Interpretation of tracer tests performed in fractured rock of the Lange Bramke basin, Germany

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Abstract Two multitracer tests performed in one of the major cross-fault zones of the Lange Bramke basin (Harz Mountains, Germany) confirm the dominant role of the fault zone in groundwater flow and solute transport. Tracers having different coefficients of molecular diffusion (deuterium, bromide, uranine, and eosine) yielded breakthrough curves that can only be explained by a model that couples the advective–dispersive transport in the fractures with the molecular diffusion exchange in the matrix. For the scale of the tests (maximum distance of 225 m), an approximation was used in which the influence of adjacent fractures is neglected. That model yielded nearly the same rock and transport parameters for each tracer, which means that the single-fracture approximation is acceptable and that matrix diffusion plays an important role. The hydraulic conductivity of the fault zone obtained from the tracer tests is about 1.5×10^{-2} m/s, whereas the regional hydraulic conductivity of the fractured rock mass is about 3×10^{-7} m/s, as estimated from the tritium age and the matrix porosity of about 2%. These values show that the hydraulic conductivity along the fault is several orders of magnitude larger than that of the remaining fractured part of the aquifer, which confirms the dominant role of the fault zones as collectors of water and conductors of fast flow.

Résumé Deux multitraçages ont été réalisés dans l'une des zones principales de failles du bassin de Lange Bramke (massif du Harz, Allemagne); les résultats confirment le rôle prédominant de la zone de failles pour l'écoulement souterrain et le transport de soluté. Les traceurs, possédant des coefficients de diffusion différents (deutérium, bromure, uranine et éosine), ont fourni des courbes de restitution qui ne peuvent être expliquées que par un modèle qui associe un transport advectif-dispersif dans les fractures à un échange par diffusion moléculaire dans la matrice. A l'échelle des expériences (distance maximale de 225 m), l'influence des fractures adjacentes a été négligée. Ce modèle a fourni pour chaque traceur pratiquement les mêmes paramètres pour la roche et le transport, ce qui signifie que l'approximation de la fracture unique est acceptable et que la diffusion dans la matrice joue un rôle important. La conductivité hydraulique de la zone de faille fournie par les traçages est d'environ $1{,}5\times10^{-2}$ m/s, alors que la conductivité hydraulique régionale de la roche fracturée dans son ensemble est de l'ordre de 3×10^{-7} m/s, selon l'estimation tirée des âges tritium et de la porosité de la matrice d'environ 2%. Ces valeurs montrent que la conductivité hydraulique le long de la faille est supérieure de plusieurs ordres de grandeur à celle de la partie fracturée restante de l'aquifère, ce qui confirme le rôle prédominant joué par les zones de failles comme drains de l'eau et comme axes d'écoulement rapide.

Resumen Dos ensayos con múltiples trazadores realizados en una de las zonas más fracturadas de la cuenca de Lange Bramke (Montes Harz, Alemania) confirman el papel dominante de la zona de fractura en el flujo de agua subterránea y el transporte de solutos. Trazadores con distintos coeficientes de difusión molecular (deuterio, bromuro, uranina y eosina) dieron curvas de llegada que sólo pueden ser explicadas mediante un modelo que acople el transporte advectivo-dispersivo en las fracturas con la difusión en la matriz. Para la escala de los ensayos (distancia máxima de 225 m), se usó como aproximación que la influencia de las fracturas adyacentes podía despreciarse. Este modelo dio lugar para cada trazador a valores muy similares de los parámetros de transporte, lo que supone que la aproximación de fractura única es aceptable y que la difusión en la matriz es un mecanismo importante. La conductividad hidráulica de la zona fracturada obtenida de los ensayos es de unos 1.5×10^{-2} m/s, mientras que la conductividad hidráulica regional para la matriz rocosa es de

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unos 3×10^{-7} m/s, valor estimado de la edad del tritio y la porosidad de la matriz es del 2%. Estos resultados indican que la conductividad hidráulica a lo largo de la fractura es varios órdenes de magnitud mayor que la correspondiente al resto de la parte fracturada del acuífero, lo que confirma el papel dominante de las fracturas como zonas conductoras preferentes de agua.

Key words tracer tests · fractured rocks · Germany · diffusion · analytical solutions

Introduction

The quantitative interpretation of tracer tests provides an estimate of relevant transport parameters for fractured-rock aquifer systems (e.g., Cacas et al. 1990; Cady et al. 1993; Himmelsbach and Maloszewski 1992; Maloszewski and Zuber 1985, 1990, 1993; Novakowski et al. 1985; Raven et al. 1988; Shapiro and Nicholas 1989), which is of particular importance for reliable modeling of contaminant transport in nonhomogeneous and anisotropic fractured rocks.

The purpose of the present study was to: (1) examine the role of a major fault of the Lange Bramke basin, Harz Mountains, central Germany, in the transport of water and pollutants through the fractured rock; (2) determine rock and transport parameters of the fault zone; and (3) demonstrate advantages of multitracer tests for determining flow and rock parameters.

Fractured rock can be approximated by a system of parallel fractures of equal aperture and spacing (Sudicky and Frind 1982). Convective–dispersive flow takes place in the fractures, whereas the matrix contains stagnant water. Exchange between mobile and immobile water is governed by molecular diffusion. For a sufficiently small scale, the influence of adjacent fractures is negligible, and a single-fracture approximation is applicable, as shown by Maloszewski and Zuber (1985, 1990, 1993) and Maloszewski (1994). In that approximation, the dispersivity depends, to a high degree, on differences in flow velocities in fractures, and the fracture aperture is an apparent value, as defined by Silliman (1989) and followed by Maloszewski and Zuber (1993).

