

Proceedings of the
➤ **8th conference on limestone
hydrogeology**

Actes du
➤ **8^e colloque d'hydrogéologie
en pays calcaire**

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Résumé : Depuis 1971, les Universités de Franche-Comté (Besançon) et de Neuchâtel organisent alternativement des colloques internationaux rassemblant des scientifiques et professionnels de l'hydrogéologie du karst. Cette 8^e édition met l'accent sur les problèmes de prospection, d'exploitation, de vulnérabilité, de gestion et de développement durable. Les actes regroupent 61 articles d'auteurs de 21 pays d'Europe, Amérique, Asie et Afrique.

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- investigations et techniques modernes de mesures,
- contaminations et microbiologie des aquifères karstiques,
- ressources en eau et changements climatiques.

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editors

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8^e Colloque d'Hydrogéologie en Pays Calcaire

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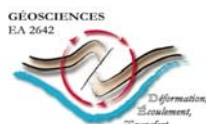
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Origine des variations de salinité sur une source karstique côtière saumâtre

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Abstract

The fresh water in coastal karstic aquifers is naturally polluted by seawater intrusion. The springs can then be brackish, thus limiting the use of this water resource. However, the water salinity generally decreases during the floods. Two phenomena explain the drop of salinity: (1) dilution by increase of the fresh water discharge, and (2) decrease of the seawater discharge by rise of the hydraulic head in the karstic conduit during the high water period. They will govern the possible exploitation mode of spring water. A new method is proposed to distinguish these two origins, using the curve of hysteresis of the seawater discharge versus salinity of the spring water. The method is illustrated by data recorded at the brackish spring of Almyros of Heraklio (Crete).

Keywords: Hysteresis, seawater intrusion, salinity, dilution, spring, Almyros

Résumé

L'eau douce des aquifères karstiques côtiers est naturellement polluée par l'intrusion saline. Les sources peuvent alors être saumâtres, limitant ainsi l'utilisation de cette ressource en eau. Toutefois, la salinité de l'eau diminue généralement lors des crues. Deux phénomènes sont à l'origine de la chute de salinité : (1) la dilution par l'augmentation du débit d'eau douce, et (2) la diminution du débit d'eau de mer lors de l'augmentation de charge hydraulique dans le conduit karstique en crue. Ils vont dicter le mode d'exploitation éventuel de l'eau de la source. Une méthode nouvelle est proposée pour distinguer ces deux origines, utilisant la courbe d'hystérésis du débit d'eau de mer en fonction de la salinité de l'eau de la source. La méthode est illustrée par les données acquises sur la source saumâtre de l'Almyros d'Héraklion (Crète).

Mots clefs : Hystérésis, intrusion saline, salinité, dilution, source, Almyros

1. Introduction

L'intrusion d'eau de mer a été démontrée par Ghyben et Herzberg à la fin du 19^{ème} siècle (Bear et al., 1999). Elle est aujourd'hui au cœur du problème de la gestion de la ressource en eau en zone côtière, et tout particulièrement dans les zones karstiques (Bakalowicz et al., 2003). L'eau de mer, plus dense que l'eau douce, entre naturellement dans les aquifères côtiers jusqu'à une certaine profondeur (z) dépendante de la charge (h) d'eau douce sus-jacente (Figure 1). Cette relation permet aussi d'expliquer la présence de sources karstiques saumâtres au-dessus du niveau de la mer si on réduit le système karstique à de simples conduits étanches (Figure 2) (Gilli et al., 2005). Dans les aquifères karstiques côtiers où les drains se développent dans une matrice fissurée, cette apparente ligne d'équilibre entre l'eau douce et l'eau de mer dans l'aquifère va être perturbée localement dans les zones de conduit karstique (Arfib, 2001; Fleury, 2005). Les sources saumâtres karstiques peuvent également être expliquées par une intrusion diffuse de l'eau de mer de la matrice entrant dans le conduit karstique (Figure 3). L'eau saumâtre circulant dans le conduit karstique se déverse finalement à l'exutoire, qui peut être une source sous-marine ou une source dans les terres.

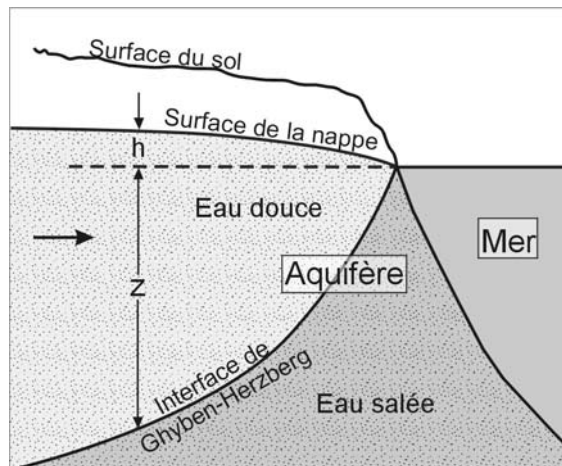


Figure 1 : Coupe schématique d'un aquifère poreux contenant une nappe libre en équilibre avec l'eau de mer entrée dans l'aquifère suivant le principe de Ghyben-Herzberg (cas hydrostatique)

Figure 1: Schematic cross-section of a porous unconfined coastal aquifer showing the intruded seawater wedge according to the Ghyben-Herzberg principle (hydrostatic)

chimiogramme et de l'hydrogramme avec les courbes d'hystérésis Débit versus Concentration (Q-C).

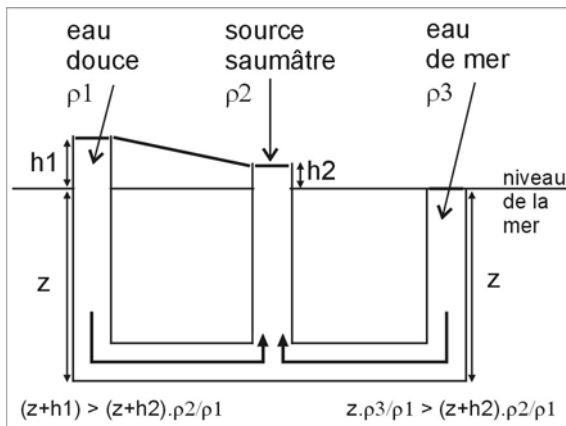


Figure 2 : Application du principe de Ghyben-Herzberg à des conduits karstiques en zone côtière (modifié d'après Gilli et al. (2005), ρ : masse volumique)

Figure 2: Application of the Ghyben-Herzberg principle to karstic conduits (modified from Gilli et al. (2005), ρ : density)

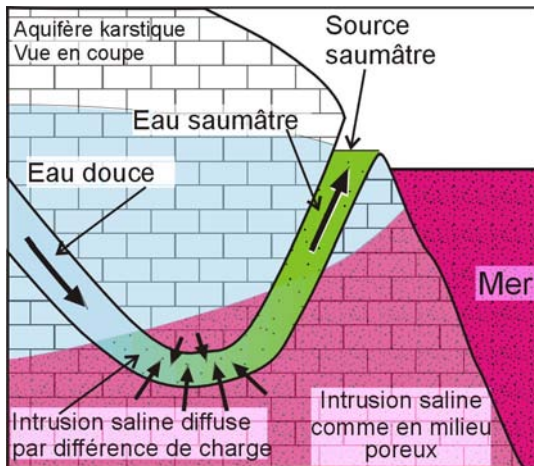


Figure 3 : Coupe schématique d'un aquifère karstique côtier dans le cas d'une intrusion saline "diffuse"

Figure 3: Schematic cross-section of a karstic coastal aquifer with diffuse saline intrusion

La salinité de la source va donc varier en fonction des proportions d'eau douce et d'eau de mer participant au mélange. L'équation de conservation de la masse du mélange s'écrit alors :

$$Q_T \cdot C_T = Q_{mer} \cdot C_{mer} + Q_{doux} \cdot C_{doux} \quad (1)$$

avec Q le débit, C la concentration (ou la salinité), et les indices "T" pour l'eau de la source, "mer" pour l'eau de la mer participant au mélange, "doux" pour l'eau douce de l'aquifère.

Dans le cas le plus classique, celui qui nous intéresse, l'évolution de la salinité est inversement proportionnelle au débit. La succession des crues caractéristiques d'une source karstique se traduit donc par une succession de chutes de la salinité. D'après l'équation 1, la chute de salinité est liée à l'augmentation du débit d'eau douce et/ou à la diminution du débit d'eau de mer participant au mélange.

Une méthode nouvelle est proposée pour distinguer l'origine de la chute de salinité à la source karstique saumâtre en crue. Elle se base sur l'interprétation graphique du

2. Matériels et méthodes

L'analyse présentée dans cet article se base sur l'enregistrement au cours du temps de la conductivité électrique de l'eau (équivalent à la salinité) et du débit de la source saumâtre karstique côtière. La mise en oeuvre de ces mesures est facilitée lors du suivi d'une source saumâtre émergeant dans les terres.

La circulation de l'eau saumâtre en conduits karstiques en charge jusqu'à la source perturbe l'interprétation des chroniques de débit et de concentration. En effet, lors d'une crue la variation de charge va se propager quasi-instantanément dans le conduit provoquant l'augmentation du débit à la source. La salinité va également varier mais avec un temps de retard proportionnel à la distance que va parcourir l'eau de salinité plus faible. Ce décalage temporel apporte de précieuses informations sur la localisation de la zone d'intrusion saline dans le conduit (Arfib, 2001a) mais ne renseigne pas sur le phénomène à l'origine de la chute de la salinité. Les courbes d'hystérésis Q-C résultantes sont alors appropriées pour faire l'analyse de l'influence du débit sur la salinité de l'eau (Evans et Davies, 1998). Valdes et al. (2005) appliquent avec succès ce type d'analyse pour étudier le fonctionnement hydrodynamique d'un aquifère karstique, en utilisant les hystérésis de Turbidité-Conductivité électrique. Dans notre cas, il s'agit d'individualiser les mécanismes engendrant la diminution de la salinité à la source karstique côtière saumâtre :

- par augmentation de la charge dans le conduit karstique qui diminue l'entrée d'eau de mer de la matrice vers le conduit karstique (diminution de Q_{mer} de l'équation 1)
- ou par dilution du fait de l'augmentation du débit d'eau douce (augmentation de Q_{doux} de l'équation 1)

Cela revient à comparer l'évolution du débit d'eau de mer (Q_{mer}) à la source en fonction de la proportion d'eau de mer participant à l'écoulement (Q_{mer}/Q_T). L'équation 1 montre que la salinité de l'eau à la source (C_T) est corrélée à Q_{mer}/Q_T . Il est donc proposé d'analyser les courbes d'hystérésis débit d'eau de mer – salinité ($Q_{mer}-C_T$) lors des crues.

3. Résultats – discussion

La méthode est appliquée sur les données enregistrées à la source saumâtre Almyros d'Héraklion (Crète, Grèce). Cette source a une salinité variable en fonction de la proportion du mélange eau douce – eau de mer, de manière inversement proportionnelle au débit. Elle a fait l'objet de recherches approfondies (Arfib, 2001a; Arfib et de Marsily, 2004). Deux ans de mesures sont disponibles, soit un total de 18 crues. C'est un aquifère karstique côtier de 305 km², drainé par la source située dans les terres à environ 1 km de la côte. La source est l'exutoire des calcaires de Tripolitza et du Plattenkalke (Figure 4). Un conduit karstique draine les eaux du Plattenkalke naturellement envahi par l'intrusion de l'eau de mer en profondeur.

La salinité de la source a été modélisée par un modèle numérique suivant le schéma conceptuel présenté en figure 3, d'acquisition de la salinité dans le conduit karstique par entrée d'eau de mer présente dans la matrice. Le milieu

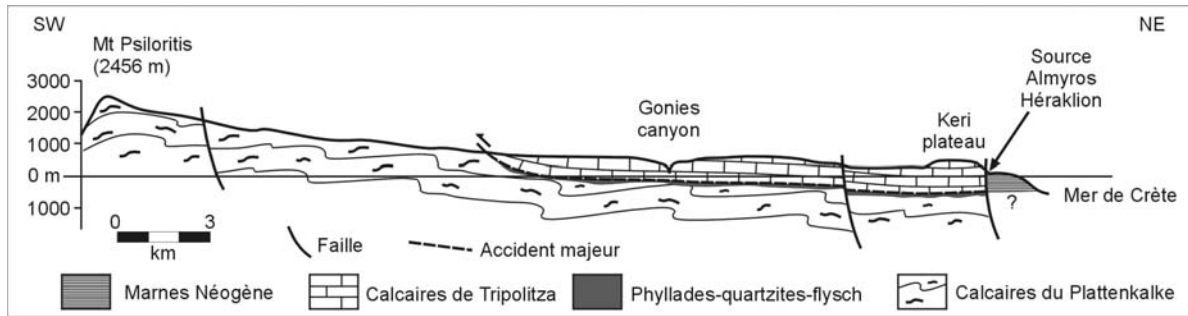


Figure 4 : Coupe à travers l'aquifère de la source Almyros d'Héraklion

Figure 4: Cross-section through the aquifer of the Almyros of Heraklion spring

calcaire est assumé équivalent à une matrice poreuse homogène traversée par un conduit karstique dans lequel l'écoulement est turbulent avec des pertes de charge quadratiques. Lors des fortes crues hivernales, la source est douce, montrant ainsi que Q_{mer} varie jusqu'à s'annuler.

La figure 5 représente la courbe Q_{mer} -Salinité, ainsi que l'hydrogramme de la crue (débit total Q_T , débit d'eau de mer Q_{mer} , débit d'eau douce Q_{doux} et proportion d'eau de mer Q_{mer}/Q_T) et le chimiogramme utilisant la salinité (C_T) sur la période du 16 au 27 janvier 2001.

La crue débute au point 1 (Figure 5) par une augmentation de débit. Jusqu'au temps n°2 (Figure 5) la salinité (C_T) n'est pas influencée par l'augmentation du débit. La courbe d'hystérésis Q_{mer} - C_T entre 1 et 2 montre une droite de pente quasiment nulle. Cette portion de courbe existe du fait de l'écoulement en charge dans le conduit karstique impliquant un décalage temporel entre l'augmentation de débit et l'arrivée de l'eau plus douce ayant à parcourir la distance qui la sépare de la source. A la source Almyros d'Héraklion, nous avons montré (Arfib, 2001b) que ce décalage temporel est observé pour les 18 crues observées, et donne lieu à l'expulsion d'un volume constant d'eau saumâtre égal à environ 770 000 m³. Ce paquet d'eau poussée a acquis sa salinité antérieurement au passage de l'onde de crue.

Du point 2 au point 3, la salinité chute à sa valeur minimale (Figure 5). La courbe d'hystérésis Q_{mer} - C_T décroît et se stabilise au point 3 à une valeur proche de zéro. Dans ce cas, l'eau de la source est douce, totalement exempte d'eau de mer. Ce phénomène est observé à l'Almyros d'Héraklion lors des crues dépassant une valeur seuil de débit supérieure à environ 15 m³.s⁻¹. L'augmentation de débit a provoqué une diminution de la salinité d'une part par augmentation du débit d'eau douce entraînant une dilution, et d'autre part par une diminution du débit d'eau de mer entrant de la matrice vers le conduit karstique du fait de l'augmentation de charge hydraulique dans le conduit en crue. Cette diminution du débit d'eau de mer entrant dans le conduit est mise en évidence par la diminution de Q_{mer} jusqu'à une valeur égale à zéro. A partir du point 3 (Figure 5) la salinité augmente progressivement avec le retour de l'eau de mer entrant dans le conduit par la baisse de charge dans le conduit en décrue. Si une autre crue n'a pas lieu, la courbe Q_{mer} - C_T reviendra au point 1.

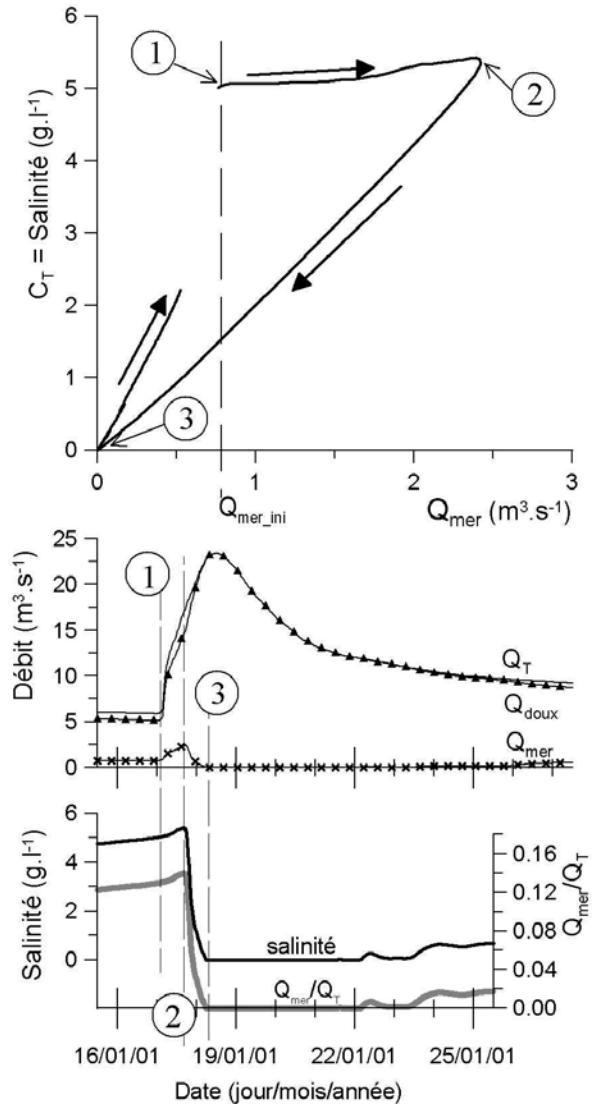


Figure 5 : Courbe d'hystérésis Q_{mer} - C_T de la crue du 16 au 27 janvier 2001 à la source Almyros d'Héraklion, et évolution au cours du temps du débit total (Q_T), débit d'eau de mer (Q_{mer}), débit d'eau douce (Q_{doux}), proportion d'eau de mer (Q_{mer}/Q_T) et salinité (C_T)

Figure 5: Hysteresis curve Q_{mer} - C_T of a flood from January 16th to 27th 2001 at the spring Almyros of Heraklion, and evolution during time of the total discharge (Q_T), discharge of sea water (Q_{mer}), discharge of freshwater (Q_{doux}), proportion of sea water (Q_{mer}/Q_T) and salinity (C_T)

La valeur du débit d'eau de mer au démarrage de la crue (Q_{mer_ini}) est une valeur critique. Elle peut être utilisée pour différencier l'origine de la chute de salinité en crue :

- Lorsque l'augmentation de charge diminue l'entrée d'eau de mer dans le conduit (cas présenté de l'Almyros d'Héraklion), Q_{mer} passera toujours sous la limite inférieure fixée par Q_{mer_ini} .

- Lorsque l'augmentation du débit d'eau douce entraîne une dilution sans diminution de l'entrée d'eau de mer dans le conduit karstique, plusieurs cas sont alors possibles :

- dans le cas où le déphasage est faible entre le débit et la salinité mesurés à la source, la courbe d'hystérésis $Q_{mer}-C_T$ restera toujours supérieure à la valeur Q_{mer_ini} .
- dans le cas inverse, Q_{mer} au point 3 sur la courbe d'hystérésis $Q_{mer}-C_T$ peut avoir une valeur inférieure à Q_{mer_ini} .

Ce type d'analyse a également été appliqué avec succès sur un autre site, la source karstique sous-marine de Port Miou (Cassis, France). Sur ce site, l'acquisition des données a lieu dans les buses d'un barrage obstruant la galerie karstique noyée 500 m avant sa sortie en mer (Potié et Ricour, 1974). Contrairement à l'Almyros d'Héraklion, il est montré à Port Miou à l'aide des courbes d'hystérésis que la diminution de la salinité sur certaines crues n'est due qu'à une simple dilution.

4. Conclusion

La méthode proposée d'interprétation des courbes d'hystérésis $Q_{mer}-C_T$ (Débit d'eau de mer – Salinité) permet de différencier les variations de salinité en crue à une source karstique côtière saumâtre en fonction de deux phénomènes :

- la dilution par l'augmentation du débit d'eau douce,
- la diminution de l'entrée d'eau de mer dans le conduit karstique alimentant la source.

Cette distinction apporte ainsi des renseignements sur les conditions de charge régnant dans l'aquifère qui régulent les échanges eau douce – eau de mer en zone côtière.

L'exploitation des sources saumâtres est quelquefois envisagée par l'intermédiaire d'un barrage à la source visant à diminuer la salinité de l'eau en augmentant artificiellement la charge. Ce type d'aménagement, en dehors des risques de perte d'eau qu'il présente, ne sera effectif qu'à condition que la salinité de l'eau diminue naturellement sous l'augmentation de charge et ne soit pas uniquement due à un phénomène de dilution. Dans le cas contraire, seule une exploitation de l'eau douce dans les terres avant sa contamination par l'eau de mer est envisageable. Le volume d'eau expulsé entre les points 1 et 2 de la figure 5 apportera alors une information complémentaire sur l'éloignement de la zone d'intrusion de l'eau de mer dans le conduit karstique.

Remerciements

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Dye Tracer Tests within the Schollberg Underground Limestone Quarry: Experimental and Hydrogeological Implications.

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Abstract

A Large scale tracer test was performed in fissured rocks in the Swiss Pre-alpine Area in an underground limestone quarry adjacent to the alluvial plain of the river Rhine valley. The artificial fluorescent dyes: Uranine, Sulforhodamine B and Naphthionate were used as tracers. The implications of this tracer test result in establishing a feasibility study for a planned underground waste disposal site to be situated within the existing limestone quarry. During an observation period of about 3 years groundwater samples were periodically taken from 10 observation boreholes. They were analysed by luminescence spectroscopy in a first step and additionally by HPLC combined with fluorescence detection in a second step. Since the results of the tracer identification and quantification were strongly dependent on the applied analytical method, it created substantial implications on hydrogeological interpretation as of configuration of flow paths and flow velocities. As a consequence of this study it is strongly recommended that every tracer test should be analysed by two different techniques. The most commonly used technique (fluorescence spectroscopy) may be used as a first screening method.

Résumé

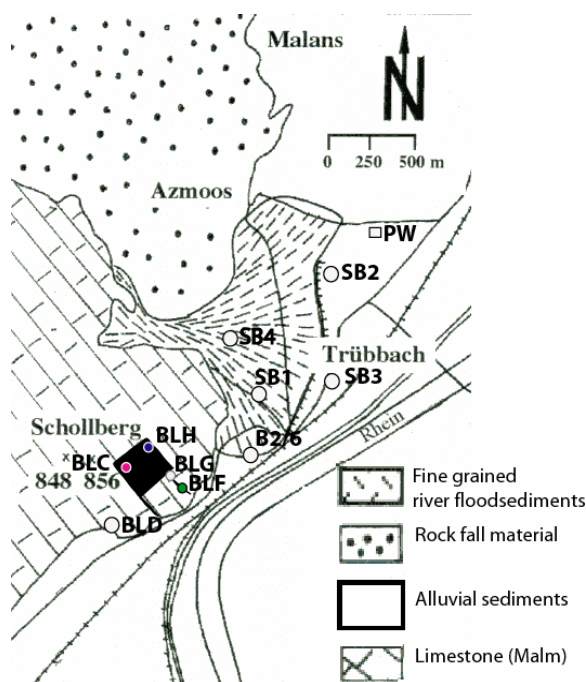
Un essai de traçage à grande échelle avec les traceurs Naphthionate, Sulforhodamine B et Uranine a été effectué dans des roches calcaires fissurées dans une carrière souterraine d'exploitation de ces roches dans la région des nappes helvétiques au Nord de Sargans (Canton de St. Gall). Ces essais de coloration étaient la phase finale d'une étude de faisabilité d'un stockage souterrain des déchets d'incinération. Pendant une période d'observation d'environ trois ans, des échantillons d'eau souterraine provenant de 10 forages ont été pris périodiquement et analysés dans une première phase par la spectrométrie de fluorescence. Ensuite des échantillons sélectionnés ont été aussi analysés par la méthode chromatographique de HPLC combiné avec la méthode de détection de fluorescence. Comme les résultats obtenus dépendent fortement de la méthode appliquée, ils peuvent ensuite aussi avoir des conséquences sévères sur l'interprétation hydrogéologique. En conséquence du cas présenté, il est recommandé fortement d'utiliser surtout pour des résultats douteux la méthode HPLC avec détection de fluorescence en plus de la méthode traditionnelle de la spectrométrie de fluorescence.

1. Introduction

In order to clarify former hydrogeological investigations (Schlegel, 1995, Weber, 1998) large scale tracer tests were performed in a fissured limestone area within the exploited part of an underground limestone quarry (Balderer, Leuenberger, 2003). For this tracer test the synthetic fluorescent dyes Uranine, Sulforhodamine B and Naphthionate were used. The aim of the tests were i) to determine fast preferential pathways in fissured rocks, ii) to discover possible connections to the adjacent alluvial aquifer in unconsolidated sediments and iii) to determine experimentally by real scale experiments the effective flow velocities and flow paths of the groundwater and to verify the range of existing hydraulic conductivities. Another aim of the test in fissured limestone was also to quantify the possible effects regarding an existing groundwater pumping station for drinking water supply (PW) within the alluvial aquifer. The outcome of these tracer experiments would have implications on the feasibility of setting up a waste disposal site in the disused part of this underground limestone quarry.

Within the investigated area (Fig. 1) a hill formed structure named Schollberg consisting of cretaceous limestone crops out (Fig. 2). This structure belongs to the helvetic nappes, to the so-called structure of the Axen-Nappe (representing the base of the nearby Saentis - Nappe). The axis of this folded structure (overturned fold) dips towards the North-East with a slope of 30°.

Fig.: 1 Geological Situation of the investigated area with the injection and observation boreholes



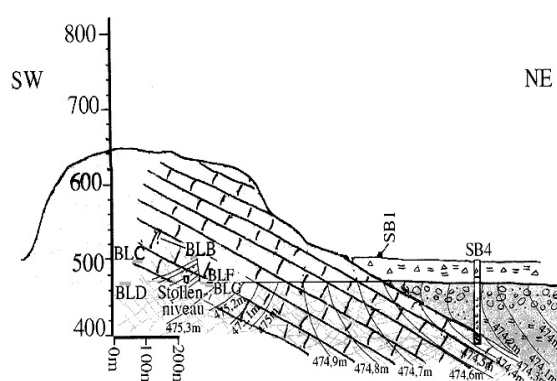


Fig. 2: Schematic cross section through Schollberg to the alluvial sediments (Weber, 1998).

Within the valley of the River Rhine this limestone bedrock is widely eroded by the combined action of glacial and fluvial erosion and the subsequent filling with fluvial-glacial unconsolidated sediments. With respect to the hydrogeological situation this limestone structure of the Schollberg (Fig. 2) in which the shafts of the underground quarry are driven represents a fractured rock of quite low hydraulic conductivity. According to the hydraulic testing (of packed intervals) in 4 drilled unlined boreholes within the limestone quarry values in the range of $1 \cdot 10^{-6}$ m/s to $1 \cdot 10^{-5}$ m/s resulted (for the borehole BLG). For the adjacent alluvial sediments the following values were found by pumping tests: $2 \cdot 10^{-3}$ m/s (borehole SB1) to $1 \cdot 10^{-2}$ m/s (for the boreholes SB2 and SB3). These values are representative of fluvial-glacial gravel aquifers. In the borehole SB4 the alluvial sediments contain more fine grained components with a somewhat lower hydraulic conductivity in the range of $5 \cdot 10^{-4}$ m/s. More adjacent to the limestone hills this gravel aquifer is overlain by fine grained river flood sediments of very low hydraulic conductivity with thicknesses in the range of 16 m (SB1) to 42 m (SB4). Below the gravel aquifer fine grained lake deposits were also encountered in borehole SB 4 (yielding a thickness of 23 m), overlying directly the limestone bed rock of the Malm (Quintnerkalk) which represents the direct continuation of the axial descending folded (nape) structure of the Schollberg hill. The chemical composition of the groundwater within the alluvial gravel aquifer is of Ca-Mg-HCO₃-SO₄ type, with a total mineralization of 422 to 526 mg / l (boreholes SB1 to SB4 and Dornau Pumping station). The water of the river Rhine is of similar water type, however, with a lower mineralization of about 390 mg/l. The water sampled in the boreholes within the Schollberg limestone Quarry shows quite varying mineralization but mainly the same chemical composition. For the water of the borehole BLD results a quite similar composition and total mineralization (420 mg/ l) analogous to that of the gravel aquifer. For the other boreholes more distant from the alluvial aquifer quite different mineralization were measured: the lowest mineralization was found in borehole BLC (187 mg/l), BLG contained 770 mg/l, and BLF had the highest values with 1400 mg/l.

Naphthionate, Sulforhodamine B and Uranine used for the tracer experiment were introduced in 3 boreholes (Fig.1) drilled into the fractured limestone rock of the underground quarry. The injected amounts of dye were as

follows: 10 kg Naphthionate into borehole BLH, 2 kg Sulforhodamine B into borehole BLC and 2 kg Uranine into borehole BLF. In order to study the propagation of the tracers by advective flow, groundwater samples were regularly collected over a period of 3 years from 10 observation boreholes: BLG within the quarry, BLD in the limestone rock just outside of the quarry and from the following boreholes situated within the alluvial sedimentary filling downstream of the Schollberg quarry B2/6, SB1, SB2, SB3, SB4 as well as from the nearby groundwater pumping station PW. (Fig.1 and Fig. 8). The water sampling out of the observation boreholes was done with a small submersible pump after renewing at least twice the water in the borehole volume.

2. Materials and Methods

Performing tracer tests in fractured rocks and porous aquifers normally yield low recovery resulting in concentrations near the detection limits of the methods.

Excitation and Emission Maxima of the Tracer Dyes			
Tracer Dye	Maximum Excitation wavelength (nm)	Maximum Emission wavelength (nm)	Detection Limit(ppb)
Na - Naphthionate	330	430	0.03
Sulforhodamine B	564	600	0.01
Uranine (Na -Fluorescein)	491	520	0.01

Tabl. 1: Instrumental Setting of the LS 50B Spectro-Fluorometer (Perkin-Elmer)

At these very low levels, natural or artificial fluorescent contaminants (originating from natural swampy waters, industrial plants or waste effluents) can mask the fluorescence intensities and therefore often yield erroneous results. Therefore it is very important to confirm the results acquired with a conventional technique such as the spectroscopic fluorescence analysis by a more substance specific method.

Consequently, the samples were first screened by fluorescence spectrophotometry (which is the conventionally used method of fluorescent dye analysis, Käss, 1992, 1998, Behrens, 1982).

In a second step they were analysed by HPLC coupled to a fluorescence detector (providing a more substance specific method, Lutz & Parriaux, 1988, Lutz, Parriaux & Tissières, 1987).

The instruments used were the following: luminescence spectrometer (Perkin-Elmer LS50B) set to measure the fluorescence intensity of the excited species at a given wavelength (details are described in tabl. 1) and a High Performance Liquid Chromatograph (DIONEX HPLC) coupled to a fluorescence detector (details are described in tabl. 2).

3. Results

Examples of measured dye concentrations over time using spectrofluorometry are shown in Fig. 3, 4 and 5 for samples from the boreholes BLG and BLD. The useful analytical range using the LS50, as represented in table 1,

was 0.01 to 2 ppb for Uranine and 0.3 to 5 ppb for Naphthionate and Sulforhodamine B using external standards. Blank signals were obtained using water taken before starting the test

Pump	P580 Low pressure gradient pump
Column	C18 Reversed phase (0.46cm x 25cm)
Flow-rate	1ml/min
Injector Volume	100µl
Eluent (for naphthionate and sulforhodamine B)	20 mM Na - Acetate, pH 5.2
Eluent (for uranine)	20 mM Na - Acetate, pH 5.8
Gradient Eluent	70% Acetonitrile
Detector	RF-2000 Fluorescence detector
Autosampler	GINA 50 with variable injection volumes
Data System	Dionex CHROMELEON software
Post Derivatization unit (for uranine)	500mM NaOH
Detection Limit of Na -Naphthionate*	≥ 0.1 ppb
Detection Limit of Sulforhodamine B*	≥ 0.1 ppb
Detection Limit of Uranine (Na - Fluorescein)*	≥ 0.1 ppb
*with reference to standards and blanks from pre-test groundwater	

Tabl. 2: Technical properties of the HPLC system

Since a separation technique prior to detection increases the specificity of the analysis, the samples showing a positive fluorescence signal using spectrophotometry were further analysed by HPLC in order to confirm the presence of the dye.

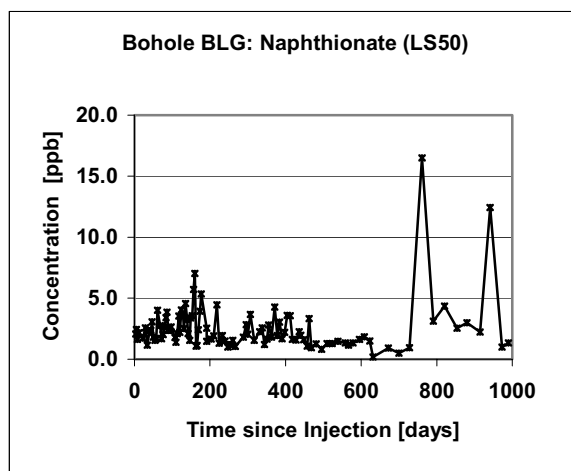


Fig. 3: Resulting concentrations of the water samples from borehole BLG by fluorescence spectrophotometry for Naphthionate

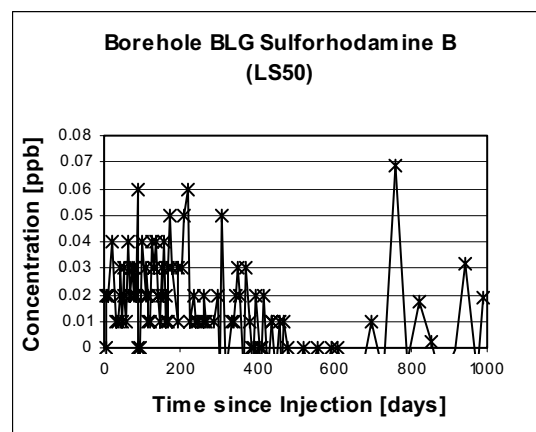


Fig. 4: Concentrations of Sulforhodamine B of the water samples from borehole BLG by fluorescence spectrophotometry over time.

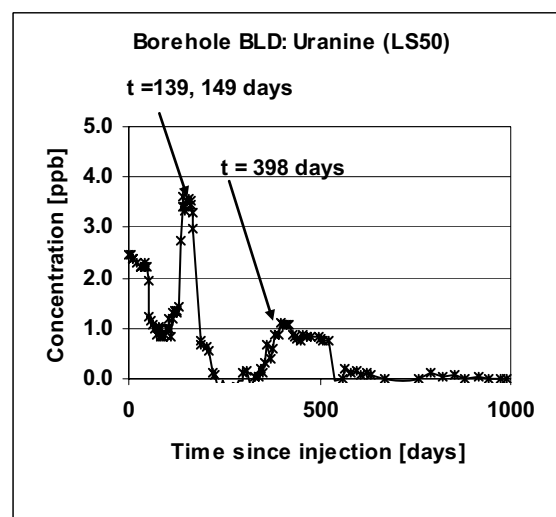


Fig.5: Concentrations of Uranine of the water samples from borehole BLD by fluorescence spectrophotometry over time

The reliability of the conventional spectrofluorometric method (using the LS50 instrument) is limited by interferences due to exogenous fluorescent substances in natural waters and not by the detection limits of the dyes of interest. Although HPLC has detection limits analogous or slightly higher than ones obtained with a spectrofluorometer it has the great advantage of providing separation of the water constituents by a chromatographic process before detection and quantification (Tabl. 2).

Fig. 6 shows an example of an HPLC Chromatogram for Naphthionate in the sample BLG72 (from 8.3.01, ie. 250 days after injection). The retention time of the comparative chromatograms of naphthionate does not coincide, proving the absence of the real dye substance in the sample.

Fig. 7 shows the overlay of the standard Sulforhodamine B curve with that of sample BLG78. On this chromatogram one observes a perfect peak match of standard and sample thus proving its presence.

If we consider the results obtained from the measured tracer concentrations using fluorescence spectroscopy, as seen on Fig.3, 4 and Fig.5, quite convincing breakthrough curves seem to be revealed from all investigated observations boreholes and even for the most distant pumping station.

However, if these results are interpreted as real flow paths of the injected tracers originating from the corresponding boreholes BLC (Sulforhodamine B), BLF (Uranine) and BLH (Naphthionate) some discrepancies with respect to the established map of lines of equal head according to the measured hydraulic heads (Fig. 8) can be recognised. This appears most drastically if the possible flow paths from the injection borehole BLF (Uranine) to the observation borehole BLD are considered, which are apparently confirmed by the diagram of tracer concentration versus time (Fig. 5) of fluorescence spectrometric measurements. In this case a tracer flow path would exist along the line of equal head which is hydraulically not realistic.

Now if we consider the results obtained by HPLC coupled to a fluorescence detector, we observe a totally

different picture. This is particularly true for samples out of borehole BLD where contrary to the positive results of the spectrophotometric analysis (Fig.5) the HPLC analyses revealed for all these samples the complete absence of Uranine.

Finally in table 3 all samples with confirmed presence of the dyes by the HPLC method are represented.

Considering these results it becomes clear, that the majority of positive break through events resulting from data obtained by spectrophotometry are in reality artefacts caused by other interfering substances present in the sampled groundwater.

Therefore these diagrams (Fig. 3, 4 and Fig. 5) reveal the astonishing fact, that even such non- identified “background substances” may show quite regular variations emerging from the measured samples like “ghost break through events” which are not or only hardly distinguishable from the behaviour of a real tracer transport phenomena and its resulting break-through curve.

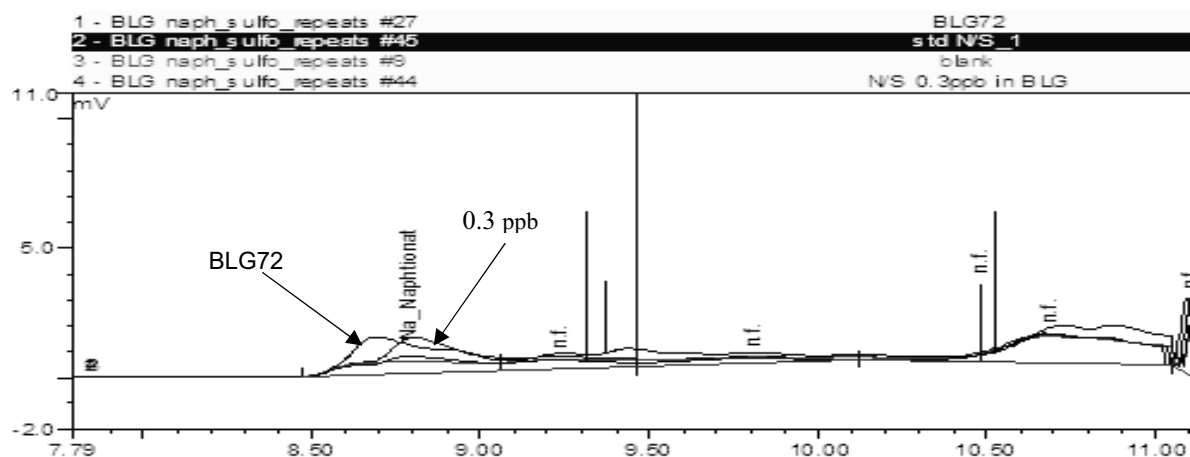


Fig. 6: HPLC Chromatogram of Naphthionate in sample BLG72 (taken 8.3.01, ie. 250 days from time of injection) overlaid by chromatograms of standards (0.1 and 0.3ppb) and blank (water without tracer)

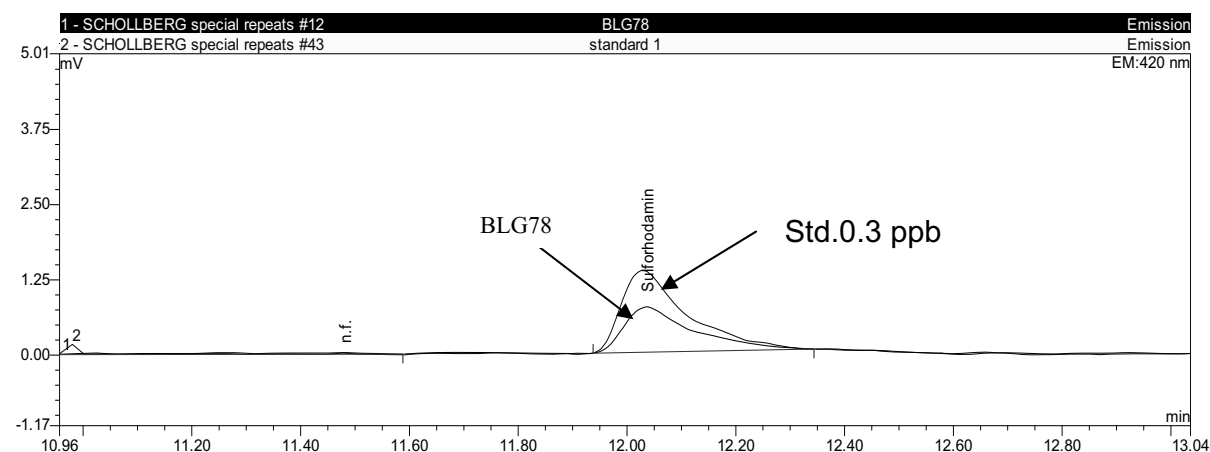


Fig. 7: HPLC Chromatogram of Sulforhodamine in sample (taken 17.4.2001, ie. 390 days after injection overlaid by a 0.3ppb Standard Sulforhodamine B sample)

Now if it is admitted that only the results of positive tracer concentration confirmed by both applied analytical methods are real observations of the tracer transport from injection to the identified observation boreholes, one has

now also to do the check of plausibility to the given hydraulic situation as revealed by the constructed maps of equal hydraulic heads established by the interpolation of the measured values on Fig. 8 (according to the head

measurements of the 25.2.1998 in all available boreholes) representative for the hydraulic situation in winter / spring season, A. Weber, 1998).

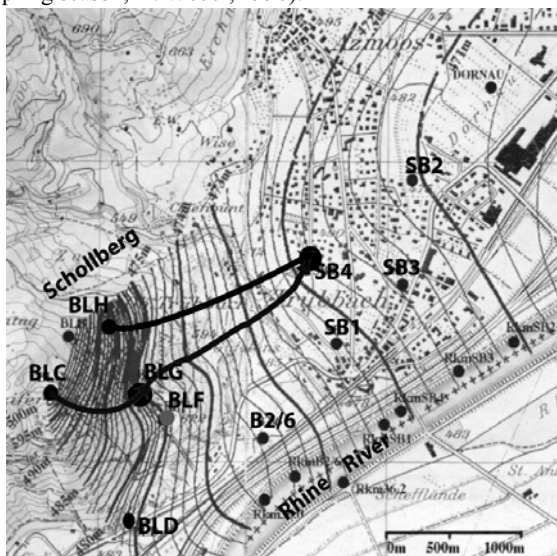


Fig. 8: Resulting flow path of the tracer according to the positive results of the two applied methods as given in table 3. On the map are also represented the lines of equal head according to measurements of the 25.2.1998 in all available boreholes, representative for the hydraulic situation in winter / spring season.

If we consider the location (Fig. 8) of the two boreholes with positive tracer recovery BLG and SB4 and its corresponding injection boreholes BLH (Naphthionate) and BLC (Sulforhodamine B) it can clearly be noticed, that these flow path according to the tracer results agree very well with the pattern of hydraulic head distribution.

4. Conclusions

As a consequence of this study we can conclude that for low to very low concentration of fluorescent dye tracers the conventional fluorescence spectrometric detection is often not specific enough, especially if other fluorescent substances such as humic acids or other fluorescent macromolecules are present. In this case a clear substance specific detection and quantification can only be achieved by the combined application of the HPLC technique with fluorescence detection.

Tracer tests are frequently applied to determine flow path in complex hydro-geological situations and for a proof or verification of existing hydraulic connections especially in karstic aquifers.

Consequently we suggested that when interpreting the results of tracer tests the following limitations should also be taken into account:

Failures in tracer identification and quantification may be due to the analytical limitations of the conventional technique of fluorescence spectroscopy. And these may lead to wrong interpretations *caused by false arrival times or even in assuming a wrong recovery of a tracer substance.*

Therefore if the hydrogeological interpretation is based only on spectrophotometric analysis, it may result in assumptions which provide:

- overestimation of flow velocities (due to too early arriving “artefact peak values”)
- underestimation of the effects of hydrodynamic dispersion (due to “high concentrations”)
- Wrong assumptions on existing flow path, and hydrodynamic connections, if preferential pathways are assumed which in reality do not exist (if an imaginary but not the real substance is identified no real positive substance specific identification is achieved).

Sampled Boreh.	Tracer Type	Inject. Bore-hole	Flow-Dist. (m)	Time since inject. (d)	Meas. conc. (ppb) by HPLC
BLG	Sulfo	BLC	290	390	0.15
SB4A	Naphth.	BLH	725	173	0.5
SB4A	Sulfo	BLC	940	427	0.1
SB4B	Naphth.	BLH	725	247	0.1
SB4B	Naphth.	BLH	725	331	0.1
SB4B*	Sulfo	BLC	940	203	0.1
SB4C*	Sulfo	BLC	940	398	0.1

Table 3: Representation of the groundwater samples with positive tracer detection with the HPLC method

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Perspectives sur l'utilisation de la fluorescence de la matière organique dissoute comme traceur quantitatif du temps de séjour dans un système karstique méditerranéen.

Application au Laboratoire Souterrain à Bas Bruit

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Abstract

This research is based the work by EMBLANCH (1998) and BATIOT (2001, 2002 and 2003) regarding the use of Total Organic Carbon (TOC) as qualitative tracer of infiltration into karstic systems. Indeed, an inverse correlation has been shown between TOC and Mg^{2+} concentrations, the latter being known as a tracer of long residence time for the system of Fontaine de Vaucluse (MUDRY, 1982).

The natural fluorescence and the Dissolved Organic Carbon (DOC) concentration have been monitored for perennial flows from the unsaturated zone of this system, which are accessible through the Laboratoire Souterrain à Bas Bruit (L.S.B.B.), and for artificial discharges from humus layer by lysimeter plates. First, due to the specific climate conditions (important precipitations followed by long dry periods), the natural fluorescence of the Dissolved Organic Matter (DOM) has shown a breakthrough of DOM-rich infiltration water in the unsaturated zone after a sudden rain. On the other hand, the global time series of the "fluorescence/DOC" ratio have shown a seasonal variation of the DOM both at the outlet and in the soil. Therefore, a more thorough characterization of this DOM should be helpful for tracing specific infiltration and estimating retention times in the hydrologic system.

Résumé

Ce thème succède aux réflexions émises par EMBLANCH (1998) et BATIOT (2001, 2002 et 2003) sur l'utilisation du Carbone Organique Total (COT) comme marqueur qualitatif de l'infiltration dans divers systèmes karstiques. Il existe, en effet, une corrélation inverse entre la concentration en COT et celle en Mg^{2+} , marqueur des longs temps de séjour (MUDRY, 1982), sur le système de Fontaine de Vaucluse (84).

Un suivi de la fluorescence naturelle et de la concentration en carbone organique dissous (COD) a été réalisé au niveau d'écoulements continus de la zone non saturée (ZNS) de ce système, accessible par le Laboratoire Souterrain à Bas Bruit (L.S.B.B.), ainsi qu'au niveau de l'horizon humifère du sol par l'installation de plaques lysimétriques. Tout d'abord, du fait des conditions climatiques de la région (fortes précipitations suivies de longues périodes sèches), la fluorescence naturelle de la Matière Organique Dissoute (MOD) nous permet d'observer, comme pour un traçage artificiel, l'arrivée d'une eau infiltrée (chargée en MOD) après une pluie soudaine au niveau des écoulements de la ZNS. D'autre part, le suivi global du rapport « fluorescence/COD » montre une variation saisonnière de la nature de la MOD aussi bien à l'exutoire que dans le sol. L'hypothèse de recherche qui en découle consiste à caractériser plus précisément la MOD, afin d'en identifier un type marqueur d'une période du cycle hydrologique et permettre donc d'approximer un temps moyen de transit à l'échelle saisonnière.

