DOUBLE POROSITY MODELS IN KARSTIFIED LIMESTONE AQUIFERS: FIELD VALIDATION AND DATA PROVISION

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ABSTRACT

Modelling results and long-term field investigations from a karst aquifer (Upper Jurassic) in southwest Germany are presented, providing further evidence and support for the Double Continuum Ground Water Modelling approach in karstified limestone areas. With water flowing at different rates through a karst aquifer, slowly through small fissures and very rapidly through large fractures and solution enlarged channels, the simulation of the flow and transport requires information on both systems and their interaction. This information is provided by investigations on a catchment scale, an intermediate and on a borehole scale. Time series analysis of rainfall events utilizing well hydrographs, spring discharge, $^{18}O/^{16}O$ ratios, temperature, electrical conductivity and turbidity of the water allow the discrimination of not only event and pre-event water, but also the quantitative distinction between the fast and the slow component of the storm derived water. The latter is an essential calibration tool for the interchange between the two systems in a double porosity model and the $^{18}O/^{16}O$ ratio serves as a regional tracer, the breakthrough of which could be modelled. Further transport parameters were provided by two different kinds of tracer tests, with the input in sinkholes (information on fast system), as well as in boreholes ("statistical" information on slower system). Hydraulic parameters were derived from slug tests. The attempt to model the aquifer with a standard porous medium model failed, primarily as a result of a different response from the two systems. Whereas peak discharge generally occurs two to three days after the storm event, the maximum in the well hydrographs lags behind by more than one month. A satisfactory representation of both the flow and the transport processes could only be achieved with a regional double porosity model, allowing flow in both the "fracture porosity" and the "matrix-porosity".

INTRODUCTION

A growing concern with groundwater contamination through agricultural chemicals and potential spills of hazardous materials calls for some method of prediction and possible tool for remedial measures. This is especially applicable in the case of a karst aquifer, which is known to be highly vulnerable to contamination. However, the lack of spatial knowledge of the karst parameters, the relative unpredictability and the extreme heterogeneity in aquifer properties has often discouraged researchers from attempting to model such aquifers.

This paper attempts to demonstrate the applicability and practical use of a double porosity model in a karst environment for the simulation of flow and regional transport. Issues concerning the provision of the specific data input and aquifer parameters specifically designed for a karst aquifer are also addressed. Special emphasis is laid on formulating methods for the quantification of the input data, required for the model simulation.

AREA OF INVESTIGATION

For the investigations, the spring catchment of the Gallusquelle was selected. It is situated in southwest Germany on the Swabian Alb, a small mountain range that stretches
in an approximate southwest-northeast direction for roughly 200 km (Figs 1 and 2). Morphologically, the project area dips gently from an escarpment (1000 m a.s.l.) in the northwest down to about 600 m in the region of the spring.

The Gallusquelle groundwater basin forms a part of the catchment (450 km$^2$) of the Lauchert River, which is fed mainly by karst springs. The River Fehla represents the boundary to the northeast. Approximately 60% of the catchment area is covered by spruce, the remaining land is predominantly agricultural. The area is well suited for the intended measuring programme. More than 25 years of continuous records of discharge, water-level fluctuations (weekly readings), and climatic data were available, as well as data from about 20 wells that allowed the construction of a water level map, even in areas far removed from the discharge point. Fifteen tracer tests helped to delineate fairly accurately the catchment boundaries which cover an area of approximately 45 km$^2$.

Geology

Geologically, the area is composed exclusively of carbonate rocks of the Upper Jurassic (Fig. 3). At the surface, predominantly the massive limestones of the Kimmeridge 2/3 (ki2/3) are exposed, which reach a maximum thickness between 90 and 140 m. They are underlain by the Kimmeridge 1 (ki1), a marly limestone sequence, with more or less expressed bedding and a thickness of approximately 50 m. The lowest relevant geological unit consists of well bedded limestones, the Oxford 2 (ox2). The whole stratigraphical succession dips southeast. The southwest border of the catchment is formed by the Hohenzollern Graben, which is tectonically still active. Another fault zone strikes north-south and borders the project area in the southeast.