Study Site

The results of conventional hydrologic, isotopic, and geophysical investigations of the Lange Bramke basin are presented by Herrmann et al. (1989). The basin, shown in *Figure 1* (0.76 km^2, 543–700 m a.s.l., 90% forested with Norwegian spruce), is situated on the windward slope of the Harz Mountains. Mean annual precipitation is 1300 mm and runoff is 700 mm ($Q = 17$ L/s). The valley is drained by the Lange Bramke Creek. Three geohydrologic units are identified, as shown in *Figures 2* and *3*: (1) The unsaturated-soil zone has an average thickness of 3.5 m and consists of residual weathering and allochthonic Pleistocene-age solifluidal materials that are rich in skeletal material. Podsolic brown soil is the dominant soil type, with vertical macropore systems, which enhance draining; (2) the fractured-rock groundwater system, which consists of folded and intensely fractured sandstones, quartzites, and some slates of Early Devonian age. Several major faults cross one another; these have partial infilling by quartz and argillaceous materials. The unsaturated upper zone of the fractured-rock groundwater system is strongly dissected, allowing ready seepage through the interface between the unsaturated-soil and the fractured-rock groundwater systems; and (3) the porous groundwater reservoir, which is located along the valley bottom and consists of gravel, including pebbles and some boulders.

A combined hydrologic and environmental-isotope (tritium and δ^{18}O) study performed during 1980–88 indicates that the mean tracer age of water leaving the basin is about 2.7 yr, of which 1.5 yr corresponds to flow in the saturated fractured part of the system (Herrmann et al. 1989). Following the approach of Zuber and Motyka (1994), the regional hydraulic conductivity is:

$$k = \frac{s(n_p + n_f)}{i t_t} \cong \frac{s n_p}{i t_t} = 3 \times 10^{-7}\ ms^{-1} \quad (1)$$

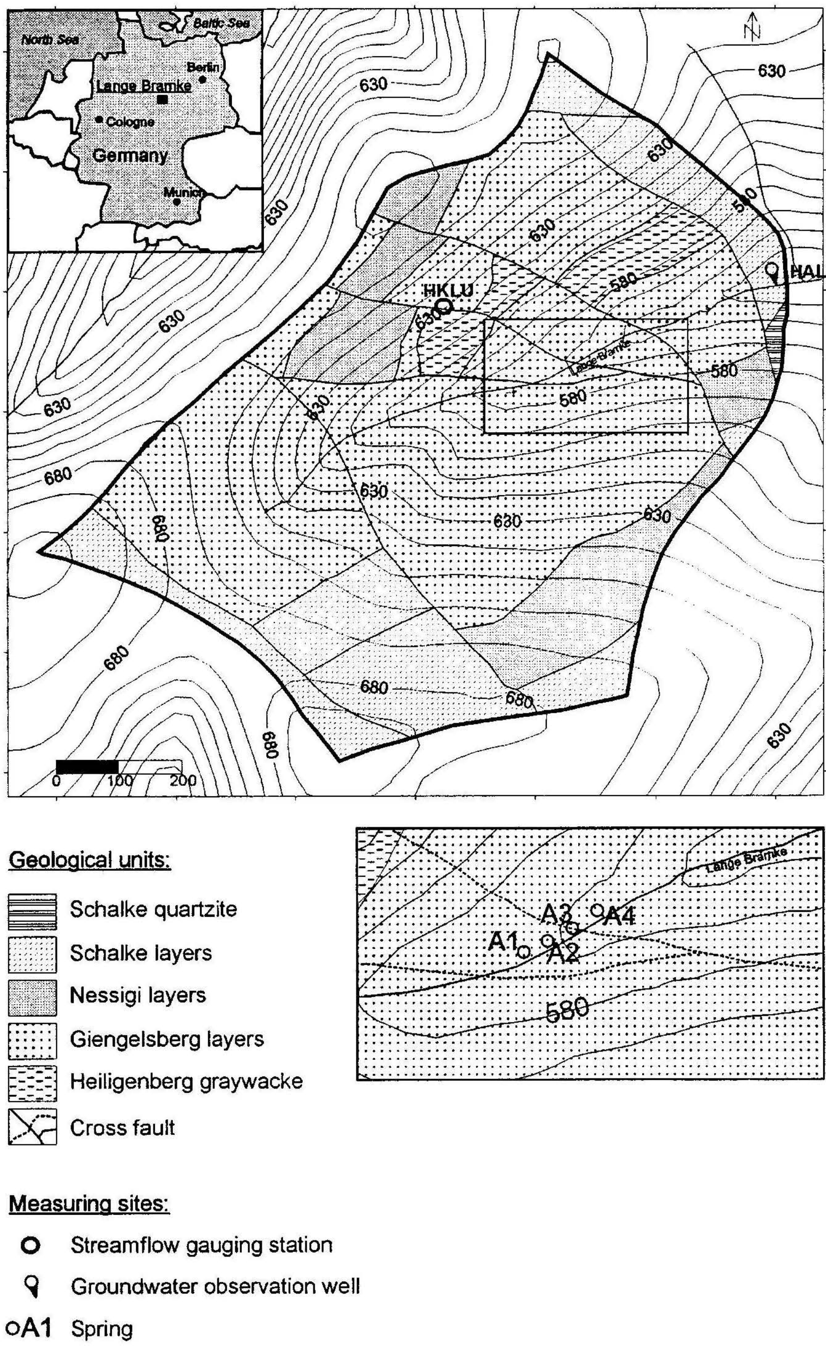
where s is the mean distance of saturated flow in the fractured system of the basin (about 150 m, as measured from the "center" of the recharge area to the creek), n_p is the matrix porosity obtained from laboratory measurements (about 0.02), i is the mean hydraulic gradient (about 0.2), and t_t is the mean tracer age (1.5 yr).

