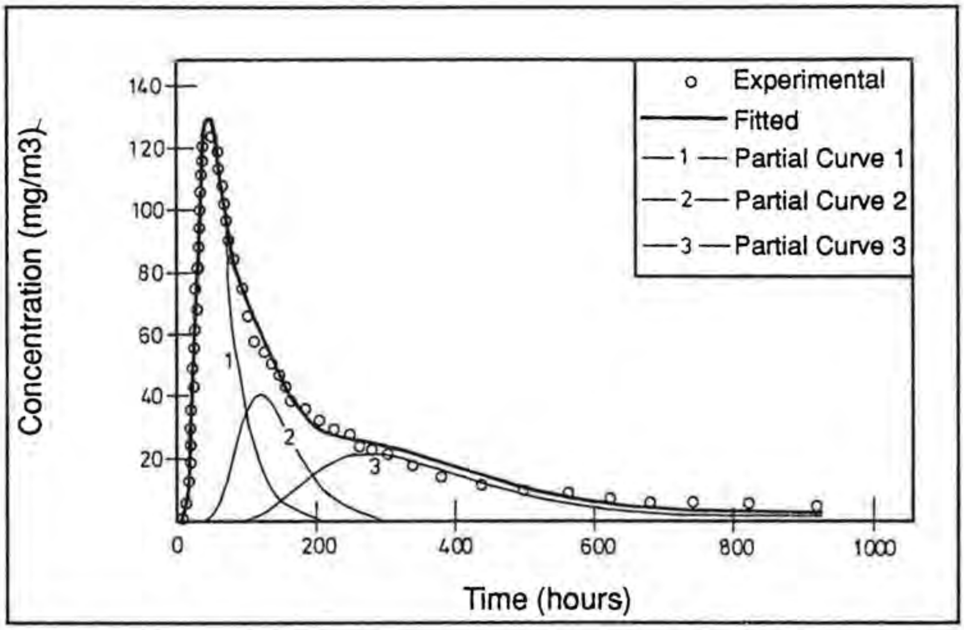

Fig. 5.1: The calculated (best fit) and observed uranine concentration curves in well C5 (distance x = 50 m).

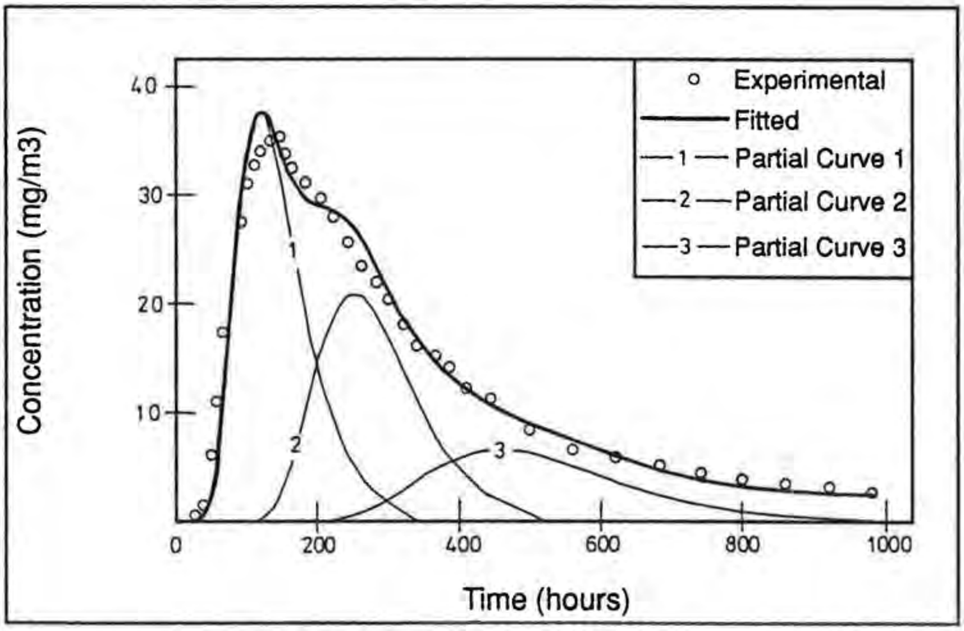

Fig. 5.2: The calculated (best fit) and observed uranine concentration curves in well D5 (distance x = 100 m).