Hydrogeology

As shown in Fig. 3, the aquifer is formed by three geological units, the massive
limestone (ki2/3) in the southeast, the marly limestone (ki1) in the centre and the Oxford 2 in the northwest. The aquifer base does not follow any stratigraphical boundary. It can be deduced from the geological logs of the boreholes, that the bottom of the active part of the aquifer runs parallel to the water table, with saturated thicknesses estimated to range between 0 and approximately 30 m, depending on the season, although the limestone sequence may be in excess of 150 m.

An increased gradient of the piezometric surface can be observed in the centre of the catchment (Fig. 2), where the water table cuts across the less permeable marly limestones. Further upgradient, the aquifer comprises of Oxford limestones. The water table constitutes the top of the aquifer. Unconfined conditions prevail in the entire catchment. The unsaturated zone is highly karstified and reaches thicknesses between 90 and 120 m.

The lateral boundaries of the catchment could be derived from numerous tracer tests; in the lower course of the River Fehla, however, it could not be clarified, whether the river is in direct contact with the aquifer or just perched above the water table. Closer to the spring, annual water level fluctuations range between 5 and 15 m and further upgradient,
between 10 and 30 m, reflecting the decrease in transmissivity and storage coefficient away from the point of discharge.

MODELLING APPROACHES IN FRACTURED AND KARSTIFIED ROCKS

Two fundamental approaches to modelling karst water resource systems have been taken: the black box model and distributed parameter model. As a result of the scarcity of spatial data, the heterogeneity of the aquifer parameters and because of its relative simplicity, a black box approach has been frequently preferred in the simulation of karst aquifers. Despite its global approach, encouraging results have been obtained by recession analysis (Atkinson, 1977), the identification of transfer/kernel functions (Dreiss, 1989b; Aiguang et al., 1988; Avias & Joseph, 1984) and also through simple regression analysis (Zaltsberg, 1984).

The inadequacies of the black box models become apparent when one attempts to model spatially variable output phenomena, like characteristic water level fluctuations, that have a definite physical basis. Frequently, geological information that could explain observed differences which in many cases is of a spatial nature has to be ignored in such models. Moreover, they fail to consider the different processes that determine flow and transport in a karst aquifer, i.e. the mechanism of groundwater recharge, the influence of the unsaturated zone and the phenomena in the aquifer itself. Furthermore, each of these factors in turn has a different influence on fast and slow flow components.

A viable alternative is the distributed parameter modelling approach. Three major methods have been used to describe the flow and transport through fractured porous media. When the fractures are narrow, evenly distributed, and if there is a high degree of connectivity an equivalent porous medium model can be applied (Pankow et al., 1986). A recent development for this kind of model has been described by Neuman (1987). The author simulates the flow in the fractured porous rocks with an equivalent porous medium model, integrating the permeability as a stochastic variable.
The second approach, the discrete fracture model implies that the effect of the matrix is neglected. Flow is simulated by considering it as flow between two parallel plates and it requires some detailed information on fracture apertures, density, orientation and connectivity (Snow, 1965; Irmay, 1984; Romm, 1966). Especially as it is very difficult, if not often impossible to obtain the required information, the fracture networks have been statistically simulated (Long et al., 1985; Smith et al., 1987).