On the basis of hydraulic assessment of the bedrock in the catchment basin, two types of zones are distinguished: (1) fractured and faulted zones, which act as fast-flow channels; and (2) much less disturbed zones, which act as barriers to the fast movement. Major cross faults are assumed to be the dominant transport pathways. Evidence for this assumption is based on the fact that drilling fluids and rock flour appeared in Lange Bramke Creek several hours after drilling the borehole HKLU, which penetrated a major cross fault, had commenced. That event gave justification for performing tracer tests aimed at evaluating the hydraulic connection between the creek and the major fault zone at the center of the basin (*Figure 2*).

Tracer Tests

Two tracer tests were performed in the major fault (*Figure 3*). For test A, a cocktail of 3.4 kg of potassium bromide, 0.1 kg of eosine, and 480 mL of HDO (100% abundance) was injected in 1.17 m^3 of water. The injection was performed on the exposed weathered/frac-

Figure 1 Location of the Lange Bramke basin in Harz Mountains, Germany, showing geology and measuring sites



tured-rock surface of 1 m^2 located at a distance of 11 m from the drilling well HKLU, to measure transport through the unsaturated and underlying saturated zones. For test B, a cocktail of 0.6 kg uranine and 1.2 kg eosine was injected in 0.06 m^3 of water. The injection was performed in the well HKLU, 11.5 m below the water table. The two tracer tests with nearby injections were performed to check if it is possible to obtain reasonable results without having to drill a specially designed injection well, and to obtain, if possible, some information on the transit time through the unsaturated zone.

Hydrologic conditions differed for the two tests. Test A was influenced by extremely dry antecedent weather conditions, resulting in a low water table (–22.6 m at well HKLU) and low discharge of Lange Bramke Creek (Q=5.5 L/s at HALB, whereas the yearly average is 17 L/s). The injection of tracer solution was followed by about 3.6 m^3 of water to keep tracers mobile for passage through the unsaturated zone. This artificial flooding (Q_{inf} about 1 m^3/h during 3.5 h) created practically saturated flow conditions within the traced section of the soil zone above the water table (for the injection area and the column height

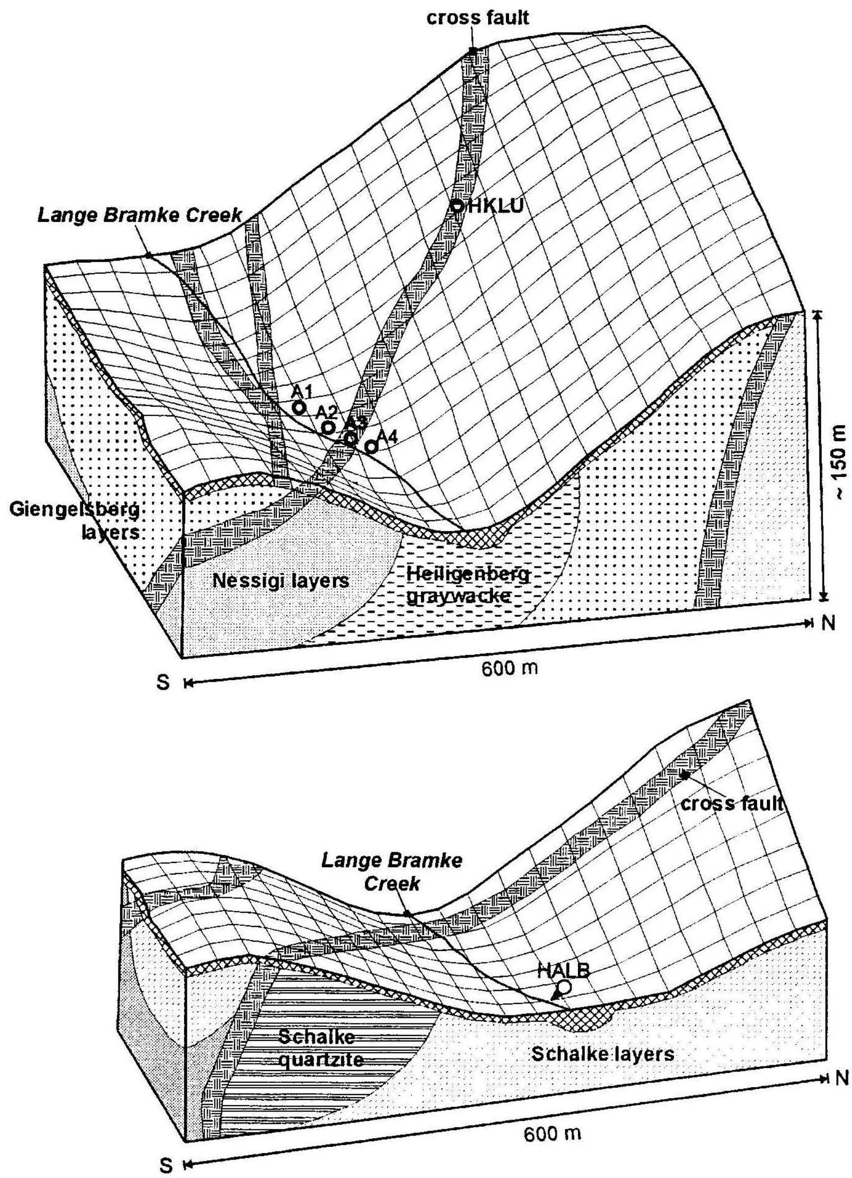


Figure 2 Three-dimensional views of the Lange Bramke basin, showing geology, major faults and experimental sites