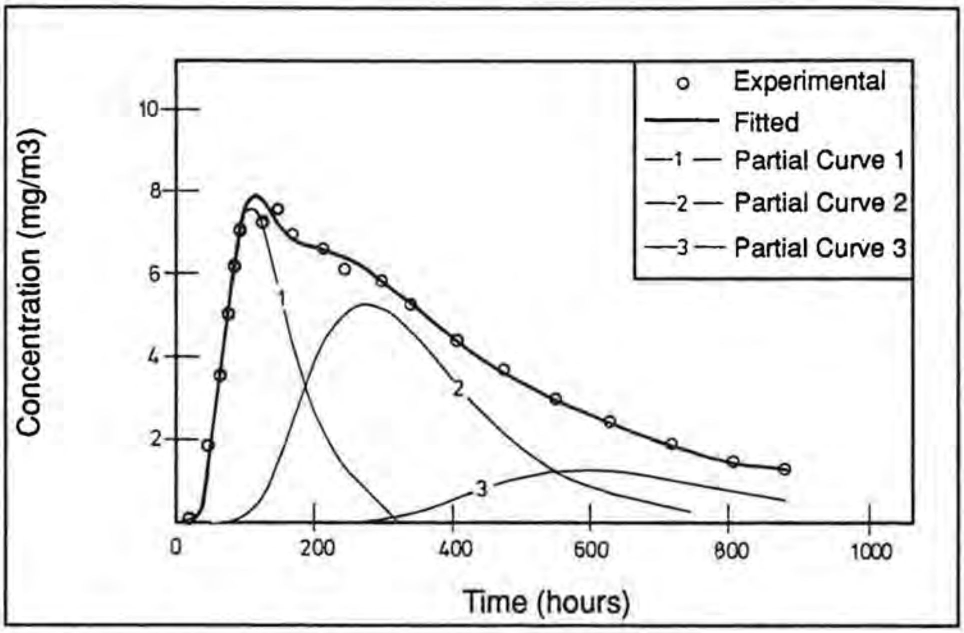

Fig. 5.3: The calculated (best fit) and observed uranine concentration curves in well E4 (distance $x = 150$ m).