The modelling of flow and transport in fractured rocks could benefit greatly from studies carried out by petroleum engineers in the field of reservoir engineering and from the research efforts in the search for safe repositories for radioactive wastes. The preferred approach in these areas of research has been the double porosity approach (Barenblatt et al., 1960; Warren & Root, 1963; Duguid & Lee, 1977). The fractured medium and the porous matrix blocks are modelled as two separate overlapping continua, each with its own flow equation. The coupling of the two media is handled with a source/sink term in each equation. The exchange of flow is controlled by the local difference in potentials. The applicability of the distributed parameter approach to karst aquifers was described in detail by Thrailkill (1986) and it's appropriateness to conduit-flow aquifers in particular was examined. Kiraly (1984), for example, could successfully apply a finite element model, representing the matrix blocks with a porous medium approach, with a superimposed discrete regular network of line elements, simulating the fast system.

For the work described, a double porosity model was chosen. It is considered suitable for the aquifer concerned, which is located at an intermediate position, possibly more towards the diffuse end member of the conduit-diffuse-flow spectrum (White, 1977), and where the physico-chemical characteristics of the spring water and tracer tests indicate that conduit flow is still an important factor. The discrete fracture approach is inapplicable, because the required data input cannot be obtained and the parallel plate assumption does not hold.

MODEL INPUT DATA PROVISION

This section demonstrates how the data for the double porosity model were obtained. The approach taken is in accordance with the concept forwarded by Smart & Friederich (1986). The aquifer system is decompartmentalized into the processes of recharge, storage and transmission, the influence of each of them is examined and the resulting responses explained in terms of the hydraulics of the aquifer and spring water chemistry. The aim is to quantify each of these phenomena, as well as to describe, where applicable, their variability with time using time series analysis.

Recharge evaluation

Groundwater recharge was computed on a daily basis, applying a soil water balance approach. Potential evaporation was calculated with the Haude (1955) method, which requires air humidity and air temperature at 14:00 as input variables. The resulting values are corrected for the varying transpiration by plants. The soil moisture balance approach was carried out according to Uhlig (1959), a procedure similar to that of Thornthwaite & Mather (1957), applying a field capacity (measured) of 70 mm. However, due to the high percentage of woodland, interception becomes important, which is not corrected for in the Haude/Uhlig method. Therefore, with the exception of drought years, a systematic overestimation of annual recharge was obtained, even after varying the field capacity within acceptable limits. A satisfactory agreement (error 10%) between the computed and the measured (cumulative spring discharge, with $\Delta S = 0$) annual recharge (period 1965-1990) could be obtained by changing the Haude factors according
to Sokollek (1983) and using a canopy interception capacity of 4.7 mm; for rainfall events below 12 mm, interception was taken as a percentage of 35% of the total daily rainfall (Benecke, 1978). The storage in snowfall was accounted for and the release of snowmelt followed a simple relationship, using a degree day factor of 6 mm.

Regarding the mode of entry, it can be shown that a high percentage (approximately 90%) of the recharge reaches the water table in the form of dispersed flow, and the remaining portion via shaft flow, which is collected in surface depressions. This figure can be substantiated by the variations in spring water chemistry examined below. Recharge, slowly released from the epikarstic zone via subcutaneous drains (Williams, 1983; Smart & Friederich, 1986) falls into the category of dispersed flow. Neither sinking streams nor impermeable cap rocks were observed.

Storage

Storage can be categorized into unsaturated and saturated storage, each of them can again be subdivided into different reservoirs, the soil zone and the epikarst (subcutaneous zone), the conduit storage and the water stored in microfractures and pores. It can be assumed that the thin cover of orthic luvisols is unable to hold large quantities of water. Even after minor rainfall events, the soil is at field capacity. Soil physical examination (Hemme, 1970), suggests a field capacity varying between 30 and 70 mm. Only in depressions, where brown earths prevail, does field capacity reach 180 mm.

The storage capacity of the subcutaneous zone is difficult to assess. Through repeated borehole logging, temperature variations have been observed in the top 5 m of the phreatic zone (100 m b.g.l.) that are closely related to the surface air temperature, even after prolonged dry periods. The only possible mode of heat transfer could be by water, temporarily stored in the unsaturated zone. These variations suggest that this type of recharge must be an important quantity, otherwise, it would have been buffered by the karst water stored in the saturated zone.