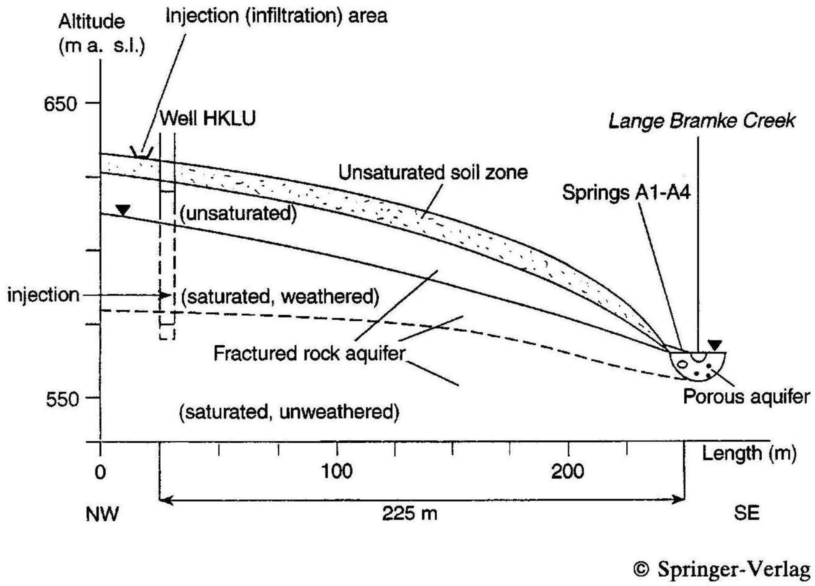


Figure 3 Simplified hydrogeological section along the fault line between the well HKLU and spring A4

of 23 m, about 3 m^3 of water is needed to achieve saturation).

Test B was preceded by two rainy days; total precipitation was 39 mm. During the following two days, 19.6 mm of precipitation fell, and precipitation continued; daily amounts were 2–4 mm until the 11th day of the experiment. As a result, the initial water table (–19.6 m) in the injection well and the discharge of Lange Bramke Creek at HALB (Q=28 L/s) were substantially greater than the long-term means.

In both tests, samples were taken automatically in spring A3 (225 m from the detection and/or injection well HKLU), and at the HALB stream-gauging station at Lange Bramke Creek (330 m from A3). During test A, samples were also taken at the well HKLU, and springs A1, A2, and A4.

Mathematical Model

The tracer is assumed to be injected into the water entering the fractures and transported along the fractures by dispersion–convection flow, with the sink term resulting from tracer losses into the matrix (Eq. 2). Tracer transport in the porous matrix between the fractures is assumed to be controlled by molecular diffusion and an instantaneous equilibrium adsorption (Eq. 3). The transit time of water in fractures is assumed to be short enough to yield the diffusion of tracer into the matrix not deep enough to be affected by adjacent fractures. The two transport equations are as follows:

$$\frac{\partial C_f}{\partial t}+v\frac{\partial C_f}{\partial x}-D\frac{\partial^2 C_f}{\partial x^2}-\frac{n_p D_p}{b}\left.\frac{\partial C_p}{\partial y}\right|_{y=b}=0 \quad \text{for } 0\le y\le b \qquad (2)$$

$$\frac{\partial C_p}{\partial t}-\frac{D_p}{R_{ap}}\frac{\partial^2 C_p}{\partial y^2}=0 \quad \text{for } b\le y\le\infty \qquad (3)$$

where C_f and C_p are the tracer concentrations in the water of fractures and the matrix, respectively; v is the mean water velocity in the fracture; x and y are the spatial co-ordinates taken in the flow direction and perpendicular to the fracture extension; D is the dispersion coefficient for the fractures; $2b$ is the apparent fracture aperture; t is the time variable; n_p is the matrix porosity; D_p is the diffusion coefficient in the porous matrix, equal to D_m/τ_p (D_m is the molecular diffusion coefficient in free water, and τ_p is the tortuosity factor for the matrix); R_{ap} is the retardation factor for the porous matrix governed by instantaneous equilibrium adsorption (for an ideal tracer, $R_{ap}=1$).

The solution to Eqs. 2 and 3 for instantaneous tracer injection described by the Dirac δ(t) function is (Maloszewski and Zuber 1985, 1990):

$$C_f(t)=\frac{aM\sqrt{Pet_o}}{2\pi Q}\int_0^t \exp\left[-\frac{(t_o-\xi)^2 Pe}{4\xi t_o}-\frac{a^2\xi^2}{t-\xi}\right]\frac{d\xi}{\sqrt{\xi(t-\xi)^2}} \qquad (4)$$

where $C_f(t)$ is the concentration measured at a given distance (X), M is the injected tracer mass, Q is the volumetric flow rate through the fracture, and ξ is the integration variable. The volumetric flow rate is well controlled only in convergent radial flow and sometimes in natural outflows. Within this work, Q was measured only at the gauging station of the creek. For samples taken in the well and springs, the Q value was considered as a fitted (sought) parameter, having the same value for all the tracers measured at a given site.

The fitting parameters of Eq. (4), called the Single Fracture Dispersion Model (SFDM), are:

$$t_o=X/v \qquad (5.1)$$

$$Pe=vX/D=X/\alpha_L \qquad (5.2)$$

$$a=n_p\,(D_p R_{ap})^{1/2}/(2b) \qquad (5.3)$$

where t_o is the mean transit time of water defined by Eq. 5.1, Pe is the Péclet number, and a is the diffusion parameter. For the known flow distance (X), the mean water velocity and the longitudinal dispersivity (α_L) are obtained from t_o and Pe. For an ideal tracer, $R_{ap}=1$, and then the fracture aperture can be calculated from Eq. 5.3, provided the matrix porosity and diffusion coefficient are known. For the fault zone, the ratio of hydraulic conductivity to the fracture porosity (k/n_f) can be estimated from Darcy's law by making use of the mean transit time of water (t_o) and the hydraulic gradient (i) between the injection and the detection points:

$$k/n_f=X/(it_o) \qquad (6)$$

The hydraulic conductivity of the fault (k_s) can also be expressed as (Maloszewski and Zuber 1993):

$$k_s=k/n_f=6.25\times10^5\,(2b/\tau_f)^2 \qquad (7)$$

where $2b$ (in meters) is the aperture of the fault fracture or the mean fracture aperture if multiple fractures exist; k_s (or k) is in ms^{-1}; and τ_f is the tortuosity factor for fractures, which is usually taken as equal to 1.5. The fracture porosity in Eq. 7 appears if, instead of a single fault fracture, the presence of multiple of fractures is assumed.