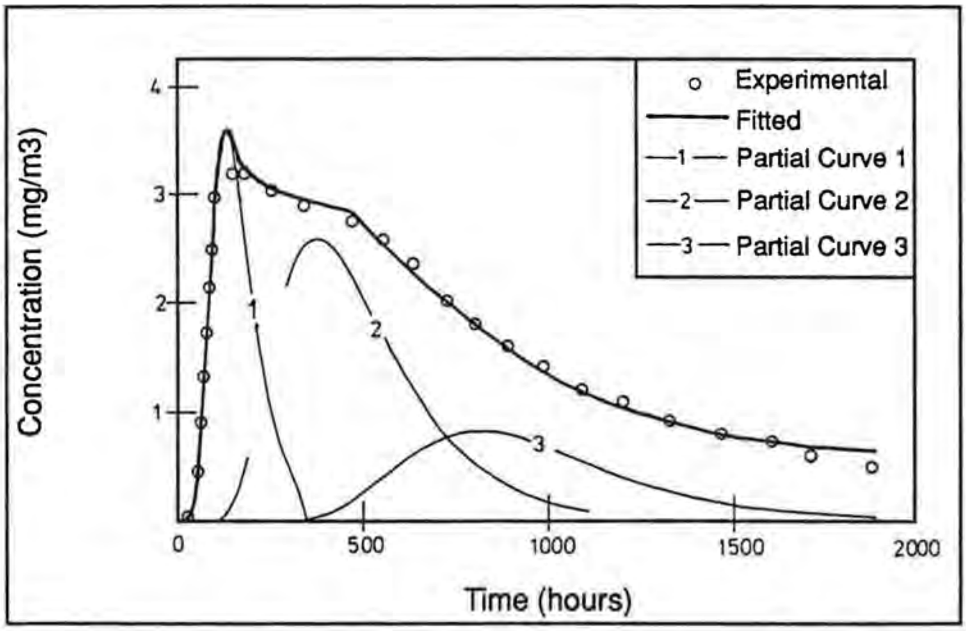

Fig. 5.4: The calculated (best fit) and observed uranine concentration curves in well F4 (distance $x = 200$ m).

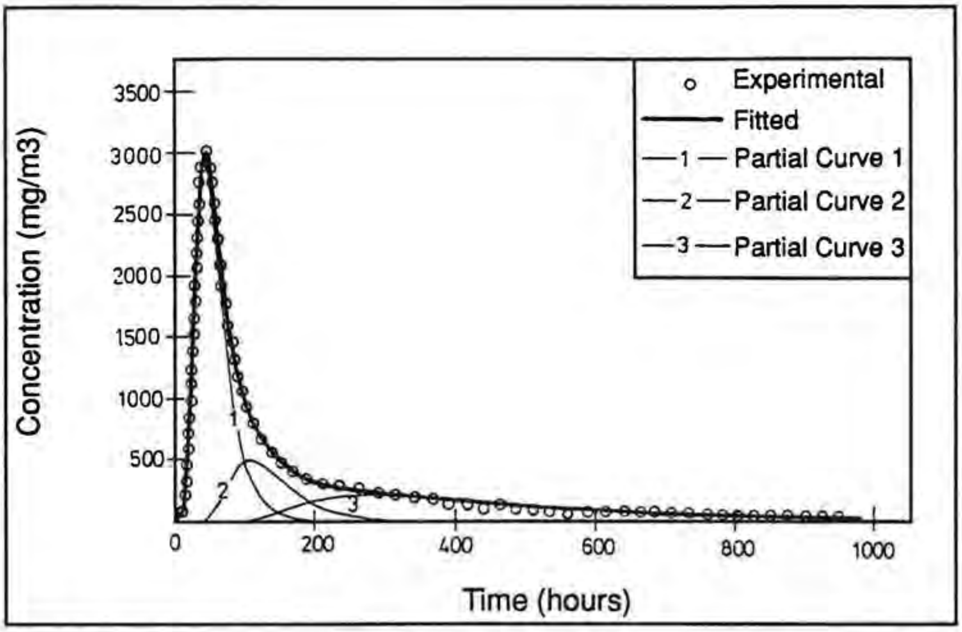

Fig. 5.5: The calculated (best fit) and observed naphthionate concentration curves in well C5 (distance x = 50 m).