Ongoing research is attempting some quantitative estimate of this type of recharge. After some recharge events, especially after prolonged dry periods, it was observed that electrical conductivity of spring water increased for a short period followed by a rapid decrease. According to Ashton (1966), the water with the higher conductivity represents so-called deep phreatic water as a result of the longer reaction time between water and rock. If however, this type of water is interpreted as water from the subcutaneous zone (Williams, 1983), which is displaced by event water and discharged only via the conduit system, some estimate of the stored quantity can be made.

Storage in the saturated zone can be subdivided into conduit and diffuse storage (Atkinson & Smart, 1981; Shuster & White, 1971; White, 1977). Conduit storage could be evaluated with the approach suggested by Williams (1983), which is a modification of Ashton (1966). The values obtained this way varied between 0.01% and 0.02%. Another method employed tracer tests, whereby the dye was injected into a doline connected to preferential flow paths (Fig. 4). The volume of groundwater discharged from the time of injection until the arrival of the dye, divided by the volume of the saturated rock produced the figure of 0.0002 ± 0.0001 for the conduit storage. The storage of the whole aquifer system, which consists mainly of diffuse storage, could be similarly evaluated by dividing the water volume discharged by the volume of the aquifer drained. The drained rock volume could be calculated using the water level changes in the boreholes. A fairly constant value of about 2% could be determined within the main aquifer body. This value is somewhat lower at very high water levels.
Flow within the aquifer

On the scale from conduit to diffuse flow, the aquifer investigated can be placed in an intermediate position (mixed-flow). Whereas the water quality parameters and the discharge react "flashy", and the response to storms is still very rapid, the determining part of the aquifer really is the matrix blocks, that store and release the water over several months. Nevertheless, it is understood that the drainage is still via a tributary network of conduits. As can be shown below, following storms, approximately 10% of the total flow reaches the spring directly via the conduit network.

Hydraulic conductivity has been frequently determined in the context of water resources developments. The values obtained range between $1 \times 10^{-4}$ m/s and $2.6 \times 10^{-3}$ m/s. Due to the higher probability of success and the lower costs for shallower boreholes, all the tested wells were drilled in valleys, and the values therefore constitute an overestimation, taking the aquifer as a whole.

Alternative approaches were taken by Villinger (1977). Attributing the average flow rate (500 l/s) to three areas of equal gradient (close to the spring, increased gradient in the centre of the project area, area near the groundwater divide in the northwest) and assuming a throughput area, Darcy's law can be used to calculate an average hydraulic conductivity for each of the sub-areas. The values varied from $2 \times 10^{-4}$ m/s in the centre, to $11 \times 10^{-4}$ m/s near the spring.

The low conductivity end of the spectrum could only be evaluated from boreholes further upgradient. Because the unsaturated thickness is very high (100 m) and the 3" casing cannot accommodate powerful pumps, it was impossible to conduct pumping tests. Instead, slug tests and constant rate injection tests were carried out. The injection tests were evaluated according to Bourdet & Gringarten (1980), an analytical type curve matching procedure, that assumes that the aquifer can be represented with the double porosity approach and which allows the determination of skin and wellbore storage effects.

The resulting hydraulic conductivity values ranged between $0.3 \times 10^{-5}$ and $4 \times 10^{-5}$ m/s (whole system). No analytical method is available to evaluate slug tests where double
porosity effects have been observed (Fig. 5) and where the boundary conditions are complicated by partial penetration (Dougherty & Babu, 1984; Barker & Black, 1983). The tests were therefore simulated with the TRAFRAP-WT code (Huyakorn et al., 1983) using a double porosity approach. This helped to account for geometrical factors and the wellbore storage. Assuming that the fractures represent 0.1 vol% of the tested area, hydraulic conductivities of $2 \times 10^{-3}$ for the fractures were obtained.