Results and Discussion

The selection of the mathematical model and the final estimation of flow and rock parameters were only possible by simultaneous consideration of the results of both experiments. Only the SFDM yielded good fits for all tracers used simultaneously. At the planning stage, it was assumed that all tracers used were conservative. Unfortunately, the dye tracers are adsorbable. The interpretation of all tracer curves was possible due to the use of two conservative tracers (bromide and deuterium), and the existence of the adsorption term (R_{ap}) in the SFDM for nonconservative tracers.

Test A

Observations at well HKLU

The three tracers used (deuterium, eosine, and potassium bromide) were jointly detected only in the HKLU well. The observed tracer concentrations, normalized by the total mass of tracer injected, are presented in *Figure 4*. All tracers have similar arrival times of the maximal concentrations; the bromide curve had a higher C/M maximal value than deuterium (0.052 and 0.033 m^{-3}, respectively). Because bromide and deuterium are nonsorbable (ideal tracers), the observed difference in breakthrough curves is explained only by diffusive exchange of tracer mass between mobile and stagnant water (double-porosity medium), because their coefficients of molecular diffusion differ (1.5×10^{-9} and 2.5×10^{-9} m^2/s, respectively). This hypothesis is supported and confirmed by successful application of the SFDM for the interpretation of experimental data.

The similar shapes of the eosine and deuterium curves, shown in (*Figure 4*), suggest that the parameters t_o, Pe, and a (Eqs. 5.1–5.3) have nearly the same values for eosine and deuterium. This in turn yields that the D_pR_{ap} values for eosine and deuterium are the same. For deuterium, $R_{ap}=1$, and D_m is 5 times larger than D_m of eosine; therefore, it was concluded that eosine is adsorbed in the matrix and that $R_{ap}\approx5$. A value of 5.4 was later obtained in the SFDM calibration.

The parameters of the SFDM are obtained by fitting Eq. 4 to the tracer concentrations measured in HKLU; these results are summarized in *Table 1*. The best fits of the SFDM to experimental data are shown in *Figure 4*. The SFDM yields the same mean transit times and Péclet numbers for the three tracers and, as expected, different values of the a-parameter. This result is consistent with the assumption about the existence of the double-porosity medium and confirms that an adequate model was used.

The values of the a-parameter can be used to estimate the diffusion coefficient of any tracer, provided that the a-value is known for at least one of the tracers (e.g., Maloszewski and Zuber 1985, Himmelsbach and Maloszewski 1992). Assuming the diffusion coefficient for deuterium $D_p=1.67\times10^{-9}$ m^2/s ($2.5\times10^{-9}/1.5$), Eq. 5.3 yields:

$$(D_p)_{\mathrm{Bromide}}=(D_p)_{\mathrm{Deuterium}}\times(a_{\mathrm{Bromide}})^2/(a_{\mathrm{Deuterium}})^2=0.94\times10^{-9}\ \mathrm{m^2/s} \tag{8.1}$$

whereas for eosine the product (D_pR_{ap}) can be calculated:

$$(D_pR_{ap})_{\mathrm{Eosine}}=(D_p)_{\mathrm{Deuterium}}\times(a_{\mathrm{Eosine}})^2/(a_{\mathrm{Deuterium}})^2=1.62\times10^{-9}\ \mathrm{m^2/s} \tag{8.2}$$

The molecular diffusion coefficient of bromide in free water resulting from Eq. 8.1 is equal to 1.4×10^{-9} m^2/s, which reasonably agrees with the value of 1.5×10^{-9} m^2/s known from the literature. The value of $(D_p)_{\mathrm{Bromide}}$ was used for the interpretation of rock parameters from the bromide concentration curves measured in the springs A1–A4. For eosine, the coefficient of molecular diffusion in free water was assumed to be 0.45×10^{-9} m^2/s, which yields $D_p=3.0\times10^{-10}$ m^2/s for this tracer. The retardation factor R_{ap} for eosine in the porous matrix is 5.4, as estimated from Eq. 8.2.

Table 1 Values of fitting parameters for test A, obtained by applying the SFDM to the tracer-concentration curves measured in the HKLU borehole

Tracer	t_o [d]	Pe [–]	a [$s^{-1/2}$]
Deuterium	0.85	6.7	4.32×10^{-3}
Bromide	0.85	6.7	3.23×10^{-3}
Eosine	0.85	6.7	4.25×10^{-3}

Observations at springs A1–A4

The modeling of the bromide concentration curves measured in the springs A1–A4 using the SFDM results in nearly the same values of fitting parameters, as summarized in *Table 2*. The best fit curve for bromide at

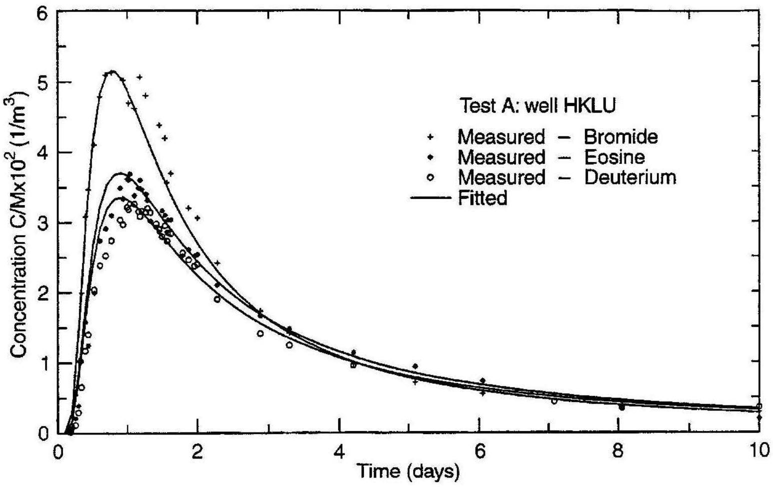


Figure 4 Normalized (C/M) bromide, deuterium and eosine concentrations measured in the well HKLU and the SFDM-fitted curves (test A; for parameters, see *Table 1*)