1. Introduction

L'estimation des temps de transit au sein des aquifères est une des étapes de l'évaluation de la vulnérabilité et de la protection de la ressource en eau souterraine. Le traçage est un outil pertinent permettant d'estimer ce paramètre. De récents travaux (EMBLANCH, 1998 ; BATIOT, 2001, 2002 et 2003 ; GARRY, en cours) ont montré l'intérêt, direct ou indirect, des MOD pour une approche qualitative (ou comparative) du temps de résidence de l'eau au sein des systèmes karstiques. Cependant, la plupart ont utilisé les Matières Organiques (MO) de façon globale à partir de l'analyse de la teneur en COT sans s'intéresser à la nature et la structure même de la MO. BATIOT (2001, 2002 et ss. presse) a remarqué des différences de fluorescence de la Matière Organique Dissoute (MOD) au cours d'un cycle

hydrologique, dénotant une variation de sa nature selon les saisons. L'objectif de cette recherche est d'utiliser la fluorescence de la MOD pour approcher quantitativement le temps de séjour de l'eau au sein d'un système karstique.

2. Présentation du site d'étude

Le système de Fontaine de Vaucluse

Notre choix s'est porté sur un système karstique déjà très étudié au L.H.A., avec plus de 120 articles et de nombreuses thèses (MUDRY, 1982 ; PUIG, 1987 ; COUTURAUD, 1993 ; EMBLANCH, 1997 ; BATIOT, 2002 ; GARRY, en cours), sur lequel un grand nombre de mesures hydrochimiques ont été effectuées : la Fontaine de Vaucluse. Au pied d'un escarpement de près de 200m de hauteur, la Fontaine de Vaucluse est la plus importante émergence karstique de France et l'une des plus imposante

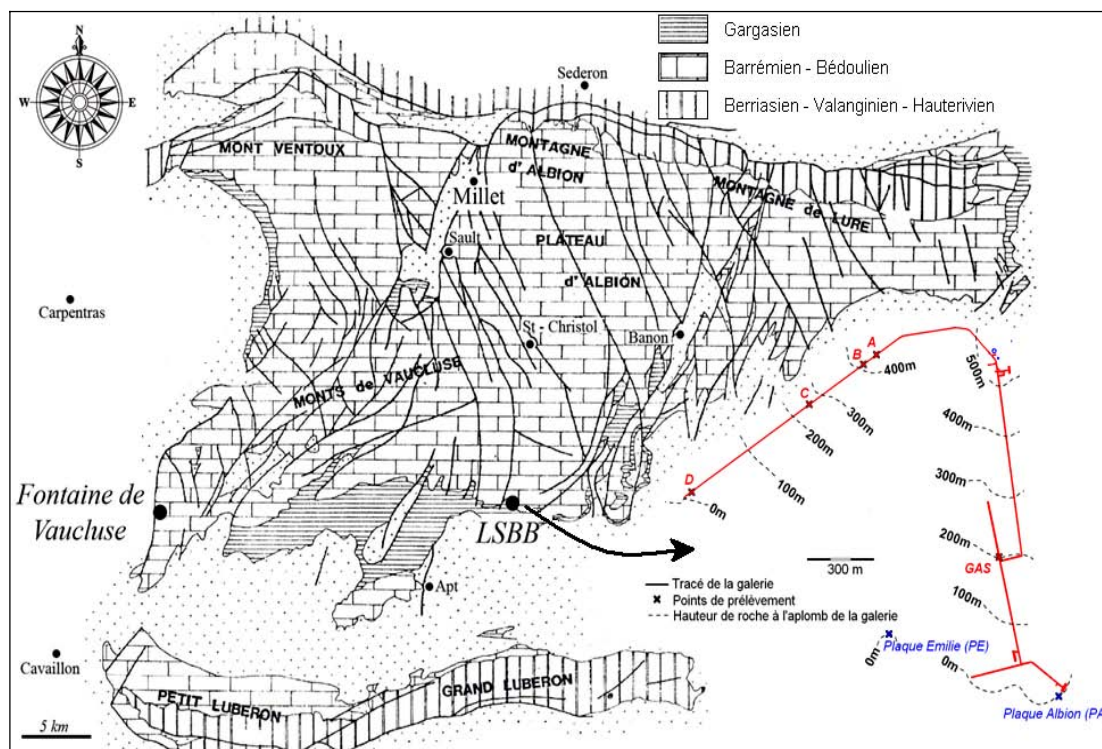


Fig. 1 : Carte géologique du système de Fontaine de Vaucluse et schéma du L.S.B.B. (PUIG, 1987 ; modifié)

au monde avec un débit moyen de $23\text{m}^3/\text{s}$ (GEVAUDAN, 2005). Elle a donné son nom au type même de sources, dont l'émergence se fait par un conduit noyé plus ou moins vertical. Située environ 30km à l'Est d'Avignon, la Fontaine de Vaucluse est à l'origine des Sorgues qui par leurs ramifications irriguent la partie la plus fertile du Vaucluse. Sans connaître précisément les limites de son bassin d'alimentation et des relations du système aquifère avec ses bordures, on peut en première approximation considérer que la Fontaine de Vaucluse est l'unique exutoire d'un massif carbonaté affleurant sur près de 1150km^2 (PUIG, 1987). Le réservoir de l'aquifère de Fontaine de Vaucluse appartient au Crétacé inférieur. Il est constitué par une série calcaire épaisse de près de 1500m (MASSE, 1972), dépourvue de matière organique, qui se développe entre les marnes du Néocomien et les marnes de l'Aptien supérieur (figure 1). Cet aquifère se compose d'une zone non saturée (ZNS) d'une épaisseur moyenne de 800m, dont le rôle hydrodynamique semble important (EMBLANCH, 2003) et d'une zone noyée (ZN) imposante de plus de 300m (PUIG, 1987).

Le Laboratoire Souterrain à Bas Bruit de Rustrel (84)

Le Laboratoire Souterrain à Bas Bruit (L.S.B.B.) est une ancienne galerie militaire (ex-poste de tir n°1 du système d'armes du plateau d'Albion) creusée dans le massif de la Grande Montagne sur le versant sud des Monts de Vaucluse (figure 1). Cette galerie pénètre dans un monoclin de pente sud-ouest et recoupe des failles de direction N-S et NE-SO (THIEBAUD, 2004). Elle offre un accès privilégié à des écoulements de la ZNS, qui, jusque-là, ne restaient accessibles que par de rares exutoires voire des avens. On observe donc des venues d'eau significatives et continues en quatre points de la galerie (A, B, C et D) et une autre non pérenne (GAS). Les propriétés hydrodynamiques de chacune montrent bien la diversité des divers types d'écoulements présents au sein d'une ZNS : ils font donc l'objet d'une thèse (GARRY, en cours). On remarquera que

tous ces points ne sont pas sous la même épaisseur de roche : A (440m), B (418m), C (350m), D (35m) et GAS (350m) (figure 1).

3. Matériels et méthodes

Tout le matériel utilisé lors des expériences a été soigneusement lavé au détergent, rincé avec de l'eau distillée présentant une conductivité de $2.5\mu\text{S}/\text{cm}$ et un COT inférieur à $0.1\text{mg}/\text{L}$ puis passé au four (pour la verrerie) à 500°C . Les campagnes d'échantillonnage ont été conduites sur le site du L.S.B.B., notamment sur les écoulements B et D de la galerie et des plaques d'infiltration PA et PE situées en surface (figure 1). Tous les échantillons sont rapportés au laboratoire dans une glacière et congelés à -15°C à l'arrivée après un passage (pour les infiltrations de sol) au travers d'un filtre en fibres de verre à $0.45\mu\text{m}$.

Travaux réalisés sur le terrain

Deux plaques métalliques (PA et PE), d'une surface de 850cm^2 , ont été placées sur deux affleurements de sol, entre l'horizon humifère de surface (A) et l'horizon d'accumulation (B). Au dessus de celles-ci, un arrosage par de l'eau distillée de même nature que précédemment a été réalisé jusqu'à l'obtention d'un échantillon de 500mL récupéré au niveau de l'orifice des plaques. Un échantillonnage mensuel de cette eau percolée à travers le sol a été effectué d'avril 2005 à janvier 2006.

Dans la galerie du L.S.B.B., l'intensité de fluorescence pour le couple « excitation/émission » $350\text{nm}/450\text{nm}$ a été suivie au niveau des écoulements B et D, depuis novembre 2003 et au pas de temps de 15min, au moyen de fluorimètres de terrain GGUN-FL30 prêtés par le département « Géosciences » de l'Université de Franche-Comté. Des échantillons de 30mL ont été prélevés au moins toutes les deux semaines.

Mesures réalisées au laboratoire

Les concentrations en carbone organique dissous (COD) sont mesurées au Laboratoire d'Hydrogéologie d'Avignon

(L.H.A.) grâce au « TOC ANALYZER model 700 » de BIORITECH, par oxydation chimique du COD et mesure du CO_2 émis par absorbance infrarouge. La précision de cet appareil a été fixée à $\pm 0.05 \text{ mg/L}$ (BATIOT, 2002). Les mesures d'intensité de fluorescence pour le couple 350nm/450nm ($I(350/450)$) des échantillons d'infiltration de sol sont effectuées, au laboratoire des sciences du sol de l'I.N.R.A. d'Avignon, avec un spectrofluorimètre LS55 de PERKIN-ELMER à lampe Xe et des fentes de monochromateurs de 10nm. La cuve utilisée est une cuve en quartz quatre faces transparentes dans laquelle est placé 1mL d'échantillon.

4. Résultats et discussion

Les travaux de EMBLANCH (1998) et BATIOT (2001 et 2002) ont montré que le COD peut être utilisé comme marqueur de l'infiltration. En effet, il existe pour plusieurs systèmes karstiques (notamment Fontaine de Vaucluse) une nette corrélation inverse entre la teneur en C.O.D. et la teneur en Mg^{2+} , marqueur qualitatif avéré d'un temps de séjour dans le système assez long (MUDRY, 1982). D'autre part, BATIOT (2002 et ss. presse) et THIEBAUD (2004) a montré qu'il existait une relation entre la concentration en COD et le $I(350/450)$. La fluorescence naturelle de l'eau peut donc être utilisée comme traceur de l'infiltration rapide d'eau chargée en COD au sein du système karstique après un événement pluvieux. Ainsi, ce traçage naturel se rapproche d'un traçage artificiel avec en plus un avantage conséquent, une injection globale du traceur (MOD du sol) au moment des précipitations et une restitution au moment de l'augmentation de l'intensité de fluorescence au niveau de l'exutoire suivi. La figure 2 nous montre ce même phénomène pour un écoulement profond (B en figure 1) après un événement pluvieux important précédé par une longue période sèche (mai à septembre 2005). Cependant, cette méthode de calcul du temps de séjour $\Delta t(1)$ possède une limite importante, à savoir la connaissance d'un temps de transit pour des événements météorologiques particuliers. Cela reste une information intéressante mais ponctuelle, à savoir valable après ce type de phénomènes météorologiques. De plus, les écoulements suivis sont de nature complexe puisqu'ils sont le résultat d'un mélange de différents types d'eau d'âges divers (GARRY, en cours). Ainsi, du fait de ces deux conditions, il est plus intéressant pour la gestion de la ressource d'obtenir un temps de séjour dont le suivi continu ne serait plus fonction de la météo.

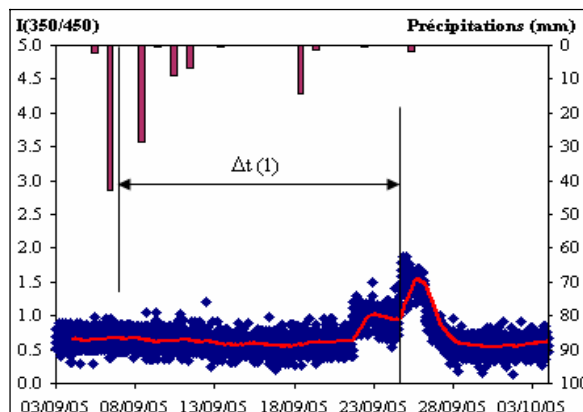


Fig. 2 : Réponse de $I(350/450)$ après un événement pluvieux au point B

C'est pourquoi, une autre approche a ensuite été mise en oeuvre. En effet, en observant de façon globale la chronique complète de l'intensité de fluorescence, il apparaît qu'il existe une variation sinusoïdale de celle-ci au niveau des écoulements B et D suivis. Afin de s'affranchir des variations en COD qui agissent énormément sur l'intensité de fluorescence mesurée, nous avons calculé le rapport « $I(350/450)/\text{COD}$ ». La figure 3 montre une même variation, malgré le passage au rapport, que l'on peut qualifier de saisonnière. Ceci démontre qu'il s'agit bien d'une variation saisonnière de la nature de la MOD au cours d'un cycle. En effet, on observe des pics, plus ou moins importants, du rapport, résultant de l'arrivée des pluies printanières et automnales.

Ceci doit être rapproché de la variation de la MOD du sol au cours des saisons. En effet, divers auteurs ont montré que le degré de dégradation de la MO du sol dépend essentiellement des conditions climatiques et biologiques du milieu (GUGGENBERGER & al., 1994 ; CHRIS et DAVID, 1996). Ainsi, pour un climat méditerranéen, marqué par de longues phases de sécheresse suivies par des phases humides fortes et soudaines (au printemps et à l'automne), la MOD du sol subit plus ou moins ce processus. En hiver, la MO reste en surface et ne subit que très peu l'activité biologique : c'est donc une MOD peu dégradée qui est entraînée dans le système au moment des pluies printanières. Du fait d'une structure moléculaire plus importante, l'intensité de fluorescence est plus faible que

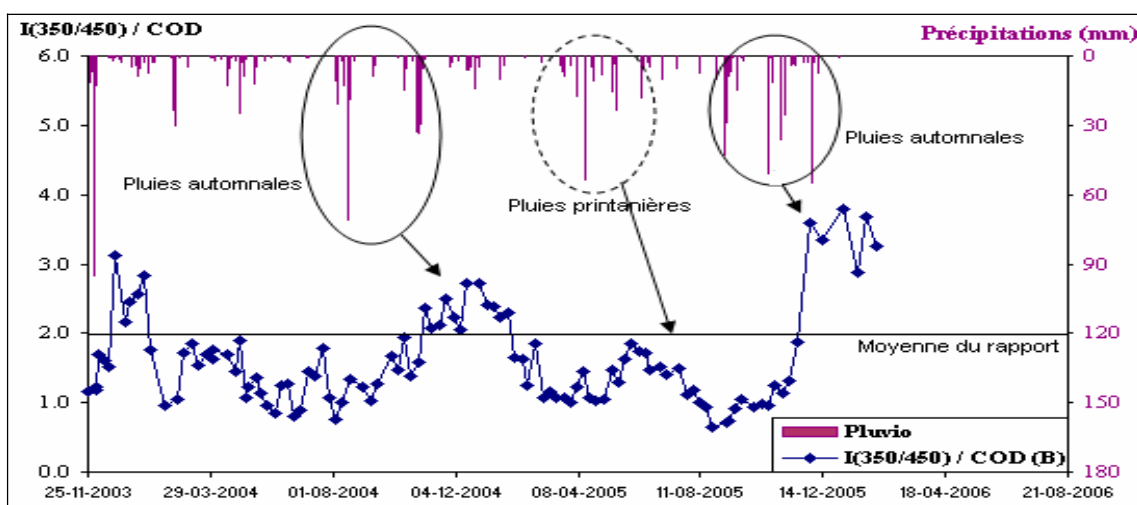


Fig. 3 : Suivi du rapport $I(350/450)/\text{COD}$ au point B

pour les suivantes (figure 4). Au contraire, en été, l'activité biologique est intense : c'est donc une MOD plus petite et très dégradée (résidus de la minéralisation des produits végétaux) qui est entraînée par les pluies automnales (HONGVE, 1999). De ce fait, l'intensité de fluorescence est grande (figure 4).

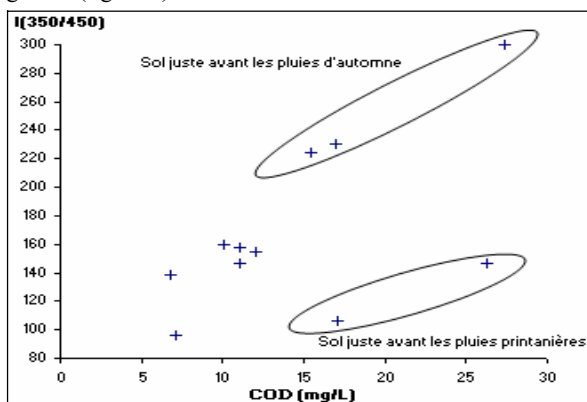


Fig. 4 : Diagramme I(350/450) – COD des échantillons PA et PE

Ainsi, il est intéressant de constater que cette variation saisonnière de la nature de la MOD du sol est encore observable au niveau d'un écoulement profond (B) (figure 3).

5. Conclusion

Nous avons donc pu remarquer que la MOD n'est pas qu'un simple marqueur qualitatif du temps de séjour. Tout d'abord, grâce à ses propriétés de fluorescence, elle nous indique l'arrivée des eaux météoriques après un événement pluvieux. Ainsi, cette caractéristique se rapproche d'un calcul du temps de transit par un traçage artificiel à injection globale auquel elle lui emprunte aussi une limite (calcul d'un temps de séjour pour un régime hydrique donné et un événement pluvieux particulier). Cependant, cette même fluorescence nous permet d'observer la variation saisonnière de la nature de la MOD au niveau d'un écoulement profond et de l'exutoire d'un système karstique (BATIO, ss. presse). Ainsi, avec une caractérisation plus fine de cette MOD, on pourrait en identifier une famille spécifique de composés responsable de l'intense fluorescence observée à certaine période du cycle hydrologique. On en tirerait donc un traceur naturel pouvant nous quantifier approximativement un temps de transit moyenne à l'échelle saisonnière.

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Practical Limits of High-Resolution Evaluation of Carbonate Chemistry within Karst Flow Systems

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Abstract

Karst flow systems experience rapid variation in flow and water chemistry conditions that govern system evolution and function, both with regard to background chemistry and anthropogenic inputs such as nutrients and bacteria from animal waste and other agricultural contaminants. Evolving methods utilizing electronic/digital data logging systems offer important tools for clarifying details of these processes. While these methods easily provide high temporal resolution, challenges remain in the quantitative evaluation of data quality resolution in terms of reported accuracy and precision.

An effort underway within Cave Spring Caverns, Kentucky USA is rigorously defining the practical limits of karst water monitoring by working under essentially ideal conditions. An epikarst waterfall is monitored for temperature, pH, and specific conductance (spC) by three independent probe/data logger systems with two-minute resolution. Redundancy reduces the probability for data loss and allows calculation of a standard deviation to quantify measurement precision. Early results show that data can be obtained within one standard deviation of <0.2°C for temperature, <4 µS/cm@25°C for spC, and <0.04 for pH for periods up to about two weeks; accuracy is similar for temperature and spC, and averaged within 0.02 pH units over seven months, although at times varied up to 0.1 unit.

1. Introduction

Recent progress has been made in understanding the details of karst flow and geochemical processes by high-resolution monitoring with electronic probes and digital data loggers (e.g. BAKER AND BRUNDSON, 2003; GROVES AND MEIMAN, 2005; LIU *et al.*, 2004). These methods are especially important for karst systems, as significant variations in karst groundwater flow and chemical conditions in response to precipitation can occur in hours and even minutes. Measurements with a resolution of minutes capture all significant structures of such hydrochemical variation, analysis of which can be used to understand aquifer structure and behavior.

FREEMAN *et al.* (2004) provided a thoughtful discussion of issues considered in the the implementation of remote hydrological monitoring sites, such as design, quality assurance, and instrument drift. They cautioned that "because of the perceived ease of installation and operation..., data provided by these systems commonly are not adequately supported by quality-assurance procedures and documentation." In the present paper we describe an effort underway to rigorously define the practical limits of reported precision and accuracy in underground karst water monitoring by working under essentially ideal conditions of easy access, equipment security and available electricity, focusing on background carbonate water chemistry, which provides the geochemical environment that in turn influences the behavior of other, often anthropogenic, water chemistry parameters.

2. Materials and methods

2.1. Field site

Cave Spring Caverns is located beneath Kentucky's Pennyroyal Plateau sinkhole plain, and consists of about two km of large, horizontal cave passages with a floor about 25 m below the ground surface. The passages are formed within the Mississippian (Lower Carboniferous) St. Louis Limestone, which locally dips to the west at about 1-2°. The top of the Lost River Chert lies occurs above the cave, and acts as a leaky perching layer, so that water tends to reach the cave at distinct locations, mainly as waterfalls emerging from the cave ceiling through fractures, draining the epikarstic zone. The closest of these waterfalls to the entrance, Barrel Falls, falls about 4.5 m from the ceiling in a recessed niche on the western side of the passage about 40 m from the cave's entrance, and serves as the focus for the data collection and analysis reported here.

The recharge area of the cave lies within the Graham Springs Groundwater Basin which flows to Graham Springs on the Barren River, about 18 km to the southwest. South central Kentucky has a humid-subtropical climate, with a mean precipitation of about 1,260 mm/yr, and the mean-annual temperature is 13°C. Precipitation is spread throughout the year, though summer and fall tend to be drier than other months, and mean-annual potential evaporation is 800 mm.

2.2. Equipment

The Barrel Falls site is equipped with sensor/loggers array and tied to a common tipping bucket rain gauge

(Campbell Scientific (CSI) TE525) resolving tips of 0.1 mm. Discharge from the rain gauge is directed into 10-mm Tygon tubing which feeds a PVC flow-through chamber (20 mm ID) mounted with a series of three Cole-Parmer double-junction industrial in-line ATC pH sensors. Each pH sensor is connected to a three-meter shielded coaxial cable and terminates in the instrument box (Pelican 1400) at a Cole-Parmer preamplifier to increase signal stability. This pH system can resolve pH to ± 0.01 SU. The pH flow-through chamber discharges into a section of 10-mm Tygon tubing where it is split into three paths, each passing through a CSI CS547A-L spC/temperature sensor. This sensor can resolve temperature to $\pm 0.1^\circ\text{C}$ and spC to ± 0.001 mS/cm. The three paths are then rejoined into a single section of tubing and positioned at an elevation approximately 40 cm higher than the sensors to assure pipe-full conditions for continuous water contact with the sensors. Each of three Campbell CR10X loggers is connected to its corresponding set of pH, conductivity (spC) and temperature sensors. There is thus redundancy in spC, temperature, pH, and data loggers that not only ensures backup in the case of malfunction, but three resulting independent measurements of these parameters allows calculation of the mean, standard deviation (SD), and coefficient of variation (CV) for each observation.

The logger takes 30-second pH, spC, and temperature values that are averaged every two minutes. To reduce redundant data, the program compares the current two-minute average values of each sensor to that of the previous two-minute average. If the absolute value of change exceeds a small preset value the current two-minute average values for all sensors are stored. If not, values are only stored once per hour. In this way we achieve two-minute resolution even during hourly recording, since we know under those static conditions the observations have not varied beyond the threshold value.

The waterfall flow data are given in tips per minute for the tipping bucket gage, but these do not yield discharge directly because some of the water, especially at higher flows, falls outside of the bucket orifice. These data thus only give a relative flow indication, but the signals give a clear indication of dry and wet conditions that correlate well with rainfall events. We are in the process of developing a rating curve relating tips per minute to actual discharge, which we measure periodically by catching the flow in a large tarp and measuring the volumetric flow rate.

3. Results

High resolution temporal data from Waterfall One (Figure 1) provide detail about flow system structure and behavior, even when comparing only relatively simple data, in this case flow and spC (GROVES *et al.*, 2005). For example, between 21 March and 10 April 2005 precipitation, waterfall flow, and spC data reflect four succeeding rain events that occurred over progressively wetter antecedent moisture conditions. Flow in the waterfall, initially under relatively dry conditions, began to increase within 2.3 hours of the onset of significant rainfall measured above the cave system, and showed a clear flow increase of about 120% that returned to the original condition within about 1.5 days with no systematic change in the spC signal following this rainfall. Approximately three days later a more intense storm occurred with obvious differences in the cave response. While the timing of the flow increase was similar to the first storm, flow rates stayed more than twice as high as the initial condition for more than four days without significant rainfall, rather than returning quickly to pre-storm levels. The spC signal from relatively dilute rainfall quickly moving through the system was also clear and corresponded to rainfall intensity, reaching a low of about $160 \mu\text{S cm}^{-1}$, or about 70% of pre-storm levels. The next storm, about four days later, was different from the first two with respect to both signals. Flow rates continued at a similarly high level without an appreciable increase, while spC dropped again in very clear relation to rainfall. In contrast to the second storm, however, spC took more than seven days to rise to the same level that had taken only two days after the previous rainfall, even though starting at a higher minimum level. Finally, a small storm about five days later, which began with waterfall flow rates at a similarly high rate, had little or no impact on waterfall behavior. The four different storm responses in Figure 1 detail significant epikarst storage capacity, responses dependent on whether that reservoir is relatively full or depleted. Different flow paths, including one or more rapid ones triggered above a threshold recharge rate, are also indicated. Though in a significantly different hydrogeologic setting, we note the similarity of these responses to those observed by PERRIN *et al.* (2003).

A design that provided three independent measurements of pH, temperature, and spC allowed us to quantify measurement variability by calculation of a mean and SD for each observation. Results show that temperature and spC precision are generally good, with SDs within 0.2°C and $4 \mu\text{S/cm}$, respectively. On the other hand, precision of the pH probe was not as good. Although the SD is within 0.01 pH unit for the first two days following calibration, it steadily increased over time. This defines the constraint that may limit their use in remote cave locations when high resolution data are required over long periods—up to two weeks the SD was still within 0.04 pH units, but continued to drift beyond this time. To test for accuracy, each pH probe was calibrated *in situ* approximately weekly. The three point (pH 4, 7, 10) calibration resulted in a linear function

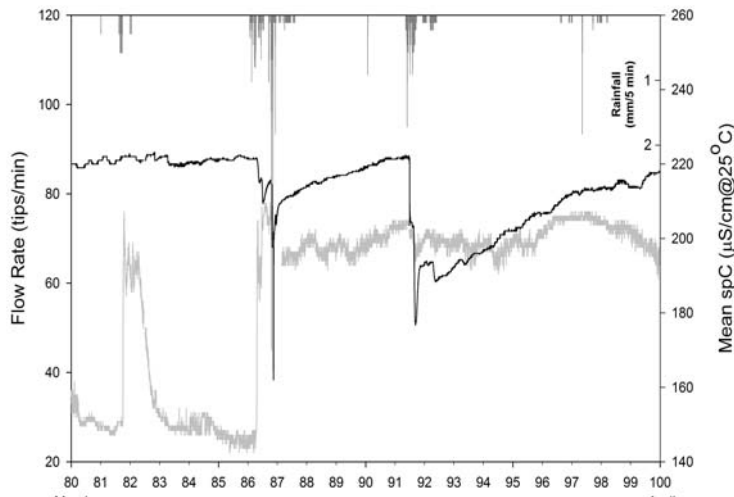


Figure 1. Early results from continuously monitored rainfall, flow, and spC at Waterfall One.

calcite dissolution rates (e.g. LIU *et al.*, 2004). Applying our measured uncertainties in temperature and pH to calculations of these parameters allowed us to quantify error propagation. Figure 2 shows the relationships of SI_{cal} , PCO_2 pressures, calcite dissolution rates, and total inorganic carbon with pH (A) and temperature (B). In the pH range of 7 to 8 and temperatures between 10 and 15°C, the conditions found at our monitoring site, along with representative values of Ca^{2+} and HCO_3^- , the rates of change of these parameters with changes in pH and temperature are shown in Figure 2 (C). Our measured uncertainty in temperature of $\pm 0.2^\circ C$, for example, resulted in an uncertainty in SI_{cal} , PCO_2 , and calcite dissolution rate, respectively, of ± 0.004 , ± 0.0001 atm, and ± 0.0008 mm/yr, whereas our maximum measured uncertainty for pH of ± 0.1 resulted in uncertainties of ± 0.1 , ± 0.0003 atm, and ± 0.05 mm/yr for the same parameters, respectively. This highlights the fact that of these pH provides the constraint on interpretation of high temporal resolution carbonate chemistry variations measured under practically available means.

4. Conclusions

In general, for data to be useful the level of noise in the system (non-meaningful variations due to random fluctuations and measurement error) must be small in relation to the signal under study. For example, the spC data are easily measured to within 5 $\mu S/cm$, while variations from storm events often exceed 40 $\mu S/c$. Another important consideration, however, is that these data can also be used in computations to obtain additional parameters other than those directly measured with the data loggers. These can include, for example, those describing background carbonate chemistry such as carbon dioxide pressures, calcite saturation indices, and mineral dissolution rates. Results here quantify the extent that pH uncertainty limits the interpretation of carbonate chemistry measurements, and that under our measurement conditions, for two minute resolution over periods of up to a few weeks we could report uncertainties in SI_{cal} , PCO_2 , and calcite dissolution rates of ± 0.1 , ± 0.0003 atm, and ± 0.05 mm/yr, respectively.

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that correlates the voltage signal (in millivolts) from each sensor with a value of pH in each calibration, typically with r^2 values of at least 0.99. The estimation errors of pH from transformation of the voltage signals of the sensors to values of pH by means of these linear equations led to average estimation errors over seven months of -0.011 (for theoretical pH = 4), +0.021 (for theoretical pH = 7) and -0.012 (for theoretical pH = 10), although at times pH errors have been as high as ± 0.1 pH.

Important information about karst systems can be obtained by simultaneously considering water/rock/gas interaction parameters such as calcite saturation indices (SI_{cal}), PCO_2 , and

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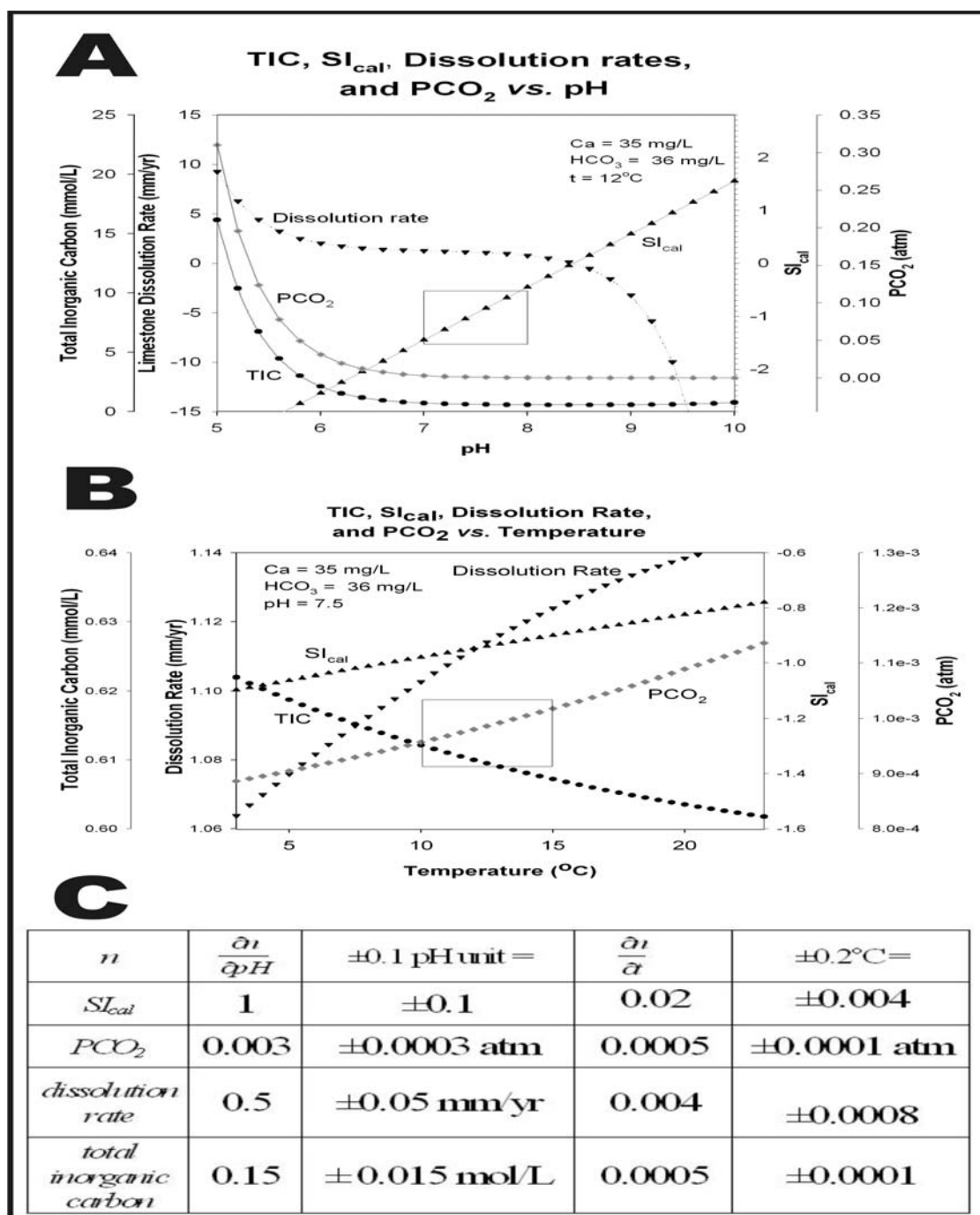


Figure 2. Sensitivity of various carbonate chemistry parameters to variations in input data including pH (A) and temperature (B). Table (C) shows output ranges based on computation error propagation for reasonable ranges of data precision and accuracy using technology described above.

Complementarities of karst hydrology and hydrogeology

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Abstract

Paper tries to answer on question: What is difference, and what is identical in karst hydrology and hydrogeology? Usual definitions of hydrogeology and hydrology are: hydrogeology is branch of geology, which deals with groundwater and especially its occurrence, while hydrology is science that deals with the processes governing the land areas of the Earth, and treats various phases of the hydrological cycle. From these definitions it is hardly possible to strictly distinguish between the two scientific disciplines. In engineering practice the division is grounded in argument that hydrology deals with surface water and hydrogeology with groundwater. However, strictly enforcing such division could have harmful consequences on development of both sciences. Synthesis of hydrogeological and hydrological approach could expedite progress in surface water-groundwater system understanding. This is especially true for the problem of water occurrence and circulation in karst. Karst is specific media where exists strong interaction between circulation of groundwater and surface water. Due to this reason progress in karst theoretical and practical investigations should be based on closed co-operation between hydrology and hydrogeology as well as many others scientific disciplines. A few examples are given in the paper.

Keywords *karst, hydrology, hydrogeology, karst aquifer, hydraulic conductivity*

1. Introduction

The growing world population, the commensurate demand on limited fresh water resources and the potential impacts of climate change or variability on catchment and its ecosystems are the main reasons for necessity of interdisciplinary scientific co-operation in water resources management. This is a basic prerequisite in finding more efficient solutions for the uncertain water future all over the world.

Karst hydrology as karst hydrogeology has long-lasting and successful experience in field of interdisciplinary scientific co-operation, which is natural consequence of complexity of karst terrain and water circulation in it.

A wide range of closed surface depressions, a well-developed underground drainage system, and a strong interaction between circulation of surface water and groundwater typify karst. It represents terrain with distinctive hydrogeology and landforms arising from a combination of high rock solubility and well-developed secondary porosity. Karst develops over about 500 millions of years, and began to form soon after the carbonate rocks were deposited. Karstification is an erosional process, which drastically changes the landscape by removing the carbonate rock at the surface and underground.

Karst is defined (FIELD 2002) as a terrain, generally underlain by limestone or dolomite, in which the topography is chiefly formed by the dissolving of rock, and which may be characterised by sinkholes, sinking streams, closed depressions, subterranean drainage and caves. The soluble rock terrains present special problems to the scientists and engineers. The varied and often spectacular surface landforms are merely a guide to the presence of unpredictable conduits, fissures and cavities beneath the ground. These subsurface features can occur even where surface karstic landforms are completely absent. In soluble rock terrains, more so than in most other terrains the unexpected should always be expected (ATKINSON 1986).

The main objective of this paper is to help in pushing forward the process of interdisciplinary scientific co-operation between hydrology and hydrogeology in field of karst water investigation.

2. Complementarities of hydrology and hydrogeology

DE MARSILY (1986) states that the study of the water cycle or hydrology in its wider sense is usually divided into three separate disciplines: meteorology, surface hydrology and hydrogeology or groundwater hydrology. UNESCO & WMO (1992) present the following definitions of hydrogeology and hydrology: hydrogeology is branch of geology, which deals with groundwater and especially its occurrence while hydrology is science that deals with the processes governing the land areas of the Earth, and treats various phases of the hydrological cycle. For KIRALY (2002) one of the principal aims of hydrogeology is to propose a reasonably adequate reconstruction of the groundwater flow field, in space and in time, for a given aquifer. Hydrology is a scientific discipline within the earth sciences. Its main focus is the terrestrial part of the hydrological cycle. Hydrology concerns the occurrence, movement and composition of water below and on the earth's surface. Hydrological science has strong links, on a wide range of spatial and temporal scales, with oceanic, atmospheric, and solid earth

sciences as well as with biological sciences. Questions are: 1) Is hydrogeology only one sub discipline of hydrology? 2) How much they are interdependent?

From these definitions it is hardly possible to strictly distinguish between the two scientific disciplines. In engineering practice the division is grounded in argument that hydrology deals with surface water and hydrogeology with groundwater. However, strictly enforcing such division could have harmful consequences on development of both sciences.

Hydrogeology generally deals with groundwater occurrence and circulation in aquifers. Aquifers are in turn geological units involved in transmission of quantities of water under ordinary hydraulic gradient. At the same time interest of hydrology is mainly focused on water balance, which is basically accounting of the inflow to, outflow from, and storage within a hydraulic unit such as a drainage basin or aquifer. Very often it is impossible and harmful to separate two above mentioned approaches, but in practice it predominantly occurs.

The main cause of this division possibly lies in different educational backgrounds. Hydrogeologists are mostly educated in geological schools, whereas hydrologists usually have geophysical or engineering background. Synthesis of hydrogeological and hydrological approach could expedite progress in surface water-groundwater system understanding. This is especially true for the problem of water occurrence and circulation in karst.

3. Complexity of karst aquifer

Karstification is primarily a geological characteristic important for water circulation and storage. It can be defined through density, frequency and number of all types of karst voids (intergranular voids, pores, joints, cracks, fractures, fissures, conduits and caves). Generally it is greatest at the surface and decreases with the depth of a karst massif. Karstification is a continuous process governed by natural and man-made interventions, so that even the most detailed models are only temporarily valid (BONACCI 2001a).

Circulation of groundwater in karst aquifers is quite different from water circulation in other non-karstic type aquifers. In karst aquifers water is being collected in networks of interconnected cracks, caverns, and channels. Hydraulic permeability of karst aquifers is essentially created by flowing water and has anisotropic character (HUNTOON 1994).

Karst aquifer is an aquifer containing soluble rocks with a permeability structure dominated by interconnected conduits dissolved from the host rock. The conduits are organised to facilitate the circulation of fluid in the downgradient direction wherein the permeability structure evolved as a consequence of dissolution by the fluid. Karst aquifers comprise a very distinct class of aquifers and present a wide variety of characters due to their different geological and geomorphologic context and history. They exhibit highly heterogeneous, anisotropic and complex structure.

Therefore the karst aquifers are some of the most complex and difficult systems to decipher. The highly heterogeneous nature of karst aquifers leads to an inability to predict groundwater flow and contaminant transport. For different scientists (hydrologists, hydrogeologists, geochemists, geographers, geomorphologists, geophysicists, hydraulic engineers, speleologists, biologists, ecologists etc.) they represent a challenging interplay of water flow and storage in large caves and conduits and very small fractures and pores. The surface and especially subterranean environment in karst provide a range of habitats for chemical and biological processes. To the biologists and ecologists, they are fragile ecosystems, hosting rare and endangered species. For the geochemists, they provide rapid transport of contamination.

Karst aquifers are generally continuous (BONACCI & ROJE-BONACCI 2000). However, numerous subsurface morphologic features in karst (caves, jamas, fractures, faults, impermeable layers etc) strongly influence the continuity of the aquifer, and commonly the aquifer does not function as a continuum in a catchment (BONACCI 2001a). One of the most important characteristics of karst aquifers is the high degree of heterogeneity in their hydraulic properties.

The root causes of karst aquifers peculiarity and heterogeneity are numerous:

- (1) heterogeneous and anisotropic surface and underground morphologic karst forms;
- (2) existence of well developed, complex, deep, and mostly unknown positions and dimensions of underground karst cracks, joints, fissures, impermeable layers, conduits etc;
- (3) strong interaction between circulation of surface water and groundwater;
- (4) high and fast oscillations of groundwater (BONACCI 1987, BONACCI 1995);
- (5) strong and direct but inadequately known links between inflow (swallow-holes) and outflow (karst springs) karst features;
- (6) generally small storage capacity of the karst medium;

- (7) fast, turbulent groundwater transport through karst conduits and at the same time slow, laminar flow through karst matrix;
- (8) natural endogenic and exogenic processes, and influence of man's induced structures and/or activities (dams, reservoirs, motorways development, water pumping, water abstraction etc) which influence fast and/or slow changes of the water regime;
- (9) different aspects of duality of water circulation in karst (KIRALY 2002).

Scale issues are particularly important for understanding and modelling karst water circulation. Conditions in karstified medium are strongly dependent on space and time scales, especially in deep and morphologically complex vadose zone. Vadose zone and karst aquifer are two-component systems in which the major part of storage is in the form of true groundwater in narrow fissures, where diffuse or laminar flow prevails. On the other hand, majority of water is transmitted through the karst underground by quick or turbulent flow in solutionally enlarged conduits. Interaction between the two above-mentioned types of flow is significant and permanently present.

Studying geometry of the karst aquifer over 1,000 m² experimental area in the vicinity of Montpellier, France, DROGUE (1980) made continual measurements of the groundwater levels and water temperatures at 19 piezometers. This study showed that groundwater level in piezometers, which were close to one another (less than 10 m), responded differently to rainfall in the catchment, primarily because of varying connections with the main karst conduits, subsurface karst features and karst springs. Groundwater level in piezometers connected to small fissures reacts much more slowly, than in piezometers connected to main karst conduit and/or spring.

According to data from the literature (DROGUE 1980, BONACCI 1999, BONACCI & ROJE-BONACCI 2000) only every second piezometer provides the necessary information on the water circulation in karst and aquifer characteristics. At least half of them are drilled in impermeable, compact or less permeable parts of the karst massif in which all dynamic processes of water flow are very slow.

DAOXIAN (1986) reported the fact that from nine wells had been excavated in the Malian valley, Waxian country, Gaungxi province (China), only two yield satisfactory results, giving a discharge of 190 and 33 l/s under 5 m drawdown. The discharges from the other seven are smaller than 0.5 l/s with a drawdown of more than 5 m. The distance between the wells was of the order of magnitude of 100 m.

DEBIECHE *et al.* (2002) showed that with a low pumping rate, the karst aquifer could be considered as an equivalent continuous medium. The karst aquifer of the Pinchinade pumping area (France) displays its actual behaviour with a dual permeability only with a high pumping rate. The critical Pinchinade investigation demonstrates that a simplification of the hydraulic behaviour of a complex fissured medium is possible only if the karst aquifer tests have been performed with sufficient duration.

Continuous measurement of groundwater level, water temperature and many chemical and physical parameters in deep piezometers represents an exceptionally important source of a wide range, necessary for all types of investigations related to the regime of groundwater in karst. These data are of special importance for investigation of the relationship between karst aquifer and karst springs.

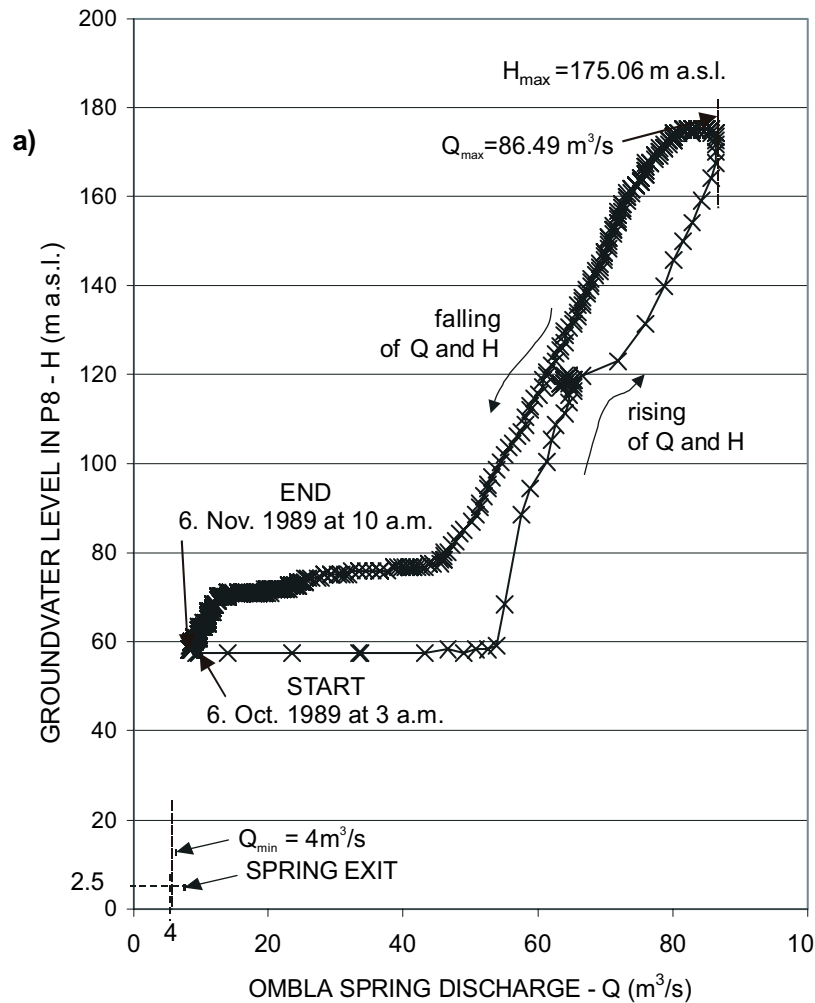
The measurements carried out from January 1988 to July 1991 on the some ten deep piezometers in the Ombla karst spring catchment helped to reach theoretical and practical conclusions (BONACCI 1995). Figure 1a gives a graphical presentation of the hourly values of groundwater level (GWL), H , in piezometer P8 and respective discharges, Q , of the Ombla Spring, which occurred from 8 Oct. 1989 at 3 a. m. to 6 Nov. 1989 at 10 a. m. Figure 1b presents discharge curves of the Ombla Spring dependent upon GWL measured once a day at 8 a. m for the above-mentioned period. The loop effect, which can be seen on Fig. 1a, was eliminated. Two discharge curves depict discharge flow under pressure for $H \leq 70.0$ m a. s. l. and $H \geq 76.0$ m a. s. l. For $76.0 \geq H \geq 70.0$ m a. s. l. discharge flow with a free surface, presumably in a karst conduit, exists.

The key for understanding and predicting of groundwater flow through karst vadose zone and karst aquifer lies in ability to accurately define the features and locations of the heterogeneity and their hydrogeologic and hydrological parameters. Hydraulic conductivity, K , is a key interdisciplinary parameter necessary for understanding and modelling karst groundwater circulation and karst aquifer characteristics. Its determination through groundwater measurements using piezometers and discharges of karst springs is particularly useful.