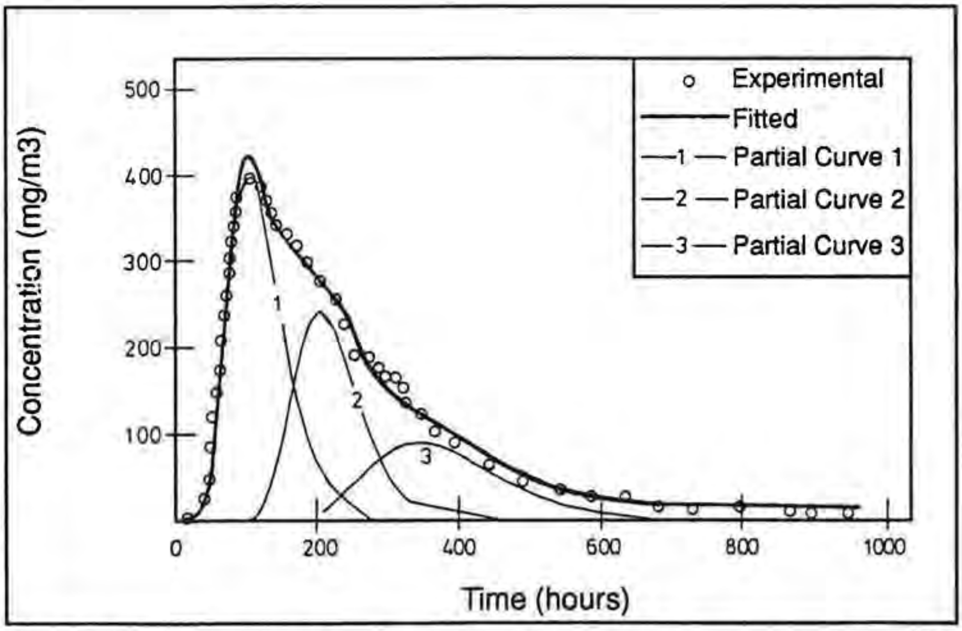

Fig. 5.6: The calculated (best fit) and observed naphthionate concentration curves in well D5 (distance x = 100 m).

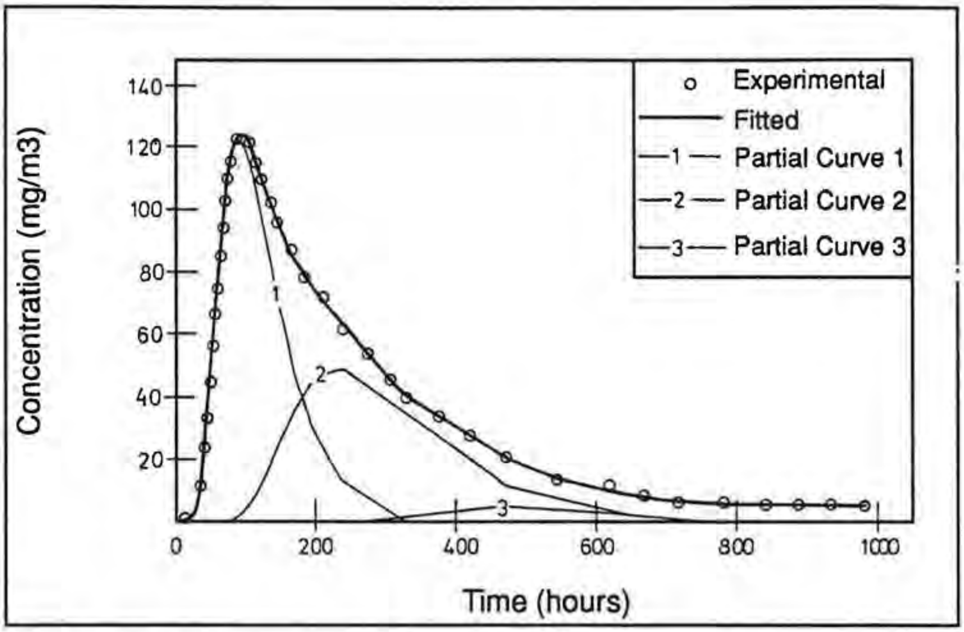

Fig. 5.7: The calculated (best fit) and observed naphthionate concentration curves in well E4 (distance x = 150 m).