Table 2 Values of fitting parameters for test A, obtained by applying the SFDM to bromide concentration curves measured in four springs, and calculated values of the mean transit times of water (t_o) and water velocity (v) in the fractured zone between the HKLU and springs (cf. *Figures 2* and *3*)

	Fitted			Calculated	
Spring	$(t_o)_s$ [d]	Pe [–]	a [$s^{-1/2}$]	t_o [d]	v [m/d]
A1	1.9	67	4.08×10^{-3}	1.05	190
A2	1.8	100	4.08×10^{-3}	0.95	223
A3	1.8	100	4.08×10^{-3}	0.95	238
A4	2.0	100	3.74×10^{-3}	1.15	180

A3 is shown in *Figure 5*. From the transit time of water $(t_o)_{HKLU}$ obtained for well HKLU (*Table 1*) and the transit time of water $(t_o)_S$ obtained for the springs (*Table 2*), one can separate the transit time of water (t_o), and the mean water velocities (v) for the flow distance between the HKLU and the spring as follows:

$$t_o = (t_o)_S - (t_o)_{HKLU} \quad (9.1)$$

$$v = X/t_o \quad (9.2)$$

In *Table 2* are compiled the parameters of the SFDM, the calculated transit times of water (Eq. 9.1), and the water velocities in the fractured aquifer (Eq. 9.2). The mean water velocity between the HKLU and springs is about 208 m/d. This value, combined with the estimated hydraulic gradient $i = 0.16$, implies $k_s = k/n_f = 1.5\times10^{-2}$ m/s (Eq. 6), which yields a fracture aperture of $2b = 232$ μm (Eq. 7). If the fault zone is not represented by a single fracture but by multiple fractures, the hydraulic conductivity depends on the assumed value of the fracture porosity within the fracture zone.

By putting the diffusion coefficient for bromide obtained from Eq. 8.1 and the known fracture aperture into Eq. 5.3, the matrix porosity of the fracture zone is about $n_p = 3.0\%$. That value reasonably agrees with 2.3% obtained from laboratory measurements on samples of unaltered rock (Motyka 1992).

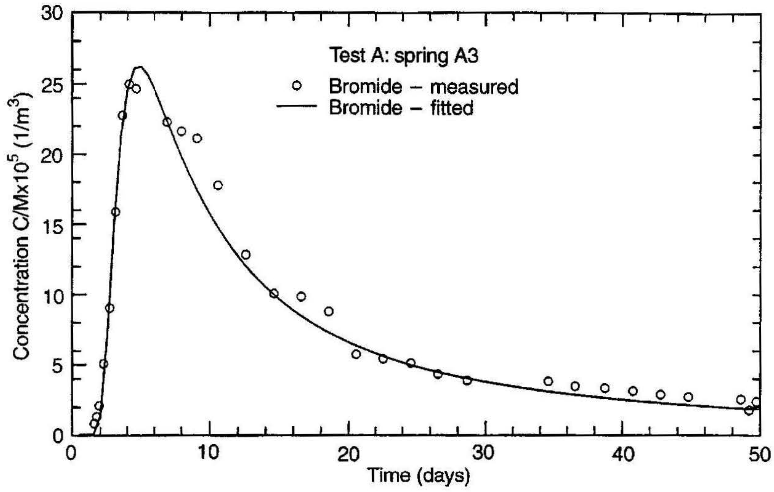


Figure 5 Normalized (C/M) bromide concentrations measured in spring A3 and the SFDM-fitted curves (test A; for parameters, see *Table 2*)

Test B

Observations at spring A3

The injection of tracers in well HKLU was monitored only in spring A3. Normalized (C/M) eosine and uranine concentrations with the curves of the best fit are shown in *Figure 6*. Values of the fitting parameters and the corresponding mean water velocities and dispersivities are compiled in *Table 3*. Results indicate that the transport parameters (t_o and Pe) agree well for both tracers. The different values of the diffusion parameter for eosine and uranine may result from different adsorption properties of the two tracers in the microporous matrix. The retardation factor for uranine is 11.6, based on considering the values of a-parameter in a way similar to that described by Eqs. 8.1–8.2, assuming that both uranine and eosine have the same diffusion coefficients, and taking $R_{ap} = 5.4$ for eosine. Thus, uranine was more strongly adsorbed than eosine, which is rather unusual. The water velocity to A3 is $v = 281$ m/d. Using a hydraulic gradient between HKLU and A3 of 0.18, the hydraulic conductivity ($k_s = k/n_f$) is 1.8×10^{-2} m/s, which yields a fracture aperture of about 254 μm. For the a-parameter for eosine and $D_pR_{ap} = 1.62\times10^{-9}$ m^2/s, as known from test A, the matrix porosity of 2.1% can be obtained from Eq. 5.3. Transport and rock parameters obtained by applying the SFDM for the interpretation of both tests are compiled in *Table 4*.

For the parallel fracture approximation, the fracture porosity is $n_f = 2b/L$, where L is the axis-to-axis distance between the fractures. The mean value of L is known from direct observations on exposed rock surfaces to be about 0.15 m (Herrmann et al. 1989). That equation, combined with Eq. 7 can be solved for $2b$ and n_f yielding, for regional hydraulic conductivity (Eq. 1) $2b = 55$ μm, and $n_f = 4\times10^{-4}$. These values provide a sense of the possible orders of magnitude of fracture parameters in the rock massif as a whole.