The matrix hydraulic conductivity varied between $10^{-5}$ m/s and $10^{-8}$ m/s. Taking into account the relative volume percentages of the fractures and the matrix, values similar to those evaluated according to the Bourdet & Gringarten (1980) method could be obtained.

![Fig. 5 - Recovery curve of a slug test, displaying double porosity effects.](image)

**Dynamic analysis of recharge and aquifer flow**

In order to be able to describe the flow mechanism and the temporal distribution of the slow and fast components of recharge and the flow within the aquifer, time series of a number of recharge events have been analysed. The parameters shown in Fig. 6, displaying the event of April 1989, have been measured. In the majority of events, it could be observed that the peak discharge occurs two to three days after the rainfall event, whereas the actual event water reaches the spring approximately 1.5 days later, which implies that peak discharge consists mainly of pre-event water, as has also been described by Williams (1983) and Dreiss (1989a). The maximum water level in the aquifer was read about five weeks after the event. The highest values in spring water turbidity were observed contemporaneously with peak discharge, i.e. the turbid water consists mainly of pre-event water.

The groundwater hydrographs show a response that suggests that the two distinct flow systems exist. Some intermediate maxima could be measured in both boreholes simultaneously with peak discharge. These might reflect the maximum potential in the fracture system, which dissipates quickly, due to the inherent high hydraulic conductivity. The bulk of the recharge water eventually arrives, as described above, five weeks later at the water table, delayed by the thick unsaturated zone (100 m) and the
lower hydraulic conductivity of the aquifer matrix. The turbidity is probably the most suitable parameter to distinguish between the fast and the slow event water component. In order to keep the clay particles in suspension, a minimum of hydraulic energy and turbulence is required, a criterion which allows the distinction between the two flow components. It also enables the determination of the time when the fraction of the fast water becomes zero.

An analysis of the electrical conductivity variations allows a semi-quantitative determination of the total fraction of event water transmitted in the fast system. It is assumed that fast percolating water does not take up any dissolved constituents on its 1.5
day passage through the unsaturated zone, whereas recharge water (slow component), in contact with the carbonate rocks for more than a month, approaches the electrical conductivity of the groundwater by the time it reaches the water table; the latter statement could be verified by continuous measurements in a cave over a period of about three months.

Assuming an electrical conductivity for the fast recharge of 50 $\mu$S/cm, which is a rather conservative estimate, and applying established component separation methods (Sklash & Farvolden, 1979; Pinder & Jones, 1969), then the fast component amounts to approximately 2.5% of the total event water.

Comparing the spring water temperature record with the electrical conductivity, the same assumption about the slow water cannot be upheld. Even after a period of 2.5 months, the spring water temperature still reveals some memory effect of the previous event. Therefore, an attempt to separate the slow and the fast components was made, using a straight line recession method (plot of temperature against log time, Fig. 7).

Fig. 7 - Temperature changes attributed to the different flow components.

With the assumption that air temperature at the end of the rainfall event corresponds to the input temperature, a figure of 12% for the fraction of the fast component can be obtained, using the same component separation method as above (Fig. 8).

A common approach to distinguish between two components is the simple recession analysis, assuming a distinct hydraulic response for each of the two systems (Maillet, 1965; Atkinson, 1977). Whereas the previously described methods quantify the event water components directly, i.e. the spring water analysed consists of event water, the recession analysis considers the change in discharge as a result of the pressure pulse induced by the new recharge water, i.e. the change in discharge reflects the arrival of the fast event water component. This feature is very useful, because it helps to describe the time variation in the input to the groundwater system (Fig. 9). The fraction of the fast component is evaluated at about 15%, assuming that the initial drop in the recession limb describes the hydraulic characteristics of the fast system over the whole aquifer thickness.
The above described methods would enable the quantification and the temporal distribution of the input of the fast component in the model, if the recharge water remained unchanged in temperature or chemical composition during its passage through the unsaturated zone.