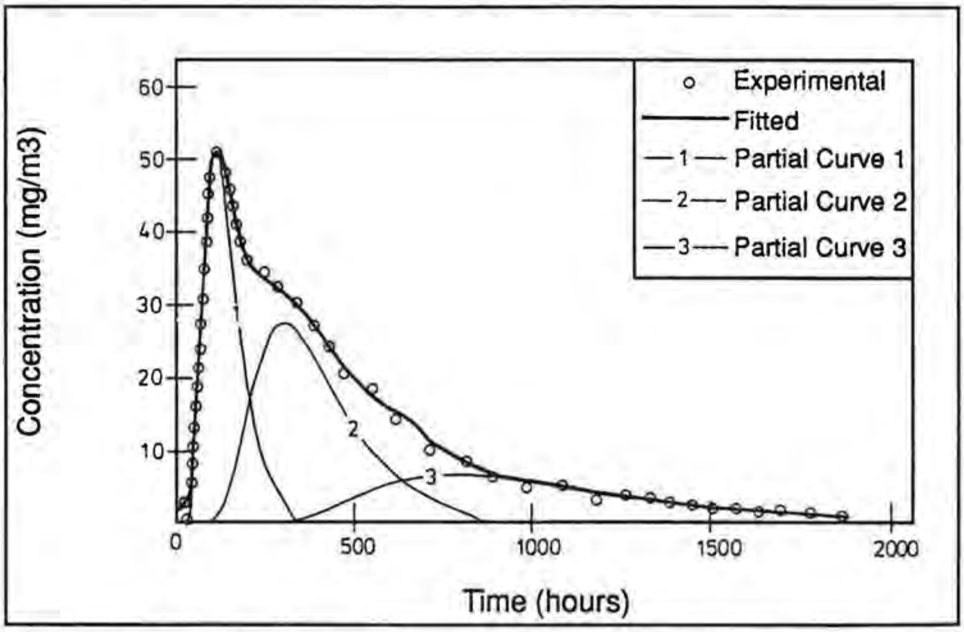

Fig. 5.8: The calculated (best fit) and observed naphthionate concentration curves in well F4 (distance x = 200 m).

Tab. 5.1: Parameters obtained as a result of modelling uranine and naphthionate concentration curves.

			URANINE			NAPHTHIONATE		
Well	Distance	Curve (j)	t_{0j} (days)	v_j (m/d)	a_{Lj} (m)	t_{0j} (days)	v_j (m/d)	a_{Lj} (m)
C5	50 m	1 2 3	3.0 6.2 16.6	16.8 8.1 3.0	6.8 3.0 4.7	2.6 6.0 17.0	19.2 8.4 2.9	5.7 3.9 5.4
D5	100 m	1 2 3	6.5 12.1 22.3	15.4 8.3 4.5	7.7 3.7 4.0	5.4 9.3 16.3	18.4 10.8 6.1	6.8 2.6 3.4
E4	150 m	1 2 3	6.5 16.0 29.7	23.1 9.4 5.1	16.3 13.1 7.1	5.8 13.6 22.0	26.0 11.0 6.8	18.1 13.4 4.7
F4	200 m	1 2 3	7.6 22.4 42.0	26.3 8.9 4.8	19.5 19.4 10.9	6.5 17.9 45.9	30.7 11.2 4.4	20.0 18.7 17.7

The hydrogeological parameters for the aquifer under consideration were calculated based on the modelling results obtained for naphthionate by using equations (5.17) to (5.19) for assumed mean aquifer thickness $H = 10$ m, mean aquifer porosity $n = 15\%$ and mean hydraulic gradient $i = 0.4\%$. The results are summarized in tab. 5.2.

The aquifer parameters obtained under above assumptions should be considered as an approximation of the real parameters.

It is due to the fact that hydraulic gradient and porosity varied in the whole experimental field "Wilerwald" and were not constant between the wells as it was assumed. The results show that the aquifer is strongly heterogen vertically and horizontally.

Tab. 5.2: The hydrogeological parameters obtained from the flow parameters found from naphthionate. The parameters values are the mean values for the flow distance between injection and observation wells.