Figure 6 Normalized (C/M) eosine and uranine concentrations measured in spring A3 and the SFDM-fitted curves (test B; for parameters, see *Table 3*)

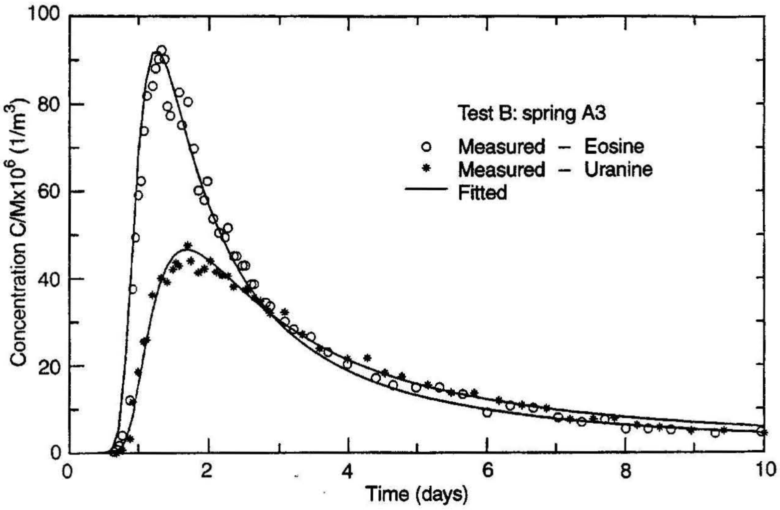


Table 3 Values of fitting parameters for test B, obtained by applying the SFDM to the eosine and uranine concentration curves measured in spring A3, and calculated water velocities (v) and dispersivities (α_L) for the fractured zone

	Fitted			Calculated	
Tracer	t_o [d]	Pe [–]	a [s$^{-1/2}$]	v [m/d]	α_L [m]
Eosine	0.8	100	3.40×10^{-3}	281	2.3
Uranine	0.8	100	4.93×10^{-3}	281	2.3

Table 4 Comparison of flow and fault parameters obtained in tests A and B for bromide and eosine along the flowpath between the borehole HKLU and spring A3 (X=225 m)

Parameter		Test A	Test B
Water velocity	v [m/d]	238	281
Dispersivity	α_L [m]	2.3	2.3
Hydraulic cond.	k/n_f [m/s]	1.5×10^{-2}	1.8×10^{-2}
Fault aperture	$2b$ [μm]	232	250
Matrix porosity	n_p [%]	3.0	2.1

Table 5 Values of fitting parameters obtained in tests A and B by applying the SFDM to the tracer-concentration curves measured in gauging station HALB

Tracer/Test	t_o [d]	Pe [–]	a [s$^{-1/2}$]
Bromide/A	0.95[a]	50	5.78×10^{-3}
Eosine/B	0.8	67	5.44×10^{-3}
Uranine/B	0.8	67	6.12×10^{-3}

[a] Corrected to the same flow length as that in test B, by applying Eq. 9.1

Observations at gauging station HALB

The SFDM was also applied to the tracer-concentration curves determined in both tests at gauging station HALB of the creek, to see if other fast-flow paths to the creek exist. The results are summarized in *Table 5*, and the best fit curves are shown in *Figures 7* and *8*. The mean transit time of water is nearly the same as for spring A3. The values of the a-parameter and the longitudinal dispersivity (X/Pe) are larger, probably due to mixing and dispersion processes in the creek.

The tracer mass recovery was examined in both experiments. The relative mass recovery (RR) is defined as:

$$RR(t) = \frac{Q}{M}\int_0^t c_f(t')dt' \qquad (10)$$

For test A, performed at a very low volumetric flow rate of the creek (5.5 L/s), the bromide recovery of 54% in HALB after 50 d agrees well with the 48% calculated from the SFDM (*Figure 7*). That result suggests that for test A, the whole mass of tracer was transported in the fractured zone to the creek. For test B, performed at a high discharge (28 L/s) and a high groundwater level, the measured recovery of eosine was only about 10%, whereas the expected value was about 68% after 20 d (*Figure 8*). This result suggests that the largest fraction of tracer was not quickly transported in the major fault but was probably temporarily stored in other connected fault zones.

Conclusions

Results of two multi-tracer tests performed at the major fault of the Lange Bramke basin, though not free of some drawbacks (only sorptive tracers were used in test B), show that for fractured rocks, the use of at least two tracers with different coefficients of molecular diffusion is necessary to obtain meaningful results.

Among the injected tracers, at least one should be conservative; otherwise, it is not possible to distinguish the influence of sorption from the influence of diffusion exchange between mobile water in fractures and immobile water in the matrix.

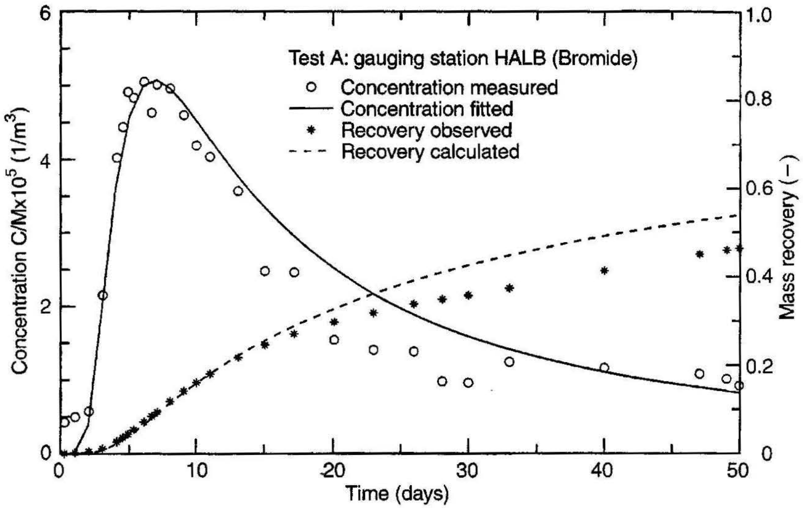


Figure 7 Normalized (C/M) bromide concentrations and recovery curves measured at gauging station HALB, showing the SFDM-fitted concentration curve and the SFDM-calculated recovery curve (test A)