However, these methods were all based on non-conservative tracers, and although they can describe the extremes and the time variation of the input, they are still somewhat arbitrary. Therefore, one further parameter was analysed, the relative abundance of the $^{18}$O isotope in groundwater and rainfall (Hess & White, 1974; Bakalowicz & Mangin, 1980). Spring water samples were taken automatically at 16-h intervals as well as daily rainfall at five stations.

The results are plotted in Fig. 10, together with the rainfall input. The figure contains two sets of information. Squares and crosses represent values of the spring water,
Fig. 10 - Oxygen isotope variations in rainfall and in spring water.

whereas the bars indicate the deviation of the $\delta^{18}O$ in the rainfall from the long-term spring average of $-10.55\%$. The shading of the bars represents the weighting of the input, corresponding to the rainfall height.

The spring data are represented by two different symbols, because it is suspected that the water discharged during abstraction periods might be slightly different in its chemical composition. The rainfall isotope data are shifted by 4.5 days to account for the observed time lag.

The dotted line indicates an interpretation of the expected bimodal breakthrough of the $\delta^{18}O$ as a response to the event of 1 April, the first peak indicating the arrival of the fast component, the second, the delayed slow event water component. The rainfall input was evaluated at $-9.5\%$. The spring water analyses display quite a pronounced scattering, which is partly due to the analytical error of $0.15\%$.

The largest variations, however, can be explained by the interferences of other rainfall events: one event on the 15-16 April, with a fairly negative input of around $-13\%$ and an event at the end of April, with values reaching almost $-17\%$. The arrival of the fast component of the latter event water probably coincides with the slow component of the event of 15-16 April, as indicated by the temperature residual in Fig. 7.

Applying the same component separation technique (mass balance of constituents) with an input $\delta^{18}O$ of $-9.5\%$ only about one third of the total event water (integral underneath the discharge curve at $\Delta S = 0$) could be recovered. The input $\delta^{18}O$ had to be varied by up to $-10.2\%$ in order to balance both totals. In this case, the fast water component amounts to about $12\%$.

This observation indicates that even the $\delta^{18}O$ is subject to change, in the present case probably due to mixing processes in the epikarstic and the vadose zone. Previous (two weeks) rainfall $\delta^{18}O$ varied between $-11\%$ and $-13\%$. The same applies to component
separation with electrical conductivity, that produced an underestimate of the fast component fraction as a result of mixing with higher conductive waters of the epikarstic zone.

The temperature of the spring water probably behaves like a more conservative tracer, as far as the fast component is concerned, because the temperature stored within the epikarst adjusts itself more to the ambient surface temperature. In sum, a proportion of approximately 10% for the fast component can be assumed with some confidence. The isotope data do not only provide an estimate of the fractions of the two recharge components, but also provide the output function for the regional transport of dissolved constituents.

**Temporal distribution of groundwater recharge**

The input to the system at the water level can be derived from hydraulic considerations for both components. The pressure pulse due to the fast water, analysed above (Fig. 10) is closely related to the spring discharge. The arrival of the slow component, delayed by the flow through the thick unsaturated zone can be described by comparing spring discharge and groundwater hydrographs (Fig. 11).

![Fig. 11 - Evaluation of the temporal distribution of the slow event water components (seepage water).](image)

The rate of seepage water (slow component) is higher than the discharge during periods of rising water level and lower when the potentials drop. At times of no head change, both quantities are equal. Together with the relative fractions of the two components and the distribution with time, the recharge input and the inputs of dissolved constituents can be described (Fig. 12).

**MODELLING REGIONAL FLOW AND TRANSPORT**

After evaluation of the hydraulic and transport input and output functions for the two flow components and estimation of the aquifer parameters, modelling of the aquifer system can be attempted.
Flow modelling

The model selected is the one described by Teutsch (1989). It is a one-dimensional finite difference double porosity model that allows flow in both continua. The code was extended to allow for the necessary selective recharge into both continua, because only then, could the distinct response of the two systems (different recession characteristics, the distinct bimodal output function, intermediate maximum in the hydrographs) be explained.