Using Dupuit's assumption BONACCI (2001b) determined hydraulic conductivity, K , in the karst massif around the Ombla Spring (Croatia). In relatively small area of about 50 km² the values of the hydraulic conductivity ranged from 0.702×10^{-3} to 26.414×10^{-3} m s⁻¹ during 262 hours of measurement. Changes in values of K , in both time and space can be attributed to differences in the position of investigated pairs of piezometers, their connection with main karst conduits, groundwater level and differences in development of

karstification process in the analysed karst massif. This points to a significant influence of the time and space scale effect upon the results obtained by investigations and measurements in karst. Another analysis leads to conclusion that piezometric relations in karst aquifer are more uniform during the descent than during rising of groundwater level. The rising phase is relatively short and lasts 5-10 % of the year. During this time all processes are rapid, flow is mostly turbulent and more non-homogenous than during descending phase. Relationships between Q , and H , during these two phases are strongly different.

MOTZ (1998) investigated vertical leakage to the upper Floridan karst aquifer for 11 karst lakes in the Central Lake District in peninsular Florida. Vertical leakage from lakes averages from 0.12 to 4.27 m yr^{-1} . The vertically averaged vertical conductance (K_v/b) (K_v is vertically averaged hydraulic conductivity of the units between the bottom of the lake and the top of the upper Floridan aquifer; b is the thickness of the hydrogeologic units between the bottom of a lake and the top of the underlying upper Floridan aquifer) was determined to range from 0.0394 to 1.00 yr^{-1} for these lakes. MOTZ (1998) showed that in water budget calculations for seepage and drainage karst lakes groundwater outflow could be a significant component of the water budget.



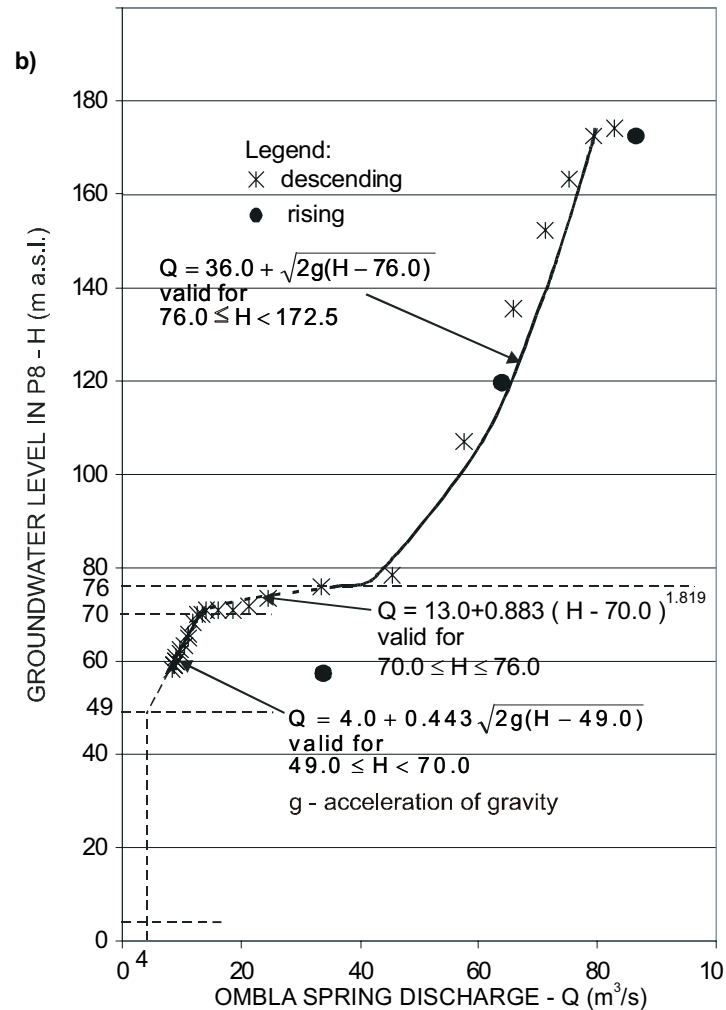


Fig.1 Relationship between hourly Ombla Spring discharge ,Q, and groundwater level ,H, measured in piezometer P8 (a), and measured one per day at 8 a.m. (b)

4. Conclusion

The karst system shows the extreme heterogeneity and variability of geologic, morphologic, hydrogeologic, hydrologic, hydraulic, ecological and other parameters in time and space. Such a complex system needs interdisciplinary approach. It is highly important to understand the interaction of groundwater and surface water in karst. Only closed co-operation between many scientific disciplines can create the best possible determination of the karst aquifer characteristics, the karst catchment areas and parameters of their water budget.

Hydrology and hydrogeology have a dual role as scientific disciplines and as bases for informed decision-making on important practical problems. It should be stressed that both of them have, at the same time, very deep scientific interests and task and an extremely important role in practice.

Progress in karst hydrology and hydrogeology is limited by a lack of data. In karst terrains processes are highly variable in space and time, and this variability exists at all scales. Data collection over a large range of scales is difficult and expensive. In the same time there is a growing tendency to minimise fieldwork in karst hydrology and hydrogeology. Investors realise that time is money and there is no more time-consuming process than fieldwork. As a result, especially karst hydrologists and hydrogeologists are asked to solve problems with computer models, remote sensing etc, rather than by direct field observations. This could be very dangerous intention.

Possibilities for overcoming karst water circulation complexity and doubts find in closed co-operation between hydrology and hydrogeology as well as with different branches of geosciences, and organisation of continuous and detailed monitoring and fieldwork. Due to particularity of water circulation in karst, role of

fieldwork using of different tracing techniques and continuous monitoring in deep piezometers are of special importance.

Karst hydrology and hydrogeology need all kinds of models and modelling, as well as the new scientific approaches, methods and technologies. In the same time it should be profoundly aware that they are only a useful tool but not a panacea. Karst hydrology and hydrogeology have a very difficult and responsible mission, which could be fulfilled on various ways, but no one understands which is the most direct, correct and proper way. First of all karst hydrology and hydrogeology should take care of their roots and develop their own practical and scientific methods and approaches in order to better explain and solve complex problems of water circulation in karst terrains.

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Etude des variations saisonnières des processus d'écoulement et de transport au sein d'un système karstique. Synclinal de Celles – Condroz (Belgique).

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Abstract

The « Condroz » is a belgian geographic area located in the central part of the Synclinorium of Dinant. This folded structure presents a general west-east direction. Its geomorphology is characteristic of an Appalachian structure: the crests correspond to famennian sandstone's anticlines and the valleys to carboniferous limestone's synclines.

The syncline of Celles is located on the south boundary of the Condroz, at 10 kilometers in the SE of the city of Dinant. It is drained by a brook called "S^t Hadelin". In the upstream part of the watershed, there are a lot of shallow-holes responsible of the existence of a 4 kilometers dry valley. These shallow-holes are located in the valley axis and in the tributary brooks coming from the sandstone hills. The "Fontaine S^t Hadelin" and the "Grande Fontaine" are the two resurgences of this karstic system.

In order to assess the flow and transport parameters characterising the karstic system, five multi-tracing tests were realized in different hydrologic conditions: high and low resurgence discharges.

Résumé

Le Condroz est une zone géographique belge occupant la partie centrale du Synclinorium de Dinant. Cette structure plissée, de direction générale ouest-est, confère au Condroz sa géomorphologie caractéristique : des crêtes topographiques localisées au droit des anticlinaux de grès famenniens alternent avec des dépressions situées dans l'axe de synclinaux de calcaires carbonifères.

Le Synclinal de Celles se situe en bordure méridionale du Condroz, à 10 kilomètres au sud-est de la ville de Dinant. Il est drainé par le ruisseau S^t Hadelin. Dans la partie amont du bassin, de nombreuses pertes, induisant une vallée sèche de plus de 4 kilomètres, sont observées. Elles se situent tant dans l'axe de la vallée qu'au niveau des petits affluents s'écoulant des crêtes gréseuses. La Fontaine S^t Hadelin et la Grande Fontaine constituent les deux résurgences du système karstique.

Afin de caractériser les paramètres d'écoulement et de transport régissant ce système karstique, cinq campagnes de multitraçage ont été réalisées dans des contextes hydrogéologiques variables : conditions de hautes et basses eaux. La comparaison des courbes de restitution a permis de quantifier la variabilité des paramètres.

1. Introduction

Le bassin versant du ruisseau de la Fontaine S^t Hadelin, d'une superficie de 30 km², se situe à 10 km au Sud-Est de la ville de Dinant (Fig.1 et 2).

En amont du village de Celles, sa géomorphologie est typique de celle du Condroz (BOULVAIN *et al.*, 1995). L'axe de la vallée se situe au niveau des calcaires carbonifères (synclinal) ; et les crêtes, de part et d'autres, sont composées de formations détritiques fammeniennes (anticlinaux) (Fig.3).



Fig.1 : Localisation de la région étudiée.

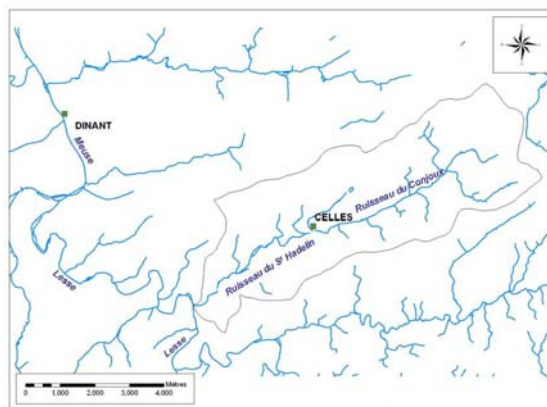


Fig. 2 : Localisation du bassin versant du Ruisseau du S^tHadelin

Les nappes logées dans ces anticlinaux se déversent dans le synclinal calcaire de manière diffuse ou par le biais des affluents du ruisseau de la Fontaine St Hadelin. Ces derniers prennent tous naissance au sein des grès et siltites famenniennes et se perdent plus ou moins rapidement lors de leur arrivée sur les calcaires. Ces nombreuses pertes, localisées dans la partie amont du bassin versant, induisent la présence d'une vallée sèche de 4 kilomètres. En période de fortes

précipitations, lorsque la capacité d'absorption des pertes est dépassée, les eaux de débordement constituent le Ruisseau du Conjoux qui s'infiltre, dans sa totalité et de manière diffuse, au niveau du lieu-dit « Petite Trussogne » ; exceptionnellement, il peut reprendre son cours superficiel sur l'entièreté de la vallée sèche (HALLET *et al.*, 2002)(HALLET *et al.*, 2003).

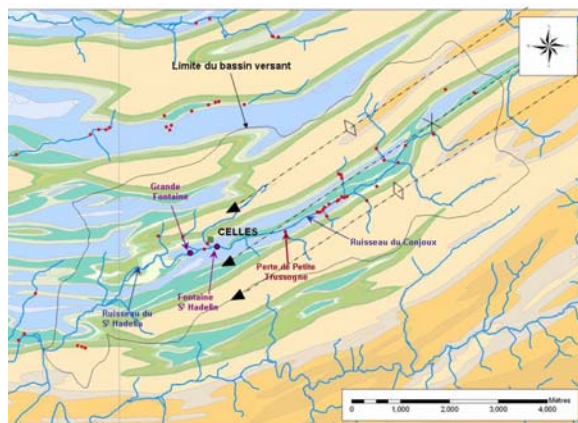


Fig. 3 : Géologie du bassin versant du Ruisseau du St Hadelin : Formations jaunes, grises et oranges : Formations détritiques fameniennes, Formations bleues et vertes : Formations carbonatées carbonifères, ● : Pertes, ● : Résurgences, — : Réseau hydrographique, — : Axe synclinal, — : Axe anticlinal

Les eaux du système karstique émergent en deux résurgences distantes de 500 mètres : l'une située à l'amont du village de Celles (Fontaine S^t Hadelin), l'autre en aval (Grande Fontaine).

Des bilans hydrogéologiques, réalisés en 1984 (HALLET, 1984) ont montré un déficit d'écoulement important (56%) au droit de la Fontaine S^t Hadelin tandis que le bilan calculé à l'aval de la Grande Fontaine ferme bien. Ceci démontre que ce système karstique ne se prolonge pas vers l'aval malgré la continuation des formations calcaires. La présence d'un petit pli anticlinal ennoyé vers l'est pourrait expliquer l'émergence des eaux du système karstique. Suite à la présence de ce pli, la stratification, orientée parallèlement à l'axe de la vallée et à pendage subvertical dans la partie amont, prend une direction perpendiculaire à l'aval de Celles induisant une diminution de la perméabilité des calcaires, créant ainsi un seuil hydrogéologique.

Ces deux résurgences sont connectées au même système karstique. La Grande Fontaine présente un débit relativement constant (250 m³/h) quelques soient les conditions climatiques et tant que la Fontaine S^t Hadelin est active ; tandis que la Fontaine S^t Hadelin voit son débit varier entre 0 m³/h et 1350 m³/h. Cette dernière joue le rôle de trop plein du réseau karstique en période de hautes eaux ou de fortes précipitations (Fig.4).

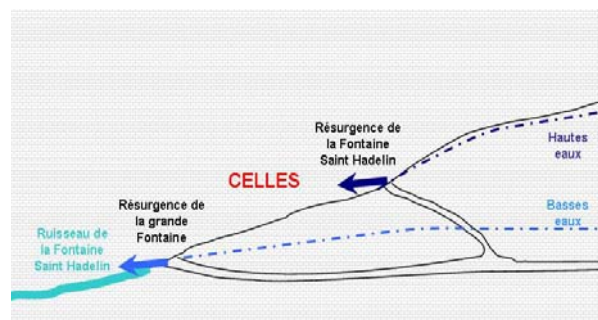


Fig. 4 : Schéma du système karstique de Celles

2. Matériels et méthodes

Afin de caractériser les paramètres d'écoulement et de transport régissant ce système karstique, cinq campagnes de multitraçage ont été réalisées sur ce site dans des contextes hydrogéologiques variables (conditions de hautes et de basses eaux). Les traceurs utilisés sont l'uranine et le naphthionate.

Deux pertes ont été sélectionnées comme point d'injection de traceur : la perte du Château S^t Martin (Ruisseau du Conneux) et la perte de Petite Trussogne (Ruisseau de Petite Trussogne) (Fig.5). Ces pertes sont respectivement localisées à 4450 m et 2000 m de la résurgence de la Grande Fontaine ; et à 3950 et 1500 m de la Fontaine S^t Hadelin.

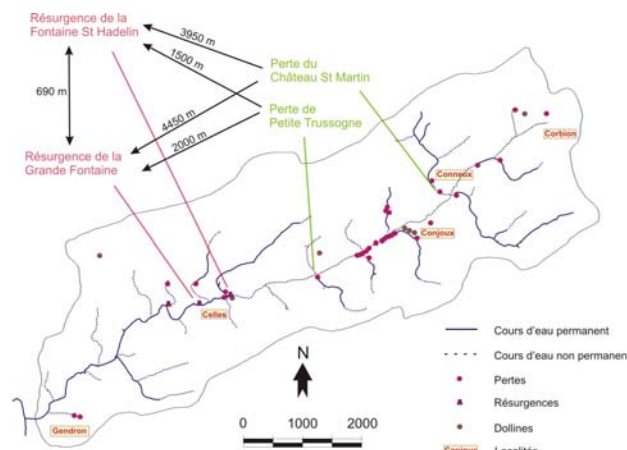


Fig. 5 : Localisation des essais de traçage.

Les sites d'échantillonnage sont la Résurgence de la Grande Fontaine et/ou la Fontaine S^t Hadelin selon les essais de traçage. Ces deux sites étaient équipés d'échantillonneurs automatiques Isco prélevant les échantillons d'eau selon un pas de temps régulier de 30 à 45 minutes selon les essais. Le tableau 1 décrit les différents essais de traçage réalisés depuis octobre 2002, les conditions hydrologiques (débits) prévalant lors des essais et les masses de traceur injectées.

Date	C.H.	Site d'injection (Débit m ³ /h)	Site d'échantillonnage (Débit m ³ /h)	Dist. (m)	Traceur utilisé	Quantité injectée (g)	1ère arrivée (h)	Vitesse max. (m/h)	Temps modal (h)	Vitesse modale (m/h)	T.R. (%)
oct-02	B.E.	CSM (12.7)	FSH (40)	3950	N	1000	37.0	107	45.5	87	6.6
		PTT (11.5)	FSH (40)	1500	U	350	17.0	88	23	65	10.8
mars-03	H.E.	CSM (85)	FSH (350)	3950	N	1000	21.0	188	25	158	22.0
		PTT (?)	FSH (350)	1500	U	350	8.5	176	11	136	42.6
nov-04	B.E.	PTT (9.5)	GF (154)	2000	U	150	22.0	91	27	74	29.4
mars-04	H.E.	CSM (26.7)	GF (262)	4450	N	700	25.7	178	31.8	140	25.5
		CSM (26.7)	FSH (344)	3950	N	700	23.8	166	29.67	133	44.1
		PTT (9.7)	GF (262)	2000	U	150	13.0	154	16	125	36.9
		PTT (9.7)	FSH (344)	1500	U	150	10.5	143	13.25	113	53.7
mai-05	B.E.	CSM (19.3)	GF (255)	4450	U	250	29.1	153	33.58	133	38.9
		CSM (19.3)	FSH (95.9)	3950	U	250	30.7	129	36.67	108	15.8

Tabl.1 : Description des cinq multi-traçages réalisés à Celles depuis Octobre 2002 à mai 2005. Légende : C.H. Conditions hydrologiques, T.R. : Taux de Restitution, B.E. : Basses Eaux, H.E. : Hautes Eaux, CSM : Perte du Château Saint Martin, PTT : Perte de Petite Trussogne, FSH : résurgence de la Fontaine Saint Hadelin, GF : Résurgence de la Grande Fontaine, N : Naphtionate, U : Uranine. **Remarque** : En novembre 04, la Fontaine Saint Hadelin était asséchée.

3. Résultats

Les courbes de restitution du naphtionate et de l'uranine aux résurgences de la Grande Fontaine et de la Fontaine Saint Hadelin sont présentées aux figures 6 et 7.

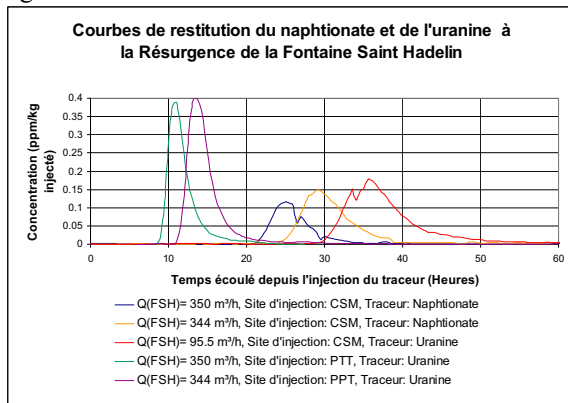


Fig.6 : Courbes de restitution à la résurgence de la Fontaine Saint Hadelin

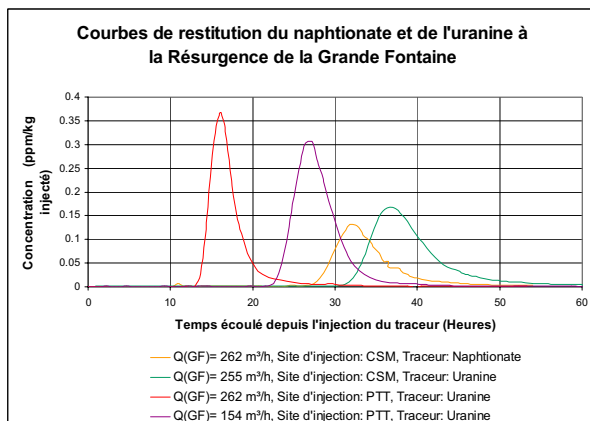


Fig.7 : Courbes de restitution à la résurgence de la Grande Fontaine

Les temps d'arrivée première et modaux, de même que les vitesses maximales et modales et les taux de restitution sont détaillés au sein du tableau 1.

Pour la Fontaine Saint Hadelin, les vitesses maximales sont comprises entre 107 et 188 m/h pour les injections à la perte du Château Saint Martin et entre 88 et 176 m/h pour les injections à la perte de Petite Trussogne. Pour la Grande fontaine, les vitesses maximales sont respectivement comprises entre 153 et 178 m/h (injection CSM) et entre 91 et 154 m/h (injection PTT). Ces vitesses maximales augmentent avec le débit ; ce qui indique que la convection, mode d'écoulement prédominant en système karstique, est bien fonction du régime hydrodynamique.

Afin de déterminer les caractéristiques hydrodispersives du système tracé, les résultats des essais de traçage ont été modélisés à l'aide du code CATTI (SAUTY *et al.*, 1992)(fig.8).

PTTFSH0303
restitution uranine

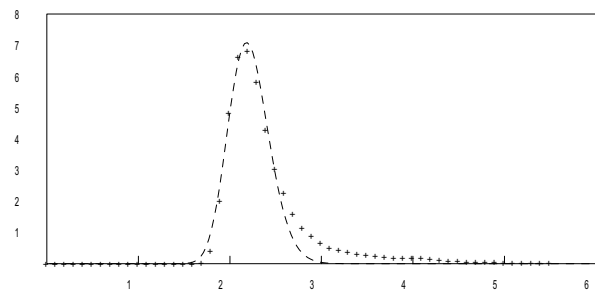


Fig.8 : Résultat de la modélisation de l'essai de traçage PTT – FSH du 03 mars 2003. Légende : Traits discontinus : courbe calculée, croix : valeurs observées, Abscisses : 1 unité = 5 heures, Ordonnées : 1 unité = 0.02 ppm

Le modèle mathématique utilisé considère un transport en écoulement monodimensionnel uniforme dans un milieu unicouche. Dans cette modélisation, la masse de traceur injectée considérée correspond à la masse de traceur récupérée et les taux de restitution

Date	C.H.	Site d'injection (Débit m³/h)	Site d'échantillonnage (Débit m³/h)	Dist. (m)	Porosité efficace	Vitesse de Darcy (m/sec)	αL (m)	Epaisseur d'écoulement (m)
oct-02	B.E.	CSM (12.7)	FSH (40)	3950	75%	0.02	7	0.75
		PTT (11.5)	FSH (40)	1500	75%	0.01	7.5	1
mars-03	H.E.	CSM (85)	FSH (350)	3950	70%	0.03	12	3.5
		PTT (?)	FSH (350)	1500	70%	0.03	7.5	4
nov-04	B.E.	PTT (9.5)	GF (154)	2000	65%	0.01	7.5	3.25
mars-04	H.E.	CSM (26.7)	GF (262)	4450	65%	0.02	13	2.5
		CSM (26.7)	FSH (344)	3950	70%	0.03	14	3.5
		PTT (9.7)	GF (262)	2000	70%	0.02	7	3.3
		PTT (9.7)	FSH (344)	1500	70%	0.02	7	5.5
mai-05	B.E.	CSM (19.3)	GF (255)	4450	70%	0.02	13	4
		CSM (19.3)	FSH (95.9)	3950	70%	0.02	12	1.5

Tabl.2 : Résultats de la modélisation des essais de traçage à l'aide du code CATTI. Légende : C.H. Conditions hydrologiques, B.E. : Basses Eaux, H.E. Hautes Eaux, Dist. : Distance, αL = dispersivité longitudinale, CSM : Perte du Château Saint Martin, PTT : perte de Petite Trussogne, FSH : Fontaine Saint Hadelin, GF : Grande Fontaine

sont égaux à 100%. Cette modélisation a permis de quantifier la porosité efficace du milieu, sa dispersivité longitudinale, la vitesse de Darcy et l'épaisseur d'écoulement. Les résultats de la modélisation sont présentés dans le tableau 2.

Selon cette modélisation, le milieu analysé présente une porosité efficace relativement homogène comprise entre 65 et 75 %. Les vitesses de Darcy sont comprises entre 0.01 et 0.03 m/sec et croissent avec la valeur du débit. Deux valeurs moyennes de dispersivité longitudinale sont obtenues: 7 à 7.5 mètres pour les essais de traçage réalisés au départ de la perte de Petite Trussogne ; et 12 à 14 mètres lors des injections à la perte du Château Saint Martin. Cela correspond une dispersivité longitudinale moyenne de 4.7 m/km pour le traçage de Petite Trussogne et de 3.5 m/km pour le traçage de Château Saint Martin. Les épaisseurs d'écoulement sont comprises entre 0.75 et 5.5 mètres en raison des variations de la section en fonction du niveau piézométrique, ce qui signifierait que le transport du traceur s'effectue dans la partie supérieure de l'aquifère. Ces épaisseurs sont maximales en période hautes eaux (entre 2.5 et 5.5 mètres) et minimales en périodes de basses eaux (entre 0.75 et 4 mètres). Une étude piézométrique couplée à une répétition des essais de traçage dans le temps permettrait de valider les épaisseurs d'écoulement calculées.

Notons que le code CATTI ne prend pas en considération les effets de retard. L'allure des courbes de restitution démontrent cependant que ceux-ci semblent relativement peu importants (Fig.8).

4. Conclusion

La modélisation CATTI des cinq multitraçages effectués au droit du système karstique de Celles a permis de démontrer la stabilité de paramètres hydrodispersifs, tels que la porosité efficace et la dispersivité longitudinale, malgré des conditions hydrodynamiques variables.

L'épaisseur d'écoulement, quant à elle, semble varier suite aux fluctuations de la piézométrie.

Afin de caractériser plus précisément la variabilité de la dispersivité longitudinale en fonction de la distance séparant les pertes des résurgences, cette étude sera approfondie par de nouveaux multitraçages au départ d'un plus grand nombre de pertes, distribuées le long du système karstique. Ces nouveaux essais seront couplés à un suivi piézométrique de regards de nappe proches du système étudié afin de valider les épaisseurs d'écoulement déterminées à l'aide du code CATTI.

De nouvelles modélisations seront réalisées à l'aide du code CATTI, mais également à l'aide du code QTRACER (MALCOLM, 2002). Ce dernier permettra d'une part de quantifier les effets de retard et d'autre part de dimensionner la section d'écoulement. L'utilisation de modèles hydrauliques sera envisagée.

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Predicting the Effect of Chabrouh Dam Reservoir on the Surrounding Karstic Hydrogeology “An Integrated Scientific Approach”

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Abstract

Located at the eastern coast of the Mediterranean Sea, Lebanon has more than 65% of its surface area covered with karst terrains. This along with its mountainous nature poses a challenge to the construction of dams. Approximately 40 km north east of Beirut city, in the western Mount Lebanon range the Chabrouh dam is under construction. The reservoir of the dam is expected to have a capacity of 8 hm³ and should cover a surface area of 0.46 km². The water level in the reservoir is planned to reach an elevation of 1613 masl. In an attempt to predict the effect of the reservoir on the surrounding karst hydrogeology of the area, the El Qana plateau located on the western side of the reservoir was the focus of structural, hydrogeological and speleological investigations. The prevailing rocks of this 8 km² plateau are mainly gently dipping limestone, dolomites and marls belonging mainly to upper Cretaceous period. Twenty small springs were identified and 3 major ones were periodically measured. Eight caves were explored and surveyed and more than 50 dolines, 1107 mesofractures, and 70 lineaments were identified and analyzed. They all indicated the structural control of the karst in the area with a dominant development in the NW-SE direction. The depth of karstification reaches an elevation of 1580 masl that is 33 m below the water level in the reservoir. This integrated scientific approach revealed the preferential groundwater flow in NW-SE karstic routes within this plateau. This will result in possible channelling of the water from the reservoir to the surrounding water outlets mainly Qana - Terrash and El Hadid Springs.

1. Introduction

Lebanon is located at the Eastern coast of the Mediterranean Sea (Figure 1). It has more than 65% of surface coverage of karstic terrains. That is about 5550 km² of surface coverage. Although Lebanon receives rainfall ranging between 200 and 2000 mm/year it is always in shortage of domestic and potable water. The water falls mostly in three to four months of the year. Its steep slopes results in high rates of runoff in a short period of time. This along with the nature of its terrain, karstic, causing high infiltration rates and results in major water shortages in high altitude areas.

Several potential dam sites were investigated since the 1960's but none of them was developed beyond the investigation stage until the year 2002. That year the Chabrouh dam project was initiated. The site allocated for this dam was in Chabrouh Valley near Faraya Village approximately 40 km north east of Beirut city, in the western Mount Lebanon range. The reservoir of the dam is expected to have a capacity of 8 hm³ and should cover a surface area of 0.46 km². The dam will be filled from snow melt and two major springs Nabaa el Laban and Nabaa el Assal (MAJDALANI, 1977) in the area that discharge an average of approximately 2.2 and 1.2 m³/s respectively (EDGELL, 1997). No recent meteorological data is available and no meteorological station is located in the vicinity of the dam site or on the El Qana Plateau. The average yearly precipitation for the area is reported to be 1400 mm (ATLAS CLIMATIQUE DU LIBAN, 1977). This does not appropriately take into account snow that might imply more yearly precipitation than actually reported. Based on the data from the ATLAS CLIMATIQUE DU LIBAN (1977) January and February are the months with most precipitation, whereas there is virtually no precipitation in July and August. Between 60 to 70% of precipitation takes place between December and February. The dam construction will be accompanied by a treatment station with a capacity of 60,000 m³/day. After its completion in the year 2006 it will feed potable and domestic water for several high altitude villages in the Kesrouane Caza.

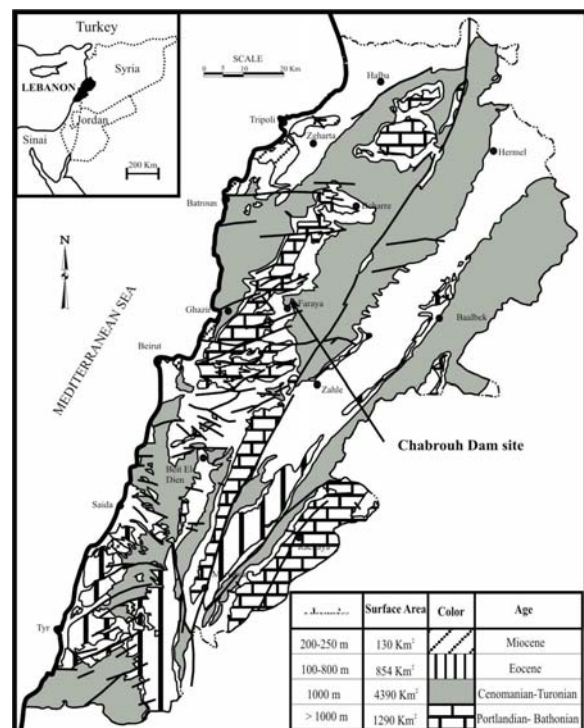


Fig. 1. Location map showing the distribution of karstic terrains in Lebanon and the location of Chabrouh Dam (EDGELL, 1997)

The water level in the reservoir behind the dam is planned to reach an elevation of 1612 masl. This level will be sustained if there was no major loss in the reservoir. Reservoir losses were a major problem in the El Qaraoun Dam reservoir which was constructed mainly on Jurassic karstic rocks. The Chabrouh Dam reservoir lies on mainly low permeability clay, shale and volcanic rocks except in its highest levels above 1500 m.

For this reason the El Qana Plateau, which forms the western borders of the reservoir, will be under comprehensive investigation.

The investigation utilized information from stratigraphy, structural geology, speleology, karstology, and hydrogeology. Those criteria proved useful in better understanding and the possible prediction of leakage of the reservoir water to the adjacent El Qana Plateau through karstic routs, possibly feeding the major outlets of the El Qana Plateau.

2. Methodology

Stratigraphy

The lithology underneath the Chabrouh Dam Reservoir mainly consists of volcanic tuffs, green to grey marls and marly limestone rocks. They are Albian in age and belong to the Hammana Formation. The lower layers are mainly volcanic tuffs and marls and the upper layers are mainly marly inter bedded with thin layers of marly limestone and limestone (Fig. 2 & 3).

The El Qana Plateau on the other hand is capped by a thick mass of limestone, dolomitic limestone, dolostone, and marly limestone. These lithologies belong to the Sannine – Maameltain Formation and are Cenomanian – Turonian and Upper Albian in age (WALLEY, 1997; Fig. 2 & 3).

The boundary between the thick Cenomanian – Turonian limestone and dolomite and the Albian marls and volcanic ash lies at approximately 1600 masl.

The bedding plane in the El Qana Plateau shows a slight inclination, less than 10° towards the NW. Some exceptions do exist especially near disturbances and in the western side of the plateau.

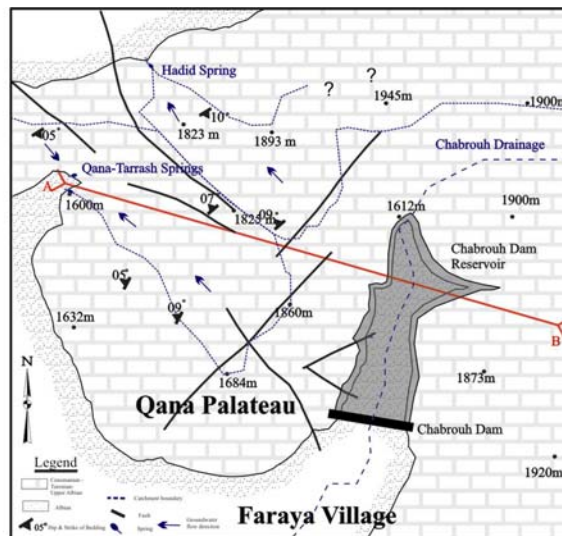


Fig. 2. Schematic hydrogeological map and cross section of Chabrouh Dam and adjacent Qana Plateau

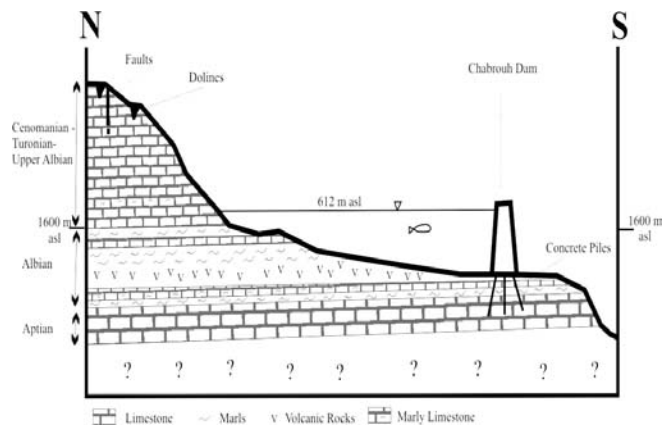


Fig. 3. N-S schematic cross section showing the stratigraphy underneath Chabrouh Dam Reservoir

Structural Geology

Approximately 70 lineaments were identified from aerial photographs (BOU JAWDEH *et al.*, 2002). When the directions were plotted on a rose diagram two dominant lineament orientations were identified these were NNE-SSW and NW-SE (Fig. 4).

Major faults were identified and two dominant orientations were observed one having a NW-SE direction and the other is NE-SW. The average inclination of the fault planes is approximately 70° with a large percentage are almost vertical. The movement of the faults in the area showed a dominant dip-slip movement.

More than 1107 mesofractures were identified and recorded including joints, faults and shear fractures. A rose diagram (Fig. 5) was constructed from those mesofractures and it revealed two dominant orientations NE-SW and NW-SE. Mainly all the fractures including the dominant ones were steeply dipping ($>70^\circ$).

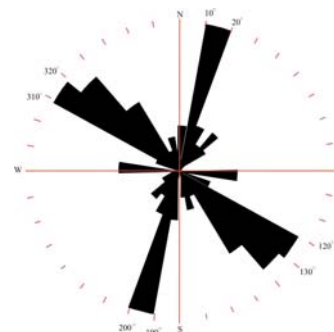


Fig. 4. Bi-directional rose diagram of 70 lineaments measured on the Qana Plateau

Speleology

Since 1998 cavers have been and are continuously discovering and mapping caves on the El Qana Plateau. Up to this date 5 vertical caves and 3 horizontal caves were discovered, explored and mapped (SCL archives). The deepest of the vertical caves located at an elevation of 1740 masl, named Houet el Qana, reaches a depth of 65 m and its direction of development was found to extend in the NW-SE direction (SCL archives). The longest of the horizontal caves has the same direction of development of approximately 1 km. This cave, named the Terrach cave, is located at the western side of the El Qana Plateau approximately 3 km from Chabrouh dam at an elevation of approximately 1630 masl (Figure 2).

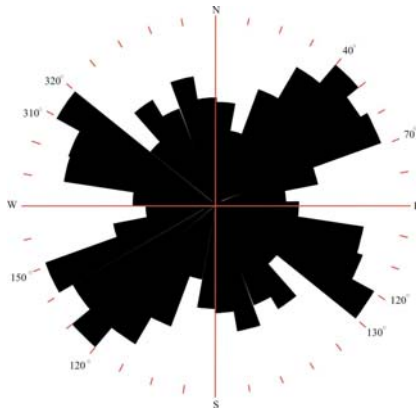


Fig. 5. Bi-directional rose diagram of 1107 mesofractures measured on the Qana Plateau

Karstology

In the El Qana plateau which covers an area of approximately 8 km² about 150 dolines were documented (BOU JAWDEH *et al.*, 2002). The size of these dolines range from 10's of meters in diameter up to 100 m. The dolines were dominantly funnel shaped with a flat bottom made up of terra-rosa red soil. Only 10 of the dolines showed 2-3 m deep ponors. About 100 of those dolines showed elongate shape while the remaining 50 are almost rounded. Measuring the orientation of the long axis of the 100 dolines and plotting it on a rose diagram revealed a dominant NW-SE direction (Figure 6).

Karstification in the Cenomanian-Turonian and Upper Albian is quite intensive and it reaches deep in this formation up to an elevation of 1600 masl close to the boundary with the Albian. However, the thinly bedded limestone in the Albian showed also some karstified beds, some of these found at an elevation of 1580 masl where disturbances were encountered.

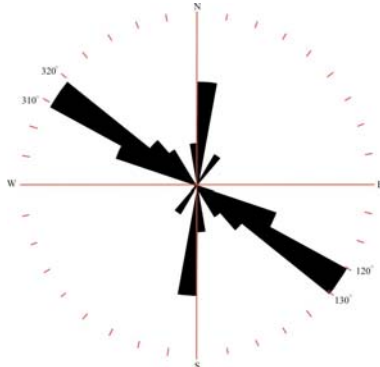


Fig. 6. Bi-directional rose diagram of the long axis of 100 dolines measured on the Qana Plateau

Hydrogeology

The Cenomanian-Turonian and upper Albian rocks form locally a significant karstic aquifer mainly due to their thickness (ca. 600m), wide exposure (i.e., catchment area) and karstic nature. The Albian rocks, on the other hand, are believed to act as an aquiclude due to its impervious marl and volcanic tuff layers.

The Cenomanian-Turonian and Upper Albian aquifer caps the top of the El Qana Plateau. Underneath this aquifer the Albian aquiclude make up the substratum. This lithological difference between the two formations and the gentle dip of the bedding towards the NW resulted in a hydrogeological boundary defined by a spring line at an elevation of 1600

masl. Those springs are the natural outlets of this karstic aquifer. More than 24 springs were identified.

More than 90% of the springs dry during the summer season, however, two form the major outlets of this aquifer. Those are the Qana-Tarrash Springs located on the eastern side of the El Qana Plateau and the Hadid spring located on the northern side at an elevation of approximately 1600 masl (Fig. 2). Between February 1999 and October 2000 the discharge of these major outlets was periodical measured. It thus revealed a steep recession curves typical of karstic springs (Fig. 7). During the snow melting season the discharge increases to as much as 1m³/s and during the dry season it declines to less than 0.01m³/s (BOU JAWDEH *et al.*, 2002). Snow between December and February of the year 2000 reached a height of more than 2 meters in some places. This caused a major access problem and flow measurements could not be taken.

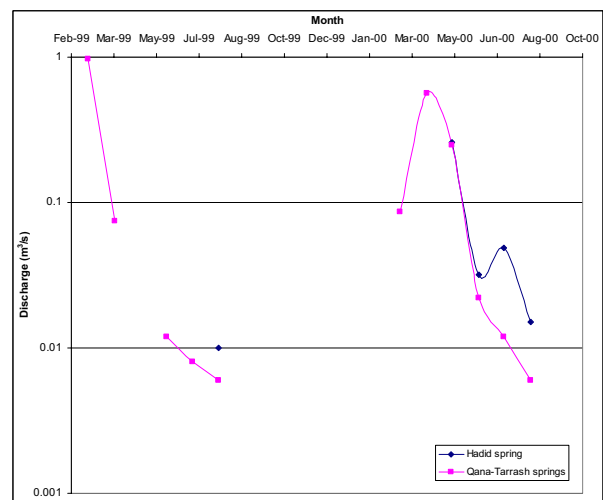


Fig. 7. Hydrographs of Hadid and Qana-Tarrash springs

Infiltration rates, in the Cenomanian-Turonian and Upper Albian of the El Qana Plateau, are expected to reach as high as 60%. This along with the precipitation data could give us a rough estimate of the catchment areas of each major spring. They are expected to be about 3-5 km² for both Qana-Tarrash springs and the Hadid spring.

The dominant direction of development of dolines, the extension of lineaments, faults, and mesofractures and the development of caves all seem to follow a NW-SE direction. This reveals that the major groundwater movement is in the NW-SE direction toward the major outlets that are the Qana-Terrash springs and Hadid springs (Fig. 2). The expected theoretical hydraulic gradient is probably in the range of 0.09-0.005.

Flow in the unsaturated section of this aquifer mainly follows downward and lateral movement along the NW features, the bedding planes respectively. However, in the saturated zone which is few 10's of meters above the 1600 masl mark, the flow is expected to be dominantly horizontal along both bedding planes and NW oriented features. The saturated zone is expected to be irregular and not continuous due to the karstic nature of the aquifer. It is expected to be relatively thick during the wet season and decrease dramatically during the dry season.

2. Effect of Reservoir

The water level in the Chabrouh Dam reservoir is expected to reach approximately 1612 masl. While karstification, as we have seen before, reaches a depth of 1600 masl in the El Qana Plateau. Therefore, the water level in the Chabrouh dam reservoir is approximately 12 m above the karstification level. That might give rise to a major change in head on the eastern side of the aquifer. This might give rise to leakage from the reservoir into the aquifer all along that 12 m band. In addition the aquiclude forming the lower hydrogeological barrier for the karstified Cenomanian-Turonian and upper Albian starts at approximately 1600 masl. This also coincides obviously with the spring line.

All this gives us a clear indication that the water above 1600 masl in the Chabrouh Dam reservoir if not properly contained would follow the gently NW dipping bedding planes and NW-SE karstified channels and feed the major outlets of the El Qana Plateau (Fig 2.).

In addition to that karstification was observed in some thin limestone beds at elevations of 1580 masl. Leakage also might take place in those thin karstified limestone beds of the Upper Albian. Flow might also follow the NW dip in bed and the NW-SE karstified channels if existing.

All these lead to the possible prediction that major leakages might occur on the western side of the reservoir through the well karstified rocks of the El Qana Plateau. The water will follow both the NW dipping karstified bedding planes and the NW-SE karstified features and will head in the direction of the Qana-Terrash springs and Hadid spring. If monitored those springs might show some rise in their discharge after the completion of the dam

3. Conclusion

The El Qana Plateau is located on the western side of the Chabrouh dam and its reservoir. Groundwater flow underneath the El Qana plateau is expedited to follow NW-SE orientations as extrapolated from the development of dolines, lineaments, faults mesofractures and caves.

Karstification is expected to reach a depth of approximately 1600 masl. The water level in the reservoir after the dam completion in the year 2007 is expected to reach approximately 1612 masl.

This will make the reservoir prone to leakage between elevation of 1600 and 1612 masl. Leakage on the western side of the reservoir might be detected from the discharge measurement of the major outlets mainly Qana – Terrash and Hdid springs.

Therefore, monitoring of the major outlets in the El Qana Plateau is important and any noticeable increase in the discharge might give a clear indication of possible leakage from the Chabrouh Dam reservoir into the aquifer. On the other hand leakage at that level might take place in the northern and eastern sides of the reservoir for similar hydrogeological conditions are expected to exit.

Therefore it is highly doubtful that the level of the water in the reservoir will be sustained at 1612 masl during low recharge periods. Further investigation and monitoring are required for better prediction and possible treatment of the problem at hand.

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Etude de faisabilité de la recharge artificielle des formations calcaires en zone semi-aride. Cas de l'oued Tagmoute, (Anti-Atlas occidental, Maroc)

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Résumé

Le bassin versant de l'oued Tagmoute a été pris comme exemple, pour étudier la faisabilité d'un barrage d'écrêtement pour la réalimentation de la nappe alluviale et par conséquent des calcaires sous-jacents à comportement karstique. L'intérêt de cette étude réside dans le fait qu'elle constitue un modèle de gestion d'eau de bassins non équipés de stations hydrométriques. L'ensemble des données de débits utilisés sera calculé par modélisation mathématique en se basant sur les observations et les caractéristiques géomorphologiques du bassin ainsi que sur les suivis des chroniques de pluies enregistrées. L'étude des paramètres physiques et hydrodynamiques des deux aquifères, alluvial en surface et calcaire en dessous, montre une importante possibilité de stockage des eaux de crues. L'écrêtement de ces crues par le barrage, permet d'augmenter le temps de contact eaux-complexe «alluvions-calcaires». L'élaboration de modèles mathématiques pour le calcul des volumes infiltrés vers les calcaires profonds prévoit un apport d'eau vers la nappe 12 fois supérieur au volume infiltré en conditions naturelles. Ce modèle type d'étude permet de prédéterminer d'avance l'intérêt et la faisabilité d'un aménagement hydraulique pour la recharge des nappes en zone semi-aride.

Mots clés : Recharge artificielle, Nappe calcaire, climat aride, barrage d'écrêtement

1. Introduction

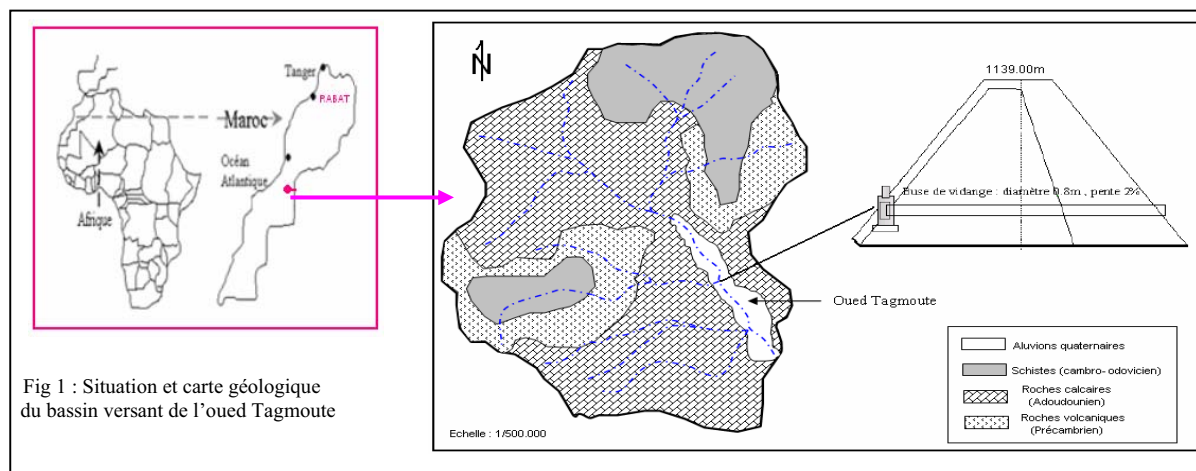
La nappe d'eau contenue dans les formations calcaires de la région aride de Tagmoute n'est alimentée que lors des événements de crues occasionnels et cette recharge ne concerne souvent qu'une partie du réservoir potentiel. Elle dépend principalement du pouvoir infiltrant des formations alluviales situées au dessus et de la durée du passage de la crue de l'oued.

L'aquifère calcaire est souvent vidé par les fuites vers l'extérieur ou par une surexploitation par pompes, il est ainsi possible d'accroître sa production en eau par une recharge artificielle graduelle par la création de barrages d'écrêtements sur le principal oued qui les draine (ROGNON, 2000).