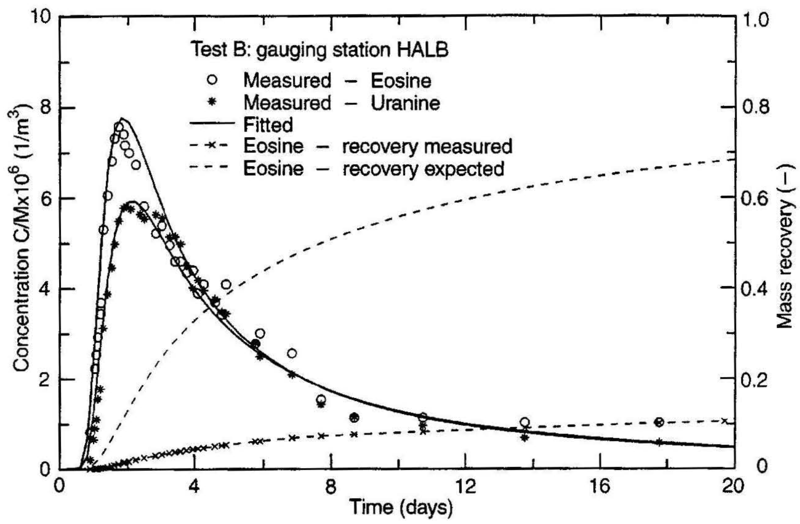


Figure 8 Normalized (C/M) eosine and uranine concentrations curves measured at the gauging station HALB, showing the SFDM-fitted curves (test B) and the observed and expected eosine mass-recovery curves

The applicability of the Single Fracture Dispersion Model to the interpretation of tracer tests in fractured rocks is demonstrated by consistent values of parameters obtained from different tracer curves. These results confirm earlier findings of Maloszewski and Zuber (1990) that matrix diffusion usually governs the transport of solutes in fractured rocks, whereas a proper representation of the fracture network is less important. The common practice of using a single tracer, or using different dye or other tracers with similar values of the coefficient of molecular diffusion, leads to incorrect interpretations, because under these conditions the reasons for tailing effects cannot be identified, and, consequently, the selection of a proper model and an unambiguous interpretation is not possible.

Results of using deuterium and bromide simultaneously with dye tracers indicate that uranine and eosine are sorbable. Therefore, tracer tests based only on dye tracers cannot be properly interpreted.

The major fault in the Lange Bramke basin is the principal pathway for the fast groundwater flow, as indicated by the flow and rock parameters that were obtained. The fault-zone parameters are substantially greater than those deduced from the environmental-tracer dating for the basin as a whole.

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References

Cacas MC, Ledoux E, de Marsily G, Barbreau A, Calmels P, Gaillard B, Margrita R (1990) Modeling fracture flow with a stochastic discrete fracture network: Calibration and validation, 2, The transport model. Water Resour Res 26:491–500

Cady CC, Silliman SE, Shaffern E (1993) Variation in aperture estimate ratios from hydraulic and tracer tests in a single fracture. Water Resour Res 9:2975–2982

Herrmann A, Koll J, Leibundgut Ch, Maloszewski P, Rau R, Rauert W, Schöniger M, Stichler W (1989) Wasserumsatz in einem kleinen Einzugsgebiet im paläozoischen Mittelgebirge. Eine hydrologische Systemanalyse mittels Umweltisotopen als Tracer (Turnover of water in a small catchment area of Paleozoic rock. A hydrological system analysis by using environmental isotopes as tracers). Landschaftsökologie und Umweltforschung 17:1–305

Himmelsbach T, Maloszewski P (1992) Tracer tests and hydraulic investigations in the observation tunnel Lindau. In: Fractured rock-test site Lindau/southern Black Forest (Germany). Steir Beitr Hydrol 43:197–222

Maloszewski P (1994) Mathematical modelling of tracer experiments in fractured aquifers. Freiburger Schriften zur Hydrologie 2:1–107

Maloszewski P, Zuber A (1985) On the theory of tracer experiments in fractured rocks with a porous matrix. J Hydrol 79:333–358

Maloszewski P, Zuber A (1990) Mathematical modeling of tracer behaviour in short-term experiments in fractured rocks. Water Resour Res 26:1517–1528

Maloszewski P, Zuber A (1993) Tracer experiments in fractured rocks: Matrix diffusion and the validity of models. Water Resour Res 29:2723–2735

Motyka J (1992) Institute of Hydrogeology, Academy of Mining and Metallurgy, Cracow, Poland (private communication)

Novakowski KS, Evans GV, Lever DA, Raven KG (1985) A field example of measuring hydrodynamic dispersion in a single fracture. Water Resour Res 21:1165–1174

Raven KG, Novakowski KS, Lapcevic PA (1988) Interpretation of field tracer tests of a single fracture using a transient solute storage model. Water Resour Res 24:2019–2032

Silliman SE (1989) An interpretation of the difference between aperture estimates derived from hydraulic and tracer tests in a single fracture. Water Resour Res 25:2275–2283

Shapiro AM, Nicholas JM (1989) Assessing the validity of the channel model of fracture aperture under field conditions. Water Resour Res 25:817–828

Sudicky EA, Frind EO (1982) Contaminant transport in fractured porous media: analytical solutions for a system of parallel fractures. Water Resour Res 18:1634–1642

Zuber A, Motyka J (1994) Matrix porosity as the most important parameter for solute transport at large scales. J Hydrol 167:19–46