This type of recharge distribution also corresponds to the physical process, that water is collected at the surface in depressions, providing a concentrated fast input, that exceeds the proportional area of fractures and conduits at the surface. The model was also modified to allow regional and time variable input and it consists of two times 22 cells. The simulation run extended over a period of 25 years, and the spring flow and the groundwater levels could be modelled reasonably well (Figs 13 and 14).

Only for the last two years, for which continuous records exist and are displayed, are there some differences in the measured and modelled flow that are mainly a result of the recharge input, that was calculated based on the rainfall of one station, assumed to be representative for the whole catchment and due to the uncertainty in the release of snowmelt from store. The differences between the modelled and the simulated hydrographs can be explained by the homogeneity of the storage coefficient in the vertical dimension that had to be assumed in the model. At the level of the observed kinks in the field hydrograph, a change in storage coefficient is suspected.

The parameters with which the fit was achieved are:

- Storage coefficient (matrix): 0.01 to 0.03
- Storage coefficient (conduits): 0.0004
- Hydraulic conductivity (matrix): $1 \times 10^{-4}$ to $5 \times 10^{-3}$ m/s
- Hydraulic conductivity (conduits): $1 \times 10^{-2}$

The lower values are generally further upgradient. The somewhat higher values for the hydraulic conductivity compared with the field values, are a result of the fact that a
constant saturated thickness of 30 m had to be assumed in the evaluation of a regional average hydraulic conductivity. The exchange parameter does not have a physical meaning in the model; the relative magnitude of the hydraulic conductivity of the matrix blocks and the exchange term can only be fixed by running the transport model (see below).

Fig. 13 - Comparison of modelled and measured spring discharge.

Fig. 14 - Comparison of modelled and measured water level fluctuations.
At low flow conditions, the observed regional steep gradient could only be maintained, if the hydraulic conductivity was reduced to one fifth of the above value at a level of about 10 m above the base of the model aquifer. This measure also prevented some model nodes from drying up, which, although physically realistic, would produce unrealistic results in a one-dimensional model.

Transport modelling

With the flow calculated by the model for the period from April to July 1989, and assuming the recharge distribution as a tracer input function, the relative change in concentration of a regional tracer was simulated with a random walk approach (Teutsch, 1989). The breakthrough curve obtained compares reasonably well with the observed change in the δ¹⁸O ratio (Fig. 15).

The porosities used corresponded with the respective storage coefficients of the flow model. A dispersivity of 15 m was used for the transport in the matrix continuum, a value which could be obtained from the tracer tests. The transport model also allowed the calibration of the hydraulic conductivity of the matrix blocks, because if the matrix/fracture exchange term chosen is too high, the tracer left the system even before the beginning of May, which does not correspond to what has been observed. With the same model, it was also possible to simulate the tracer test "Bitz", for which the bulk of the tracer was injected directly into the conduits via a sinkhole.

The "B8" test (Fig. 16), however, was somewhat problematic to simulate. Due to the fact that the dye was injected directly into the groundwater and because the tracer probably travelled for only part of the time within the matrix before reaching a major conduit, the model described is unsuitable for the simulation of this kind of transport. Matrix diffusion processes, which the model did not include, are also likely to determine the shape of the breakthrough curve.
CONCLUSION

The presented results demonstrate that even a karst aquifer with its extreme flow systems can be successfully modelled with a one-dimensional double porosity model, which can therefore provide a valuable tool in the understanding of the flow and transport processes. Some of the discrepancies between modelled and measured data can probably be overcome by including a second vertical dimension.

However, the data input requires substantial time and efforts, and the continuous data records necessary to calibrate the models are not always available. In most karst groundwater basins, large numbers of boreholes, somewhat removed from the spring are not available either.

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