La présente note technique a pour but de fournir un exemple de méthodologie d'étude de faisabilité d'un aménagement de réalimentation artificielle de la nappe des calcaires drainés par l'oued Tagmoute, pris comme exemple dans cette étude, adaptable aux conditions climatologique et hydrologique du site choisi.

2. Bassin versant de l'oued Tagmoute

L'exemple de la méthodologie d'étude proposée a été expérimenté sur le bassin versant de l'oued Tagmoute appartenant à une zone (Anti-Atlas occidental) à climat aride (fig. 1).



D'orientation générale NO-SE, ce bassin présente une superficie de 310 km² (fig. 1). Les altitudes y oscillent entre 1020 et 1870m avec une longueur du cours principal de l'ordre de 18.7 km et une pente moyenne $P_{moy} = 4.56\%$. Les précipitations annuelles sont généralement faibles et varient entre 35 et 420mm. Ces précipitations sont caractérisées généralement par leur violence et leur spontanéité qui engendrent, dans la plupart des cas, des écoulements avec de fortes vitesses de crues qui empêchent largement l'infiltration des eaux au niveau des formations alluviales et calcaires profondes. La nécessité d'intervenir sur le milieu naturel (DILLON, 2005) devient donc obligatoire pour faire décroître ces vitesses et permettre une recharge graduelle dans le temps des deux nappes qui s'abouchent en profondeur.

Le meilleur type d'ouvrage permettant d'atteindre ce but est la réalisation d'un barrage d'écrêtement type filtrant (TUINHOFF *et al.* 2003), qui réduira au maximum la vitesse de l'écoulement de l'oued en aval du barrage et une importante infiltration en amont de celui-ci dans la cuvette. L'importance d'un tel aménagement sera mise en relief en comparant les 2 volumes d'eau simulés infiltrés pendant le passage de la crue, en présence et en absence du barrage à réaliser.

3. Modélisation des écoulements de l'oued

Le dimensionnement du barrage d'écrêtement doit passer par une simulation des différents débits pour différentes périodes de retour. En absence des données de débits de l'oued, on s'est référé à une modélisation par la méthode du Curve-Number des précipitations efficaces pour la reconstitution de l'hydrogramme unitaire. Ce modèle a été mis au point par les géologues du Soil Conservation Service.

Les données des pluies maximales journalières de la station d'Ighrem (plus proche station/bassin), ont été ajustées suivant la loi statistique de distribution des probabilités de Gumbel permettant de calculer les différentes pluies correspondant aux différentes périodes de retour (Tab. 1). L'estimation aux temps de retour 50 et 200 ans a été faite à l'extrémité supérieure de l'intervalle de confiance (95%) avec ($P = \mu + 1.96\sigma$).

Temps de retour	2 ans	5 ans	10 ans	25 ans	50 ans	200 ans
Valeur centrale μ en mm	23	36	45	56	64	80
Ecart-type σ en mm	3.1	5.3	7.1	9.6	11.5	15.3
Pluies en mm	23	36	45	56	87	110

Tableau 1 : Pluies journalières calculées par adaptation statistique de Gumbel.

L'application du modèle du Curve-Number pour ces pluies journalières a montré que les averses de fréquence 2 et 5 ans ne produisent aucun écoulement, ce qui nous a poussé à recourir pour celles-ci au modèle cinématique de l'hydrogramme triangulaire qui se base sur le calcul du temps de concentration des eaux évalué au moyen des formules de Kiprich et de Giordotti. La figure 2 illustre les débits calculés pour la période de retour (T_r) de 2 ans.

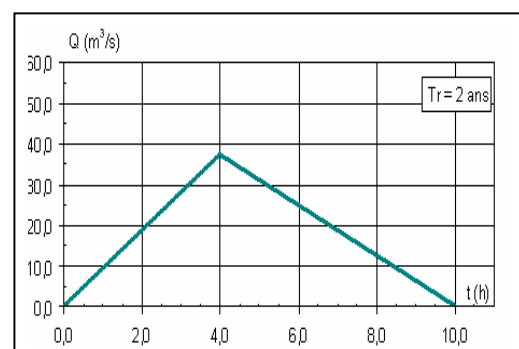


Fig. 2 : Hydrogramme de crue $T_r = 2$ ans.

Pour les autres périodes de retour, les calculs des pluies efficaces et des volumes infiltrés et évaporés ont été réalisés à l'aide du logiciel CNUH qui applique les formules empiriques du modèle Curve-Number. La figure (3) montre l'exemple de cette application pour la période de retour de 10 ans.

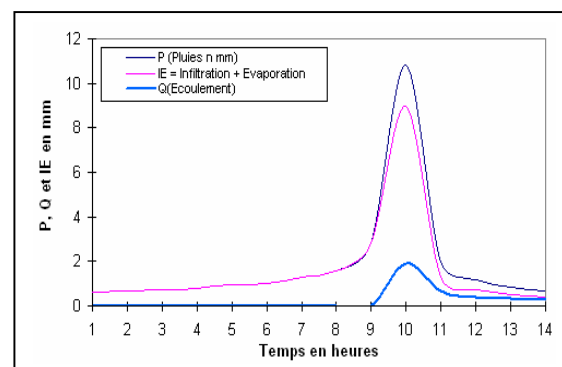


Fig.3 : Pluies efficaces calculées par la méthode du Curve-Number

Une addition des hydrogrammes unitaires appliquée à chaque terme de la pluie efficace permet d'établir la courbe des débits de la crue pour la période de retour (200 ans) adoptée pour le dimensionnement du barrage (Fig. 4).

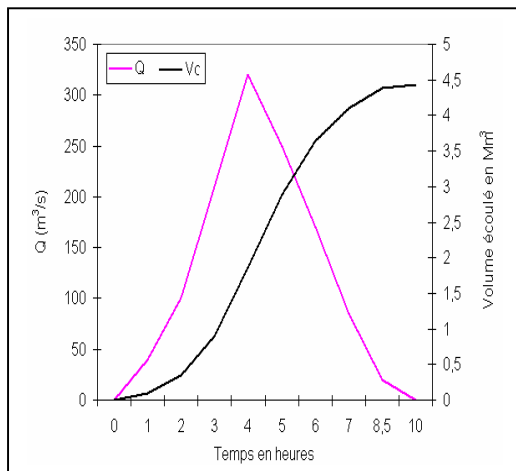


Fig. 4 : Graphique présentant la crue adoptée pour le projet (200 ans) et le volume correspondant total écoulé.

Capacité de recharge des aquifères alluvionnaire et calcaire

L'analyse de la carte géologique du bassin versant de l'oued Tagmoute montre l'abondance des formations calcaires de l'Adoudounien (CHOUBERT, 1963) qui occupent plus de 50% de la surface totale du bassin versant et dans lesquels l'oued a déposé ses produits d'érosion sous forme d'alluvions dont les dépôts épousent les lits actifs de l'oued. Les deux nappes alluviale et calcaire se retrouvent en contact direct en profondeur.

Les investigations de terrain réalisées au cours de cette étude, nous ont permis de déceler un site favorable situé à 10 km en amont de la gorge de l'oued pour la réalisation du barrage d'écêtement des eaux qui permettra d'augmenter la durée de contact entre l'eau et les formations perméables.

Il présente des caractéristiques géomorphologiques (berges constituées de dalles calcaires compactes avec des altitudes assez importantes, déversoir naturel, altitude importante des berges), qui permettent de minimiser au maximum le coût financier de l'aménagement.

Les formations calcaires et alluvionnaires, qui feront l'objet de cette recharge d'eau, concernent celles de la cuvette du barrage et celles situées entre le barrage et la gorge de l'oued en aval. Leurs capacités ont été approchées par deux méthodes : 1) la prospection géophysique surtout électrique pour le calcul des épaisseurs et 2) le traitement des photos aériennes et relevés topographiques pour le calcul des superficies. L'épaisseur moyenne calculée est estimée à 22m correspondant à un volume total de stockage des alluvions de l'ordre de 10,5 hm³. Les essais de pompage réalisés dans les formations calcaires et mêmes alluvionnaires montrent une valeur moyenne du coefficient de perméabilité de l'ordre de $K = 2,4 \cdot 10^{-3} \text{ m/s}$ et un coefficient d'emmagasinement de

l'ordre de $S = 0,20$. Une estimation du volume total des eaux qui peuvent recharger l'aquifère alluvial est de $10,5 \text{ Hm}^3 \times 0,20 = 2,1 \text{ Hm}^3$. Ce volume montre une possibilité de stockage des eaux importantes du complexe aquifère alluvions-calcaires.

4. Paramètres hydrologiques de dimensionnement du barrage

Le proportionnement du déversoir et de la retenue du barrage est basé sur l'extraction de la courbe Hauteur – Superficie – Volume, caractéristique de la cuvette, elle même calculée sur la base des relevés topographiques au 1/1000 réalisés. La cote de 1139 m ($H_0=1127\text{m}$), adoptée pour la retenue permet une extension de la cuvette à partir du barrage sur une longueur totale de 1000m au nord permettant d'avoir un volume d'eau emmagasiné supérieur à 4,435 hm³ calculé pour une crue de période de retour de 200 ans adoptée pour le projet du barrage. Le calcul des différents volumes d'eau, associés au réservoir, ont été réalisés par le logiciel ECRET qui utilise les différentes formules hydrauliques relatives au calcul des débits de déversoir et de la buse de la vidange du barrage.

5. Impact du barrage sur la nappe des calcaires

Volume des eaux infiltrées en conditions naturelles en absence du barrage

La durée du contact Eau – Complexe «alluvions-calcaires» en absence du barrage est égale à la durée de la crue. Le calcul du volume d'eau infiltré intéressera essentiellement la zone de la cuvette du futur barrage et le tronçon compris entre le site du barrage et la gorge en aval. La longueur totale de ce tronçon est de 2200 m, sa superficie est de 474.000 m² pour une largeur moyenne de 215m. L'infiltration s'opérera principalement dans le lit actif de l'oued dont la largeur ne dépasse guère les 50m. La superficie A du contact eau-complexe «alluvions-calcaires» atteint 110 000 m² ($= 2200 \times 50$).

Pour le calcul des volumes d'eau infiltrés vers la nappe alluviale, on s'est référé à la crue de période de retour de 2ans dont la durée de passage est estimée à 10 heures avec un débit de pointe de 37 m³/s correspondant à un volume total d'eau écoulé de 0,67 Hm³. Le volume d'eau effectivement infiltré a été calculé par la formule suivante (Loi de Darcy) : $\text{VOL.inf} = A.t.V_i$.

A : superficie du contact eau-terrain,

V_i : vitesse d'infiltration de l'eau dans le terrain,

t : durée du contact.

Le volume d'eau effectivement infiltré est de 0,044 hm³ correspondant uniquement à 6,6% du volume total de la crue (Fig. 6).

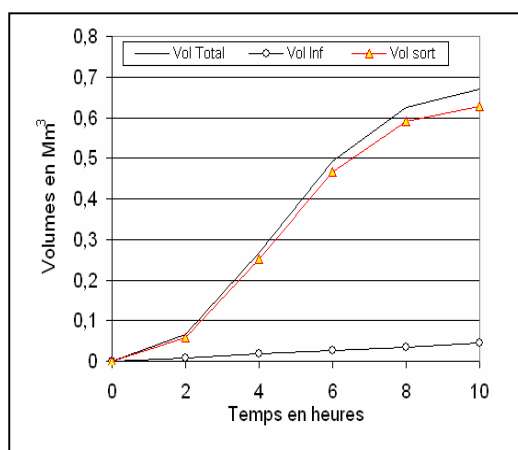


Fig 6 : Volumes d'eau infiltrés et écoulés en absence du barrage.

Volume d'eau infiltré en présence du barrage

Une fois le barrage réalisé, l'infiltration de l'eau dans les alluvions sera effectuée aussi bien dans la cuvette en amont du barrage qu'en aval de celui-ci dans le tronçon compris entre le barrage lui-même et la gorge de l'oued. L'utilisation du logiciel ECRET pour le calcul des volumes d'eau en présence du barrage montre que la durée de passage de l'onde de crue, qui était de 10h en absence du barrage, a été prolongée de 14 fois (150h), permettant un temps de contact plus long entre l'eau et les alluvions (fig. 7).

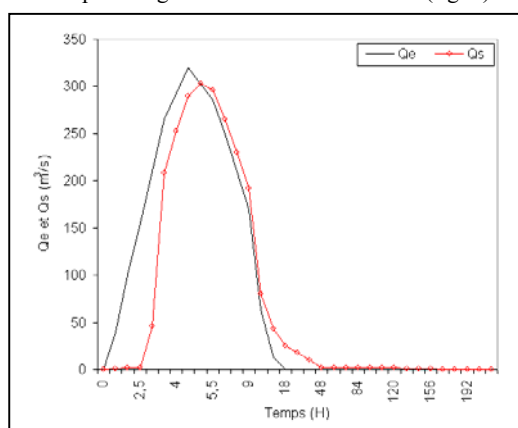


Fig. 7 : Ecrêtement de la crue ($T_r = 200$ ans) par effet du barrage.

Au total un volume d'eau de 0.37 hm^3 s'infiltré dans la cuvette et à 0.30 hm^3 d'eau sortent du barrage. Une partie de ce dernier volume s'infiltrera au niveau des alluvions dans le tronçon compris entre le site du barrage et la gorge de l'oued en aval. En adoptant les mêmes formules que pour le calcul des volumes infiltrés en absence du barrage, on a pu estimer les volumes d'eau infiltrés en aval de celui-ci jusqu'à la gorge de l'oued, il est de l'ordre de 0.17 hm^3 . La somme des deux volumes d'eau infiltrés dans la

cuvette et en aval du barrage est de 0.54 hm^3 , correspondant à 12 fois le volume infiltré en conditions naturelles (fig.8).

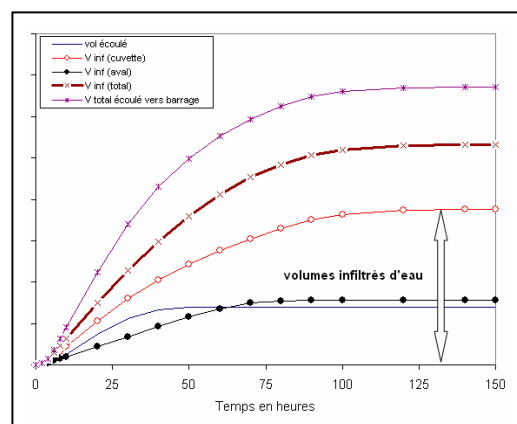


Fig 8: Graphique montrant les volumes infiltrés par effet du barrage d'écrêtement.

6. Conclusions

L'étude de faisabilité de la recharge artificielle du complexe «Alluvions-Calcaires» drainé par l'oued Tagmoute est basée sur une modélisation mathématique des débits à différentes périodes de retour. Le calcul des volumes infiltrés en absence et en présence du barrage a été fait par le logiciel ECRET qui nous a permis d'estimer l'apport en eau vers la nappe. Le prolongement du temps de contact entre la nappe. Le prolongement du temps de contact entre le complexe «alluvions-calcaires» et les eaux de la crue, par l'écrêtement du barrage a induit, un volume infiltré 12 fois plus important qu'en conditions naturelles. Cette étude a permis d'évaluer l'intérêt et la faisabilité d'un barrage d'écrêtement en zone semi-aride.

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Apport de la combinaison entre le traitement des images satellitaires et la prospection géophysique à la détermination des écoulements préférentiels en zone karstique

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Résumé :

Malgré le grand nombre de puits et forages d'eau réalisés dans les différentes régions karstiques du Maroc, de nombreuses inconnues subsistent encore concernant les critères d'implantation et le coût final des ouvrages. Afin d'optimiser les campagnes de forages à venir, une méthodologie de prospection multicritères a été adoptée dans la région karstique nord-orientale du Maroc. Elle s'articule selon trois axes à savoir :

- 1) un traitement visuel des images satellitaires pour la détermination des linéaments,
- 2) Une application de la géophysique électrique par a) sondages électriques pour la détermination des épaisseurs des niveaux géologiques et b) par trainés électriques pour la localisation précise de la zone de passage des linéaments,
- 3) et finalement une application de la résonance magnétique des protons pour la vérification de la présence d'eau. Les différents résultats de sondages mécaniques réalisés à la suite de cette campagne de prospection montre un taux de réussite très satisfaisant qui a atteint 75% de réussite.

1. Introduction et problématique

Satisfaire d'une manière urgente et immédiate les besoins en eau des populations, surtout des régions karstiques à déficit hydrique, conduit à une précipitation dans la réalisation des forages. Ce mode d'action justifie de nombreuses insuffisances : tâtonnements fréquents, pertes énormes de temps, d'argent et de matériel. La présente note technique a pour but l'élaboration d'une méthodologie d'étude qui permet de cibler des sites répondant aux deux critères suivants :

1- meilleur potentiel aquifère souterrain localisé à des profondeurs accessibles et à des coûts compatibles avec les budgets attribués;

2- recherche d'un nombre suffisant de cibles pour sécuriser les investissements et la gestion durable de la ressource sur le long terme.

Cette réponse prend la forme d'une étude combinée entre un outil cartographique d'aide à la décision (traitement d'image satellitaire) accompagnée d'une prospection géophysique.

2. Méthodologie adoptée

La démarche scientifique multicritère utilisée repose sur les outils et techniques de la télédétection, de l'analyse morpho structurale, de la pédologie, des formes de végétation et de la géophysique notamment

Les deux méthodes de prospection électrique et de résonance magnétique des protons (RMP).

Ces techniques apportent des informations pertinentes sur la présence d'une ressource exploitable dans les 100 premiers mètres de profondeur.

La première phase consistant au traitement visuel de l'image satellitaire permet de détecter les linéaments

pouvant drainer les eaux infiltrées au niveau du plateau calcaire de Deboud, très karstifié, vers la plaine de Tafata. Le but essentiel est de localiser les drains karstiques sous les formations alluviales de la plaine. Ces linéaments se caractérisent au niveau des images satellitaires par des éléments rectilignes ou curvilignes (HOBBS 1904, O' LEARY, ROWAN et al., 1974; SIEGEL & ABRAM, 1976) que l'on peut percevoir à la surface de la terre et qui traduisent des phénomènes plus profonds comme par exemple des failles, des fractures ou des contacts géologiques.

Il s'ensuit en seconde phase, une prospection géophysique pour la vérification, d'une part de la zone précise du passage du linéament par sondages et trainés électriques, et d'autre part la capacité de drainage de ce linéament par la Résonance Magnétique des Protons.

3. Contexte géologique de la zone prospectée

La plaine de Tafata, région choisie pour l'application de cette méthodologie, se présente sous forme d'une dépression synforme située à l'extrémité NE du massif du Moyen-Atlas.

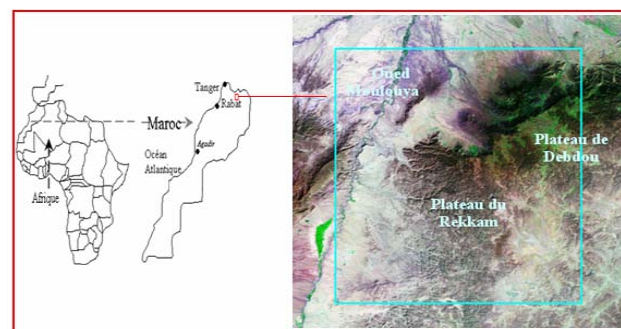


Fig. 1 : Image satellitaire de la zone d'étude

Ce dernier constitue la branche septentrionale du Haut-Atlas qui vient s'envoyer dans ce secteur sous les formations tertiaires du remplissage du bassin.

Les formations géologiques, présentes d'âge Jurassique, sont caractérisées par une lithologie à dominante calcaréo-dolomitique constituée de la base au sommet par 1-des argiles rouges à grises probablement attribuées au Toarcien, 2-des marno-calcaires attribuées à l'Aalénien et 3-des dolomies massives attribuées au Bajocien (fig. 2).

Âge	Log	Lithologie	Épaisseur (m)
Quaternaire		Limons, argiles, cailloutis, conglomérats et croûtes calcaires lamellaires.	<100
Plio-Villafranchien		Calcaires lacustres ou conglomérats.	<50
Mio-Pliocène		Continental. Grès et conglomérats, marnes à gypse de couleur variée et bancs de lignite.	?
Miocène (Tortonien)		Marin transgressif. Affleure à l'VO et au S de la plaine. Marnes blanches gypseuses.	>500
Kiméridgien, Lusitanien, Callovo-Oxfordien		Calcaires dolomitiques.	
Dogger		Grès, Dolomies grises, calcaires sub lithographiques gris avec au sommet des intercalations marnées. Faciès récifal du Bouloulou à madréporaires.	300 à 400
Lias (Domérien)		Concordant sur le Lias inf. Calcaires rés.	100
Lias inf.		Calcaires noirs et dolomies massives avec quelques bancs marneux.	
Trias		Argiles rouges bariolées à sel et gypse. Niveaux isolés à sables dolémitiques. Primaire et Trias = substratum imperméable.	
Primaire		Schistes violacés ou brunâtres à bancs de quartzites. Présence d'un batholite de granite hercynien.	

Fig. 2 : Log stratigraphique de la zone d'étude

D'une manière générale, la série du Jurassique supérieur repose en contact discordant sur le socle constitué de schistes et de calcaires massifs fortement fracturés ayant subi d'intenses phénomènes d'érosion.

Au nord de la plaine, les reliefs du jbel Lahlowa sont constitués de séries structurées en un pli complexe probablement déversé vers le SE au contact d'un accident inverse de direction atlasque (N60°E).

Le socle affleure dans le vaste massif de Debdou structuralement déconnecté des hauts plateaux calcaires par la vallée de Debdou. Celle-ci de direction ENE/WSW, souligne probablement un accident majeur de même direction. Sur les hauts plateaux, l'observation de l'image satellitaire met bien en évidence une variation de faciès dont on peut situer la limite aux abords du linéament profond évoqué plus haut. La cuesta principale, formant la falaise dominant le cirque de Rhida et la vallée de Debdou, présente un pendage faible vers l'ouest et le sud, de telle sorte que la totalité des couches se retrouvent au dessous de leur côté topographique observé sur le plateau.

Le versant abrupt N de la cuesta dolomitique est affecté par des failles de direction générale N60°E responsables de la déconnexion de panneaux constitués de la série marno-calcaire de l'Aalénien et dolomies du Bajocien, puis de leur effondrement dans la dépression. Ce secteur correspond à la bordure faillée du Môle sur lequel reposent les plateaux calcaires.

La présence d'aquifères potentiels dans ces formations pourrait s'expliquer par le piégeage d'eau dans les faciès réservoir calcaréo-dolomitiques (ou à la limite Jurassique/Socle). Les écoulements préférentiels seraient alors guidés par les failles et pourraient alimenter les blocs adjacents par connexion latérales ou verticales. L'observation rapide de ces affleurements a révélé une fracturation complexe qu'il convient de ne pas négliger : 1- grandes fractures ouvertes et/ou partiellement colmatées par des produits d'érosion marneux en surface, 2- présence de failles normales mettant en contact des bancs calcaires sub-lithographiques et des dolomies karstifiées et 3- des fractures orthogonales ouvertes puis entièrement colmatées par une précipitation de la calcite. Ces failles jouent un rôle important dans le drainage des eaux à partir des plateaux calcaires vers la plaine. Un traitement numérique de l'image satellitaire couvrant cette zone pourrait préciser les zones du passage de ces failles.

4. Résultats obtenus

Apport du traitement de l'image satellitaire

La première étape consiste en une interprétation morpho structurale de l'image satellitaire couvrant la zone d'étude. Celle-ci, de type Landsat (1/50 000), a été obtenue à partir du site de la Nasa. L'interprétation de cette image a été réalisée de manière visuelle et a fait l'objet d'une structuration en Système d'Information Géographique (SIG). Une carte des linéaments a été dressée à la suite de l'interprétation visuelle sur écran de l'image satellitaire (fig. 3).

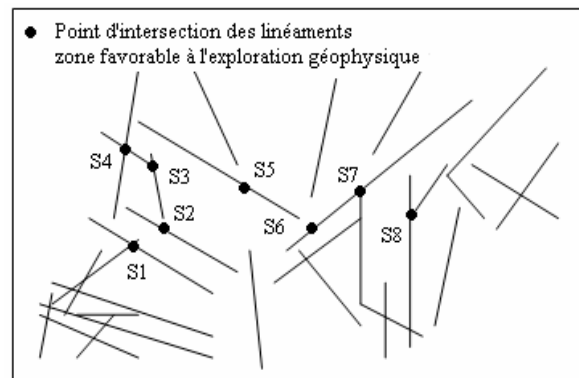


Fig. 3 : Carte des linéaments majeurs.

Elle montre un réseau de fractures de différentes directions dont la plus prépondérante est la direction N50 qui correspond relativement à la même direction N50-70 relevée au niveau des affleurements (fig. 4). Plusieurs cibles ont été choisies au croisement de linéaments importants (fig. 3) ou dans la prolongation de certains linéaments masqués par le remplissage des formations récentes (cas des cibles S5 et S6). Il faut signaler qu'au total 128 linéaments ont été détectés dont les plus importants ont été cartographiés (32).

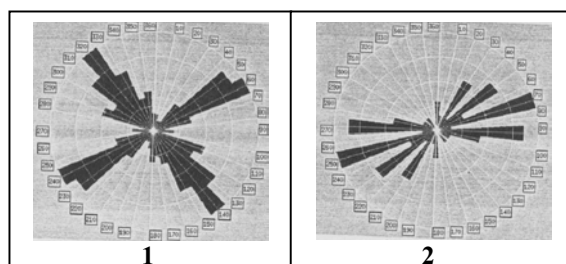


Fig. 4 : Rosaces de directions des :
1- linéaments (Nombre total : 128),
2- fractures (Nombre total : 300/ 9stations).

Elles ont fait l'objet d'une prospection géophysique par trainés électriques pour le positionnement exact du passage du linéament et de la résonance magnétique des protons (RMP) pour la vérification de la présence d'eau en quantité importante.

Sur les 8 points prospectés, tous ont présenté des anomalies conductrices correspondant au passage de linéaments. Cependant, seule la cible S1 a fourni un fort signal RMP correspondant à la présence d'eau en grande quantité en profondeur. Pour les besoins de cette note, un intérêt particulier sera porté pour détailler les résultats de cette cible.

Prospection électrique

Un sondage électrique vertical a été réalisé à l'aplomb de la cible S1 déterminée à partir du traitement visuel de l'image satellitaire. Ce sondage a permis de déterminer l'épaisseur des différents niveaux géologiques en profondeur. L'interprétation par optimisation du modèle introduit à 5 couches du sondage montre la succession des couches géoélectriques suivantes (fig. 4 et tab. 1) :

Couche n°.	Résistivité (ohm.m)	Epaisseur (m)	Prof. toit (m)	Cote du toit (m)
R _m : 1	800	1.60	0.00	716.63
C ₁ : 2	333	4.90	-1.60	715.03
C ₁ : 3	107	4.30	-6.50	710.13
R ₁ : 4	337	10.30	-10.80	705.83
R ₁ : 5	1230	27.70	-21.10	695.53
C ₂ : 6	100	inf	-48.80	667.83

Tableau 1 : Résultats du sondage électrique S1

- un dépôt résistant superficiel (R_m) d'une épaisseur de 2m qui pourrait correspondre aux dépôts colluvionnaires,
- au dessous, un ensemble conducteur (C₁) d'une puissance de 9m correspondant aux argiles rouges,
- un second dépôt résistant (R_{1a}) dont le toit se situerait à une profondeur de 21m qui pourrait correspondre à une des formations marnocalcaires,

- un niveau plus résistant (R_{1b}) correspondant vraisemblablement aux calcaires sains du Dogger,
- et finalement un conducteur (C₂) qui pourrait correspondre aux calcaires fracturés recherchés.

L'identification géologique des différentes couches s'est basée essentiellement sur les observations des formations géologiques sur le terrain.

Les résultats du traitement mathématique de ce sondage ont permis de préciser les longueurs fixes (200m) des lignes d'émission du courant utilisés pour les trainés électriques. L'utilisation de ces derniers, par un balayage latéral, permet de localiser avec précision la position des fractures : failles et contacts anormaux (ASTIER, 1971). Les trainés à répétition avec une double longueur de ligne permettent d'estimer le pendage de ces anomalies en profondeur.

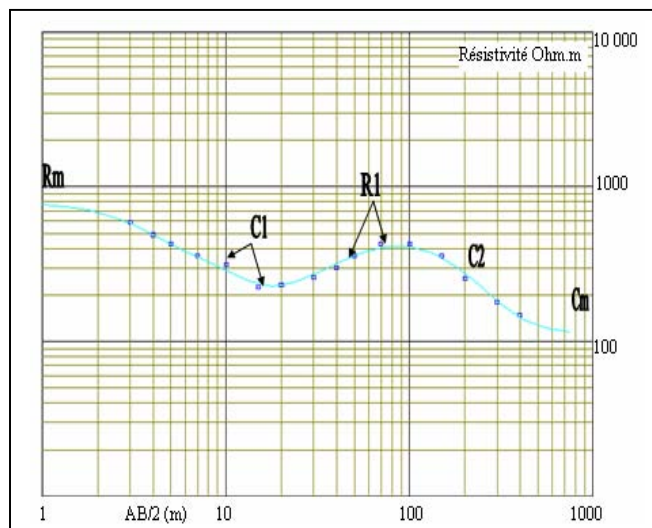


Fig. 5 : Sondage électrique au droit de la cible S1.

Un trainé électrique avec une longueur d'émission AB (200m) et des MN (20m) est réalisé à l'aplomb de la cible satellitaire S1 selon une direction orthogonale au sens du linéament permettant de couvrir la totalité de la zone où il passe. Il présente une longueur totale de 610m avec une direction de déroulage E-W (fig. 4).

Il montre une anomalie conductrice bien marquée au niveau de la station S17-18 avec des résistivités qui chutent depuis des valeurs de 1800 Ohm.m à la station S15 à 1000 Ohm.m au droit de l'anomalie.

Pour tester l'aptitude du drainage d'eau en profondeur par cette faille, un sondage RMP a été réalisé au niveau de cette anomalie.

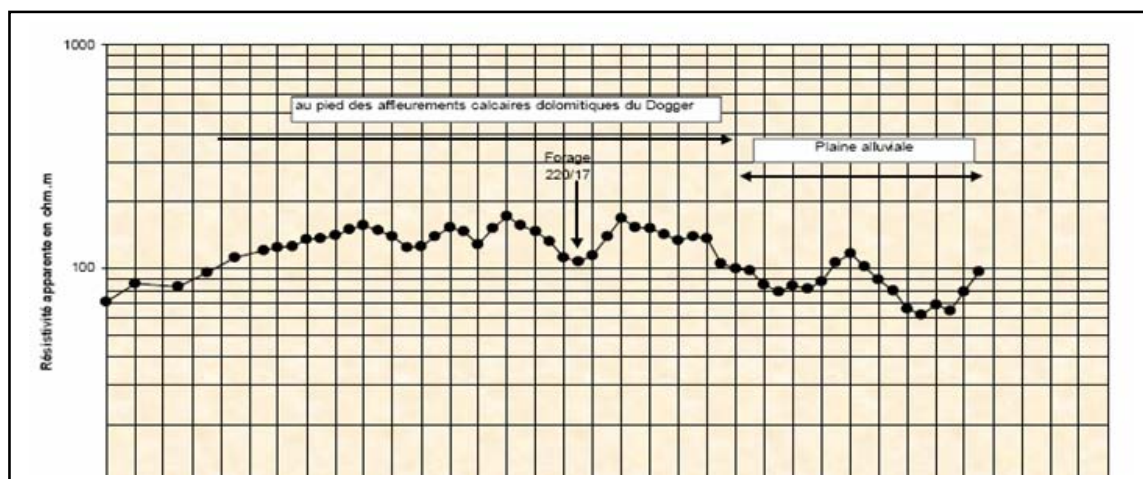


Fig 6 : Trainé électrique réalisée au droit de la cible S1

Résonance magnétique des protons (RMP)

Avant d'exposer les résultats de la RMP, il convient de rappeler son principe d'application ainsi que l'équivalence hydrogéologique des différents paramètres physiques mesurés.

Principe de la RMP

La RMP est une méthode de détection directe de l'eau à partir de mesures effectuées en surface. Son principe repose sur l'analyse du signal de résonance des noyaux d'hydrogène (ou protons) contenus dans les molécules d'eau en réponse à un signal électromagnétique de fréquence donnée (LEGCHENKO *et al.*, 2002). Le signal d'impulsion est créé par la circulation, dans un câble disposé en boucle au sol (fig. 5), d'un courant très puissant, pouvant atteindre plusieurs centaines d'ampères et produit par une tension électrique de plusieurs milliers de volts. Il crée un champ magnétique qui modifie l'équilibre énergétique des protons des molécules d'eau présentes dans le sous-sol. La coupure du courant au bout d'un temps bref - quelques dizaines de millisecondes - provoque un retour à l'équilibre des protons qui renvoient alors un signal de relaxation sous la forme d'un champ électromagnétique. Celui-ci, souvent faible, correspond à une différence de potentiel mesurable, de l'ordre de quelques nano volts. Son amplitude est directement proportionnelle à la quantité d'eau présente dans le sol (LUBCZYNSKI *et al.* 2003 ; LUBCZYNSKI *et al.* 2004). Le temps mis par le signal RMP pour disparaître (le temps de décroissance T2) fournit des indications sur l'environnement direct des nappes : il est d'autant plus long que les protons sollicités sont ceux d'une eau peu enserrée dans la roche, donc d'une nappe à potentiel hydrodynamique élevé. Les mesures du courant RMP sont sensibles au bruit électromagnétique (lignes haute tension, antennes radios, pompes hydrauliques...) et sont gênées par la présence de roches magnétiques.

Résultats de l'application de la RMP

Dans notre cas, un sondage RMP a été exécuté avec une boucle de 75x75m au droit de l'anomalie enregistrée au

niveau du trainé électrique.

Lors de l'acquisition et après inversion mathématique du sondage, un signal RMP a été observé entre la 13^{ème} et la 16^{ème} impulsion avec une amplitude maximale de l'ordre de 124 nV enregistrée à la 15^{ème} impulsion (fig. 6). Sur la totalité des autres sondages RMP exécutés dans cette étude, aucun n'a dépassé les 40 nV comme en témoigne les quelques sondages représentés dans la figure 7.

La courbe de variation du signal montre une forme en cloche avec une augmentation progressive des valeurs enregistrées entre la 9^{ème} et la 11^{ème} impulsion. Elles indiquent vraisemblablement l'humidité rencontrée à l'approche de l'eau (fig. 7). Le signal RMP correspondant à la présence d'eau, est enregistré entre la 13^{ème} et la 16^{ème} impulsion.

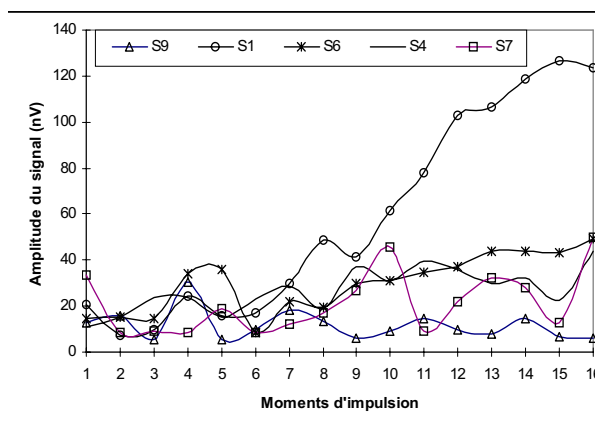


Fig 7 : Variation de l'amplitude du signal de quelques sondages RMP au cours des impulsions.

Les fréquences répétitives (1776 Hz) entre la 11^{ème} et la 16^{ème} impulsion, correspondant à la fréquence Larmor calculée à partir du champ magnétique local, montrent que ce signal correspond à une présence d'eau en profondeur.

Les valeurs positives et rapprochées de la phase (fig. 8) confirment ce résultat. La constante de temps T2 enregistre des valeurs importantes (400-450 ms) aux alentours des

impulsions (13-16) enregistrant le signal impliquant ainsi des conditions hydrodynamiques favorables :

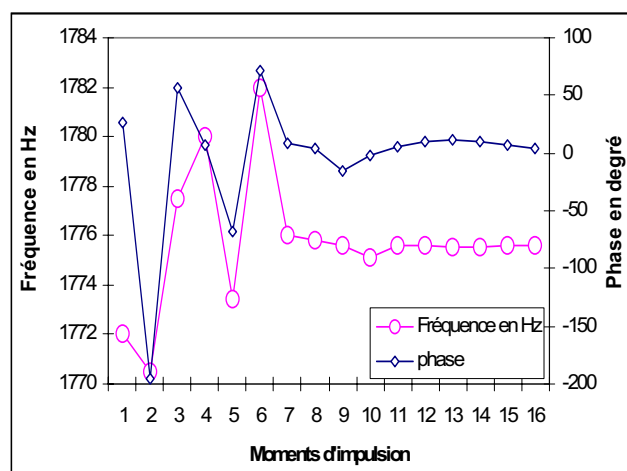


Fig 8 : variation de la fréquence et la phase lors de l'acquisition du sondage RMP S1

(Perméabilité de fissure importante dans cette tranche de profondeur (fig. 9)). Il faut signaler que les conditions d'acquisition du sondage sont idéales puisque les valeurs enregistrées du bruit ambiant pouvant altérer l'enregistrement du signal d'eau, sont relativement inférieures à 200 nV (fig. 9).

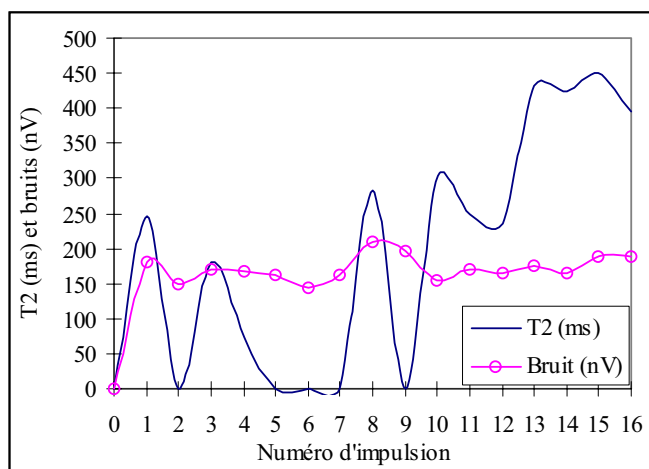


Fig 9 : variation de la constante de temps T2 et du bruit lors de l'acquisition du sondage RMP S1

L'inversion mathématique du sondage RMP (en utilisant la matrice calculée à partir du sondage électrique vertical réalisé) montre une tranche d'eau, correspondant au signal, dont le toit est situé à 44m de profondeur (fig. 10). Le volume d'eau calculé, par unité surfacique, correspondant à ce signal est de $V_{eau} = 3.2 \text{ m}^3/\text{m}^2$. Ce sondage a fait l'objet d'un forage mécanique de 80m de profondeur dont le débit après essais de pompage a atteint (144m³/h). Les résultats de ce forage corroborent ceux de la RMP puisque la venue d'eau a été observée à 42m de profondeur.

Cette étude montre l'importance de l'utilisation de la résonance magnétique des protons pour l'optimisation du choix des cibles pouvant faire l'objet de futurs forages positifs. Cependant, l'existence d'une seule cible présentant un potentiel d'eau important laisse penser que celle-ci est située dans une zone de recharge du linéament qui est très favorable.

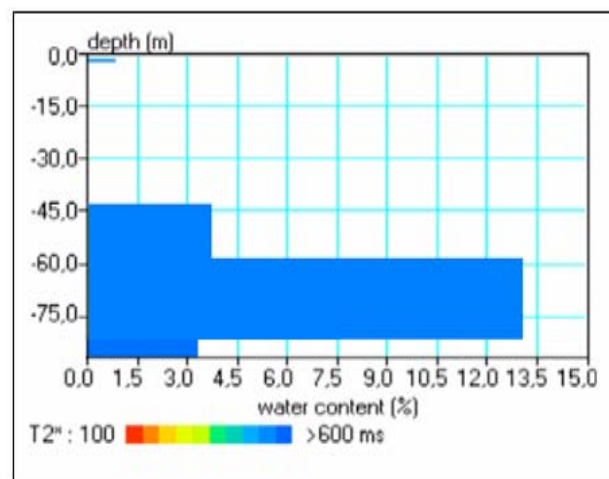


Fig 10 : Résultats de l'inversion mathématique du sondage RMP S1

Zone de recharge pour la cible S1

D'où vient l'eau alimentant la zone de la cible S1 ?

- Elle ne peut venir de la plaine de Tafrata qui se trouve sur un compartiment effondré.
- Elle ne peut venir non plus de la partie ouest puisque le socle affleure et forme un écran imperméable pour l'infiltration des eaux.
- Les quantités d'eau ruisselée et infiltrée sur la retombée du plateau sont trop faibles pour fournir une alimentation compatible avec les volumes issus des essais de pompage réalisés au cours de la foration du forage S1.

La seule origine viable est un drainage souterrain par un accident bordier WSW-ENE, déjà déterminé par le traitement de l'image satellitaire. Ce collecteur constitue un drain pour les eaux des oueds descendant du plateau de Rekkam sur une vingtaine de km du sud vers le nord.

L'un des critères les plus remarquables ressortant de l'interprétation des données satellitaires et la bonne recharge qui existe en périphérie des plateaux. La zone de ruissellement et d'infiltration de l'amont est de l'ordre de 400 km², elle est très peu végétalisée et ne présente pas d'utilisation agricole. Le potentiel drainé vers les aquifères est donc actuellement intacte. Par ailleurs, du fait de la présence d'un minimum annuel de pluie tombée (350 mm), on estime qu'à la profondeur explorée (100m), on se situe bien dans une zone de recharge annuelle et non pas dans un aquifère fossile.

En absence de budgets alloués à la réalisation d'études similaires dans cette zone, la meilleure stratégie d'implantation de futurs points est la proximité des accidents bordiers du plateau calcaire, tout en restant dans les compartiments formant les horsts. C'est ainsi que l'on a

le plus de chances d'atteindre une cible répondant aux exigences rapides et urgentes de différentes localités.

Conclusion :

Une méthode multicritères pour l'optimisation des sites d'implantation des forages a été appliquée dans la région de Guercif appartenant à la zone nord orientale du Maroc. Le traitement visuel de l'image satellitale a permis de cartographier l'essentiel des linéaments présents au niveau de cette région. La mise en évidence d'anomalies

conductrices au niveau des traînés électriques réalisés au droit des points de jonction des linéaments permet de localiser ceux-ci. L'inversion mathématique du sondage à résonance magnétique des protons réalisé au droit de l'anomalie électrique S1 montre un important signal correspondant à la présence d'eau en profondeur. Elle indique la présence de tranche d'eau dont le toit se situerait à 40m de profondeur. Le sondage mécanique réalisé confirme ces résultats puisque la venue d'eau a été rencontrée à 42m au niveau des formations calcaires fracturés. Cette méthodologie a prouvée sa fiabilité d'utilisation dans les régions hydrogéologiques discontinues à déficit hydrique.

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Bacteriophages retardation in the epikarstic zone

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Abstract

The purpose of the research was to study the infiltration and migration of health-hazardous human viruses, such as enteroviruses, in the unsaturated zone of fractured and karstified rock, since these rocks present important aquifers in Slovenia. As a possible model for behavior of health-hazardous viruses, we used the salmonella bacteriophage P22H5. After injection, bacteriophages remain in the fractures (channels) and microfracture systems of the unsaturated zone and are rinsed by subsequent larger precipitation events even up to several months after the injection. The field experiments have shown different flow patterns depending on the fractured rock structure. In the research area some fast conduits (large fractures or faults) exist where water runs faster than in the total conductive part of the rock. On the other hand the tracer delay in microfracture system areas was observed.

1. Introduction

Bacteriological pollution resulting from agriculture is the most important problem in groundwater protection, especially in karst springs discharging high karst plateaus. Solute tracers, particularly fluorescent dyes like uranine, are commonly used to characterise the fate and transport of solute contaminants, while colloidal tracers are appropriate to simulate the transport of microorganisms and contaminants that are attached to suspended colloids. Thus colloidal tracers like bacteriophages and fluorescent polystyrene microspheres are used to evaluate the hygienic risk for groundwater resources.

The purpose of the research was to study the infiltration and migration of health-hazardous human viruses, such as enteroviruses, in the unsaturated zone of fractured and karstified rock, since these rocks present important aquifers in Slovenia. As a possible model for the behaviour of health-hazardous viruses, salmonella bacteriophage P22H5 (400-800 nm) and fluorescent microspheres (yellow green, 1 µm) were used. Phages have served as useful models for the behaviour of human enteric viruses in water treatment processes because of their similarity to enteric viruses in structure, size, and resistance to inactivation (HEDBERG & OSTERHOLM, 1993).

The bacteriophage P22H5 is a virulent mutant that propagates in mouse typhoid fever bacteria *Salmonella typhimurium* and rarely occurs in waters (SEELEY, 1982). From previous tracing experiments (BRICELJ, 1986) it is well known that coliphages are a common constituent of faecally polluted waters and for this reason are not a suitable tracer, especially in the case of very high dilutions of the tracer, when a high or low background of coliphages may interfere with the tracer curve. The phage tracer P22H5 was injected at ten locations in 14 tracer experiments in running water and into the unsaturated zone in a karstic area where no background of phages for its host bacteria were present (BRICELJ, 2003).

The multi-tracer experiment was carried out in the subsurface zone, since the rate of microbial activity is assumed to be the highest in the upper parts of the unsaturated zone. This multi-tracer experiment was performed in the frame of the common project of the Association of Tracer Hydrologists (ATH), where several tracers were used: deuterium (90 %), potassium bromide, lithium chloride, zinc sulphate, sulfonic acid, pyranine, naphthionate, uranine, Sulforhodamine B, micro spheres and bacteriophages P22H5. The microspheres were counted at the University in Karlsruhe (GÖPPERT *et al.*, 2005) whereas the bacteriophages were counted at the National Institute of Biology in Ljubljana. The other tracers were analysed by colleagues from the ATH. The aim of this multi-tracer experiment was to quantify fast preferential flow and slow flow through microfracture system and to determine different behaviour of conservative tracers, which was observed in previous tracer experiments.

2. Research field site Sinji Vrh

A multi-tracer experiment with bacteriophages was performed on the research field site Sinji Vrh (ČENČUR CURK, 1997), Slovenia. It is located in the unsaturated zone of fractured and karstified Jurassic limestone at the edge of the Trnovski Gozd plateau (mean altitude 900 m a.s.l.), which is an overthrust of carbonate rock over Eocene flysch (Fig. 1). The subvertical Avče fault with a Dinaric direction NW-SE and several parallel faults cross this territory. These faults are interwoven with numerous connecting faults extending in the general direction N-S. Their intensity varies from open wide fractured zones to crushed and broken zones (JANEŽ, 1997).