Well	Distance	Layer (j)	R_j	t_0 (days)	k (m/s)	k_j (m/s)	h_j (m)
C5	50 m	1 2 3	0.56 0.20 0.24	6.7	3.2×10^{-3}	8.2×10^{-3} 3.6×10^{-3} 1.3×10^{-3}	2.2 1.9 5.9
D5	100 m	1 2 3	0.45 0.32 0.23	9.2	4.7×10^{-3}	8.0×10^{-3} 4.8×10^{-3} 2.7×10^{-3}	2.7 3.2 4.1
E4	150 m	1 2 3	0.50 0.45 0.05	10.1	6.4×10^{-3}	11.1×10^{-3} 4.8×10^{-3} 2.9×10^{-3}	2.9 6.0 1.1
F4	200 m	1 2 3	0.30 0.43 0.27	22.0	3.9×10^{-3}	13.2×10^{-3} 4.8×10^{-3} 1.9×10^{-3}	1.0 3.5 5.5

Vertically, the differences in hydraulic conductivities varies 3 to 8 times between the layers. Horizontally, in the first layer one observes increase of the hydraulic conductivity (k) after first 100 m whereas in the second and the third layer already after 50 m. The third layer shows in addition a decrease of the parameter k after 150 m.

The average aquifer hydraulic conductivity increases nearly linearly till 150 m and after that decreases.

6. Conclusions (Ch. LEIBUNDGUT, A. DE CARVALHO DILL, I. MÜLLER, P. MALOSZEWSKI, J. SCHNEIDER)

Tracer tests under natural conditions are a useful tool in order to obtain transport and aquifer parameters respectively and in order to calibrate flow models. However, the evaluation and interpretation of tracer tests has to be done very careful. Crucial points of evaluation are the assumptions concerning the aquifer properties of a largely unknown aquifer. An aquifer must be considered as largely unknown even then if data from boreholes are available. Due to the point character of these data the knowledge of the spatial distribution of the aquifer parameters is mostly insufficient.

The aquifer parameters received from the tracer tests reflect these features. As well the knowledge of the aquifer's properties, especially the variability of the hydraulic conductivity in horizontal and vertical direction, as the assumption of a three layer model had to be reconsidered after getting the results of the tracer tests. The distribution of the hydraulic conductivity has been checked with the VLF technique and a reinterpretation by using a different approach of flow modelling has been carried out in order to obtain improved results and understanding of the flow behaviour respectively in the test field's aquifer.

As could be seen in the Transmissivity Map (Fig. 4.3) the localisation of the injection point on a "watershed" in a less permeable domain plays an important role in the explanation of the differences within tracer experiments. These findings are documented in two profiles (P1 and P3, s. fig. 4.2). What consequently became evident is the existence of a large depression in the aquitard, west of the injection point EP1. The information given by the VLF-R, leads to an extended interpretation of tracer transport and by using the obtained maps, block diagrams and profiles gives a more complet and clear concept of the structure and permeability distribution of the aquifer. The existence of true channels has been proved. Further it shows the crucial importance for the localisation of observation wells and injection well for sampling and interpretation.

In order to reinterprete the first serie of tracer tests a multilayered transport model has been used. The application of the model led to three clearly different layers within the aquifer. Compared with the first evaluation of the tracer test and especially with the geophysical experiment the result is unexpected.

One has to be aware of the fact that the applied model gives mean values of parameters for the flow distance between the injection well and the several well galleries. The combined use of tracer methods and the geophysical investigations give satisfactory information about hydrodynamic conditions and aquifer heterogenity.

With the todays knowledge of the aquifer the goal of a natural laboratory is reached. However, the implementation of further sampling points (wells) according to the VLF results (s. chap. 3.) is needed. With such an improved infrastructure tracer tests should be carried out for and in the different parts of the aquifer, e.g. in the high permeable channel. Also the assumption for the choice of the adequate flow model will be easier.

Summary

A test field for tracer experiments is available in a porous aquifer near Berne in Switzerland. In the study on hand the evaluation of the test field "Wilerwald", the installation of the observation well network, the tracer tests with their interpretation by mathematical modelling and the geophysical prospection is described. The results show a very heterogenous aquifer in a three layer structure interpersed with a channel and zones of higher permeabilities.