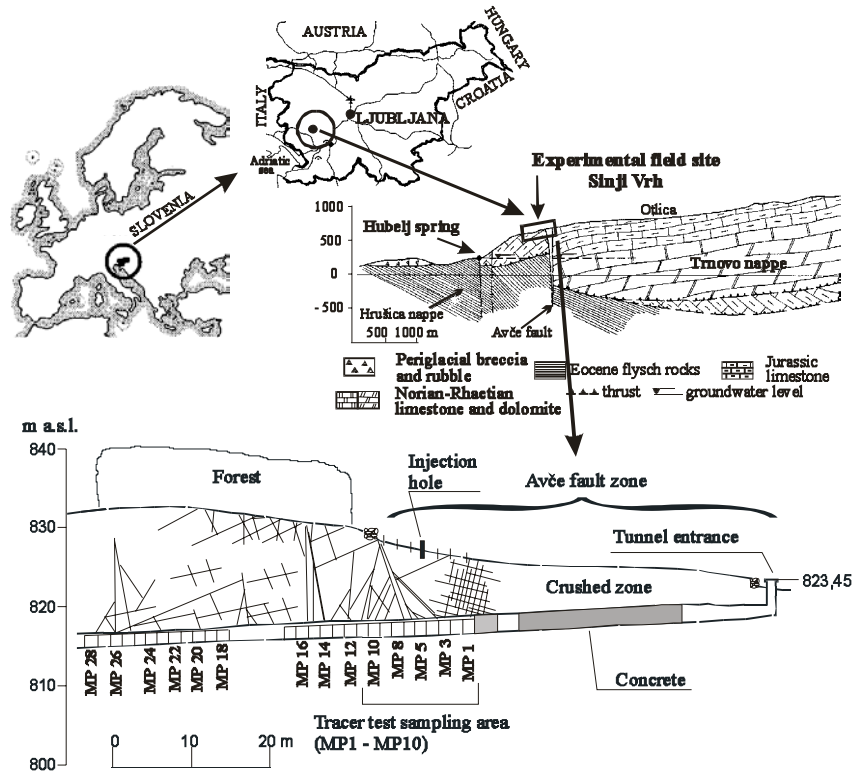


Figure 1. Location of the experimental field site Sinji Vrh (EFS Sinji Vrh) with geological cross section of Trnovo plateau after JANEŽ (1997) and VESELIČ et al. (2001). Below: Longitudinal section of the EFS Sinji Vrh with tracer tests area and tracer test sampling points in the research tunnel: MP1 – MP10 (after VESELIČ & ČENČUR CURK, 2001).

The groundwater horizon lies extremely deep and appears on the surface at the lowest point of the impermeable flysch border (Fig. 1) at the karstic Hubelj spring.

The experimental field site Sinji Vrh presents a 340 m long artificial research tunnel, 5 to 25 m below the surface (Fig. 1). An agrometeorological station has been installed on the surface near the tunnel entrance, where precipitation, evaporation, air temperature, air moisture, solar radiation, wind speed and direction (both at two levels) are continuously measured. A tracer experiment area (Fig. 1) was chosen close to the tunnel entrance on the north-western part. The main dip direction of fractures is NNE-SSW with subvertical dip because of the location of the area within a crushed zone of Avče fault. In the broken zone the tunnel is supported by concrete (Fig. 1). The Jurassic limestone of the tracer experiment area is composed of 99% calcite and has a south-westerly dip direction and a gentle dip (of 5° to 30°). The unsaturated fractured and karstified limestone has a negligible matrix porosity and very high fracture density with some greater conduits (VESELIČ & ČENČUR CURK, 2001).

The injection hole was drilled through the soil cover in order to avoid tracer retardation because of sorption. A special construction for collecting water penetrating through the rock was developed. The water seeping from the ceiling of the research tunnel is gathered in 1.5 m long segments (MP1 - MP10; Fig. 1) with a gathering surface of 2.2 m².

3. Material and methods

Bacteriophage P22H5 was produced on salmonella mouse typhoid fever bacteria NIB22 (LT2 w.t. strain) on the Department of Molecular Biology, Biotechnical Faculty, University of Ljubljana. The propagation of phages to obtain crude bacteriophage lysates was done by the method described in dissertation thesis of BRICELJ (1994). The nutrient media Brain Heart Infusion Broth and Nutrient Agar were from Biolife, Milano. Water samples and phage suspensions were titrated for viable phages (pfu) according to the agar layer method of ADAMS (1966), using host bacteria as the indicator strains.

On 29th September 2003 10.4 liters of tracing solution, composed of 11 tracers was poured into the new drilled borehole to the depth of 0.9 m. There were $1.2 \cdot 10^{15}$ plaque forming units of bacteriophage P22H5 and 10 ml of 1 µm microspheres as a colloidal part of tracer cocktail.

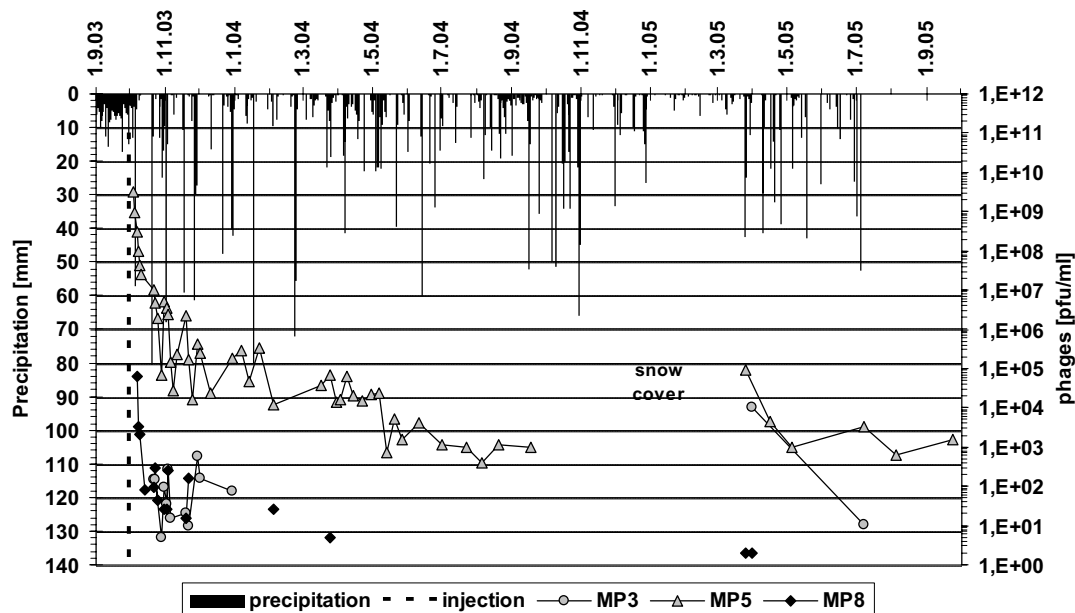


Figure 2. The presence of bacteriophage P22H5 at selected sampling points MP3, MP5 and MP8.

4. Results

The phage tracer appeared first at sampling point five (MP5) after 4.1 days with the peak value of $3.1 \cdot 10^9$ pfu. One day later a positive result was obtained at MP4, the phage appeared with the peak value of $1.1 \cdot 10^8$ pfu (Fig. 2). The tracer appeared at all sampling points within 22 days. Peak values only occurred at the time of appearance at MP2 and MP8. At MP3 the tracer appeared after 8 days, but the peak value was not reached until 50 days.

At the furthest sampling points from the injection hole (MP1, MP2 and MP9, MP10) the peak values were also within the lowest pfu values. The sampling of water was completed at MP4 after 324 days and at MP5 after 347 days. At the time of the last sample, there were still $4.2 \cdot 10^2$ pfu in MP4 and $9.8 \cdot 10^2$ pfu in MP5 in 1ml of water sample. The recovery value was calculated only for MP4 and MP5 and was 0.95% of the injected quantity at both sampling points (0.04% for MP4 and 0.91% for MP5). The results of appearance of the phage tracer on different sampling points are summarized in Table 1.

Table 1. Appearance and presence of the phage tracer at MP1 to MP10. Marked as →: there were still phages in the sample after finishing sampling on the 726th day (after two years).

Sampling point	first appearance of tracer	peak value	appearance of peak value	last positive result
	days after injection	pfu/mL	days after injection	days after injection
MP1	7.0	$2.1 \cdot 10^3$	11.1	64
MP 2	7.0	$2.9 \cdot 10^2$	7.0	648
MP 3	22.1	$5.7 \cdot 10^2$	50.0	648
MP 4	5.0	$1.1 \cdot 10^8$	5.0	726 →
MP 5	4.1	$3.1 \cdot 10^9$	4.1	726 →
MP 6	8.0	$4.3 \cdot 10^4$	11.1	676
MP 7	13.4	$1.4 \cdot 10^4$	30.9	550
MP 8	7.0	$6.5 \cdot 10^4$	7.0	550
MP 9	22.1	$1.7 \cdot 10^3$	24.9	214
MP 10	8.0	$4.9 \cdot 10^3$	40.0	178

The microspheres were observed in all sampling points (MP1-MP10) as well (GÖPPERT *et al.*, 2005). For those sampling points that are well connected to the large fracture below the injection borehole, especially MP5 and MP4, the highest amount of microspheres was observed in the first samples taken after the rainfall. For the sampling points that are discharging the microfracture system, microspheres are not observed until several days later.

5. Discussion

After the injection of bacteriophages and microspheres, they remain in the fractures (channels) and microfracture systems of the unsaturated zone and are rinsed by subsequent larger precipitation events even up to several months after the injection. The results were compared (Fig. 3) with the results from tracer tests in soil and gravel at the Wagna lysimeter station (Austria). The phage tracer was very quickly eliminated from the water trickling through 1 m soil and 0.5 m gravel and positive results were concluded after nine or fourteen days respectively, in the left and right lysimeter (WLL and WRL). The results in the lysimeter are consistent with the findings of VAN ELSAS *et al.* (1991) and POWELSON *et al.* (1991).

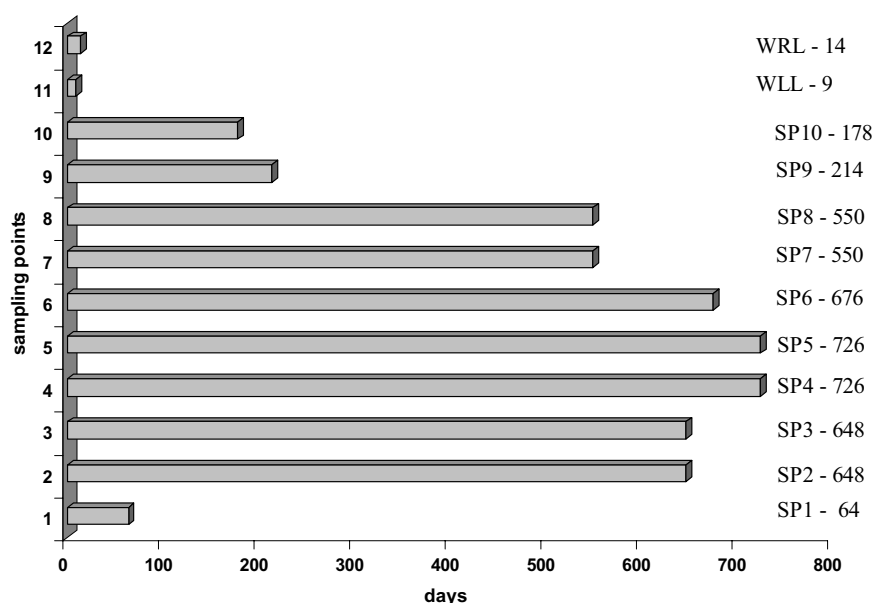


Fig. 3. The presence of bacteriophage P22H5 (duration of the breakthrough) at the sampling points MP1 – MP10 in Sinji Vrh. The sampling points WRL and WLL, represent the percolation of phage tracer through right and left soil lysimeter at Wagna experimental field near Graz.

The results from Sinji Vrh have shown that the unsaturated zone in the fractured and karstified rocks plays an important role in pollution retardation and storage. The rinsing of pollutants to deeper parts of the karst aquifer depends on the saturation rate of the soil and the unsaturated zone (precipitation events). The field experiments have shown different flow patterns depending on the fractured rock structure. In the research area some fast conduits (large fractures or faults) exist where water runs faster than in the total conductive part of the rock, as in the case of MP4 and MP5. Tracer delay in microfracture system areas was also observed, especially at MP3, MP7, MP9 and MP10. At these points the appearance of the peak value was delayed for 50, 31, 25 and 40 days respectively. A very low recovery rate is due to the dispersion of the tracer in directions where it could not be sampled and the decay of tracer, dependent upon removal mechanisms such as filtration, sedimentation and irreversible adsorption (SINTON *et al.*, 1997).

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Stormwater Modeling in Karst Areas

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Abstract:

Caves are natural drainage systems that bear many similarities to engineered stormwater systems. However, some of the differences can cause numerical instabilities in modeling the hydraulics of flows in karst. The Storm Water Management Model (SWMM) was used to model three karst watersheds in Bowling Green, Kentucky, U.S.A., but numerical instabilities gave unreasonable results. The instabilities were caused by rapid filling of small systems fed by larger systems. The smallest time step available in SWMM could not overcome the instabilities. These instabilities were eliminated by using higher order differential equation solutions with smaller time steps. Longer time steps were used to calculate runoff from large watersheds, but finer time steps were used with the smaller hydraulic systems that would fill quickly. Crest gages were used to provide model calibration data. The models, though applied to combined engineered-natural systems, have some application to all natural systems, with ramifications for speleogenesis of high energy caves.

1. Introduction

The U.S. Environmental Protection Agency Storm Water Management Model (SWMM) is software that can be used for calculating the hydrology and hydraulics of watersheds and stormwater management systems. CAMPBELL & SULLIVAN (2002) recognized that caves are natural stormwater management systems and used SWMM to model the hydraulics of flow in a cave system. A newer version SWMM 6.0 was used to model some watersheds of Bowling Green, Kentucky, U.S.A. Bowling Green lies in the Pennyroyal sinkhole plain with very few surface streams.

The discharge point of most Bowling Green stormwater systems is one or more injection wells. City regulations require that these wells be drilled at least 30 cm into fissures in the karstic limestone. These injection wells are often located in the ubiquitous sinkholes of the area. CAMPBELL (2005) studied flooding of these sinkholes and found a very small correlation between volume of water held by the sinkholes and the area of the watershed draining directly to the sinkhole. Consequently, the sinkholes overflow in the largest storms. He developed a procedure to delineate systems of sinkholes that overflowed into each other.

This study focuses primarily on one of the larger sub-basins of Campbell's 2005 study. Fig. 1 shows the watershed that drains to the Glendale Baptist Church culvert. This culvert is a 230 cm by 150 cm elliptical pipe. The figure also shows areas approximate areas of flooding during a 1998 storm. Each of the flooded areas shown is either a sinkhole flood zone or an area

of overflow between sinkholes. The Nahm sinkhole drains an area of more than 90 hectares comprised of mostly residential neighborhoods and commercial areas. Runoff from approximately 65 hectares flows to a large stormwater collection system behind the Glendale Baptist Church. During very heavy rainfall, this system is overwhelmed and flow continues over the surface to the Nahm sinkhole. The discharge captured by the system flows through a 230cm by 150 cm elliptical pipe (DINNING, ET AL. 1980). This pipe discharges into a large box manhole (300 cm by 180 cm by 370 cm). The invert (low point of a cross-section) of the pipe is located approximately 160 cm above the invert of the box. Two other pipes connect to this large manhole. The invert of the larger of the two goes from the box to a sinkhole that has been enlarged into a detention pond with a volume of approximately 55000 m³. Fig. 2 shows a plan view of the system at Nahm sinkhole. The photograph at the upper left shows the sinkhole. The picture at the lower left is the top of the large junction box, and the picture at the lower right shows the author holding the recording stick from the crest gage installed in the injection well manhole.

The injection well consists of a manhole installed over a bedrock cavity (the author is standing on bedrock in the picture). The capacity of the injection well is no more than 0.7 m³/s while the peak flow to the system could be more than 20 m³/s. In heavy flooding, debris clogging of all pipes and the injection well is possible. Very high heads can be expected in both the injection well manhole and in the junction box manhole. This is similar to

situations observed in parts of many cave systems.

The goal of this study was to model the hydrology and hydraulics of a classic karst area for the purpose of understanding the mechanisms of flooding in a hydraulically complicated system. Existing software such as SWMM restricts the modeling time step to a minimum of

The goal of this study was to model the hydrology and hydraulics of the Nahm watershed and hydraulic system which is a combined engineered-natural drainage system. Originally, SWMM was used for modeling, however the 1 second minimum time step of SWMM does not allow sufficient time resolution for this situation. According to DINNING, ET AL. (2005), the peak flow to the system during a 100-yr flood would be 23 m³/s. The total volume of the large

1 second. This is not small enough for this and many other systems. It is particularly not a useful limit for studying the hydraulics of natural cave systems where large conduits feed junctions with very little storage.

2. Materials and Methods

junction box was 20 m³. Consequently, the entire box would be filled in a single time step. This clearly leads to numerical instabilities so that a new model with smaller time steps was needed. None was available so new software was developed. This model described below provided some insights into not just the Nahm watershed, but also into cave hydraulics and speleogenesis for high energy caves.

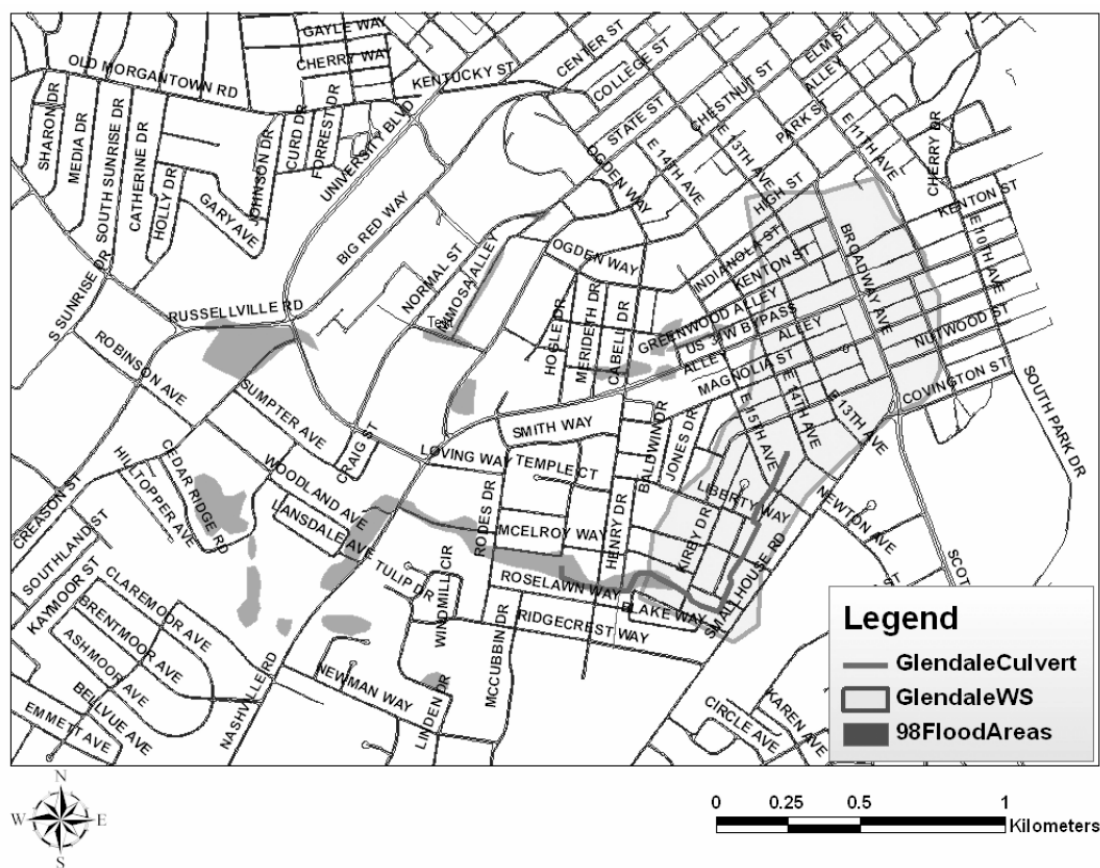


Figure 1: Watershed draining to the Glendale culvert and approximate areas of flooding during the April 1998 flood

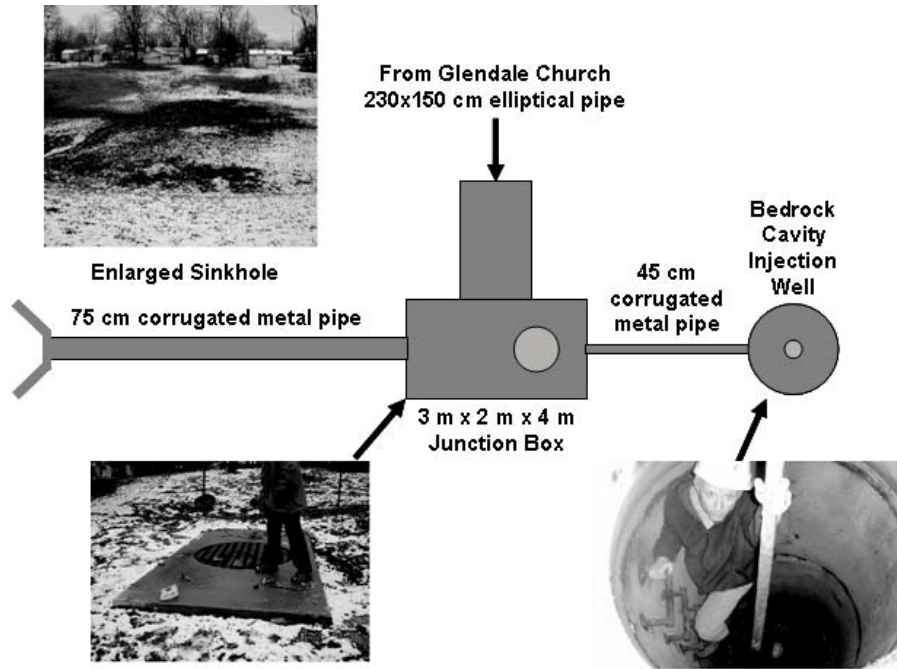


Figure 2: Plan view of the stormwater receiving system at the Nahm sinkhole

A set of differential equations based on the simplified continuity equation was used for modeling. For each node (junction box, manhole/injection well, and sinkhole) the depth of water was modeled with the following equation.

$$\frac{dy}{dt} = \frac{Q_{in} - Q_{out}}{A(y)} \quad (1)$$

where y is the depth of water above datum, t = time, Q_{in} = total inflow into the node, Q_{out} is the total outflow out of the node, and $A(y)$ is the surface area of the sinkhole at a given elevation y . In the equation, the inflow is a function of time only and was calculated as runoff from the watershed using standard hydrologic modeling software (HEC-HMS). The outflows from the nodes were modeled as a combination of orifice and weir flows. The modeling allowed the flows to be either positive or negative.

Inflow and outflow for many devices and cave systems can be modeled as a combination of weir flow and orifice flow. This is the way that flow into inlets and grates are modeled (NEENAH, 2006). CAMPBELL & SULLIVAN (2002) found that stream capture by caves exhibited the characteristics of weir flow rather than the orifice flow they expected. However, the water depths from their calculations were not high enough for the stream capture to function as

orifices. The hydraulic calculator provided by Neenah Foundry Company uses equations similar to the ones below for weir and orifice flow, respectively.

$$Q_{weir} = C_{dw} \sqrt{2g} P H^{1.5} \quad (2)$$

$$Q_{orf} = C_{do} A_{orf} \sqrt{2gH} \quad (3)$$

where Q_{weir} = weir flow rate, C_{dw} = the weir discharge coefficient, g = the acceleration of gravity, P = the open perimeter of the inlet, H = the depth of water above the inlet, Q_{orf} = the orifice discharge, C_{do} = the orifice discharge coefficient, and A_{orf} = the open cross-sectional area of the orifice.

Equations (2) and (3) are modified from the Neenah form in that they are dimensionally homogeneous (not dependent on the system of units used). Notice that weir flow depends on water depth raised to the 1.5 power, while the orifice equation shows a dependence with depth to the 0.5 power. For a general inlet, the perimeter in equation (2) is determined as the length of open area around the outside of the grate. The area is the sum of the areas of the holes in the grate. For flow at low depths, inlet grates behave as weirs and equation (2) is used to calculate the flow. For greater depths, grates behave as orifices. In between, transitional flow

occurs when the flow swirls into the inlet. Fig. 3 illustrates inlet flow in this transitional regime.

The Neenah calculator is founded on laboratory measurements and appears to operate on the following principles. Weir flow is equated to orifice flow and the equation solved for the depth H . Transitional flow is assumed to occur from 0.8 times this depth to 1.2 times the depth. In the transitional regime, the flow is taken as 80 percent of the average of weir flow and orifice flow. The swirling at the inlet during transition reduces the inflow from that expected from either orifice or weir flow.

This method can be applied to many different phenomena. These include grate and curb inlets, stormwater injection wells, flow through cave passages and discharge into a swallet. Injection wells are drilled to bedrock and consist of a riser with holes cut with a welding torch. Water can flow into these wells from the holes in the side of the riser or into the top of the well. According to BROWN, ET AL. (1996) the discharge coefficient in equation (3) is given by 0.6 for smooth holes and 0.4 for rough holes. For cave or swallet flow, smaller values may be used depending on orifice shape and debris blockage.

Outflow in equation (1) was calculated using the orifice equation. This accounted for flow between the nodes shown in Fig. 2. For flow in long, straight conduits (pipes or caves), the head loss is given by the Darcy-Weisbach equation.

$$h_f = \frac{f L}{D} \frac{v^2}{2g} \quad (4)$$

where h_f = friction head loss, f = the Darcy Weisbach friction factor, L = the length of the conduit, D = the conduit diameter, and v = the average velocity in the conduit.



Figure 3. Grate inlet in transition between weir flow and orifice flow

The friction factor f can be determined from the Moody diagram or calculated from an equation given by HAALAND (1983). Equation (5) is the Haaland equation.

$$f = \frac{1}{\left\{ 1.8 \log_{10} \left[\left(\frac{\varepsilon/D}{3.7} \right)^{1.11} + \frac{6.9}{Re} \right] \right\}^2} \quad (5)$$

where ε is the equivalent sand grain roughness of the conduit, Re is Reynolds number = $\rho v D / \mu$, ρ is the fluid density, μ is the fluid viscosity, v is the average fluid velocity in the conduit, and D is the conduit diameter.

This equation has errors less than two percent for turbulent flow. Note that the friction factor is derived for long, straight sections of conduit. Using very large values of f (up to 100) as is commonly done in the karst literature is inconsistent with the definition of friction factor (for example, JEANNIN [2001]). Jeannin recommends abandoning the Darcy-Weisbach form of the loss equation. The very high losses commonly observed in cave flow are caused by minor losses that occur each time a fluid is accelerated such as at a bend, a contraction, or an expansion. While the Haaland equation can be used to calculate friction factor, in many practical karst situations f is small compared to the so-called minor losses. In all but the rare cases when a karst conduit is very long and straight, it would make more sense to replace fL/D in equation (4) with a “minor” loss coefficient K as shown in equation (6).

$$h_f = K \frac{v^2}{2g} \quad (6)$$

where K is the “minor” loss coefficient or the sum of minor loss coefficients, v is average velocity in the conduit, and g is the acceleration of gravity. For expansions and contractions, v is normally taken as the highest velocity.

In cases when a map of water carrying passages is available, it would be possible to calculate f using the Haaland equation and estimate all of the minor losses from handbook values of bend, expansion, contraction, and other loss coefficients. Because handbook values are for engineered systems, the value of K should be developed from measurements whenever possible for a karst system.

The above discussion applies to the hydraulics of cave and karst flow. The inputs to these hydraulics systems are provided by the hydrology of contributing watersheds. The inflow in equation (1) was calculated using SCS (Soil Conservation Service, now known as the Natural Resources Conservation Service) methods and was cut and pasted into Visual Basic programmable spreadsheet software. Solutions for equation (1) were obtained using Runge-Kutta 4th order methods.

Since a very complex watershed was being modeled as a lumped parameter system with several highly uncertain parameters, calibration was critical. A crest gage was installed in the injection well manhole. The gage is a standard type described by the U.S. Geological Survey (USGS). This one consists of 5 cm diameter galvanized pipe with end caps. Several holes were drilled in the bottom end cap with a single vent hole in the side of the top end cap. Inside the pipe is a wooden rod with a small metal cup containing granulated cork. As the water rises the cork floats out of the cup. As the water level falls, the cork adheres to the rod leaving a clear high water mark. This high water mark was used to develop a calibrated model. Unfortunately, the nearest rain gage was about 3 km away from the Nahm sinkhole so that the calibration would not be as accurate as desired. However, the resulting model is much better than an uncalibrated model.

Most of the Nahm sinkhole watershed runoff is provided by the Glendale system. This system transfers flow from a box culvert to the large, elliptical pipe system that terminates at Nahm sinkhole as shown in Fig. 2. Flow entering the grate drops about 1 m into the elliptical culvert system. Consequently, the large grate acts hydraulically like a weir unless water elevations become very high. The length of the weir would be the sum of the lengths of the two sides. The discharge coefficient would account for the presence of the grate, flow turning into the grate, and debris plugging.

3. Results

Several storms occurred during the study, but none nearly as intense as a 100-yr flood. If DINNING, ET AL.. (1980) are correct, 130000 m³ would flow into a sinkhole that holds only 55000 m³. In such a flood, the junction box would fill very quickly and overflow into the sinkhole from both the 75 cm pipe and from the grate on top of the box. This condition would continue for some time until the sinkhole filled

and overflowed. In a very intense storm, it is possible that the heavy grate could be forced off the junction box.

The most recent, intense storm occurred in 1998. The sinkhole overflowed and caused extensive flooding (Fig. 1). One street had knee-deep water with large standing waves. Many houses in the area were flooded. Residents provided many observations regarding this storm. Subsequent to the 1998 flood, the city of Bowling Green enlarged the sinkhole significantly and modified the pipes feeding the injection well. Consequently, the observations by residents would not be relevant to the existing system.

The Dinning study estimated the peak flow to the sinkhole during a 100-yr flood to be 23 m³/s. The large elliptical pipe could carry no more than 14 m³/s assuming that the collector system could deliver this much flow to the pipe. The remainder of the flow would travel above the surface to Nahm sinkhole. The model did not account for losses to the several injection wells that exist in the watershed. Also, some significant infiltration would occur if the cave systems below did not fill during the flood. Bubbles have been observed in the Nahm sinkhole during heavy rain. Presumably, this occurs because of flooding of cave systems below trapping air that is then forced under pressure through the soil above. Generally, flood elevations in Bowling Green are calculated assuming that sinkholes don't drain. This is probably a good assumption for many sinkholes. However, one drainage well drilled recently at the foot of a hill frequently discharges into, and overflows its detention pond.

The Nahm junction box receives water from the large elliptical pipe system. The maximum flow from the box to the sinkhole for a 3.7 m head is 2.3 m³/s. The pipe to the injection well manhole could carry no more than 0.7 m³/s. This leaves an excess of 11 m³/s to build head inside the box. With a total volume of 20 m³, it would take a maximum of 2 seconds to fill. The arrangement of manholes and sinkhole is similar to situation encountered in many cave systems where very high heads can be expected in heavy rain events. QUINLAN & EWERS (1989) observed that water levels in Mammoth Cave rose more than 30 m during a 20-yr event that occurred in 1984. LUNDQUIST & VARNEDOE (1991) reported that Peterson's Cave, a 23 m pit in Newsome Sinks in north Alabama completely filled during rains in December 1990. The entrance room in Saturday Cave near Huntsville,

Alabama regularly fills to a depth of 6 m and overflows to bypass passages. Many other examples could be offered. It is clear that the Nahm sinkhole system and many cave systems develop very high heads and that these heads and debris clogging may have a significant impact on the formation of secondary passages after the formation of the main passage by forcing chemically aggressive stormwater into joints and bedding planes.

4. Conclusions

Commonly available software such as SWMM have severe limits placed on the minimum time step. These limits assure that the code cannot be used to model many engineered and natural systems without significant numerical instabilities. The only recourse available to model cave and stormwater hydraulics is to develop models without these arbitrary limits.

A simplified model of the Nahm watershed and collection system was developed from scratch. Differential equation forms of continuity equations were solved using the Runge-Kutta 4th order method. A reduced time step produced results without numerical instability. The model showed that very high hydraulic heads developed quickly in manholes following the 100-yr storm. This physical phenomenon is very similar to that observed in some high-energy cave systems and probably has ramifications for speleogenesis in these systems. The high heads, chemically aggressive stormwater, and debris clogging of existing passages would provide a driving force for the development of secondary conduits after the formation of the primary passages.

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Ressource karstique côtière en Méditerranée : les sources sous-marines de Port Miou et du Bestouan (Marseille – France)

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Abstract

In the region of Marseilles, the two submarine springs of Cassis (Port Miou and Bestouan) correspond to the outlets of a vast and deep karstic aquifer of more than 500 km². As often noticed around the Mediterranean coast, these littoral karstic springs are brackish, due to a marine contamination of the karst freshwater.

New measurements are in progress in both springs to explain the mechanisms responsible for the seawater intrusion at a distance up to 2 km from the shore. The time series of pressure, discharge, electrical conductivity and temperature of the springs of Port Miou and Bestouan are correlated with the rainfall data and the sea level variations.

Correlation and spectral analysis of the different time series and the time lag between the increase in spring discharge and the related variations in salinity are used to specify the factors influencing the water salinity.

Keywords: submarine spring/coastal karst/seawater intrusion/spectral analysis

Résumé

Dans la région de Marseille, les sources sous-marines de Cassis (Port Miou, Bestouan) constituent le principal exutoire d'un vaste et profond aquifère karstique de plus de 500 km². Comme il est souvent constaté en Méditerranée, ces sources karstiques littorales sont saumâtres, du fait d'une contamination de l'eau du karst par l'eau de mer.

Aujourd'hui de nouvelles mesures sont en cours aux sources pour expliquer les mécanismes responsables de l'intrusion d'eau de mer à grande profondeur dans le système karstique sur plus de 2 km de distance. Les variations de pression, de débit, de conductivité et de température de l'eau suivies en continu aux sources de Port Miou et du Bestouan sont confrontées aux données pluviométriques ainsi qu'aux variations du niveau marin.

L'analyse corrélatrice et spectrale des différentes séries chronologiques et le décalage temporel entre la montée du débit lors des crues et les variations de conductivité associées sont utilisées pour préciser les facteurs influençant la salinité de l'eau.

Mots clés : source sous-marine / karst littoral / intrusion saline / analyse spectrale

1. Introduction

La connaissance des massifs karstiques littoraux est récemment rendue possible par l'étude des résurgences karstiques côtières. Ces sources sont fréquemment polluées par l'eau de mer et le mécanisme de contamination reste difficile à comprendre. C'est le cas des sources sous-marines littorales de Port Miou et du Bestouan localisées dans la région de Marseille, à proximité de Cassis. Ces deux émergences sont l'exutoire d'un vaste et profond aquifère karstique de plus de 500 km². La rivière souterraine de Port Miou a été explorée sur plus de 2 km à l'intérieur des terres par une galerie pouvant atteindre 400 m² de section et celle du Bestouan sur près de 4 km par une galerie de taille plus modeste (Fig. 1). Des travaux récents ont démontrés que l'eau saumâtre des sources résulte essentiellement d'un mélange entre une eau douce karstique et de l'eau de mer (Blavoux et al., 2004). Des prélèvements d'eau effectués dans le puits terminal de Port Miou (-140 m NGF M. Douchet en 1992, -170 m NGF par J. Meynier en 2005) ont confirmé que ce mélange s'effectuait à grande profondeur. Toutefois, le mécanisme à l'origine de la contamination du réseau par de l'eau de mer reste à ce jour inconnu. De nouveaux enregistrements ont été mis en place aux sources pour permettre de mieux comprendre les différents processus à l'origine de la contamination saline des sources. Cet article montre les premiers résultats partiels obtenus sur six mois depuis l'installation du matériel. Les données sont interprétées à l'aide des analyses corrélatrices et spectrales

et de l'observation au cours du temps des décalages de variation de salinité et de débit aux sources.

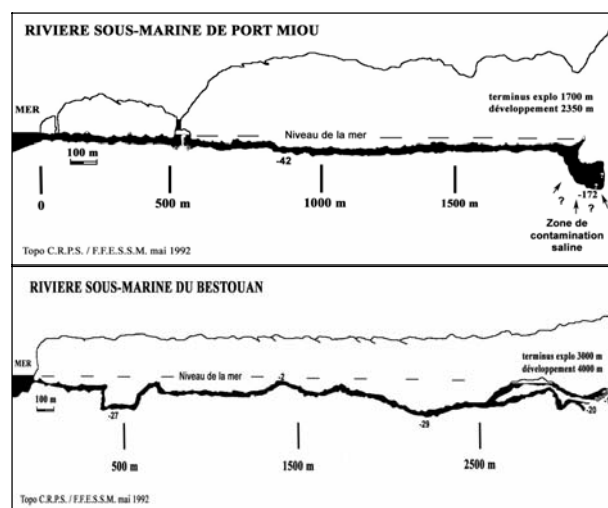


Figure 1 : Coupes développées des rivières souterraines de : a) Port Miou ; b) Bestouan (modifié d'après (Fage, 1993))

Figure 1: Cross-section of underground rivers of: a) Port Miou; b) Bestouan (modified by (Fage, 1993))

2. Matériels et méthodes

L'analyse présentée dans cet article se fonde sur l'enregistrement au cours du temps de la conductivité électrique (rapportée à la salinité), la pression, la température et le débit des sources sous-marines saumâtres. Les capteurs (courantomètres et sonde CTD) ont été placés directement dans les galeries sous-marines. A Port Miou la présence d'un barrage à 530 m du littoral obturant la galerie a facilité l'installation des appareillages. Au Bestouan les appareils ont été placés dans la galerie à une distance de 50 m de l'émergence en mer. Ces données sont complétées avec les chroniques de suivi des niveaux marins enregistrés au marégraphe de Marseille, et de la pluie sur le massif à la station pluviométrique du mont St Cyr située à une dizaine de km au nord-ouest des sources.

L'analyse descriptive de l'évolution des différents paramètres au cours du temps est complétée par l'utilisation de techniques d'analyse corrélatrice et spectrale. L'analyse des corrélogrammes croisés permet de vérifier l'interdépendance entre deux paramètres. L'approche spectrale permet quant à elle de repérer des phénomènes par rapport à leur fréquence (Mangin, 1984). Cet outil d'analyse est courant dans l'étude des systèmes karstiques (Padilla & Pulido-Bosch, 1995 ; Larocque et al., 1998 ; Rahnamaei et al., 2005). Deux fonctions sont utilisées ici d'après les formules de Jenkins et Watts (Jenkins & Watts, 1968) :

- la fonction d'auto-corrélation r_k obtenue à l'aide du corrélogramme C_k :

$$r_k = \frac{C_k}{C_0} \quad (1)$$

$$C_k = n^{-1} \sum_{i=1}^{n-k} (x_i - \bar{x})(x_{i+k} - \bar{x}) \quad (2)$$

où k est le pas de temps ($k = 0, 1, 2, \dots, m$), n la longueur de la chronique et m la troncature.

- le spectre de densité de variance S_f obtenu par la transformée de Fourier de la fonction d'auto-corrélation du corrélogramme :

$$S_f = 2 \left[1 + 2 \sum_{k=1}^{k=m} D_k r_k \cos(2\pi F k) \right] \quad (3)$$

avec $F = j/2m$ la fréquence considérée, j étant le pas choisi ($j = 1 \text{ à } m$), D_k une fonction de pondération.

Les analyses sont réalisées à partir du logiciel STOCHASTOS 3.2.

La période d'étude, de juin à décembre 2005, a été divisée en deux :

- de juin à août 2005 : une période d'étiage marquée par l'absence de précipitation
- de septembre à décembre 2005 : une période de crues de faible intensité avec un retour au débit d'étiage après chaque événement

Au cours des crues, l'augmentation de débit est suivie d'une diminution de la salinité (ou conductivité) enregistrée aux sources. Un décalage temporel existe donc entre le transfert de pression quasi instantané dans le conduit, et l'arrivée de l'eau plus douce à la source. Ce décalage dépend de la vitesse d'écoulement de l'eau dans le conduit et de la distance séparant la source de la zone à l'origine de la baisse

de la salinité. Cette diminution de la salinité est consécutive à la modification des proportions du mélange eau douce – eau salée lors de la crue, par augmentation de la quantité d'eau douce, et/ou diminution de la quantité d'eau de mer. A chaque crue, un certain volume d'eau saumâtre s'écoule à la source avant l'arrivée de l'eau plus douce, indiquant que le conduit se comporte comme une conduite étanche sur ce dernier tronçon. Si on admet une section de galerie constante, ce volume permet d'estimer l'éloignement minimal de la zone de mélange.

3. Résultats - Discussion

Sur la période d'étiage, les pressions enregistrées dans les galeries souterraines des résurgences évoluent de manière identique aux oscillations du niveau marin. Les variations de la pression à Port Miou sont toutefois atténuées par rapport à celles du Bestouan. Les pertes de charges étant très faibles dans ces galeries les capteurs enregistrent donc les variations du niveau de la mer local avec une atténuation selon la distance du capteur à la mer (530 m à Port Miou contre 50 m au Bestouan).

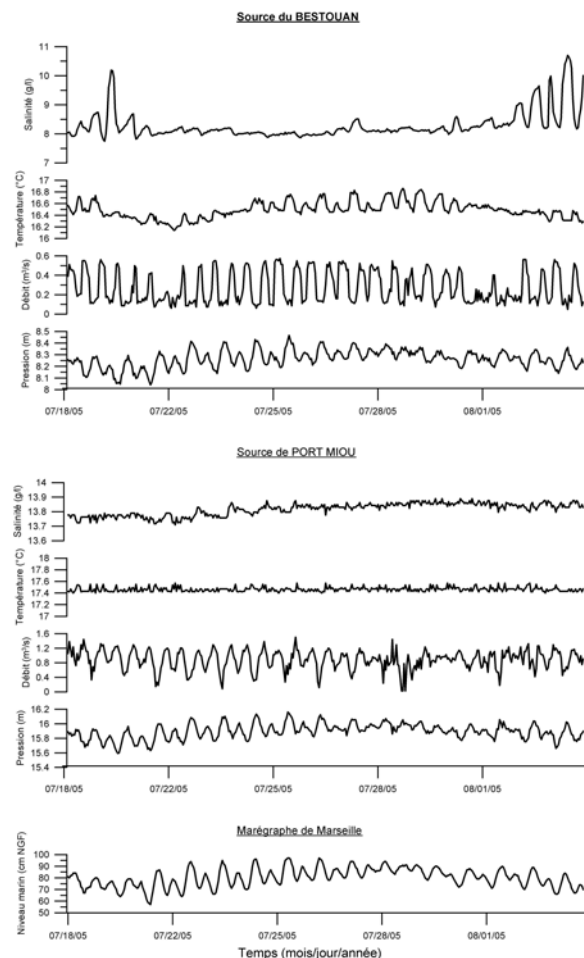


Figure 2 : Evolution, du 18 juillet au 4 août 2005, de la salinité, température, débit et pression des sources sous-marines de Port Miou et du Bestouan et du niveau marin au marégraphe de Marseille

Figure 2: Salinity, temperature, discharge and pressure at Port Miou and Bestouan submarine springs and sea level at Marseilles versus time from July 18th to August 4th 2005

A Port Miou la température et la salinité augmentent au cours de l'été. La température de l'eau varie de 17,3 à 17,6°C pour une salinité de 13,5 à 14 g/L. Au Bestouan la température et la salinité présentent de plus fortes variations. Ainsi, la température varie de 16,1 à 16,9°C pour une salinité de 7,3 à 11,7 g/L (Fig. 2). Ces fortes oscillations ne s'observent pas tout au long de l'étiage, elles s'accroissent à certaines périodes.

Le débit des sources présente de fortes oscillations au cours de la période d'étiage. Les débits maximums atteints sont de 0,6 m³/s au Bestouan et de 1,6 m³/s à Port Miou. Par moment les débits deviennent quasi-nuls mais ils ne s'inversent jamais.

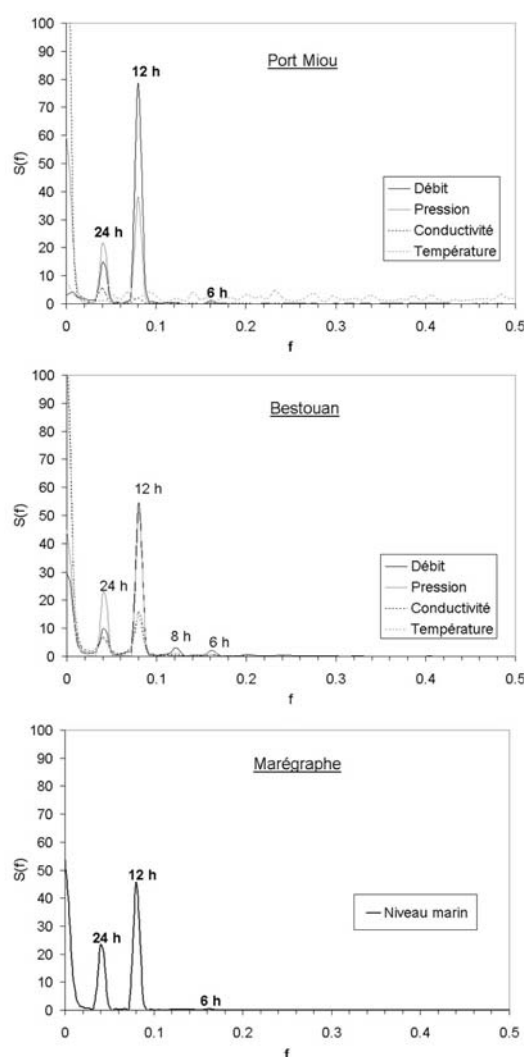


Figure 3 : Spectres de densité des paramètres enregistrés au cours de la période d'étiage 2005 pour : a) la source de Port Miou ; b) la source du Bestouan ; c) le niveau marin

Figure 3: Spectral density functions of recorded parameters during the low water period at: a) Port Miou spring; b) Bestouan spring; c) sea level

L'analyse des spectres de densité des paramètres enregistrés sur les sources lors de la période d'étiage 2005 (juin-août) met en évidence l'influence importante des phénomènes cycliques diurnes (24h) et surtout semi-diurnes (12h) de la marée. Les pics caractéristiques à 12 et 24 h (Fig. 3) sont

très prononcés sur l'enregistrement des pressions et des débits. Par contre pour la conductivité et la température, les pics sont peu marqués sur les enregistrements du Bestouan voire presque inexistant sur ceux de Port Miou.

L'analyse spectrale des paramètres sur la période de crue montre que l'influence de la marée reste présente et atténuée sur les débits et les pressions des sources et qu'elle disparaît complètement à Port Miou pour la conductivité et la température.

L'analyse spectrale révèle donc que les débits des sources sous-marines sont régulés tout au long de l'étude par les cycles des marées. Lorsque la marée est haute, la différence de charge hydraulique entre l'aquifère (conduit karstique) et la mer diminue, entraînant une chute de la vitesse dans le conduit et du débit à l'émergence. Cette régulation bi-journalière s'effectue de manière d'autant plus importante en période d'étiage du fait que la charge hydraulique dans les conduits est faible et proche du niveau marin.

L'analyse croisée apporte une information supplémentaire sur les variations des débits et de la salinité (ou conductivité) des sources. Le corrélogramme croisé entre le débit en entrée et la conductivité en sortie au Bestouan montre un pic de r_k négatif, avec une symétrie par rapport à l'axe des ordonnées, indiquant que les deux paramètres sont corrélés entre eux de manière inversement proportionnelle (Fig. 4). En d'autres termes lorsque le débit de la source du Bestouan augmente la salinité diminue. Cette relation apparaît de manière moins évidente à Port Miou à l'étiage. Ce phénomène peut s'expliquer par l'intrusion d'eau de mer dans la galerie du Bestouan directement par l'émergence pendant les périodes de chute de débit de la source (correspondant à la marée haute), alors que celle-ci est bloquée physiquement à Port Miou par le barrage obturant la galerie.

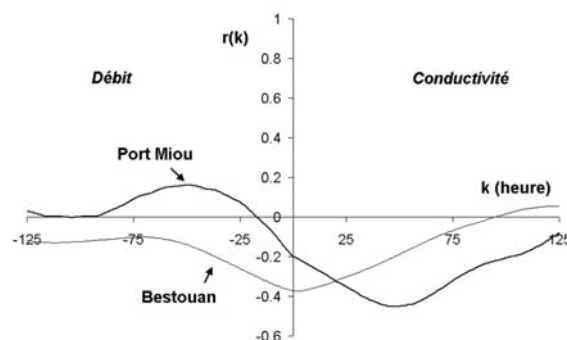


Figure 4 : Corrélogramme croisé entre le débit et la conductivité des sources de Port Miou et du Bestouan durant la période d'étiage 2005

Figure 4: Cross-correlation functions of spring discharge as input and electrical conductivity as output at Port Miou and Bestouan during the low water level (year 2005)

Au cours des mois de septembre à décembre 2005 la région de Marseille et Cassis a enregistré une série d'épisodes pluvieux. La réponse aux pluies a été une augmentation immédiate du débit des sources. Le plus souvent, pour les deux sources, la crue débute dans les premières 24h succédant le début des précipitations. A chaque augmentation de débit est associée une chute plus ou moins franche de la conductivité et de la température. Cette

réaction, inversement proportionnelle au débit, se fait toujours avec un certain décalage dans le temps. Pour chacun le volume moyen d'eau saumâtre expulsé par les galeries souterraines a été calculé. A Port Miou, sept crues ont pu être exploitées, au Bestouan seulement cinq. A chaque volume d'eau saumâtre expulsé est associée une distance théorique de galerie "étanche" dépendante de la section moyenne de cette galerie. Les récents relevés spéléonautiques de ces galeries (Fage, 1993) permettent d'évaluer une section moyenne à Port Miou en amont du barrage de 300 m² et au Bestouan de 15 m².

Les résultats montrent que le volume d'eau saumâtre débité par les sources n'est pas constant au cours du temps. A Port Miou ce volume varie d'environ 80 000 à 300 000 m³ et au Bestouan de 12 000 à 48 000 m³. Ces volumes représentent alors des distances de galeries très variables de 400 à 1400 m à Port Miou et de 800 à 3200 m au Bestouan à partir des stations de mesure.

Pour la source de Port Miou sur laquelle le débit est mesuré avec précision, une analyse complémentaire a été menée portant sur l'évolution de la proportion d'eau de mer participant à l'écoulement au cours du temps. Cette analyse montre que seul le débit d'eau douce augmente lors des crues de septembre à décembre 2005 (Arfib et al., 2006), provoquant ainsi la diminution de la salinité. Le volume calculé indique donc la distance séparant la source sous-marine d'une zone où de l'eau douce supplémentaire arrive en crue. L'eau de mer est quant à elle acquise encore plus en amont, ce que confirment les prélèvements faits par les plongeurs. Ceci appuie l'hypothèse d'un développement profond du système karstique au Messinien à plus de 200 m de profondeur sous le niveau de la mer actuel (Gilli, 2001). Les variations du volume expulsé au début de chaque crue pourraient donc traduire l'existence de conduits karstiques favorisant l'infiltration rapide de l'eau douce de la surface du sol vers la zone saturée en connexion avec les galeries émergeant aux sources. La salinité de l'eau est acquise en profondeur dans l'aquifère et n'a pas ici de relation directe avec ce volume calculé. Ces variations peuvent aussi être liées à une organisation spatiale du système très complexe. Les relations entre le Bestouan et Port Miou restent ainsi énigmatiques et la géométrie profonde des réseaux reste totalement à définir.

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4. Conclusion

L'analyse descriptive de l'évolution de la pression, de la conductivité, de la température et du débit des sources sous-marines de Cassis au cours du temps, complétée par les techniques d'analyse spectrale et corrélatrice, a permis de mettre en évidence :

- un débit des sources régulé par les cycles marins, notamment pendant la période d'étiage
- un phénomène d'intrusion marine directe préférentielle plus marquée sur la source du Bestouan
- une intrusion saline à grande profondeur et grande distance des sources sous-marines

Ces observations soulignent le fait que l'aquifère karstique de Port Miou est fortement influencé quantitativement et qualitativement par les fluctuations du niveau marin au cours de son étiage. Le barrage sous-marin construit à Port Miou est un exemple de protection contre ce type de pollution directe par l'émergence, mais il n'a pas permis de diminuer la salinité acquise en amont à grande profondeur.

L'analyse des déphasages entre la réponse des débits et les chutes de conductivité lors des crues ne convient pas pour déterminer l'éloignement précis de la zone de contamination en eau de mer des sources sous-marines de Port Miou et du Bestouan. En outre, les variations des temps de déphasages au cours d'une période de recharge peuvent constituer une information intéressante sur l'état de karstification d'un aquifère.

Ces premiers résultats, en accord avec la géologie régionale et la géométrie de cet aquifère karstique côtier, nécessitent d'envisager un mécanisme de contamination complexe qui ne pourra être détaillé qu'avec le suivi sur le long terme des sources sous-marines et en particulier l'analyse des crues en période de hautes eaux.

Remerciements

Les données des sources de Port Miou et du Bestouan sont acquises grâce au soutien financier de la Société des Eaux de Marseille et de l'Association Nationale de la Recherche Technique (contrat de Convention CIFRE), et avec le concours du Conservatoire du Littoral.

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Reliability of fecal coliforms and fecal enterococci as indicators of microbial contamination of groundwater in carbonate aquifers

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Abstract

A research is in progress since January 2001 in order to analyze the effect of grazing and manure spreading on microbial pollution of groundwater in carbonate aquifers. The research is carrying out through laboratory and field experiments in different test sites of Southern Italy. The results suggest that fecal enterococci are more reliable than fecal coliforms as indicator organisms of microbial contamination of groundwater. In carbonate Southern Apennines this difference in reliability is strongly conditioned by meteorological and pedological factors. The soil medium, which has a pyroclastic origin, causes a retention of fecal coliforms higher than that of fecal enterococci. This phenomenon is non-uniform at field scale and can reach a difference of an order of magnitude. It often produces the absence of fecal coliforms in spring water samples characterized by low concentrations of fecal enterococci. Both freezing and freeze-thaw intervals will cause a significant decrease in population (3 orders of magnitude, at least) of fecal coliforms while fecal enterococci are temporary inhibited. Each water year this phenomenon determine the absence of fecal coliforms in spring water samples contaminated by fecal enterococci during winter and spring, before the beginning of the new seasonal grazing.

1. Introduction

Historically, fecal coliforms and *E. coli* have been utilized as indicator microorganisms of water quality (CLESCERI *et al.*, 1998). Recent studies have been carried out to evaluate the reliability of fecal enterococci as indicators of surface water in recreational areas (KINZELMAN *et al.*, 2003). As per the groundwater, fecal enterococci did appear a more reliable indicator than fecal coliforms for the detection of microbial pollution in a carbonate aquifer of Southern Italy (CELICO *et al.*, 2004). The main purpose of the present study was to compare fecal coliforms against fecal enterococci as indicators of microbial contamination of groundwater in carbonate aquifers of Central-Southern Italy. The source of pollution is a widespread cattle grazing. The test sites are located within the carbonate Southern Apennines (Italy). They are mainly made of limestones. The rocks have very low primary permeability but are extensively fractured and subordinately karstified. In pasture areas the topsoils can be related to pedogenetic processes acted on volcanic soils coming from eruptive centres of Campanian district. From the hydrogeological point of view, the test sites are similar to other carbonate aquifers of Southern Italy where the groundwater preferentially flows through the fracture network (CELICO, 1986).

2. Materials and methods

Precipitations and air temperature were monitored hourly through a meteorological station.

The discharge of springs was measured weekly or daily by means of a current meter, while the groundwater level was monitored in different piezometers through pressure transducers and a water level meter. The microbiological quality of spring water was monitored weekly or daily. Taxonomic classification of fecal enterococci was performed by use of API 20 Strep fermentation strips (bioMérieux, Marcy l'Etoile, France) and by sequence analysis of one of the 16S rRNA genes amplified with two universal oligonucleotides: P1 (5'-GCGGCGTGCCTAATACATGC) and P2 (5'-CACCTTCCGATACG GCTACC), annealing to nucleotides 40 to 59 and 1532 to 1513, respectively, of *B. subtilis* *rrnE*. Six intact soil blocks of *Epilepti-Vitric Andosols* (*Mollic*) (181.36 cm square by 11 cm deep) were extracted randomly from the study site in the pasture area, in order to carry out column tests, by using the experimental procedures developed by CELICO *et al.* (2004). The interaction between faecal bacteria and soil blocks was analyzed through the utilization of a collection strain of *E. faecalis* (ATCC 29212), nalidixic acid resistant, and a collection strain of *E. coli* (ATCC 10536), resistant to nalidixic acid and ampicillin. No fecal coliforms and fecal enterococci resistant to both antibiotics were observed in the natural background of soil blocks collected in pasture areas. Physical and index properties of six soil samples have been analysed by means of standard laboratory tests: water content, organic matter, grain size and specific gravity of particles.

3. Results

The cattle grazing produced a significant, but discontinuous microbial contamination of spring water in the carbonate aquifers (see example of Fig. 1). The time dependence of pollution was strongly conditioned by the distribution of precipitation versus time. In many cases, no fecal coliforms were detected in contaminated spring water samples (more than 1 CFU/100ml of fecal enterococci). The comparison of field data (hydrogeological and microbiological monitoring) with the results of column tests in intact soil blocks collected in a same test site demonstrated a significant influence of the soil medium on the transport of both fecal coliforms and fecal enterococci towards the groundwater. On the whole, both indicators were highly retained. The simulation of the 50% of the mean annual effective infiltration in the study site did cause the outflow of less than 10% of the inoculated *E. faecalis* cells and less than 0.5% of the inoculated *E. coli* cells (Table 1). Mainly due to their lower retention within the soil, fecal enterococci represent indicators better than fecal coliforms in order to show the existence of microbial contamination of spring water caused by cattle grazing. *E. coli* cells were always more retained than *E. faecalis* cells, even though just in 2 of 6 soil samples this difference can be considered statistically significant. The different retention of *E. faecalis* cells at core scale is not significantly conditioned by the soil texture and the organic matter, which have comparable characteristics in all soil blocks (Table 2). Therefore, there should be significant differences in terms of soil structure and distribution of macropores at block scale. Taking into account the random extraction of soil blocks from the study site, the column tests demonstrated the existence of a uniform retention of *E. coli* cells at field scale. On the contrary, *E. faecalis* cells are characterized by a non-uniform retention at the same scale. Decay should play a secondary role because of the short lengths of transport and the high permeability of carbonate rocks in the study site. Differences may be expected in larger aquifers, where the survival of bacteria into the groundwater could play a primary role, even though the higher decay rate of fecal enterococci should emphasize their reliability as indicator microorganisms.

Soil sample	<i>E. faecalis</i> cells recovered (%)	<i>E. coli</i> cells recovered (%)	Ratio of <i>E. faecalis</i> to <i>E. coli</i>
1	0.46	0.32	1.4
2	0.11	0.08	1.3
3	0.21	0.18	1.1
4	0.45	0.33	1.4
Mean	0.31	0.23	
Stand. Dev.	0.18	0.12	
5	6.04	0.41	14.5
6	8.13	0.19	42.0
Mean	7.09	0.30	
Stand. Dev.	1.48	0.16	
1 to 6			
Mean	2.57	0.25	
Stand. Dev.	3.56	0.12	

Table 1. Percent of *E. faecalis* and *E. coli* cells recovered at the end of each column test.

Soil Samples	Organic Matter (%)	Specific Weight (g/cm ³)	Sediment Type (%)			U (D ₆₀ /D ₁₀)
			Sand	Silt	Clay	
1	27,12	2,06	80,8	15,9	3,3	3,60
2	32,50	2,32	75,0	23,0	2,0	3,21
3	19,78	2,36	83,9	11,1	5,0	9,00
4	22,69	2,23	81,0	17,0	2,0	2,90
5	29,33	2,28	78,9	19,1	2,0	3,33
6	33,80	2,21	73,0	25,0	2,0	4,86

Table 2. Soil blocks properties.

In another test site, the precipitation which produced effective infiltration caused microbial contamination of groundwater, unless the temperature range was characterized by freezing and/or freeze–thaw intervals (Fig. 1). The effect of low temperatures was different for fecal coliforms and fecal enterococci. Concerning fecal coliforms, a few weeks characterized by freezing and/or freeze–thaw intervals were highly lethal.

Taking into account their maximum concentration in contaminated water samples (a few hundreds CFU/100ml), their absence in 1000 ml of spring water showed a decrease of 3 orders of magnitude, at least. This significant decrease in population was confirmed by the absence of this indicator in spring water samples until the beginning of the new seasonal grazing, even if the precipitation did produce effective infiltration. Concerning fecal enterococci, the same temperature range caused their temporary inhibition, and a slighter decrease in population, due to bacterial cell death. The inhibitory effect on fecal enterococci is in agreement with the detection of these microorganisms in spring water samples between the end of freeze-thaw intervals and the beginning of the new seasonal grazing (June 2002 in Fig. 1). Hence, fecal enterococci detected within this lap of time must be absolutely correlated to the fecal matter disposed in the pasture area before the freezing period. This is also confirmed by the observation that there were no significant changes on the set of enterococcal species before, during and after this period.

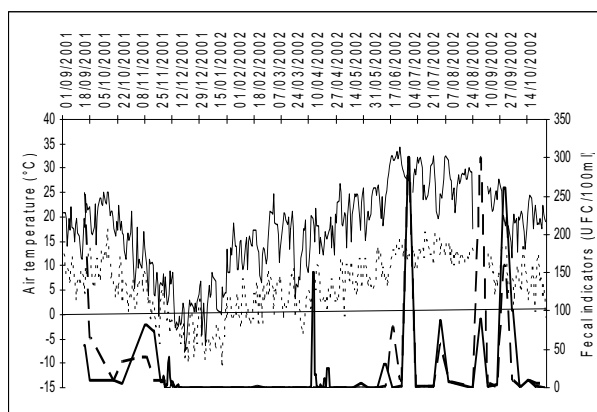


Fig. 1 – Correlation between maximum (thin line), minimum (thin dashed line) daily air temperature and concentration of fecal enterococci (thick line) and fecal coliforms (thick dashed line), at a spring.

Other factors, such as the ratio of fecal coliforms to fecal enterococci in animal feces and the differences in their properties for attachment to solid surface within soil and aquifer, can not play an important role on this phenomenon. In fact, these factors are constant throughout the water year and can not cause the absence of fecal coliforms just between the end of winter and the beginning of the new seasonal grazing.

4. Conclusion

Fecal enterococci seems are an indicator better than fecal coliforms where the limestones underlie topsoils of pyroclastic origin and in case of environmental stresses induced by freezing and freeze-thaw intervals.

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Apport de l'hydrochimie à l'étude des perturbations hydrauliques liées au cycle sismique.

Exemple de deux aquifères karstiques Jurassiens

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Résumé

A la suite du séisme de Besançon (5.1 M_L), une perturbation de la minéralisation de l'eau a été enregistrée à la source de Fourbanne (Basse vallée du Doubs, France) et à la source de la Milandrine (Ajoie, Suisse). Nous attribuons ces variations de minéralisation à différentes perturbations mécaniques liées au cycle sismique, qui sont la variation de contrainte crustale, et la déformation élastique de l'aquifère, liée au passage de l'onde sismique. Les différentes perturbations mécaniques imposées par le cycle sismique perturbent le fonctionnement hydraulique des aquifères étudiés et sont à l'origine du signal hydrochimique mesuré. Cette étude montre la pertinence de l'hydrochimie dans l'étude des phénomènes mécaniques qui affectent les aquifères fracturés discontinus, ainsi que dans l'étude et le suivi du cycle sismique.

1. Introduction

Plusieurs types de réponses hydrogéologiques post-sismiques ont été observées et mesurées depuis des décennies dans différentes régions du globe (Muir-Wood et King, 1993; Toutain et Baubron 1998)

Le lundi 23 février 2004 à 18h31 (heure locale) un séisme de magnitude 5.1 M_L (<http://renass.u-strasbr.fr>) s'est produit dans la vallée du Doubs entre Baume-les-Dames et Besançon (47.3°N, 6.3°E) (Fig. 1). Ce séisme a été ressenti très largement en dehors du Doubs dans tout l'Est de la France, en Suisse et dans le Sud-Ouest de l'Allemagne. Le massif du Jura est une région à faible sismicité notamment sur sa bordure ouest, ce séisme constitue donc un événement géologique rare pour la région.

Entre minuit et 01h00 le 24 février, le dispositif de mesure (pression, température, pH, conductivité enregistrés toutes les 15 minutes) installé à la source karstique de Fourbanne (Fig. 1), situé à 3 Km de l'épicentre, enregistre le début d'une augmentation anormale de la minéralisation de l'eau. Cette augmentation de conductivité électrique de $50\mu S.cm^{-1} \pm 1\mu S.cm^{-1}$ peut être qualifiée d'anormale étant donné l'historique de mesure que nous possédons à la source de Fourbanne et les connaissances acquises au cours de l'étude de ce système karstique.

En effet, le comportement de l'aquifère vis-à-vis des événements hydrométéorologiques est bien contraint et il paraît difficile d'expliquer cette augmentation brutale de minéralisation de l'eau par une interprétation hydrogéologique classique (CHARMOILLE, 2005). Cette augmentation de conductivité n'est accompagnée d'aucune variation synchrone des autres paramètres surveillés (hauteur d'eau, température) permettant d'expliquer habituellement une variation de minéralisation. Au niveau de cet aquifère karstique seul une augmentation du temps de séjour de l'eau lors des périodes d'étiage provoque une augmentation de minéralisation. Ce type d'augmentation est lent. Lors des épisodes pluvieux on observe des dilutions et non des augmentations de minéralisation. Ajouté à ceci, au même moment, des variations de conductivité du même type ont été enregistrées sur le réseau de la Milandrine (Fig. 2) (Données fournies par MFR Géologie-Géotechnique® SA) dans l'Ajoie (Jura suisse). Le signal anormal de conductivité électrique enregistré à Fourbanne n'est donc pas propre à cet aquifère. Ceci exclut toute origine anthropique de ce signal en provenance du bassin versant, ou biais de mesure au niveau de la station d'enregistrement, pouvant être à l'origine de cette augmentation de conductivité électrique de l'eau à Fourbanne.

Après avoir décrit les signaux de conductivité électrique enregistrés dans la vallée du Doubs et en Suisse, nous déterminerons l'origine hydrogéologique du signal enregistré à Fourbanne, puis fournirons une hypothèse quant au mécanisme sismo-tectonique mis en cause. Nous testerons enfin notre interprétation de ce signal chimique particulier en le confrontant à celui enregistré au niveau du réseau karstique de la Milandrine.

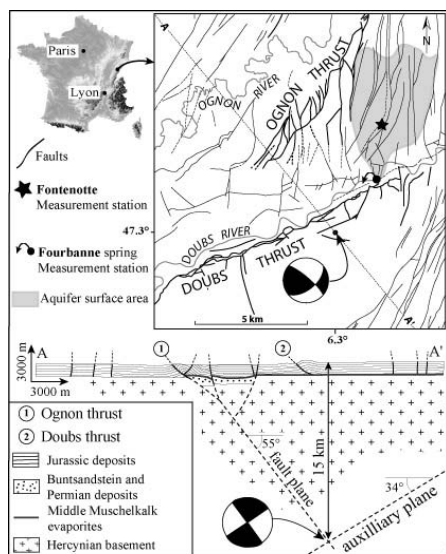


Fig. 1 : Contexte géologique de la source de Fourbanne (CHARMOILLE et al., 2005)

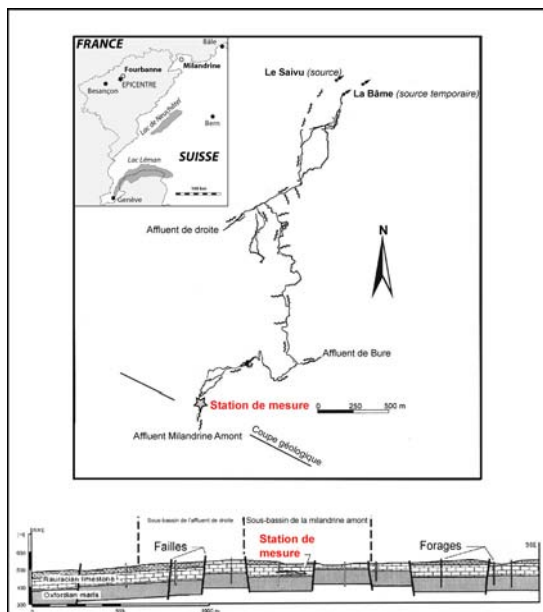


Fig. 2 : Contexte géologique du réseau de la Milandrine. (d'après (JEANNIN, 1996; PERRIN *et al.*, 2003)

2. Résultats

Du 12/02/04 au 06/03/04, une augmentation significative de la minéralisation de l'eau est enregistrée à la station de mesure de Fourbanne (Fig. 3). Cette augmentation atteint une valeur maximale de $640 \mu\text{S}\cdot\text{cm}^{-1}$. Elle intervient 7 heures après le séisme et va s'étendre pendant 12 jours. Le signal enregistré peut être divisé en trois parties. Une première phase d'augmentation de $3 \mu\text{S}\cdot\text{cm}^{-1}$ par heure est enregistrée pendant 12 heures. Cette augmentation de conductivité est ensuite suivie du 26/02/04 au 03/03/04 d'une dilution induite par un épisode pluvieux, puis d'une nouvelle augmentation de conductivité. Ensuite, jusqu'au 06/03/04 la conductivité décroît régulièrement pour retrouver sa valeur initiale à l'occasion d'une crue. Durant toute cette période, le niveau d'eau à la source ne varie que de quelques centimètres. Cette variation correspond au passage de l'onde de crue induit par l'épisode pluvieux.

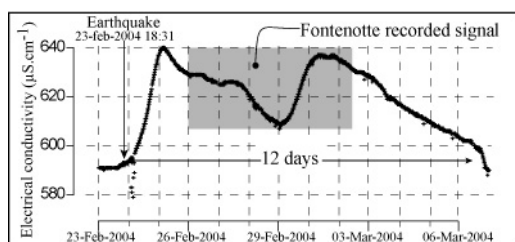


Fig. 3 : Signal post-sismique enregistré à Fourbanne (CHARMOILLE *et al.*, 2005)

Concernant l'aquifère de la Milandrine, trois augmentations de conductivité successives d'une durée d'environ 24 heures chacune sont mesurées après la secousse sismique (Fig. 4). La valeur de conductivité maximale atteinte est de $65 \mu\text{S}\cdot\text{cm}^{-1}$ supérieure à la valeur mesurée avant le séisme. Trois jours après le séisme, la conductivité électrique se stabilise à une valeur de $10 \mu\text{S}\cdot\text{cm}^{-1}$ plus élevée que sa valeur initiale et conserve cette valeur jusqu'à la prochaine crue.

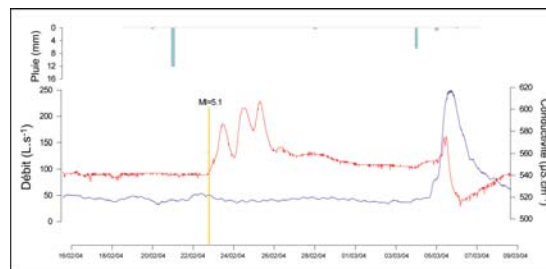


Fig. 4 : Signal post-sismique enregistré dans le réseau de la Milandrine (CHARMOILLE, 2005).

3. Discussion

L'analyse de l'enregistrement de la station de Fourbanne révèle la présence d'une augmentation de conductivité électrique de l'eau sans lien hydrogéologique avec les événements météorologiques et le comportement hydrodynamique naturel de l'aquifère (CHARMOILLE, 2005). Il faut préciser que la température montre, une évolution en accord avec l'évolution des hauteurs d'eau, ce qui ne permet donc pas de corréliser ce paramètre avec l'évolution de la conductivité électrique de l'eau.

Cette augmentation de minéralisation ne peut pas être expliquée par des interprétations hydrogéologiques classiques. En effet, une augmentation de conductivité significative ($50 \mu\text{S}\cdot\text{cm}^{-1}$) et durable (12 jours) comme celle mesurée après le séisme à Fourbanne, ne peut être attribuée à une infiltration d'eau de pluie ou une fonte de neige. Nous ne pouvons donc attribuer l'augmentation de conductivité électrique, enregistrée à Fourbanne qu'à l'événement sismique qui précède.

- Origine du signal hydrochimique et mécanisme sismo-tectonique mis en cause

L'onde sismique produite par le séisme de Besançon a induit une déformation élastique de la croûte terrestre. La déformation co-sismique liée à l'onde sismique peut être une cause de variation de perméabilité des zones peu perméables et de leur participation accrue à l'alimentation du conduit. Cependant, une telle variation de perméabilité induite par la déformation élastique des fractures de l'aquifère de Fourbanne, liée à la propagation de l'onde sismique, est un mécanisme de courte durée. Par conséquent, l'arrivée à la source de cette eau plus minéralisée ne devrait pas s'étendre pendant une durée supérieure à une cinquantaine d'heures. En effet, les différentes expérimentations réalisées au niveau de l'aquifère de Fourbanne nous montrent que le transit dans la zone saturée ne dépasse pas cette durée (CHARMOILLE, 2005). Ceci n'est pas en accord avec le signal enregistré à Fourbanne qui s'étend sur 12 jours. Le signal de conductivité mesuré à Fourbanne nécessite un changement de perméabilité durable de l'aquifère et une alimentation prolongée du conduit par les zones peu perméables.

Le passage de l'onde sismique aurait pu produire un tel signal dans le cas de la création de nouvelles fractures ou micro-fractures qui auraient amélioré la perméabilité des zones peu perméables (ROJSTACZER & WOLF, 1992; ROJSTACZER *et al.*, 1995). Un tel scénario au niveau de l'aquifère de Fourbanne peut être écarté. En effet, le séisme de Besançon n'a causé aucune rupture en surface. De plus, la distance entre l'hypocentre et l'aquifère superficiel de Fourbanne est de 17 Km. A de telles distances, la déformation induite par un séisme d'une magnitude de 5,1

M_L reste dans le domaine élastique pendant le passage de l'onde (KENNETT, 1983). Pour cette magnitude, la création de nouvelles fractures est limitée à la région hypocentrale et concerne un volume d'environ 1 Km³ centré sur l'hypocentre (SCHOLZ, 2002).

MUIR-WOOD & KING (1993) montrent que le modèle de déformation élastique co-sismique peut expliquer une grande variété de réponses hydrogéologiques d'origine sismique. Ils suggèrent que, suivant l'orientation des discontinuités, les fractures préexistantes vont se fermer ou s'ouvrir en réponse à la variation de contrainte crustale, associée au cycle sismique. Ce scénario peut être appliqué à l'aquifère de Fourbanne. Pendant la période inter-sismique, l'eau est stockée dans les fractures peu perméables. Lorsque le séisme intervient, l'eau s'écoule des discontinuités peu perméables, dont la perméabilité a été améliorée par le relâchement des contraintes crustales. Le retour de la conductivité électrique à sa valeur initiale n'implique pas un retour des discontinuités peu perméables à leur ouverture hydraulique initiale. En effet, l'eau à long temps de séjour présente dans ces discontinuités a pu être remplacée par de l'eau moins minéralisée récemment infiltrée dans l'aquifère. La persistance du signal durant 12 jours confirme l'existence d'un changement de perméabilité durable en effet habituellement les modifications hydrochimiques ou hydrauliques intervenant dans l'aquifère de Fourbanne sont évacuées après une vingtaine d'heures.

Le modèle de déformation élastique co-sismique explique bien l'augmentation de conductivité électrique, mesurée dans l'aquifère de Fourbanne, ainsi que la persistance durant 12 jours de ce signal à la source. Dans notre cas, à la différence du modèle proposé par MUIR-WOOD & KING (1993), les volumes d'eau d'origine profonde ne jouent aucun rôle au niveau de l'aquifère de Fourbanne, étant donné la présence des marnes du Lias à la base de l'aquifère. Pour que ces modèles puissent s'appliquer à l'aquifère de Fourbanne, il est nécessaire que le volume de croûte continentale soumis à la contrainte permanente, qui a généré le séisme, affecte un volume de roche que nous estimons à 250 km³. Ce volume est calculé en fonction de la distance hypocentre-aquifère et de la surface de l'aquifère. Cette estimation rentre dans les ordres de grandeurs définis par MUIR-WOOD & KING (1993), qui estiment le volume de croûte affecté de 10² à 10³ km³.

L'arrivée d'eaux profondes en surface peut être écartée. En effet aucune augmentation de température n'a été mesurée à la source. Les études réalisées à l'échelle de la vallée du Doubs (CELLE-JEANTON et al., 2003; CHARMOILLE, 2000; CHARMOILLE, 2005; MUDRY et al., 2002; ROBBE, 2001) montrent que l'eau prélevée dans les forages à 150 m de profondeur sont de 4 à 6°C plus élevées en température au printemps et en été. Une arrivée d'eau d'une profondeur comparable ou supérieure, devrait donc s'accompagner d'une augmentation de température, synchronisée avec l'augmentation de conductivité électrique.

Les études préalablement réalisées au niveau de cet aquifère (CELLE-JEANTON et al., 2003; CHARMOILLE, 2005) nous amènent donc à affirmer que l'augmentation de conductivité mesurée à Fourbanne reflète une participation post-sismique plus importante des zones peu perméables saturées, à l'alimentation du conduit induite par une variation de la contrainte crustale (MUIR-WOOD & KING, 1993). Ces eaux ayant un temps de séjour prolongé avec la roche (perméabilités faibles) leur minéralisation est plus importante.

- Application à l'aquifère de la Milandrine

Nous disposons de peu de données sur ce site. Cependant, les travaux réalisés sur cet aquifère karstique (JEANNIN & GRASSO, 1995; JEANNIN, 1996; PERRIN et al., 2003) nous permettent d'interpréter l'augmentation de conductivité électrique mesurée. Le conduit drainant cet aquifère n'est pas saturé, Jeannin (1996) a identifié trois sous bassins qui alimentent le conduit karstique où la station de mesure est installée (Fig. 2). Ces trois bassins sont le sous bassin de l'affluent de droite, le sous bassin de l'affluent de Bure et le sous bassin de la Milandrine-amont. La station de mesure est installée dans le sous bassin de la Milandrine-amont. Le conduit karstique s'est développé dans un graben (limité par deux horsts). Mis à part en période de crue, le niveau de la rivière souterraine parcourant le conduit correspond à la surface affleurante de la zone saturée. Le signal enregistré à la Milandrine, après le séisme, montre trois augmentations de conductivité successives. La structure de l'aquifère permet d'expliquer la forme particulière de ce signal (Fig. 5).

Si l'on considère comme à Fourbanne que l'augmentation de conductivité électrique de l'eau est le résultat d'une participation plus importante des zones peu perméables à l'alimentation du conduit, nous pouvons fournir l'explication suivante (Fig. 5). L'eau à long temps de séjour est mobilisée dans la zone saturée (t_0). La première augmentation de conductivité correspond au signal chimique en provenance de la zone saturée sous-jacente au conduit (t_1). Les deux autres augmentations correspondent à l'arrivée différée dans le temps du signal hydrochimique produit par les zones saturées des horsts présents au nord et au sud du conduit (t_2). L'augmentation de minéralisation enregistrée à la Milandrine sous forme de trois augmentations successives de minéralisation est donc le résultat du découpage tectonique de l'aquifère.

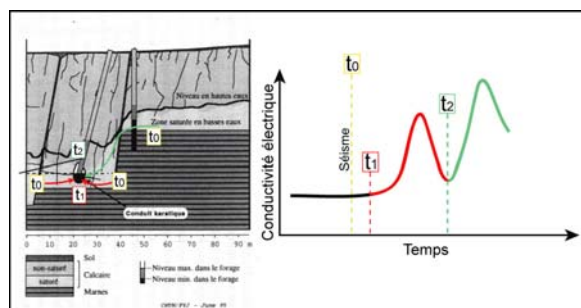


Fig. 5 : Genèse du signal post-sismique dans l'aquifère de la Milandrine (CHARMOILLE, 2005)

La différence avec l'aquifère de Fourbanne réside dans le mécanisme sismo-tectonique mis en cause. En effet, la durée du signal enregistré à la Milandrine n'atteint pas la durée du signal enregistré à Fourbanne. De plus, la majeure partie de la persistance temporelle du signal de la Milandrine s'explique par le temps de transit de l'eau très minéralisée, depuis les différents compartiments de l'aquifère, vers le conduit (Fig. et Fig. 5). En effet, le signal enregistré dans le réseau de la Milandrine correspond à une succession de diarcs, qui sont la manifestation d'une expulsion d'eau d'une durée limitée dans le temps. Nous en déduisons que la déformation élastique des zones peu perméables liée au passage de l'onde sismique explique le signal physicochimique enregistré à la Milandrine.

Ceci est en accord avec les travaux de MUIR-WOOD & KING (1993). En effet, pour que la variation de contrainte

crustale affecte l'aquifère de la Milandrine, la volume de croûte qui doit être considéré ($>10^3 \text{ km}^3$) dépasse largement le volume affecté par la variation de contrainte crustale, dans le cas d'un séisme de magnitude 5.1 Ml.

Les deux aquifères étudiés ont donc été sollicités par deux phénomènes mécaniques différents amenant à la restitution d'eau plus minéralisée à la source. Dans le cas de Fourbanne le changement de perméabilité est durable, tandis qu'à la Milandrine il est temporaire (co-sismique).

4. Conclusion

A la suite du séisme de Besançon (23/02/04), une augmentation de minéralisation post-sismique a été enregistrée à la source de Fourbanne. Cette réponse hydrochimique est interprétée comme le résultat d'une amélioration de perméabilité des discontinuités peu perméables présentes dans la zone saturée. Cette amélioration de perméabilité a induit une alimentation plus importante du conduit (30% en plus) par les zones peu perméables de la zone saturée. Le volume excédentaire en provenance des zones peu perméables a été estimé, par décomposition de l'hydrogramme enregistré à Fourbanne, à $4,3 \cdot 10^3 \text{ m}^3$ (CHARMOILLE et al., 2005).

La mise en évidence de l'effet d'une variation de contrainte crustale sur l'aquifère de Fourbanne, montre que même des événements sismiques de magnitude modérée peuvent avoir des conséquences détectables sur des aquifères superficiels (LEONARDI, 1998). Ceci confirme l'hypothèse de MUIR-WOOD & KING (1993) qui affirment que "même les petits événements sismiques peuvent modifier les conditions hydrogéologiques de surface". Malgré la distance importante entre l'aquifère et l'hypocentre, le modèle de déformation élastique co-sismique permet d'expliquer l'excès de minéralisation mesuré à Fourbanne. Cette interprétation est en accord avec les caractéristiques géologiques, hydrogéologiques et hydrauliques mises en évidence au cours de l'étude de cet aquifère.

Au contraire l'aquifère de la Milandrine ne semble, lui, n'avoir uniquement été affecté que par la déformation élastique liée au passage de l'onde, ce qui est en accord avec les estimations proposée par d'autres auteurs (MUIR-WOOD & KING, 1993). Dans ce cas l'amélioration de conductivité électrique a été temporaire et est intervenue de manière co-sismique.

Il apparaît que l'enregistrement de la conductivité électrique couplé à une bonne connaissance des caractéristiques géologiques et hydrauliques de l'aquifère est un outil très sensible et ce même dans les zones à faible sismicité. En effet, sans le suivi de ce paramètre aucune réponse de l'aquifère à la suite du séisme n'aurait été détectée. En l'absence de libération de volumes d'eau importants en réponse à un séisme, les caractéristiques chimiques de l'eau sont un outil de détection adaptée dans le suivi de la déformation crustale post-sismique (TOUTAIN & BAUBRON, 1999). Cette étude démontre donc que l'hydrochimie est un outil approprié pour l'étude des processus mécaniques se déroulant au sein des aquifères karstiques et plus généralement des aquifères fracturés.

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Apports conjoints de suivis hydroclimatologique et hydrochimique sur le rôle de filtre des aquifères karstiques dans l'étude de la problématique de changement climatique *Application au système de la Fontaine de Vaucluse*

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Abstract

Many studies have shown the complexity of the hydrodynamic behaviour of the Fontaine de Vaucluse karst system (southeastern France). Information on the water-storage zones and their participation on discharge has been obtained from a long term rainfall-runoff database (period 1877-2004) and a hydrochemical survey carried out at the outlet of the system. The rainfall-runoff data have shown the large impact of the climatologic variations on the hydrogeological system. Unexpected high flow rates have been recorded during successive dry years when occurred after a long period of high rainfall amount. This behaviour is assumed to arise from the contribution of a storage zone characterised by a very slow recharge kinetic (20 to 30 years). The magnesium content observed in the water during low flows in summer has proved the role of a reservoir with a yearly-based filling process. The contribution of this reservoir to the flow would depend on rainfall amount over the 2 or 3 previous years. Improvement of our knowledge of such reservoirs is an important point to deal with when studying the consequences of a possible climatic change on the hydrodynamic behaviour of this karst system.

Résumé

De nombreuses études ont démontré la complexité du comportement hydrodynamique du système karstique de Fontaine de Vaucluse (Sud-Est de la France). L'étude, menée conjointement sur la base de données historique de pluie et de débit disponible sur ce site (période 1877-2004) et sur les suivis hydrochimiques effectués à l'exutoire du système, fournit des informations sur les zones de stockage du système et sur leur participation aux écoulements. La base de données pluie-débit montre que le système est soumis à une grande variabilité des conditions hydroclimatologiques. Un soutien des écoulements apparaît lors d'une succession d'années de pluviométrie déficitaire ayant eu lieu après une longue période d'excédent pluviométrique, soutien s'expliquant sans doute par la participation d'une zone de réserve à cinétique de recharge très lente (20 à 30 ans). L'étude de la teneur en magnésium dans les eaux de l'exutoire au cours d'étiages estivaux démontre le rôle d'une réserve se rechargeant annuellement, dont la participation aux écoulements dépend de l'historique pluviométrique des 2 à 3 années précédentes. L'existence de réserves dont les cinétiques de recharge se comptent en années ou même en décennies sera d'une importance notable dans le cadre de l'étude des conséquences d'un éventuel changement climatique sur le fonctionnement de ce système karstique.

1. Introduction

Les études réalisées par MANGIN (1975) sur les réponses du système karstique de Fontaine de Vaucluse aux sollicitations pluvieuses ont montré la complexité du fonctionnement hydrodynamique de ce système. En regard de sa large zone d'alimentation (1130 km²), ce système répond relativement rapidement aux événements pluvieux (réponse 24 à 72 h après la pluie), mais les hydrogrammes de crue sont aussi caractérisés par des décrues lentes prouvant la participation aux écoulements d'une importante zone de stockage.

Suite aux travaux réalisés dans les années soixante (prospections géologiques et traçage artificiel, PALOCH (1969)), PUIG (1987) a proposé, à partir de l'étude du bilan hydrologique et de l'utilisation du traçage naturel des eaux, une délimitation de la zone d'alimentation. L'auteur a aussi mis en évidence le rôle de la zone non saturée sur les écoulements à l'exutoire du système. Depuis, de nombreuses études, fondées sur l'utilisation du traçage naturel des eaux, ont fourni des informations complémentaires sur les zones de stockage d'eau du système et sur leur participation aux écoulements (MUDRY, 1987, BLAVOUX *et al.*, 1992, LASTENNET *et al.*, 1995, EMBLANCH *et al.*, 2003, BATIOT *et al.*, 2003). Si toutes ces études ont confirmé le rôle primordial de la zone non saturée, des questions subsistent encore sur l'identification

de l'ensemble des zones de stockage et sur leur dynamique de recharge.

Nous proposons ici de fournir des informations complémentaires sur ce sujet en exploitant d'une part la base de données historique de pluie et de débit (127 années de suivi) disponible sur ce site, et d'autre part le résultat des suivis hydrochimiques effectués sur les eaux de la Fontaine de Vaucluse au cours de cinq étiages estivaux.

2. Matériels et méthodes

La Fontaine de Vaucluse, située dans le sud-est de la France à 30 km à l'est d'Avignon, est l'unique exutoire d'un système karstique dont la zone d'alimentation du système, d'une superficie de 1130 km², s'étage entre 84 et 1909 m d'altitude (Fig. 1). Avec un débit moyen d'environ 20 m³/s, la Fontaine de Vaucluse est la première émergence karstique de France. Les débits mesurés les plus faibles sont de l'ordre de 4 m³/s. Les débits maximaux sont évalués à 100-120 m³/s. Ce système karstique s'est développé dans une puissante série carbonatée du crétacé inférieur, d'une épaisseur de 1500 m (MASSE (1968)). Ce système est aussi caractérisé par une importante zone non saturée, dont l'épaisseur varie entre 600 et 800 m (PUIG (1987)).

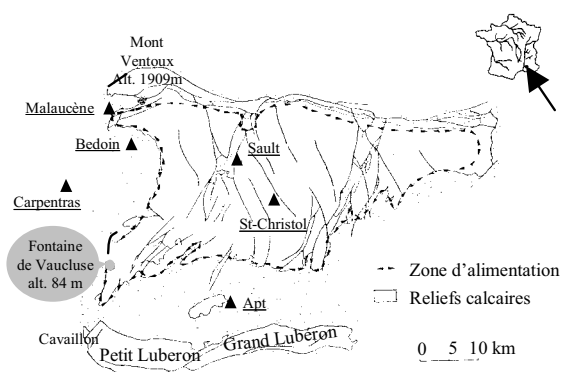


Figure 1 : localisation de la zone d'étude, d'après PUIG (1987)
▲/▲/▲ poste pluviométrique

Le régime pluviométrique, de type méditerranéen, montre une distribution irrégulière des précipitations dans l'année, avec deux saisons humides (printemps et automne) et deux saisons sèches (hiver et été). La répartition des précipitations est caractérisée par une forte variabilité interannuelle. La variabilité spatiale de la pluviométrie est largement pilotée par le relief.

Base de données hydrologiques

• Débits

Grâce à un suivi régulier des débits de la Fontaine de Vaucluse réalisé depuis 1877, nous avons pu travailler sur une base de données continue de débits mensuels sur 127 ans (période 1877-2004). Entre 1877 et 1964, les débits étaient estimés à partir des hauteurs d'eau de la rivière Sorgues alimentée par la Fontaine, hauteurs mesurées 5 km en aval de l'exutoire. Depuis 1965, les débits sont déduits des hauteurs d'eau dans la vasque. L'étude de la base de données montre que les débits estimés à l'aide de la première station de jaugeage (période 1877-1964) sont largement surestimés en comparaison de ceux mesurés depuis 1965. Cette surestimation des débits de la période 1877-1964 a été largement corrigée (COGNARD-PLANCQ *et al.* (2006)), mais il subsiste toujours un biais dans la base de données, ce qui oblige le plus souvent à traiter séparément les deux parties de la base de données.

• Précipitations

Une base de données de précipitations mensuelles, synchrone avec celle des débits, a été construite à partir des données de six postes pluviométriques localisés sur la figure 1. L'effet orographique a été pris en compte grâce au gradient altitudinal des précipitations mis en évidence sur la zone par PUIG (1987).

Suivi hydrochimique

Quatre campagnes de suivi hydrochimique ont été réalisées entre 1981 et 2004 sur le système de Fontaine de Vaucluse, au cours desquelles ont été réalisés régulièrement des prélèvements d'eau à l'exutoire (MUDRY (1987), EMBLANCH *et al.* (1998), BATIOU *et al.* (2003)). Nous disposons donc sur ces périodes d'un suivi régulier des caractéristiques chimiques des eaux. Pour l'étude du fonctionnement du système, et plus particulièrement pour identifier le rôle de la zone saturée, l'utilisation des suivis

des concentrations en magnésium est particulièrement intéressante. BLAVOUX & MUDRY (1983) ont en effet montré, à partir des suivis hydrochimiques réalisés entre 1981 et 1982, que des concentrations élevées en magnésium pendant les étiages marquent les eaux de réserve du système (zone noyée). Pour tenter de caractériser le rôle des zones de réserve sur les écoulements du système, nous nous intéresserons donc aux concentrations en magnésium enregistrées au cours des étiages estivaux de 1982, 1996, 2000, 2001 et 2004.

3. Résultats

Les chroniques historiques de données annuelles de précipitations et de débits sont présentées en figure 2. La chronique des précipitations illustre la variabilité du climat méditerranéen. La pluviométrie moyenne annuelle est d'environ 1100 mm, mais varie entre 641 mm pour l'année la plus sèche (1953) et 1740 mm pour l'année la plus humide (1977). La série est quasi-stationnaire, à l'exception de la période 1934-1936 particulièrement arrosée. La chronique des débits est elle aussi marquée par une forte variabilité interannuelle. Pour la période 1877-1964, le débit annuel est en moyenne égal à 25 m³/s, mais varie entre 10 et 53 m³/s, respectivement en 1905 et 1915. Pour la période 1965-2004, le débit moyen annuel est d'environ 19 m³/s, mais varie entre 7,6 et 35,5 m³/s, respectivement en 1990 et 1978. La chronique des débits présente cinq périodes stationnaires (Tabl. 1). La cinquième période correspond à la seconde partie de la base de données (suivi des débits à partir des hauteurs d'eau dans la vasque de la fontaine), caractérisée comme nous l'avons déjà indiqué, par des débits nettement plus faibles que ceux estimés sur la première partie de la base de données.

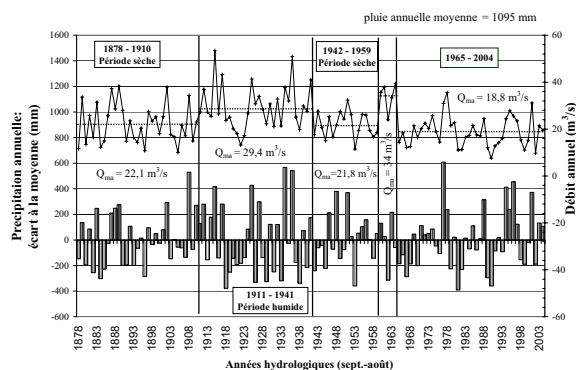


Figure 2 : Chroniques des débits et précipitations annuelles. Les précipitations sont présentées sous forme d'écart à la valeur moyenne de 1095 mm. Les lignes en pointillé correspondent aux débits moyens sur chaque période de stationnarité (Q_m).

Périodes de stationnarité (années hydrologiques)	Débit annuel moyen (m ³ /s)	Pluie annuelle moyenne (mm)
1878-1910	22,1	1030
1911-1941	29,4	1146
1942-1959	21,8	1000
1960-1964	34,0	1269
1965-2004	18,8	1132

Tableau 1: périodes de stationnarité de la chronique des débits annuels.

L'analyse conjointe des chroniques de pluie et de débit montre dans un premier temps un lien marqué entre les

débits annuels et la précipitation annuelle de la même année (coefficient de corrélation entre les 2 variables égal respectivement à 0,80 et 0,85 pour les période 1877-1964 et 1965-2004). Le débit annuel est aussi en partie expliqué par la pluie de l'année précédente. Toutefois, si nous étudions l'évolution de la réponse du système lors de 4 périodes caractérisées par une succession de plusieurs années de déficit pluviométrique, caractérisées en gris sur les figures 3 et 4, nous constatons pour les 3 premières périodes que le déficit pluviométrique entraîne un déficit marqué sur les écoulements. Au cours de la quatrième période de déficit pluviométrique, le déficit d'écoulement est beaucoup moins marqué (Fig. 3 ; Tabl. 2). Au cours de cette période, les écoulements ont donc bénéficié d'un soutien particulier, expliqué par un contexte pluviométrique plutôt excédentaire au cours des 17 années précédentes. Nous pensons mettre en évidence ici le rôle d'une zone de stockage très inertielle. Cette réserve ne participe pas, ou peu, au cours des trois premières périodes de déficit, car les réserves de cette zone ne sont sans doute pas assez reconstituées. En effet, ces trois premières périodes de déficit accru de pluviométrie prennent place dans un contexte climatique globalement sec.