References

BEAR, J. (1972): Dynamics of Fluids in Porous Media. – Amer. Elsevier, New York.

FISCHER, G., B.V. LE QUANG & I. MÜLLER (1983): VLF Ground Surveys, a Powerful Tool for the Study of Shallow Two Dimensional Structures. – Geophysical Prospecting, **31**, 977–991.

FISCHER, G., P.-A. SCHNEGG, J. MA, I. MÜLLER & M. BURCKHARD (1987): Étude VLF-R du remplissage quaternaire de la Vallée de Gastern (Alpes Bernoises, Suisse). – Ecolgae géol. Helv., **80**, 773–787.

KOLL, J. & I. MÜLLER (1989): Elektromagnetische Very Low Frequency-Resistivity (VLF-R) Prospektion zur Erkundung von Grundwasserleitern im paläozoischen Mittelgebirge am Beispiel des Oberharzes. – Steir. Beitr. z. Hydrogeologie, **40**, 103–122, Graz.

KREFT, A., A. LENDA, B. TUREK, A. ZUBER & A. CZAUDERNA (1974): Determination of effective porosities by the two-well pulse method. – Isotopes Techniques in Groundwater Hydrology, 295–312, IAEA, Wien.

LEDERMANN, H. (1978): Erläuterungen zu Atlasblatt 72, Geologischer Atlas der Schweiz 1 : 25.000, Blatt: 1127, Solothurn. – Schweizerische Geologische Komission.

LENDA, A. & A. ZUBER (1970): Tracer dispersion in groundwater experiments. – Isotope Hydrology **1970**, 619–641, IAEA, Wien.

MÄGDEFESSEL, J. (1990): Zum Fließverhalten von Tracern und Schadstoffen im Porengrundwasser. – Unpubl. Diplomarbeit, Univ. Freiburg i.Br.

MALOSZEWSKI, P. (1981): Computerprogramm für die Berechnung der Dispersion und der effektiven Porosität in geschichteten porösen Medien. – GSF-Bericht **R 269**, GSF, Neuherberg.

SANSONI, M., B. SCHUDEL & T. WAGNER (1987): Tracermethodischer Beitrag zum Stofftransport im Porengrundwasser. – Unpubl. Diplomarbeit, Univ. Bern.

SANSONI, M., B. SCHUDEL, T. WAGNER & Ch. LEIBUNDGUT (1988): Aquiferparameterermittlung im Porengrundwasser mittels fluoreszierender Tracer. – Gas – Wasser – Abwasser, 141–147.

SCHNEIDER, J. (1991): Tracer Experiments in a Porous Aquifer in Switzerland. – Unpubl. Diplomarbeit, Univ. Freiburg i.Br.

SCHOTTERER, U. & I. MÜLLER (1986): Investigation of Groundwater Flow Systems in a Quaternary Sandstone Complex near Berne (Switzerland) by means of indirect methods of Environmental Isotopes and Geophysiques. – Proc. of the 5th Internat. Symp. on Underground Water Tracing, Athens 1986, 403–412, Institute of Geology and Mineral Exploration, Athens.

STOBER, I. (1988): Dispersion als Hinweis auf den Karsttypus. – Deutsche Gewässerkundl. Mitteilungen, H. **4**, 107–110.

THIERRIN, J. & I. MÜLLER (1988): La méthode VLF-Résistivité Multifréquence, un exemple d'exploration hydrogéologique dans un Scientifique de l'Université de Besançon, Geologie. – Mémoire Hors Série n°6. (Quatrième Colloque d'Hydrogéologie en Pays Calcaire, Sept. 1988), 17–25.

ZUBER, A. (1974): Theoretical possibilities of the two-well pulse method. – Isotopes Techniques in Groundwater Hydrology, 277–293, IAEA, Wien.