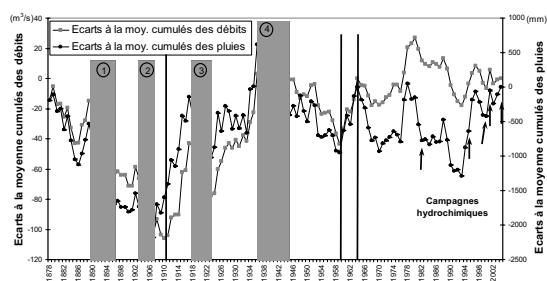


Figure 3 : Chroniques des écarts cumulés à la moyenne des débits et des précipitations annuelles ; période 1877-1964 : écarts par rapport aux moyennes de pluie et débit sur cette période (respectivement 1079 mm et 25,3 m³/s) ; période 1965-2004 : écarts par rapport aux moyennes de pluie et débit sur cette période (respectivement 1132 mm et 18,8 m³/s)

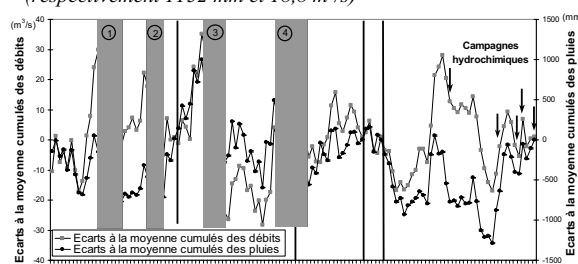


Figure 4 : Chroniques des écarts cumulés à la moyenne des débits et précipitations annuelles ; écarts calculés et cumulés sur chaque période de stationnarité des débits.

Période	D _c P (mm)	D _c Q (m ³ /s)	D _c Q (mm)	D _c Q / D _c P (mm/mm)
① (1890-1896)	- 1162	- 49,2	- 1373	1,2
② (1903-1907)	- 643	- 43,5	- 1214	1,9
③ (1918-1923)	- 880	- 34,6	- 966	1,1
④ (1937-1945)	- 1028	- 3,6	- 100	0,1

Tableau 2 : Comparaison des déficits cumulés de pluie et d'écoulement lors de succession d'années à pluviométrie déficitaire ; D_cP : déficit cumulé de pluie ; D_cQ : déficit cumulé de débit ; utilisation des déficits présentés en fig. 3.

L'analyse des concentrations en magnésium mesurées dans les eaux des exutoires karstiques pendant les périodes d'étiage estival fournit des informations sur les caractéristiques des réserves des systèmes et sur leur participation aux écoulements. L'évolution classique des concentrations est illustrée par la figure 5a. La concentration en magnésium augmente à mesure que diminue le débit, pour finalement se stabiliser en étiage à une valeur plafond. Cette stabilisation sur des valeurs élevées de magnésium démontre la participation aux écoulements d'étiage d'une zone de réserve constituée d'eaux à temps de séjour long, temps de séjour permettant des échanges chimiques avec la matrice rocheuse. A Fontaine de Vacluse, en période de pluviométrie plutôt bénéficiaire (année 1995 et 2000), les étiages estivaux sont caractérisés par des débits de l'ordre de 9 à 10 m³/s (Tabl. 3). Au cours de ces étiages, l'évolution des concentrations en magnésium dans les eaux de la fontaine de Vacluse suit l'allure décrite en figure 5a, les concentrations atteignant une valeur maximale voisine de 9 mg/l. A contrario, au cours des étiages estivaux plus sévères (1982, 2000, 2001), la concentration en magnésium ne se stabilise pas, les valeurs restant sur la partie ascendante de la courbe, selon des évolutions illustrées par la figure 5b. Cela se traduit dans le tableau 3 et sur la figure 6 par des intervalles de variation importants de la concentration en magnésium.

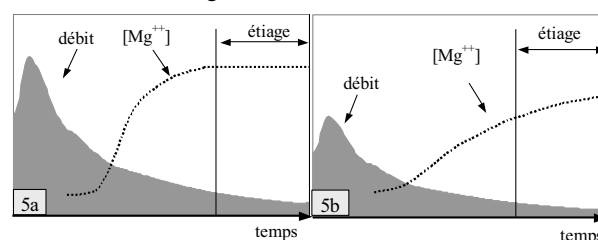


Figure 5 : évolution de la concentration des eaux en période d'étiage estival.

année	débit moyen d'étiage (m ³ /s)	concentration en Mg ⁺⁺ (mg/l)		
		concentration moyenne	concentration minimale	concentration maximale
1982	5,6	7,1	5,9	8,3
1995	9,7	9,2	9,1	9,2
2000	6,9	7,6	7,0	8,1
2001	9,2	8,8	8,7	8,9
2004	5,9	8,3	8,0	8,4

Tableau 3 : débit moyen et concentration en magnésium en étiage estival (périodes concernées : 24/06-15/10/82 ; 04/09-05/10/1995 ; 20/07-19/09/00 ; 29/08-27/09/01 ; 01/09-27/10/04)

Le tableau 3 indique clairement que la prise en compte du débit d'étiage n'est pas un paramètre suffisant ni même pertinent dans l'analyse de l'état hydrique du système, en effet l'étiage le plus sévère (1982 avec un débit moyen de 5,6 m³/s) ne correspond pas à l'écoulement des eaux au plus long temps de séjour moyen (moyenne à 9,2 mg/l en 1995) ni même au plus long temps de séjour mesuré (concentration maximale de 8,3 mg/l en magnésium en 1982, alors qu'en 1995, la valeur maximale est de 9,2 mg/l). Ici aussi la prise en compte d'un ensemble de paramètres hydroclimatiques semble nécessaire (pluviométrie des années précédentes et probablement le type de précipitations).

Il apparaît que, plus le déficit pluviométrique est marqué sur les 3 années précédant l'étiage, plus le signal

magnésien reste éloigné de la stabilisation (Tabl. 4). Nous voyons donc ici s'illustrer le rôle d'une réserve sensible au contexte hydroclimatologique à plus court pas de temps que la zone de réserve précédemment évoquée.

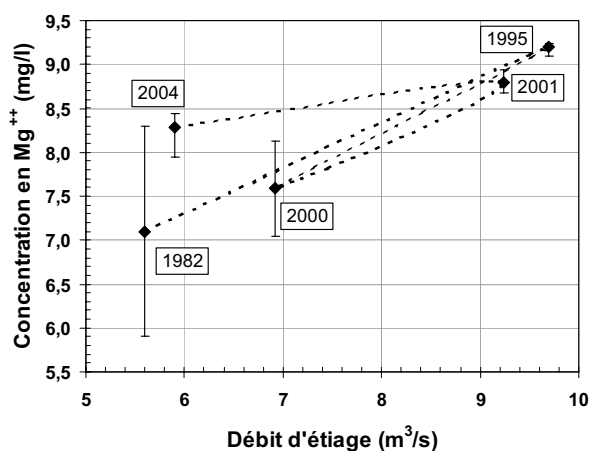


Figure 6 : relation entre débit d'étiage et concentration en magnésium ; I : intervalle de variation

	EcP* (mm)		EcP (mm)		EcP (mm)
1980	+22	1998	-154	2002	-188
1981	-390	1999	-189	2003	+132
1982	-230	2000	-16	2004	+106
cumul	-598		-359		+50

* EcP : écart pluviométrique à la moyenne ; valeur négative en cas de déficit

Tableau 4 : contexte pluviométrique sur les trois années précédant les étiages avec suivi hydrochimique

4. Conclusion

L'étude conjointe de longues chroniques de pluies et de débits nous a permis d'identifier l'existence d'une probable zone de réserve très inertielle. La participation de cette zone aux écoulements est contrôlée par les conditions pluviométriques ayant régné sur les 20 à 30 années précédentes et n'apparaît de manière notable qu'après de longues périodes de déficit pluviométrique. L'analyse des données hydrochimiques a montré que la « signature » magnésienne des eaux d'écoulement de la Fontaine de Vaucluse en étiage estival est très différente suivant le contexte pluviométrique des 2 à 3 années précédentes. Ce dernier point met en évidence le rôle d'une seconde zone de réserve, dont le cycle de recharge serait sensible à la variabilité interannuelle du climat.

Il apparaît donc, à l'issue de ce travail, que l'utilisation conjointe de données hydroclimatologiques et hydrochimiques est d'une grande utilité pour l'étude des transferts et des stockages dans un système karstique. Les résultats de notre étude, avec en particulier la mise en évidence de zones de réserve dont la constitution et la participation aux écoulements du système dépendent des caractéristiques pluviométriques des décennies précédentes pose la question de l'impact d'un possible changement global du climat sur les écoulements du système karstique de Fontaine de Vaucluse. Notons qu'au regard de la chroniques de pluies, aucune baisse significative des précipitations n'est pour l'instant détectée sur notre zone d'étude. Nous prévoyons toutefois d'enrichir la base de données hydroclimatologiques d'information sur le suivi de

la température, autre paramètre « clé » pour juger d'un éventuel changement climatique.

Remerciements :

Tous nos remerciements à toutes les personnes ayant travaillé à l'acquisition des précieuses données utilisées dans le cadre de cette étude ; les observateurs bénévoles qui ont régulièrement relevé les hauteurs d'eau de la Fontaine de Vaucluse et participé aux mesures de précipitations ; les chercheurs qui se sont attachés à l'étude de ce fascinant karst de Fontaine de Vaucluse.

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Mapping a shallow large cave using a high-resolution Very Low Frequency Electromagnetic method

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Abstract

It is always challenging to map and circumscribe shallow cave morphology for vulnerability and risk assessments in karstic aquifers (speleology prospecting). Thus, a high resolution VLF-EM (Very Low Frequency Electro Magnetic) survey was carried out on a small portion of the karst aquifer of Verne-Fourbanne (En Versenne karst system), Fontenotte, Doubs (25), France. We selected a special area above a 50 m-deep, 60 x 25 x 25 m buried cave. This cave has been extensively investigated by the Speleological Group of Clerval – Baume-les-Dames.

VLF profiles have been measured over 4 hectares at 4 frequencies at 10 m intervals, in order to highlight specific structural features related to the location of the cave and consequences of such an extremely resistive air volume on electric currents. The imaginary part (outphase) of the tipper (Hz/Hx) is interpolated over this dense network. It shows features that are consistent with the previously known geological information.

The results show two regional superimposed geological structures which are in good agreement with the main local faults. These structures run parallel to the cave walls. Room delineation seems possible based on the 77.5 kHz outphase map, but not on the 23.4 kHz map. Two-D forward modelling of this structure reveals that the room, despite its great volume and resistivity, does not affect the model response to a noticeable extent. Results of a 2-D inversion approach quite well with the field data.

Résumé

La délimitation précise en surface de conduits karstiques peu profonds constitue un des enjeux majeurs en matière de vulnérabilité du milieu calcaire. Une campagne de géophysique VLF-EM (Very Low Frequency Electromagnetic) haute résolution a été menée à l'aplomb d'une salle souterraine (60 x 25 x 25 m, localisée à 50 m de profondeur) de la partie amont (système karstique d'En Versenne) de l'aquifère karstique de Verne-Fourbanne, Fontenotte, Doubs (25), France. Cette structure a été abondamment documentée par le Groupe Spéléologique de Clerval – Baume-les-Dames.

Espacés de 10 m, sur environ 4 hectares, des profils VLF-EM ont été enregistrés à 4 fréquences différentes afin de mettre en évidence les caractéristiques géologiques liées à l'emplacement de la grotte et d'estimer l'impact d'un tel volume d'air (infiniment résistant) sur les champs électromagnétiques. L'outphase (partie imaginaire du tipper Hz/Hy) a été interpolé sur cette surface et montre une parfaite cohérence par rapport aux données connues.

Les résultats révèlent 2 directions structurales, dont une (NNE-SSW) est parallèle au réseau et aux parois de la salle du Grand Vatel. La délimitation de la salle semble éventuellement possible sur la carte de l'outphase à 77.5 kHz, mais elle n'apparaît pas sur la carte à 23.4 kHz. Les modélisations 2-D effectuées montrent que, malgré son volume conséquent, le vide recherché, résistif, n'affecte que très légèrement les résultats du modèle. Les résultats d'une inversion 2-D approchent les données brutes.

1. Introduction

Extreme vulnerability of karstic aquifers to contamination (ZWAHLN 2004) has conducted to a diversity of investigation approaches in calcareous environments.

Groundwater pathways in limestone media are intimately linked to fractures and fissures which are preferentially enlarged by dissolution. Such features are often sealed by clays in variable quantity, considerably increasing their electrical conductivity and making them easily detectable by EM techniques such as the VLF-EM method (TURBERG & MÜLLER 1992, BOSCH & MÜLLER 2001, BOSCH & MÜLLER 2005).

This study aims at investigating a subsurface void embedded in a resistive medium and structural conductive features related to this karstic network using an electromagnetic method. The strong resistivity contrast between air (isolator), surrounding media (resistive) and

fractures (conductive) is being considered. Previous studies used VLF-R (resistivity) (GUERIN & BENDERITTER 1995, OGILVY *et al.* 1991) for detecting air-filled galleries. Their success was limited by the size and depth of the embedded target. It is to be expected that sufficiently large and not too deeply buried caves would be detectable due to electric field anomalies associated to the cave and the surrounding rocks. The forced deviation of the primary current flow lines around a 3D void should induce a secondary electric field and increase the apparent resistivity (OGILVY *et al.* 1991).

2. Framework

The study area is situated in the tabular Jura close to Baume-les-Dames (Doubs (25), France) in the town of Fontenotte (Fig. 1). The Verne-Fourbanne karstic network (V.F.) expands in the northwestern part of Baume-les-Dames over about 9 km. This network has been investigated

by the Speleological Group of Clerval – Baume-les-Dames and is divided into an upstream part (known as En Versenne cave) and a downstream one only known by tracer tests discharging at the Fourbanne spring (Fs), at the edge of the Doubs river. Its watershed (recharge area) of 37 km² has been precisely delimited in a recent hydrogeological study (GSCB 2001, CHARMOILLE 2005). Groundwater flow rates range from 28 l/s to 15 m³/s in exceptional spates.

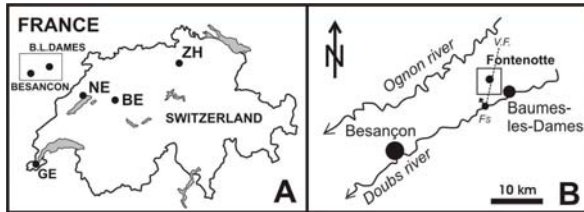


Fig. 1: General overview (A) and close-up (B) of the the study area.

In 1999, the GSCB dug a shaft in a depression (anticlinal valley) below the village of Fontenotte for the purpose of facilitating access to the En Versenne cave, at a short distance (a few hundred meters) from one of the largest rooms of the system (Grand Vatel). The study area is situated above this room. The landuse consists of (uncultivated?) grass crossed by only one road.

The En Versenne karstic network expands NNE-SSW between the Ognon valley and the Doubs valley where numerous faults run reveal a similar orientation. At Fontenotte, the network is embedded in the permeable middle Jurassic (Aalenian, early Bajocian), overlaid by the impermeable Liasic marls that form an aquiclude (Fig. 2).

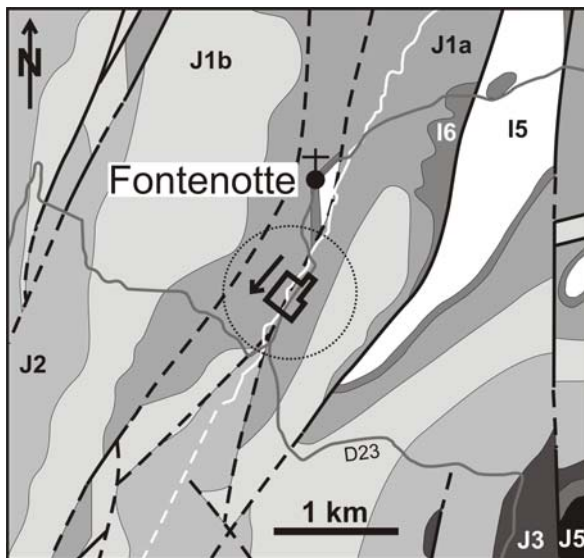


Fig. 2: Geological map of the study area, modified from BRGM (1957). I5-6, J1a, J1b, J2, J3, J5: geological units. The black thick and dashed lines show main faults. Grey lines are roads. The white line is the schematically projected En Versenne network. The black polygon in the centre is the investigated zone with a black arrow indicating the orientation of the VLF-EM maps (Fig. 4).

The general morphology of the Grand-Vatel room is elongated with an upside down truncated triangular section. The room dimensions are 60 m length, 25 m width and 25 m height in its highest part (Fig. 3). The room is located 50 m below the land surface.

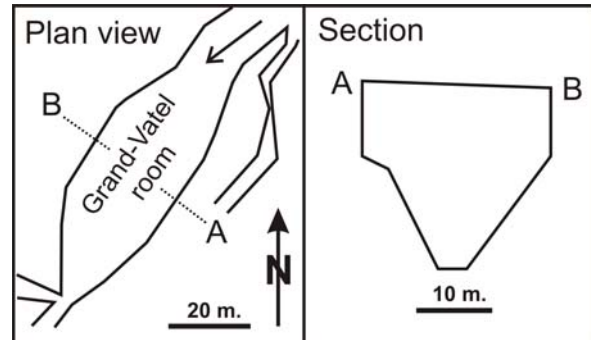


Fig. 3: Morphology of Grand-Vatel room. Plan view and section.

3. VLF-EM method

The VLF-EM method has been extensively used over the past 20 years at the Centre of Hydrogeology of Neuchâtel (CHYN) within the framework of the geophysical research of karst aquifers (BOSCH & MÜLLER 2005).

Including both very low (VLF, 3-30 kHz) and low (LF, 30-300 kHz) frequency ranges, this technique uses signals of public radio transmitters situated around the world, mainly used for navigation and military purposes.

The transmitted primary EM field consists of a vertical electric field component E_{pz} and a horizontal magnetic field component H_{py} perpendicular to the propagation direction x . At a source distance exceeding several wavelengths, the primary EM field can be considered as a plane wave.

While penetrating into the earth, the primary EM field induces a much smaller secondary field. The intensity of this field depends on the conductivity of the ground. The secondary EM field consists of a horizontal electric field E_{sx} , and a magnetic field H_s , which can be subdivided into horizontal and vertical components. One part of the secondary field oscillates “in-phase” and the other part oscillates “out-of-phase” with respect to the primary field (TURBERG & MÜLLER 1992, BOSCH & MÜLLER 2005). The penetration depth of the primary field depends on the transmitter frequency and the electric resistivity of the ground. It can be estimated by the skin depth δ_{skin} at which an EM wave is attenuated by a factor of $1/e$. The skin depth (in metres) is given by:

$$\delta_{skin} = 500 \times \sqrt{(\rho / f)}$$

with (ρ) is the electrical resistivity in Ωm and (f) the frequency in Hz.

The investigated parameters are the horizontal and vertical local resultant magnetic field components for a chosen transmitter frequency f .

According to many authors (KAROUS & HJELT 1983, TURBERG & MÜLLER 1992), radio transmitters have to be oriented as closely aligned with the strike direction as possible (N-S and NNE-SSW) in order to get the best electromagnetic response of the structure. Therefore, the following four transmitters were selected in the available frequency range: 23.4 kHz (0°N), 77.5 kHz (40°N), 128.9

kHz (35°N) and 183.0 kHz (0°N). The results of both 128.9 and 183 kHz are not presented in this paper. To carry out the investigation, the field area was subdivided into a 10 m grid by sticks and the measurements were done by foot with the VLF-EM device equipped with a data logger. The profiles have been recorded in both directions to cross-validate the results.

4. Results

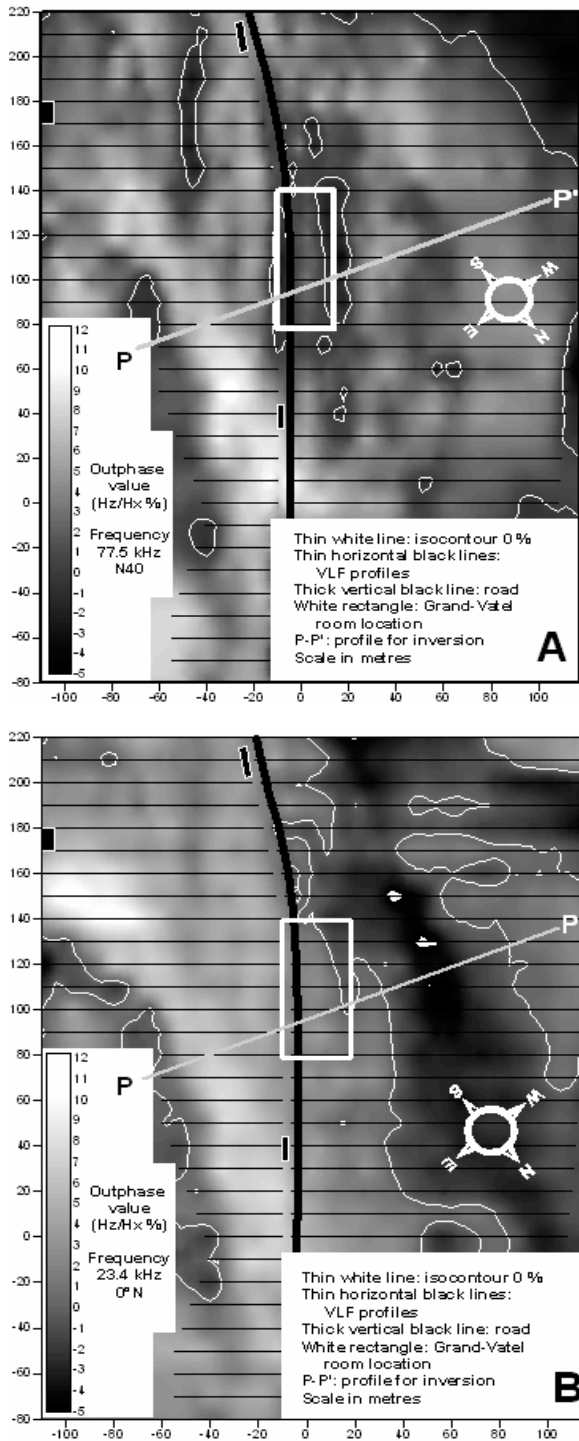


Fig. 4: Quadrature (outphase) contour plots of 77.5 kHz (40°N) (A, above) and 23.4 kHz (0°N) (B, below) radio transmitters.

The outphase variations could easily be interpolated from one profile to the other due to the high and regular data density. The results show that VLF-EM provides very consistent data. The 77.5 kHz contour map (Fig. 4.A) reveals the two main structural regional directions N-S and the associated joint pattern NNE-SSW (Fig. 6) that can also be observed on the field.

According to the location of the Grand-Vatel room (GSCB 2001) and the position and size of the two parallel outphase anomalies, we delimited the Grand-Vatel room in the white rectangle. While morphologies of the room entrance and exit are not well constrained, the width and approximate length matched very well with the outphase anomalies. Uncertainty on its location is estimated at ± 10 metres in both directions of the room.

N-S oriented structures rotating NNE-SSW northwards are also very well imaged (Fig. 4.B) by the outphase value at 23.4 kHz, while the central NNE-SSW joint pattern is hardly distinguishable. With the 23.4 kHz the penetration depth is increased so that the measurements penetrate into the Liassic marls located below the cave.

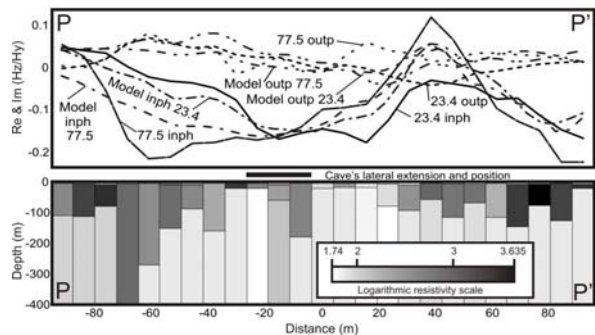


Fig. 5: Field data and model responses for both frequencies (inphase & outphase) and resistivity model.

2-D forward modelling (WANNAMAKER *et al.*, 1987) was carried out in order to attempt to reproduce the field data. Making use of local geological information (BRGM 1957, CHARMOILLE 2005), we constrained the resistivity models, starting with low resistivity values (50 Ω m) below the Grand Vatel room due to the presence of marls and higher resistivity values (400 Ω m) for the fractured media. The field data was simulated for both frequencies (23.4 and 77.5 kHz) (only) separately. The data was inverted for both frequencies (Fig. 5), using an inversion code combining an optimization routine (SCHNEGG 1993) and a finite element solution (WANNAMAKER *et al.* 1987). In this way, both inphase and outphase for the two frequencies could be reasonably fitted. The resistivity model indicates that the cave is located in a more conductive zone (that extends toward P' and corresponds to the middle of Fontenotte anticlinal valley) where strong contrasts of resistivity co-exist and seems to highlight faults.

Problems of misfit may stem from the azimuth of the two different fields in both strike directions (N-S and NNE-SSW) (FISCHER *et al.* 1983, BEAMISH 1998, MCNEILL & LABSON, 1991), the lack of resistivity data for constraining the initial model and/or the complex, possibly 3D geological framework at shallow depth.

Figure 6 shows the interpreted structural map of the studied area. The fractures pattern may be the result of consecutive periods of extension (Rhine graben, Oligocene) and compression (Alps, Mio-Pliocene) that affected the external Jura (HOMBERG & AL. 1994). The first structural orientation of N-S faults was extremely favourable for the development of such a NNE-SSW joint pattern, dictating karstification.

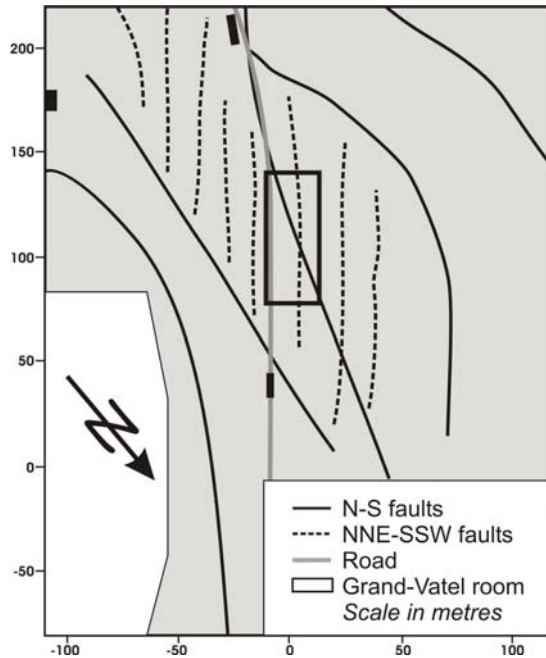


Fig. 6. Interpreted structural sketch of the study area.

5. Conclusions

The preliminary results of this study show that, in such a specific structural context, large caves at shallow depth seem to be delimited by conductive faults, which may be highlighted by the VLF-EM method.

Easy interpolation between profiles show the accuracy of high-resolution VLF-EM mapping. This method accurately highlighted the two regional directions (N-S and NNE-SSW) of the main combined faults. Inherited complex tectonics seem to have played a significant role in the development of the karstic network.

2-D forward modelling shows that even a large air volume does not seriously affect the electric current distribution. 2-D inversion processing allowed simulation and fitting of the field data. The results reveal that the Grand-Vatel room is located above a more conductive zone than what was previously believed, within a complex structure.

A first attempt of radio-location of the Grand-Vatel room was carried out in May 2005 with the GSCB, but unfortunately failed because of too restricted transmitter capacities of 1 GHz. Further geophysical investigations are going to be carried out in this area, mainly new radio-location, electrical tomography and Radio-Magnetotellurics to confirm these results and to better constrain the modelling. In addition, we would like to apply this technique on the southern unknown part of the Verne-

Fourbanne network with the aim of mapping the unknown parts.

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Modelling flow in karst aquifers: coupling turbulent conduit flow and laminar matrix flow under variably saturated conditions

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Abstract

A finite element model for simulating variably saturated, coupled conduit-matrix flow is presented. Turbulent flow in the conduits is governed by the one-dimensional diffusive wave approximation of the Saint-Venant equations. Laminar flow in the fissured rock matrix is described by the Richard's equation. Coupling is based on combining the discrete forms of these non-linear equations into one matrix system and by assuming continuous hydraulic heads on common nodes. The discrete continuum approach is used to represent the structure of the karst aquifer.

Simulations on a hypothetical karst system show the capabilities of the numerical model.

1. Introduction

The discrete continuum approach in which the conduits are represented by discrete elements within the limestone continuum is very well suited to represent the overall structure of a karst aquifer. Models based on this approach have been described by Kiraly (1998), Annable & Sudicky (1998), Clemens et al. (1996) and Kovacs et al. (2005).

However, these models only reflect the essential features of a karst system under restricted hydrodynamic conditions, assuming either laminar flow in the conduits and/or fully saturated conditions.

This study presents an innovative discrete continuum finite element model in which the turbulent flow in the conduits and the laminar flow in the fissured limestone are coupled under variable saturated conditions.

Turbulent conduit flow is described by the one-dimensional diffusive wave approximation of the Saint-Venant equations. Two different forms of this approximation are implemented in the model to deal with free-surface and pressurized flows in the conduits. The head-based form of the Richard's equation describes the variably saturated flow in the rock matrix.

As shown by VanderKwaak (1999), the one-dimensional diffusive wave approximation of the Saint-Venant equations and the head-based form of the Richard's equation can be coupled fully implicitly. By assuming continuous heads on the common nodes, the equations can be combined into one matrix system. A Picard iteration scheme accounts for the non-linearity of the partial differential equations.

The illustrative simulation example shows that: 1) the discharge at springs can be highly influenced by short term storage in conduits above the mean phreatic zone 2) the model can simulate a tailing-effect at springs related to long term storage in the matrix.

2. Conceptual model

Karst aquifers are characterized by a high permeable conduit network embedded in a low permeable fissured limestone matrix. The concentrated discharge of karst systems at springs is a direct consequence of this structure. Groundwater flow in the fissured rock is laminar. The flow

regime in the conduit network is turbulent. Depending on local hydrodynamic conditions, conduits can be dry, partially filled or pressurized.

Depending on relative head differences, the conduits can recharge or drain the fissured rock matrix. Since changes in head occur more rapidly in the conduits, conduits can successively recharge and drain the matrix after infiltration events.

3. Governing equations

3.1 Matrix flow

The variably saturated flow in the fissured rock matrix is described by the head-based form of the Richard's equation:

$$C(\theta) \frac{\partial h}{\partial t} = \nabla \cdot (K(\theta) \nabla h)$$

where C is a capacitive term, θ the volumetric water content, h the hydraulic head, t the time, K the conductivity tensor. For saturated flow the capacitive term is defined by:

$$C = S_s = \rho g (\kappa_a + \phi \kappa_w)$$

where S_s is the specific storage coefficient, ρ the density of water, g the gravity, κ_a the compressibility of the aquifer, κ_w the compressibility of water and ϕ the porosity. For unsaturated flow the capacitive term is defined by:

$$C(\theta) = \frac{\partial \theta}{\partial p}$$

where p is the pressure head.

The hydraulic head can be used as independent variable by using the relationships of Van Genuchten:

$$\theta(p) = \theta_r + (\theta_s - \theta_r) \left[\frac{1}{1 + (\alpha p)^n} \right]^m$$

$$K(\theta) = K_s \left(\frac{\theta - \theta_r}{\theta_s - \theta_r} \right)^{\frac{1}{2}} \left[1 - \left(1 - \left(\frac{\theta - \theta_r}{\theta_s - \theta_r} \right)^{\frac{1}{m}} \right)^m \right]^2$$

where K_s is the conductivity tensor under fully saturated conditions, θ_r is minimum volumetric water content, α the inverse of the air-entry pressure and n an empirical parameter. The parameter m is defined by:

$$m = 1 - 1/n$$

Alternatively $\theta(p)$ and $K(\theta)$ can be derived by using by an exponential model:

$$\begin{aligned} \theta(p) &= \theta_r + (\theta_s - \theta_r) e^{-\alpha p} \\ K(\theta) &= K_s e^{-\alpha p} \end{aligned}$$

3.2 Conduit flow

Variably saturated turbulent flow in the conduits is described by the diffusive wave approximation of the Saint-Venant equation:

$$C(p) \frac{\partial h}{\partial t} = \frac{\partial}{\partial s} \left(\frac{K^*(p)}{\sqrt{\left| \frac{\partial h}{\partial s} \right|}} \frac{\partial h}{\partial s} \right)$$

where C is a capacitive term, s the space coordinate in the direction of flow and K^* the conveyance factor. Following Manning-Strickler the conveyance factor is given by:

$$K^* = \frac{1}{\eta} A R_H^{2/3}$$

where η is the Manning's coefficient for friction, A the cross-sectional area of flow and R_H the hydraulic radius. The hydraulic radius is defined as:

$$R_H = \frac{A}{P}$$

where P is the so-called wetted perimeter. The equation for conduit flow can be used for free-surface flow as well as for pressurized flow. For free-surface flow the capacity term is defined by:

$$C(p) = W(p)$$

where $W(p)$ is the top width of the flow. For pressurized flow the capacity term is defined by:

$$C(p) = \rho g A \kappa_w$$

If the conduit flow is not fully saturated, the capacitive term and conveyance factor depend on the pressure head. Expressions for the conveyance factor as function of pressure head can be derived for a variety of conduit

geometries. Worked out examples are provided by Panday & Huyakorn (2004).

4. Numerical implementation

The equations for matrix flow and conduit flow are solved by a finite element scheme. The conduit network is represented by linear discrete elements. The discretization of the matrix continuum into linear tetrahedral elements provides flexibility for fitting in a conduit network.

Fully implicit and Crank-Nicholson time marching procedures may be used in the model. Mass lumping is implemented to improve the stability of the scheme.

Because the two flow regimes are described by partial differential equations of a similar form, all the conductance and conductivity matrices on the element level can be combined into one matrix system by assuming continuous heads on the common nodes. This provides a robust, fully implicit coupling of the two flow regimes. The matrix system is solved by a conjugate gradient solver.

A standard Picard iteration scheme is used to account for the non-linear terms. Adaptive time-stepping is implemented, by making the timestep dependent on the number of Picard iterations.

At each iteration the pressure heads are needed to determine the saturation and to evaluate the capacitive and conductive terms. For the conduits the saturation is checked by comparing the pressure heads with the diameter of the conduits. If negative pressure heads occur in conduits, the conductive and capacitive terms are given zero values.

Without introducing a special treatment of the wetting process in the conduits, the wetting of conduit nodes takes too much time. This can result in serious mass balance errors. To speed up the wetting process, small artificial pressure heads are defined on conduit nodes ahead of a wetting front. This procedure decreases the mass balance errors.

The model offers several options for defining boundary conditions at the springs, generally used in the field of modelling channel flow (Panday & Huyakorn, 2004). Fixed heads (for under water springs), zero-depth gradients and critical depth conditions can be imposed on the springs. The zero depth gradient boundary condition is given by:

$$Q = -\frac{1}{\eta} A R_H^{2/3} \sqrt{S_0}$$

where Q is the volumetric discharge and S_0 is the bed slope at the boundary. The critical depth boundary condition is given by:

$$Q = -\sqrt{\frac{g A^3}{W}}$$

5. Simulation example

Figure 1 shows the simulated hypothetical karst system. The simulations illustrate the capabilities of the model. Mass lumping and an implicit time marching procedure are used. Values for the saturation and the conductivity are calculated by the exponential model.

The conduit network is discretized into 1D linear elements of about 0.5 m length. The conduit network has one

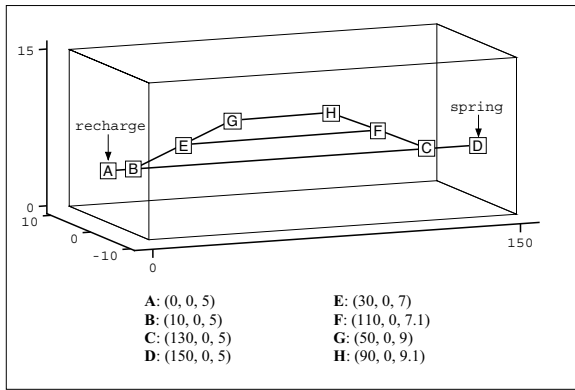


Figure 1: The simulated hypothetical karst system (dimensions in meters).

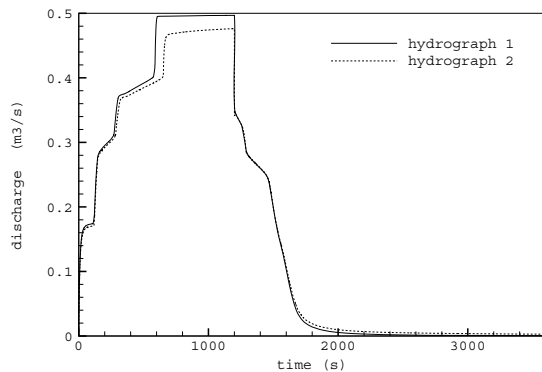


Figure 2: The two simulated hydrographs at the spring.

recharge point and one spring. The matrix is discretized into 64476 tetrahedra. Except for one recharge point and a fixed head boundary on the plane containing the spring no-flow boundaries are imposed.

Conduits EF and GH have a radius of 0.5 m and the other conduits have a radius of 0.25 m. The Manning's coefficient for friction is $0.05 \text{ s m}^{-1/3}$.

The initial heads are 5.5 m and the karst system is recharged with $0.5 \text{ m}^3/\text{s}$ for 20 minutes.

Figure 2 shows two simulated spring hydrographs. The first hydrograph is simulated with $K_s = 0.1 \cdot 10^{-5} \text{ m/s}$ and $\phi = 0.1 \cdot 10^{-2}$. The second hydrograph is simulated with $K_s = 0.1 \cdot 10^{-4} \text{ m/s}$ and $\phi = 0.5 \cdot 10^{-2}$.

It is clear that during the recharge period water is stored in the karst system and that this stored water is drained towards the spring after the recharge period.

The form of the hydrographs is mainly influenced by the geometry of the conduit network above the initial water table. The two initially dry, horizontal conduits have a pronounced effect on the hydrograph.

The first two periods where the rate of change of the discharge is decreasing are related to the filling up of the horizontal conduit EF. The sudden increase of the discharge between these two periods happens if conduit CF is filled up. The third period where the rate of change of the discharge is decreasing is related to the filling up of the horizontal conduit GH. If the whole conduit network is filled up, all conduits are pressurized and accordingly the discharge at the spring suddenly increases.

After the recharge period, the conduit network is depressurized and the two periods where the rate of change of the dis-

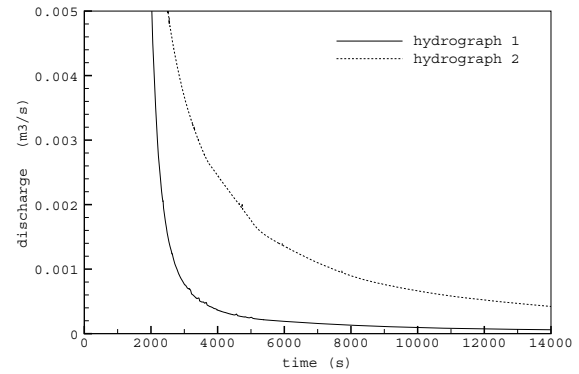


Figure 3: The simulated tailing effects in more detail.

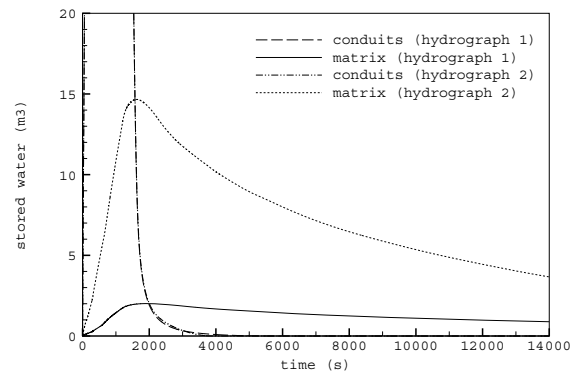


Figure 4: The amount of stored water in the conduit network and the fissured matrix during the simulations.

charge is decreasing are related to the drainage of the two horizontal conduits EF and GH.

During the period that the conduit network is pressurized the hydrograph clearly indicates that water is being stored in the fissured matrix. More water is stored in the fissured matrix if K_s and ϕ have higher values.

The hydrographs shown in figure 3 show the tailing effects at the spring in more detail. Figure 4 shows the amount of water being stored in the conduit network as well as in the matrix during the simulation. As can be observed from figure 4 the long-term tailing effects at the spring are related to the drainage of the fissured matrix by the conduit network.

6. Conclusions

An innovative, physically based model for simulating coupled conduit-matrix flow in variably saturated karst systems has been discussed. The structure of the karst system is well represented by the discrete continuum approach and its hydrodynamic behaviour is well captured by the implemented equations. The model fully accounts for the two different flow regimes in karst systems under variably saturated conditions: turbulent flow in conduits and laminar flow in the rock matrix. The fully implicit coupling of conduit-matrix flow is considered physically correct as well as robust.

It has been shown that the model is able to simulate the effects of temporal storage in variably saturated karst systems on spring hydrographs.

The model is considered a powerful tool applicable to various topics in karst hydrogeology. Simulations will help to find answers on questions concerning the storage of groundwater in karst systems and the availability and evolution of water resources in the matrix for various degrees of karstification. The model will also be important to improve the inferences on aquifer parameters from spring hydrographs and to test the applicability of existing models.

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Hydrology and Rock/water Interactions of an Alpine Karst System: Spring Creek, Mineral King, Sequoia National Park, California

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Abstract

The hydrology of a steep, alpine karst spring with dominant allogenic recharge was examined using high resolution data determined from measured water levels. Chemical data was used to calculate the denudation rate for the carbonate rock in the basin. The discharge showed as many as six patterns. Seasonal discharge patterns show no response to snowfall events, but a strong response to rainfall in late summer. Daily discharge patterns appear to be snowmelt pulses that arrive at the spring at varied times due to varied retention times. 12-hour patterns that vary with the seasons are interpreted as due to increased plant transpiration. The denudation rate in the alpine basin is high.

Introduction

The Spring Creek research site was established in 2001 at an elevation of 2425 meters in the Mineral King Valley of the Sierra Nevada of California to produce high-resolution data documenting water level, temperature, electric conductivity, and pH for this karst spring. High-resolution data sets and hydrographs from karst aquifers and springs are unusual, and this data set allows an examination of this aquifer in great detail (GROVES & MEIMAN, 2000).

The extent of the recharge basin was determined by SCHULZ & TINSLEY (1996) using dye and conductivity tracers. The spring drains a basin of 7.8 km². Transit time to the spring from the White Chief sink was 3.5 days during near base-flow conditions.

Vertically-bedded, metamorphosed marble covers 0.65 km² of the basin and hosts the Spring Creek Spring. The marble is part of a larger pendent of Mesozoic marine rocks composed mostly of volcanic sediments and surrounded by Sierran granitics (BUSBY-SPERA, 1982). Due to the limited extent of the carbonate rock, most recharge is allogenic, including the primary sinking streams of the basin, Eagle and White Chief creeks. The topographic apex of the basin is Vandever Peak at 3642 m.

Karst hydrology in the upper basin is complex with numerous springs, sinking streams, sinkholes, and caves formed within the carbonate unit. White Chief Creek flows through four separate caves before sinking a final time 2.8 km upstream of Spring Creek Spring (Figure 1) at an elevation of 2930 m. A series of sinkholes parallels Eagle Creek before it sinks into the marble at 2700 m. amsl and 1.7 km above the spring, which lies at 2,500 m. (SCHULZ, 1996). Gradients from the sinking streams to the resurgence are 15.3% and 11.75% respectively.

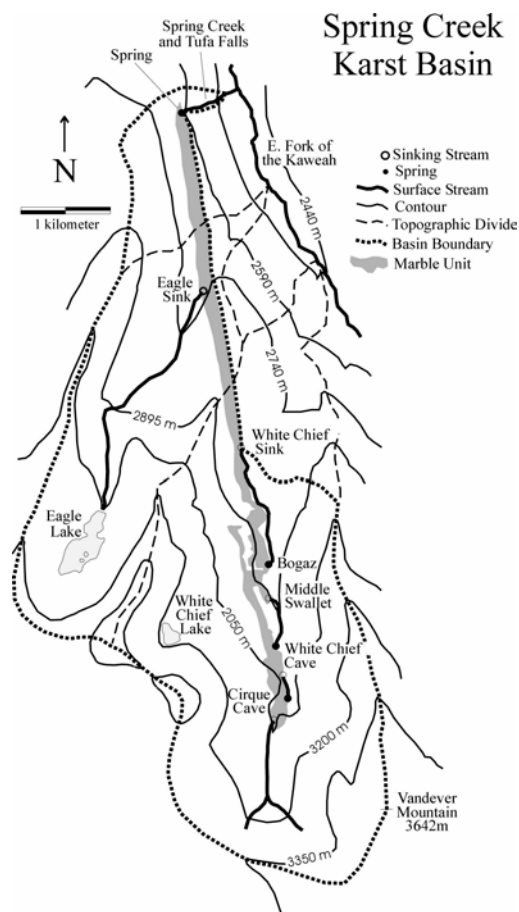


Figure 1: Spring Creek watershed

The climate in Mineral King is Mediterranean, with warm dry summers and wet cool winters. Most precipitation falls as snow, which is generally retained until warmer spring temperatures promote runoff. Average peak snowpack on the floor of Mineral King Valley (below the basin) is 51.05 centimeters of water content (BLACK, 1994). Mean average precipitation at the closest meteorological station (Lower Kaweah, 25 kms. northwest and at 1900 m. elevation) from 1980 through 2003 is 94 cm. Remnant snow fields remain throughout the summer months on north-facing slopes at high elevations.

Methods

Water level data for the spring was collected in an instrumented pool 200 m. below the spring orifice using a pressure transducer. Periodic tracer-velocity-area discharge measurements throughout the study period allowed the transformation of the water level data to discharge. An exponential growth regression using 12 calculated discharges produced a fit of 0.977 and a line formula of:

$$Y=e^{13.3964 \cdot x}$$

While data were collected from late 2001 through 2003, the 2002 data set has the fewest omissions or errors and also captures that year's spring runoff, therefore the 2002 data set will be considered in this paper. Basic statistics were applied to the data.

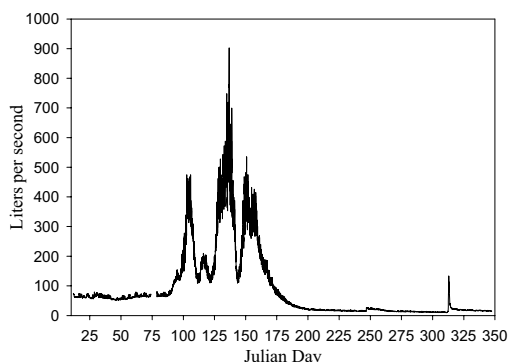


Figure 2: Spring Creek Spring discharge for day 11 to 348 2002.

Mean daily discharge data for the East Fork of the Kaweah are collected by Southern California Edison Company (SCE) below Mineral King where water is diverted into a flume and at the power generating station where the flume ends. This data is furthered compiled in California water resources data reports. (FREEMAN, 2003 and ROCKWELL 2004).

Temperature data is collected every hour by the National Park Service at the Lower Kaweah meteorological and air quality monitoring station, located in Giant Forest, Sequoia National Park. This data for 2002 was purged of omissions and daily means were generated for this study.

Precipitation data, including snowfall converted to actual water content, were also collected at the Lower Kaweah site.

Results

Larger-scale examination of the discharge data reveals six periodic trends: 1) An annual pattern with peak discharges for approximately 20% of the year and longer periods of base flow; 2) a 24-hour pattern; 3) a 12-hour pattern; 4) a five-day pattern; 5) trends that develop over several weeks; and 6) an irregular three- to 15-day pattern of approximately 36-hour periods of uncharacteristically low variance and little or no response to the usual diurnal pattern.

The seasonal pattern shows extended periods of base flow, a single spring run-off event with three peaks and a strong response to late-season rainfall events.

At time scales of several days, daily patterns and the 12-hour patterns are clear. However, the two vary. During the winter months, low flow at the spring occurs close to midnight. In later months, following high runoff, high flows are often seen at night (Figs. 3 and 4). Eighty six percent of the year, the daily peak discharge is at or near midnight, correlating with Tinsley's low-flow transit time of 3.5 days.

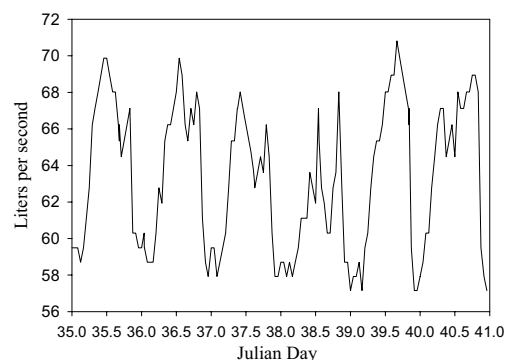


Figure 3: Day 35 to 41 2002 discharge values for Spring Creek showing low flows near midnight.

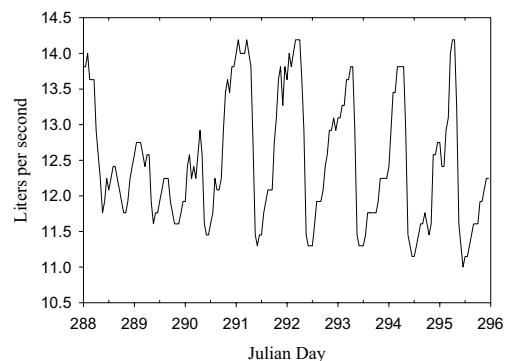


Figure 4: Discharge values for Spring Creek day 288 to 296 2002 showing high flows near midnight.

Daily discharge variances were calculated for three 10-day time periods for mid-winter, spring run-off, and late summer base flow (Table 1). The lowest variance (and lowest discharge) was seen in late summer when snow and ice in the basin are at their annual minimums, and, thus, have the least effect on the daily change in discharge.

Table 1: Daily discharge variance during three 10-day periods in 2002

Julian Days	Variance
53 to 62	13.017
127 to 136	280.305
260 to 269	3.429

A moderate variance was seen in late winter when snow covers the basin, but cold temperatures preclude large-scale melting. The highest values were seen for the period of spring run-off, when higher temperatures and longer, sunny days lead to a large daily variance in flow. Twelve-hour patterns are more subdued and difficult to discern during winter months.

Comparison of the discharge hydrographs from the Spring Creek Spring and the East Fork of the Kaweah for 2002 reveal parallel patterns of discharge (Fig. 5). Spring runoff begins about day 55 in both graphs. This appears to

have been initiated by the first days of 2002 with mean temperatures above 10° C on day 52 and 53. Mean temperatures above 12° C on days 90, 91, 92, 103, and 104 added to the melt and rising discharges. The three-spiked apex of high seasonal flows is seen in both discharge graphs. This pattern reflects regional changes in temperature and is not basin specific. The decline in discharge following runoff is parallel in both graphs as are the extended periods of base flow.

There are two late season anomalies seen in both discharge graphs – the rise in discharge on approximately day 245 and the spike in discharge for the spring and the large drop in discharge for the river on approximately day 312. The day 245 event is localized rainfall event in the East Fork basin, but not in the Kaweah watershed as a whole, as is common in the region due to summer monsoonal moisture. The day 312 values resulted from a rainfall event that produced more than 20 cm of precipitation, generating a 50-year flood event throughout the Kaweah watershed. Debris in the river and mechanical problems with SCE equipment resulted in incorrect low discharge data values for this time period. Actual discharge after the storm can be seen by day 324.

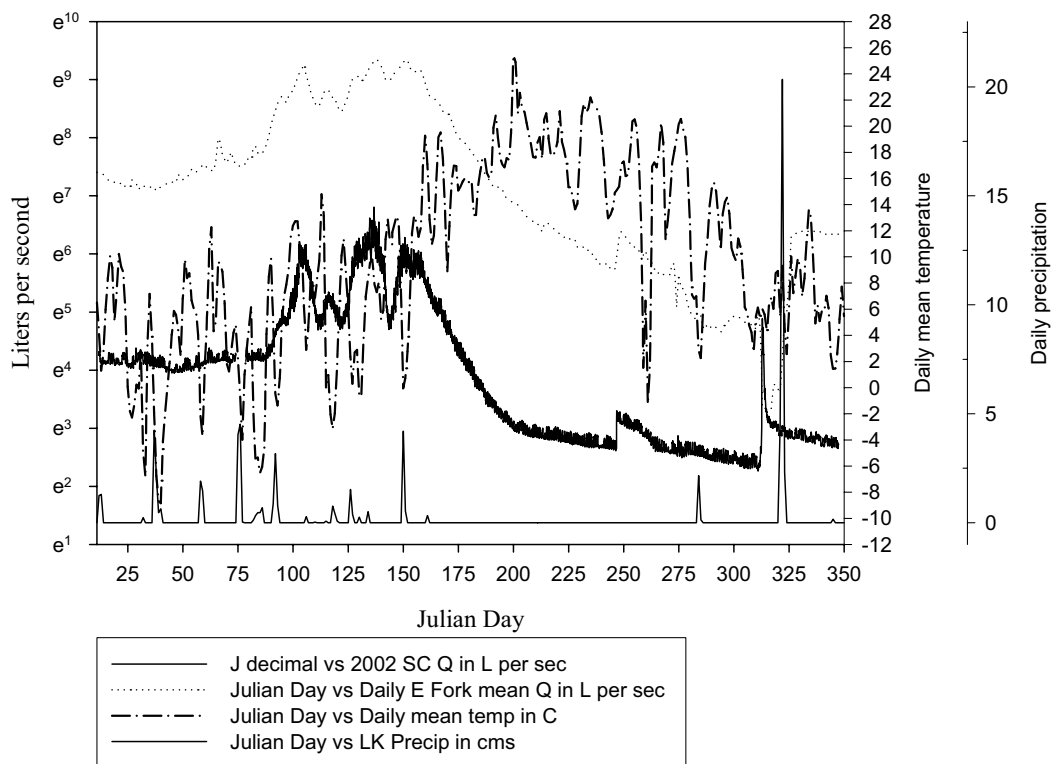


Figure 5: Discharge for Spring Creek and the E. Fork of the Kaweah, daily mean temperature and precipitation from Lower Kaweah for 2002.

Other relationships can be seen between rising temperatures and rainfall events versus discharge. Lower temperatures during storms on days 104 to 108, 117 to 125, and 139 to 145 correlate with lower discharges during the peak spring runoff period. This effect created the varied discharge in these sections of the hydrographs. Discharge during run-off correlates negatively with storm events that bring lower temperatures, increased cloud cover, and less snow melt.

Ions in solution in the spring water were calculated using high-resolution conductivity measurements calibrated by 14 water samples analyzed at the Mammoth Cave National Park water chemistry lab. Regressions for each ion were used to generate continuous data throughout the 2002 sample period. Rainfall values for ions from Lower Kaweah for the time period were very low – Ca averaged .0558 mg/L for 2002 – and so, were not included in this analysis. 1,688,023 mols or 62.57 m² of CaCO₃ were removed from the basin in solution. This produces a denudation rate of 148.6 mm/1000 years.

Conclusions

Seasonal discharge patterns show little or no response to snowfall events in the basin, but a response to lower temperatures and cloud-cover, and a strong response to rainfall events in later summer. Daily discharge patterns appear to be snowmelt pulses that arrive at the spring at varied times due to varied retention times within the hydrologic system. Twelve-hour patterns that vary with the seasons are interpreted as increased plant transpiration during warm summer afternoons. Denudation rate in the alpine basin is high.

This may be due to allogenic recharge, prevalent granitic sediments in basin streams, low pH runoff from alpine meadows, anthropogenic acid-rain effects, the exposed carbonate bedrock, or other causes.

In this alpine setting with alleogenic recharge largely from plutonic rocks and steep gradients we see a quick response to storm events and spring runoff and an overall discharge that closely reflects that of the greater basin. The effects of snow cover, temperature and cloud cover all play a role in discharge patterns, which are many and complex.

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Climate change and water resources in limestone and mountain areas: the case of Firenzuola Lake (Umbria, Italy).

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Abstract

The Firenzuola reservoir (catchment 24 km²) is located on the River Marroggia, in a limestone mountain area in the Umbria-Marche Apennines of Central Italy. The dam, completed in 1962, was built for purposes of flood control and irrigation-water storage. The hydrogeological budget shows that part of the water reaches a regional karstic system: some of the water of the catchment supplies a very large high-salinity karstic spring about 30 km away from the dam. The climatic data indicate that rainfall is decreasing, consistently with the Western Mediterranean regional climatic trend. The water budget indicates that the current water yield is lower than predicted with reference to the climate data at the time of planning the reservoir. Some future scenarios have been developed on the basis of the hydrogeological set up and the detected trends, considering the hypothesis that they will continue for the next few decades: all the scenarios indicate that the yield of the basin is likely to decrease. In order to minimize the impact of the foreseeable climate change, it is necessary to review in depth the water and reservoir management plans as well as those for exploitation of the groundwater in the area.

Introduction

The area considered in this paper (Fig. 1) is located in the Umbria-Marche Apennine, in S-W Umbria, Central Italy.

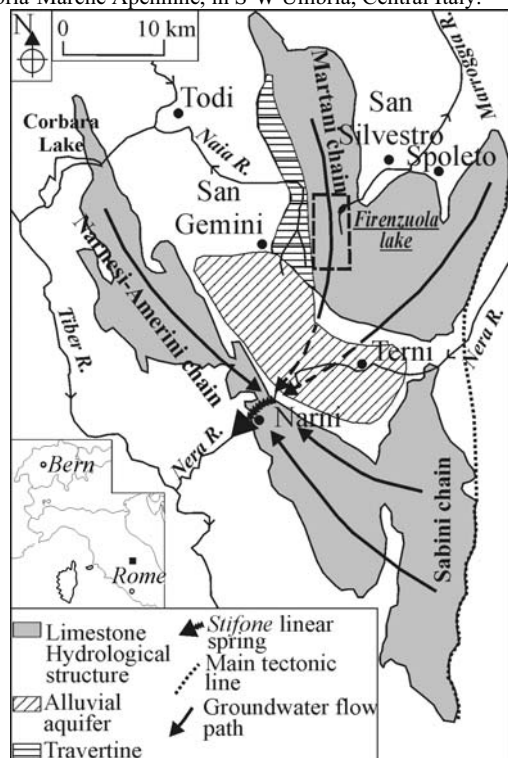


Fig. 1 – Location of the study area and layout of the water system feeding the Stifone linear spring (layout adapted from BONI, 2000). The alluvial and the limestone aquifers are divided by an aquitard (showed in the figure by the outlined limestone groundwater flow path).

Since the second half of 19th century, this area has been affected by a climatic trend, very probably to be linked to Global Climatic Change (cf. for example DRAGONI, 1998; SINGER, 1999; IPCC, 2001; BARAZZUOLI *et al.* 2002; COTECCHIA *et al.* 2003). According to elementary physics and Global Circulation Models (GCMs), on a global scale the increase in temperature should be associated to an increase in evaporation and therefore in rainfall; nevertheless there are local situations in which the rainfall could decrease. This seems to be the case of Central and Southern Italy of which many climatic datasets have been analysed in order to verify if they had a trend. The presence and intensity of trends vary according to the statistical technique employed, but the general picture is rather clear:

not all of the stations have a statistically significant trend, but – in practice – all of the stations having a trend show a decrease in rainfall and an increase in temperature, with linear rainfall gradients ranging from -2 mm/year up to -6 mm/year, and temperature gradients up to about 1 °C/century (DE FELICE & DRAGONI, 1994; DRAGONI, 1998; TRENBERTH, 1999). Furthermore it has been detected that in many meteorological stations there is a significant linear relationship between mean annual temperature and mean annual rainfall, and such a relationship indicates that an increase in temperature is associated with a decrease in rainfall, with a linear gradient ranging between - 40 mm/°C and -130 mm/°C. No stations present temperature/rainfall data indicating an opposite significant relationship.

This situation is consistent with what has occurred during the recent past: the paleoclimatic studies carried out in Central-Southern Italy and Sicily show that during the upper Holocene (i.e. in the last 3000 years) the climate alternately shifted between slightly warmer or slightly colder conditions (within a range probably no wider than ±0.5-1 °C), at intervals of a few hundreds of years, and that warmer periods were drier than cooler ones (DRAGONI, 1998).

Temperature increase and rainfall decrease inevitably lead the water yield of hydrological systems to be reduced: indeed the analysis of the discharge data series of some hydrogeological systems located in Central Italy showed a negative trend. Although for some systems it is not easy to define in which proportion this trend is to be related to climatic variations or to over-exploitation, there are some systems – particularly mountain catchments and springs – which can be considered uninfluenced by human activity and which show a decrease in yield in the last forty – fifty years (DRAGONI, 1998; CAMBI & DRAGONI, 2001).

In the area here considered, a peculiar hydrogeologic characteristic is the groundwater yield of the Narnese-Amerina limestone massif (fig. 1), which is more than 80% larger than what is to be expected in terms of difference between rainfall and actual evapotranspiration over the massif's area (BONI *et al.* 1986; BONI, 2000): indeed the Nera river, in the gorge below Narni, increases its discharge of about 10-15 m³/s (fig. 1), through a set of karstic springs, the larger of which is located close to the Stifone village and has high salinity (in the following, for sake of simplicity, we shall call collectively all the springs “Stifone Springs”). According to BONI (2000), such a high discharge can be explained only assuming a connection between the Stifone springs and the limestone Martani Mts. (fig. 1), located about ten kilometres north of the Narnese-Amerina chain. A recent investigation carried out by the present authors for the Umbria Geological Regional Office, in spite of some differences in the total yield, confirms that a large percentage of the Stifone discharge cannot originate from

rainfall over the Narnese-Amerina chain, and that the Stifone Spring is the main outflow of a regional karstic groundwater circuit.

In the framework of the situation just described, the present paper reports the results of an investigation regarding the water balance of the Firenzuola reservoir, located in the Martani Mts., in the basin of Marroggia River (fig. 1) and its connections with the karstic and salty Stifone Springs. These issues are important for the following reasons:

- The Firenzuola dam was planned in the 1950s and, at the time, the expected water yield was based on the rainfall of the 1920-1950 period, higher, at regional scale, than the present one and of the one that can be expected for the next decades: the research investigates the existence at local scale of climatic trends, aiming to give an idea of the present water balance and that to be expected in the next future.

- In the area there is the need for new supplies of high quality groundwater, but most of the springs are already exploited as well as the alluvial aquifers. The large Stifone Springs are useless because of their high salinity: it would be a good management strategy to exploit the groundwater drained by the Stifone Springs before it becomes salty. Based on the water budget of the Firenzuola reservoir, the present paper reports a first estimate of the quantity of non salty water entering, from

the area where the Firenzuola reservoir is located, in to the regional flow feeding the Stifone Springs.

Geological and hydrogeological characteristics

Figure 2 shows a geological outline of the area where the Firenzuola reservoir on the Marroggia River is located; figure 2 is based on the studies by COLACICCHI *et al.* (1970); PIALLI (1976); MARIOTTI *et al.* (1979); COLACICCHI *et al.* (1988); SERVIZIO GEOLOGICO NAZIONALE (1991); CATTANEO (1995), LITI (1995) and CARLUCCI (2002) and on some new geological mapping by the present authors. The rocks of the hydrological basin of the Firenzuola dam belong to the Umbria-Marche sequence: a detailed description of the geology of the area can be found in the above given bibliography, while in the following only a short synthesis is given.

The Umbria-Marche sequence begins with the Raetic dolomitic limestone *Raetavicula Contorta* (medium-low permeability), which does not outcrop in the Firenzuola basin, but outcrops in the Narnese-Amerina Chain, just N of the Stifone Springs (fig. 1).

Most of the Firenzuola catchment as well as the Martani Mountains (fig. 1 and 2) are made up by the “complete” Umbria-Marche sequence, and the Hettangian-Sinemurian *Calcare Massiccio* (CM, high permeability karstified limestone) is the more ancient outcropping term (cf. fig. 2, stratigraphic column A).

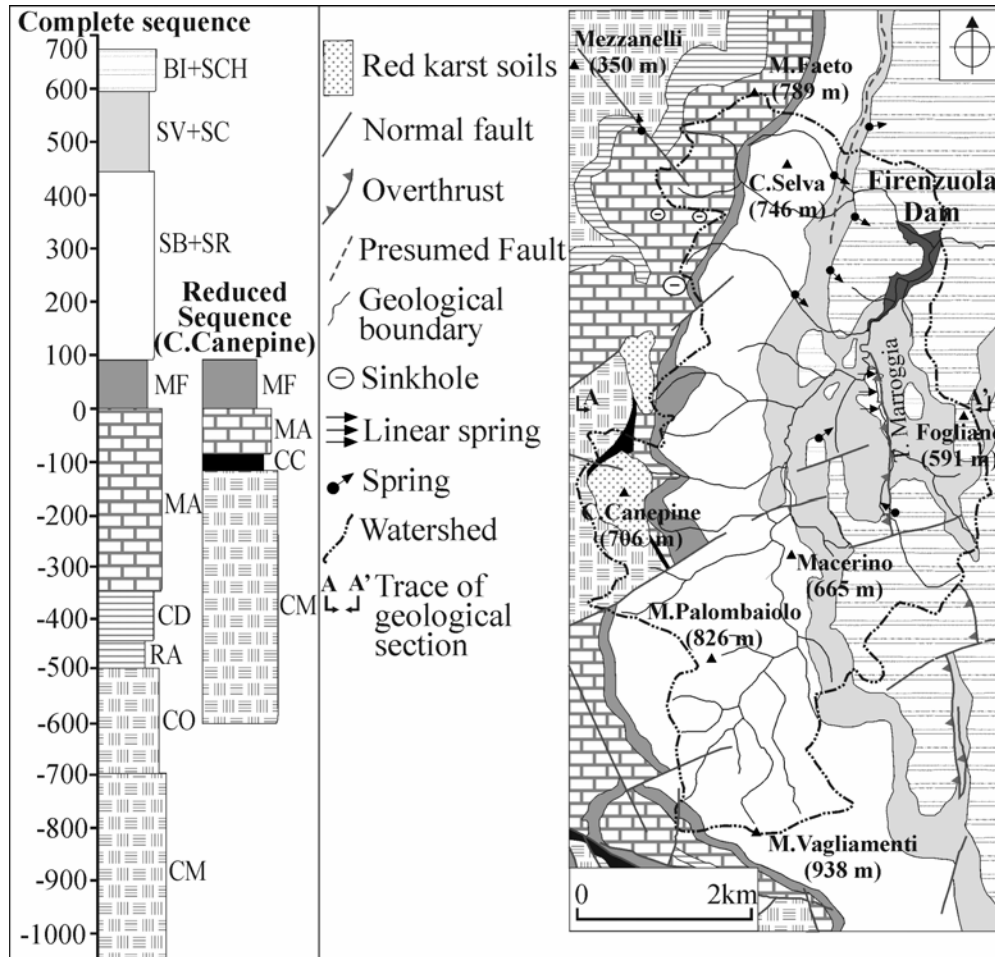


Fig. 2 – Geological layout of the hydrographic basin of the Firenzuola dam (the stratigraphic columns are simplified after BARCHI, 1991). CM = Calcare Massiccio (banks of crystalline limestone); CO = Corniola (grey micritic limestone); RA = Rosso Ammonitico (red or greyish-yellow marls); CD = Calcarei Diaspri (regularly stratified flint-limestone); CC = C.Canepine formation (modular limestone); MA = Maiolica (well-stratified white micritic limestone); MF = Marne a Fucoidi (marls and marly limestone); SB+SR = Scaglia Bianca e Scaglia Rossa (limestones well stratified); SV+SC = Scaglia Variegata e Scaglia Cinerea (marly limestones and calcareous marls); BI+SCH= Bisciaro e Schlier (marls and siliceous-marly limestone).

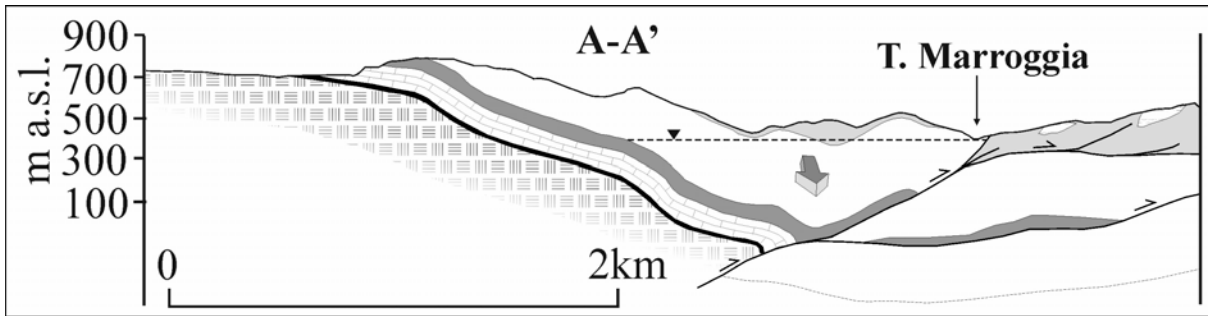


Fig. 3 – geological section (trace is represented in fig. 2).

In a small area (around C. Canepine village) the sequence is “reduced” (fig. 2, stratigraphic column B): here the Jurassic *Corniola* (CO, high permeability), the *Rosso Ammonitico* (RA, low permeability) and *Calcari Diasprigni* (CD, low permeability) formations are replaced by a few metres of *Nodular Limestones* (CC, low permeability) and the *Maiolica* (MA, high permeability limestone) is only a few tens of metres thick.

The Umbria-Marche sequence rests over Triassic anhydrides, not outcropping in the Firenzuola basin, but located everywhere below the sequence: when the groundwater reaches the anhydrides, salinization occurs, as in the case of Stifone Springs. From a structural point of view, the hydrographic basin of the Firenzuola dam lies within the “box fold” of the Martana chain that tends to roll over in the eastern part onto a thrust fault of variable depth, running in a WSW to SSW direction (BARCHI, 1991; BROZZETTI & LAVECCHIA, 1995). In the study area it is possible to note the secondary plays associated with the thrust fault present further eastwards (cf. geological cross-section in Fig. 3).

The Cretacic formation of the Maiolica is affected by karstic phenomena, clearly highlighted by the presence of sinkholes, mainly in the north-western part (Fig. 2). From the hydrogeological point of view, with reference to the stratigraphic columns of fig. 2, within the Umbria-Marchigiana sequence there are lithotypes featuring very different relative degrees of permeability, from the very permeable or relatively low permeability of the mainly marly lithotypes (*Rosso Ammonitico*, *Calcari Diasprigni*, *Nodular Limestones* of C. Canepine, *Marne a Fucoidi*, *Scaglia Variegata-Scaglia Cinerea*, *Bisciaro and Schlier*) to the medium-to-high or very high permeability of the fractured limestone calcareous lithotypes (*Calcare Massiccio*, *Corniola*, *Maiolica*, *Scaglia Bianca-Scaglia Rossa*). In order of importance, the *Scaglia Variegata-Cinerea* (called here SV+SC) has a predominant role in local water circulation, being a no-flow boundary of the fractured aquifer of the *Scaglia Bianca-Scaglia Rossa* (called here SB+SR), as shown by the small non-perennial springs located at the contact between the two units (fig. 2). These springs are situated in the northern part of the hydrographic basin, where the altitude of stratigraphic contact between the SB+SR and the SV+SC is about 550-600 m a.s.l. The hydrogeological setting along the River Marroggia (cf. geological cross-section in Fig. 3) is a key point of the river and Firenzuola reservoir: the displacement of the thrust fault associated with the erosion caused by the river has caused the aquifer of the SB+SR to surface in a strip comprised between 400 and 450 m a.s.l. Along this stretch of the river, which is dry in summer, there are a series of linear springs that feed the river from the hydrographic left. These springs, which are the main source of groundwater for the river, constitutes a local groundwater system, whose yield is made up by a kind of overflow of the deeper, regional systems (cf. fig. 3). Here the term “overflow” is not referred to a specific kind of karst springs as defined by BÖGLI (1980, p. 121), but is meant as the outflow of the upper part of a deep aquifer, also feeding a

regional flow. These “local springs” are particularly vulnerable to the decrease in recharge, as this would correspond to a decrease of the piezometric head, so that they would be much more severely affected by a recharge decrease than the water bodies fed by deep and regional flow.

Climatic features

Climatic characterisation of the basin of the Firenzuola reservoir was carried out by examining the rainfall and temperature data for the period 1992-2002, for which data concerning the level of the lake and the volumes of water released through the dam were available (CARLUCCI, 2002). It should be pointed out that there are unfortunately no stations at altitudes higher than that of the dam. For this reason, the average rainfall on the catchment was calculated on the basis of the correlations between altitude and precipitation for the stations closest to the basin, that is to say San Gemini (337 m a.s.l), Spoleto (317 m a.s.l), San Silvestro (387 m a.s.l) and Terni (170 m a.s.l), and of the rain-gauge at the dam (410 m a.s.l), located in figure 1. On analysing the data it was found that the figure of the rainfall provided by the rain-gauge at the dam was very close to that of the station in Terni, about 330 m of altitude lower. The analysis (using the “double mass” method) for the consistency of the data showed that data recorded at the dam were reliable, indicating thus the probable existence of particular local microclimatic conditions. Fig. 4 shows the correlations between altitude and precipitation. At the average altitude of the lake (398.9 m), an average yearly rainfall of 0.883 m/year is found, while at the average altitude of the hydrographic basin the rainfall is 0.990 m/year.

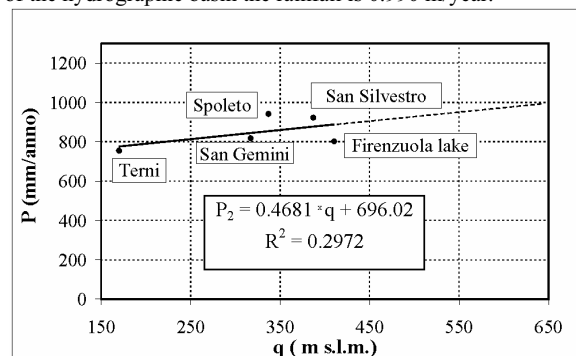


Fig. 4 – Correlation between elevation and precipitation used to define the climatic characteristics of the study area. (data supplied by Servizio Idrografico - Regione Umbria).

The yearly average air temperature at the lake was obtained by a thermometer at the dam (13.6 °C) while that at the average altitude of the basin was estimated by assuming the theoretical vertical thermal gradient of -0.006 °C/m . The existence of a decreasing trend of precipitations is clearly noticeable on analysing the yearly rainfall at the Spoleto station (Fig. 5),

showing a significant drop by -3 mm/year over the period 1920-2002 (F-test significance greater than 95%). Similar results are given also by the shorter data set of Terni.

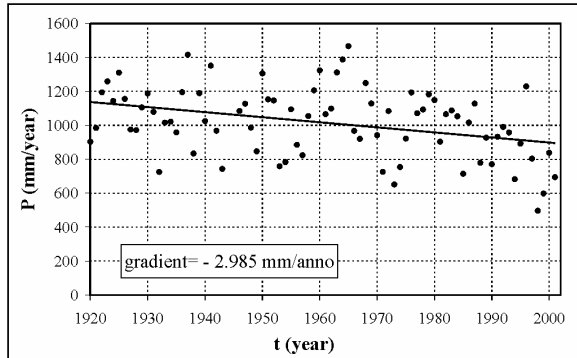


Fig. 5 – Precipitation trend at the Spoleto Weather Station for the period 1920-2001.

Hydrogeological budget

The hydrogeological budget of the Fiorenzuola basin was calculated for the 1992 – 2002 period, using the following: $S = P_b - ETR_T$ (1)

In equation (1) the letters have the following meaning:

S = average yearly water yield;

P_b = average rainfall P_b over the basin;

ETR_T = actual evapotranspiration, estimated by applying Turc's formula (TURC, 1954).

Turc's method, whose validity is often questioned, was chosen because the lack of specific weather and climatic data did not enable the use of sounder methods like, for instance, the Penman-Monteith (cf. WARD & ROBINSON, 1999). It must be noted, however, that in temperate climate areas as Europe, Turc's method is widely used and appears to give good results. In particular, in Central and Northern Italy (in zones where there is a soil cover, as in our case), Turc's method compares well with the "measured" data, as it will be shown by the following examples. In catchments located in areas of low permeability rocks, i.e. where the groundwater flow can be considered nil, on an average yearly basis the stream's discharge corresponds to S , and if P_b and S are measured, the "true" evaporation ETR can be computed by equation (1) as the difference between P_b and S . Table 1 reports a set of data regarding a few streams located in Central and Northern Italy; the table shows how the evapotranspiration computed by means of Turc's equation compare well with that computed using equation (1), and how the largest differences are probably similar to the approximation of the estimates of rainfalls and river discharges.

Another example confirming the validity of Turc's formula in Central – Northern Italy can be obtained indirectly from a recent paper by LO RUSSO *et al.* 2003. These authors, in an irrigated test site, located in the alluvial Po plain (NW Italy), estimated the infiltration during the year 1998 - 1999, by using the steady-state chloride profile method. The authors assumed no surface runoff as, at their site, this is negligible: in such a condition S of equation (1) corresponds to the infiltration. Table 2 reports the results obtained by LO RUSSO *et al.* 2003, and ETR_T computed by the present authors: the value of ETR_T is practically identical to that obtained by the highly reliable chloride profile method. Other examples confirming the validity of Turc's method in temperate areas can be easily found in literature (cf., by way of example, BONO, 1993; DE FELICE *et al.* 1993; STRZEPEK & YATES, 1997; KRÜGER *et al.* 2001; BONACCI, 1999).

River and time period	Area (km ²)	P_a (mm)	T_a (°C)	ETR_T (mm)	ETR (mm)
Reno at Pracchia 1956-1962 1965-1969 1970-1997	41.0	2040	10.2	586	590
Reno at Molino 1956-1961 1965-1969	89.2	2063	10.0	580	577
Limentra 1956-1960 1970-1973	64.6	1709	8.8	529	488
Nestore 1986-1992	106.3	700	13.4	528	524
Rigalto 1987-1992	17.0	737	13.6	545	556
Rio Acquina 1985-1988	10.7	974	11.9	568	543

Table 1 – Evapotranspiration computed by using Turc's method (ETR_T) and as difference between rainfall and river discharge (ETR), in the case of low permeability basins, with negligible groundwater flow (DRAGONI & VALIGI, 1994; VALIGI, 1995).

P_b (mm)	T (°C)	S_{Cl} (mm)	ETR_{Cl} (mm)	ETR_T (mm)
712	11.8	205	507	502

Table 2 – Comparison between the evapotranspiration estimated by the chloride profile method (ETR_{Cl}) and Turc's method (ETR_T): here the rainfall P_b has been increased by adding the irrigation water; T is the average temperature of the air, S_{Cl} is the infiltration computed according to the chloride profile method (LO RUSSO *et al.*, 2003).

In equation (1) the numerical values of the symbols are:

$P_b = 0.990$ m/year; $ETR_T = 0.582$ m/year; $S = 0.408$ m/year. Considering that the area of the basin (excluding the lake surface) is $A_b = 23.4$ Mm², S corresponds to a water volume $W_B \approx 10$ Mm³/year. The volume W_B so obtained was compared with the average volume of water actually received by the lake in the same period (W_L). Indeed if we apply the budget equation to the lake, making the inputs positive and the outputs negative, we obtain:

$$\Delta V = W_L + P_L \cdot A_L + EV \cdot A_L + Q_{ril} + PR \quad (2)$$

where:

- $\Delta V = -0.2$ Mm³/year, mean yearly change of water stored in the lake in the interval considered. This was calculated as the difference between the volume of the lake at the end and at the beginning of the period 1992–2002, using the level-volume curve of the lake.
- $P_L = 0.883$ m/year, average yearly rainfall on the lake surface, resulting from the regressions in Fig. 3.
- $A_L = 0.23$ Mm², average yearly area of the lake surface, based on the level-area curve of the lake.
- $EV = -1.054$ m/year, average yearly evaporation from the surface of the lake. EV was estimated using the formula proposed by DRAGONI & VALIGI (1994), adopting the temperatures measured at the dam (average yearly temperature = 13.6 °C).
- $Q_{ril} = -1.18$ Mm³/year, average yearly volumes released through the dam.
- $PR = -2.5$ Mm³/year, average yearly volume taken directly from the lake.

By replacing the letters with the numbers in equation (2), we obtain $W_L \approx 3.50$ Mm³/year, which is only 37% of the total yield W_B produced by the hydrographic basin: it has to be concluded that about 63% of W_B feeds the regional karst groundwater flow that has its main outflow through the Stifone

Springs and the alluvial plane aquifer of Terni (fig. 1). If we assume the water surplus of the Martani Mountains area (238 km²) to be the same as over the Firenzuola catchment (0.408 m/year) we obtain, under the present climatic conditions, a total yield of about 97 Mm³/year.

Nevertheless it has to be taken into account that the mean elevation of the Martana chain is higher than the mean elevation of the Firenzuola watershed and that the formations of Calcare Massiccio and Maiolica are generally more permeable than the Scaglia Bianca and Rossa. These last formations cover about 50% of the Firenzuola catchment. The higher permeability of the Calcare Massiccio and Maiolica is mainly due to karst phenomena (sinkholes etc.) which do not develop in the Scaglia Bianca and Scaglia Rossa. Karst structure can be observed in the Firenzuola catchment only on the north-western part and in C.Canepine karst plane (fig. 2). On the contrary Calcare Massiccio and Maiolica outcrop over a wide area of the Martani Mountains. In the light of these considerations it is likely that the percentage of water surplus which infiltrates in the Martani chain is higher than the 63% estimated for the Firenzuola catchment. Indeed, according to the literature, the average infiltration in fissured limestones (as in the case of Firenzuola basin) is between 50% and 85%, whereas for karst limestones is somewhere between 75% and 100% (BONI *et al.* 1986; CELICO, 1988; CIVITA, 2005). It can therefore be assumed that a high percentage of the 97 Mm³/year water surplus estimated for the Martani Mts. infiltrates, contributing to the discharge of the high salinity Stifone Springs (BONI, 2000; CHIODINI *et al.* 1999; MINISALE *et al.* 2002). The water salinization of these springs occurs when the groundwater reaches the anhydrides, below the Umbria-Marche Sequence (fig 6).

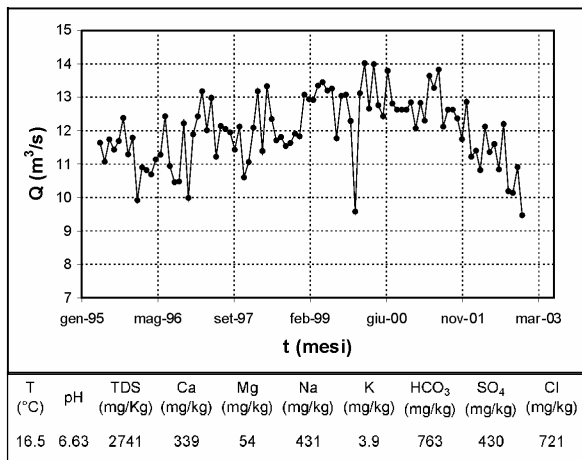


Fig. 6 – Discharge of Stifone Spring for 1995-2002 period (data collected by ENDESA Italia) and its chemical quality (modified after MINISALE *et al.* 2002).

Such a situation suggests that investigations aimed at siting some water wells exploiting the regional groundwater flow in the Martani mountains or below the alluvial Terni Valley, before salinization occurs, could satisfy the water demand in the area.

From the point of view of the water resources which will be available for the Firenzuola reservoir over the next decades, it is possible to have some indications about the decrease which should be expected, if we assume that for the next fifty years the decreasing trend of 3 mm/year detected in the 1921-2001 interval (fig. 5) will go on. Under this assumption, and assuming a cautionary constant temperature (the temperature trend in the region seems to be rather small and quite different from station to station) for the next fifty years we should expect an average rainfall of 0.915 m/year, corresponding to an average total yield S of about 0.350 m/year, which is about

12% less than the present total yield. If it is assumed that, as at present, only 37% of water surplus will feed the reservoir, we obtain an average of 3.1 Mm³/year of water for the dam against the present 3.5 Mm³/year: it may be interesting noticing that the last figure is about 25% lower than expected with reference to the climate data at the time of planning the reservoir.

Discussion and conclusions

In spite of the bad approximation of data and results, it is possible to make the following considerations:

1. The data 1992-2002 indicate that about 63% of the total yield produced by the Firenzuola catchment feeds the regional groundwater flow, which has its main outflow through the high salinity Stifone Springs. At the moment, due to the lack of hydrogeological data, it is not possible to define the volumes entering into the alluvial aquifers and the volumes entering the karstic system feeding the Stifone Springs.

2. In order to satisfy the water demand, the outlined situation suggests carrying out investigations aiming at siting some water wells to exploit the regional groundwater flow in the Martani mountains or below the alluvial Terni Valley, before salinization occurs.

3. The climatic data indicate, as for other places of central Italy, a decrease in rainfall of about 3 mm/year. If the present trend continues, in the next 50 years we should expect a total yield about 12% lower than the present one. Such a decrease, on an average basis, may seem small: however it should be pointed out that the general average decrease of the rainfall will raise the demand for irrigation, so that some reshape of the present land use and management plans should be considered.

4. It has to be noted that part of the water coming to the reservoir arrives as overflow of the higher part of the regional circuit (fig. 3): if the recharge is going to decrease, also the average piezometric level will decrease, so that the overflow of the regional groundwater circuit towards the reservoir will decrease more than the regional flow (CAMBI & DRAGONI, 2000). This implies that probably the yield towards the reservoir will be even smaller than the estimated 3.1 Mm³/year. It has also to be noted that the above computed total water yield decrease is the average decrease for the next 50 years: if the rain's quantity of the second part of the five decades is considered, it appears that, on the whole, the future situation is by far worse than the mere estimates of an average decrease of 12% suggests.

5. In the Italian Region, besides the decreasing average rainfall, the climate shows a tendency towards an increase of the intensity of rainfall (BRUNETTI *et al.* 2004): this implies an increase of flood frequency and intensity, which should be considered while defining the future reservoir's management plans (MILLY *et al.* 2005).

6. The considerations and the suggestions here presented are based on a set of data which is not large and reliable as necessary for the importance of the issues discussed. So it is imperative that a reliable data net for the measurement of evaporation, piezometric heads, rain, temperature, river and spring discharges is established.

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Le système karstique hydrothermal constantinois (Algérie nord orientale) : structure et fonctionnement.

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Résumé

Le karst hydrothermal constantinois se développe dans les carbonates de plate-forme, structurés en horsts et grabens. Il est encadré, au nord par l'accident majeur E-W de M'Cid-Aicha-Debar, et au sud par l'accident majeur SW-NE Meimel-Oum Settas. Les affleurements de l'aquifère comprennent quelques horsts émergeant essentiellement des terrains miocènes, considérés jusque-là comme une couverture imperméable. La participation de la couverture miocène à la recharge est plus ou moins précisée en contexte hydrodynamique à partir de l'analyse des courbes des débits classés et de l'analyse des courbes de récession d'Ain Zouaoui, principale sortie du système à Hamma Bouziane (800 à 900 l/s avant captage par forages). Parmi les horsts, seul le massif de Constantine possède quelques exutoires au Fort Sidi M'Cid (28 à 29°C) ; la décharge de l'aquifère s'effectue à travers les terrains de couverture et des travertins produits par les sources, à proximité nord et ouest du rocher de Constantine, précisément entre Hamma Bouziane et Salah Bey, où on observe un grand nombre de sources thermales à température variable (20 à 36°C). Ceci s'explique par un drainage des eaux du sud vers le nord, le long du fossé d'effondrement Ain Smara-Constantine. Le faible thermalisme des eaux en contraste avec la profondeur notable de l'aquifère, indique que la zone saturée du système est le siège d'une circulation d'eaux thermales profondes mêlées aux eaux de recharge. Cette situation est corroborée par le contenu isotopique en carbone-13 ainsi qu'en oxygène-18 des bicarbonates dissous et par les données des équilibres calco-carboniques de quelques échantillons d'eau. L'ensemble des paramètres hydrogéologiques a permis de proposer un modèle conceptuel du fonctionnement du système.

Abstract

Jurassic-cretaceous age carbonates and clastics around Constantine, NE. Algeria have been block faulted into separate massifs and mantled by Mio-Pliocene cover. Upstanding carbonate horsts have developed karst draining to adjacent infilled valleys. Springs emerge from faults and clastics and have formed travertine terraces. They show elevated temperatures (20-36°C), supersaturation with respect to dolomite and calcite and high pCO₂ and δ¹⁸O (+0.7 ‰ SMOW). Hydrograph analysis of the d'Ain Zouaoui, the primary outlet of the Hamma Bouziane aquifer shows very sustained recession, attributed to retarded recharge through the karst and cover rocks. Overall drainage is from the south, with mixing of meteoric and hydrothermal waters.

1. Introduction

La série carbonatée jurassico-crétacée, subsidente (500 à 1000 m), structurée en grabens, supporte des unités telliennes marneuses à marno-calcaires, des unités de flyschs gréseux numidiens, et les dépôts tertiaires post-nappes, d'épaisseur (300 m) et d'extension notable, de type laguno-continentale, (VILA, 1980). Les massifs carbonatés, Salah, Kelal, Constantine, Chettabah, Feltene, Sekoum, Ouled Salem, et Oum Settas, affleurent en horsts émergeant de la couverture néogène à mio-pliocène principalement continentale (COIFFAIT, 1992 ; ARIS, 1994).

De nombreuses sources thermales émergent à proximité nord et ouest de Constantine, pour la plupart de la couverture miocène. Parmi les horsts, le rocher de Constantine est le seul massif d'où émergent des sources (29°C). C'est entre Hamma Bouziane et Salah Bey que se trouve le plus grand nombre d'émergences au travers de la couverture néogène, parfois avec un très fort débit comme à Ain Zouaoui (900 l/s). La température de l'eau des sources entre 20 à 36°C varie selon les conditions d'émergences ; les sources à basse température émergent des terrasses de travertins de Hamma Bouziane, occupant une superficie notable, témoins d'anciennes activités de sources thermales. La structuration des carbonates, l'hypothermalisme, et les conditions d'émergence ont conduit à analyser le fonctionnement du système, à travers les caractéristiques hydrodynamiques et hydrogéochimiques.

2. Matériels et méthodes

Ain Zouaoui est la seule émergence équipée d'une station de jaugeage depuis 1995 (ANRH). Son débit actuel s'est réduit à 200 l/s du fait du grand nombre de forages, alors qu'il était de 800 à 900 l/s auparavant. Les caractéristiques hydrodynamiques du système ont été établies par analyse des débits classés et des courbes de récession (MANGIN, 1975), sur des chroniques de débits moyens journaliers du cycle 1995-1996.

La méthode des débits classés consiste à établir un histogramme de fréquences de classes de débits. Les débits des classes cumulées s'alignent suivant une droite par un changement de variable approprié. Un fonctionnement par trop-pleins présente des ruptures de pente de la droite.

Les courbes de récession consistent en un hydrogramme de la fonction sortie après une crue se prolongeant par une période d'étiage. L'analyse porte sur la décrue et le tarissement de la courbe : $Q(t) = Q_{R0}e^{-\alpha t} + q_0 \frac{1-\eta t}{1+\varepsilon t}$

Le retard à l'infiltration $Y = (1 - \eta t) / (1 - \varepsilon t)$, avec $0 \leq t \leq t_i$, t_i étant le moment où l'infiltration devient négligeable et où le tarissement débute, informe sur l'organisation des vides de la zone sous-saturée du système.

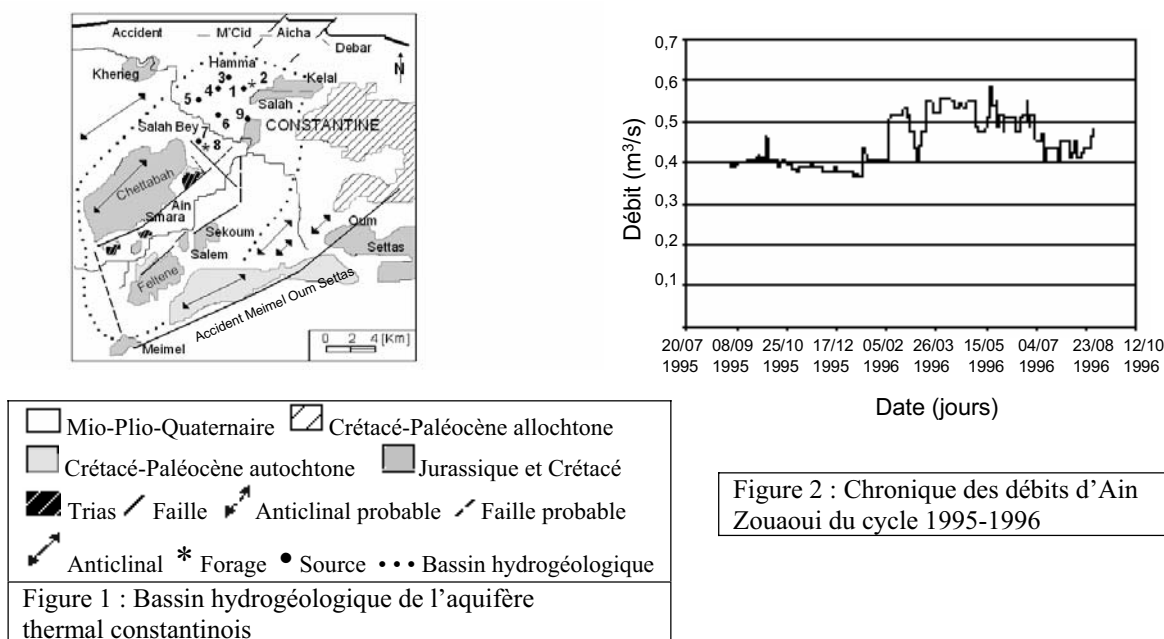
Les réserves du karst noyé sont évaluées par le volume dynamique : $V = \int_0^x Q_{R0} \cdot e^{-\alpha t} = CQ_{R0} / \alpha$

Les caractéristiques chimiques ont été fournies par quelques prélèvements d'eau de sources et de forages aux sites Hamma Bouziane, Constantine et Salah Bey. L'analyse isotopique du carbone-13 et de l'oxygène-18, contenus dans les bicarbonates des échantillons d'eau de sources Ain Zouaoui, Ain Salah Bey et Sidi M'Cid, est effectuée au spectrophotomètre de masse. Ce procédé permettra de corroborer l'origine profonde du CO_2 gazeux, en relation avec le contexte structural des carbonates, ou d'apports hydrothermaux (BAKALOWICZ, 1994).

Dans le système CO_2 - H_2O -carbonate, les indices de saturation de la calcite, SIc, et de la dolomite, Sid, et la pCO_2 équilibrante, sont calculées à partir des données chimiques. Ils informent sur les conditions d'écoulement dans l'aquifère, de même que l'indice de saturation du gypse, SIg (BAKALOWICZ, 1994, 1996). Ces paramètres des équilibres ont été établis pour quelques prélèvements d'eau de sources et de forages aux sites Hamma Bouziane, Constantine, Salah Bey. Ca^{2+} , Mg^{2+} , HCO_3^- caractérisent les eaux karstiques et Na^+ , K^+ , Cl^- , SO_4^{2-} sont fournis par les eaux météoriques et les évaporites. Ces données permettent d'établir le degré de karstification de la zone noyée et les particularités du fonctionnement du système.

3. Résultats et discussion

Le bassin hydrogéologique de l'aquifère thermal constantinois est défini structuralement au nord par l'accident majeur M'Cid-Aicha-Debar et au sud par l'accident majeur Meimel-Oum Settas (Fig. 1). Les caractéristiques hydrodynamiques sont étudiées grâce à l'hydrogramme du cycle 1995-1996 de Ain Zouaoui (Fig. 2), cycle au cours duquel les prélèvements des forages étaient encore négligeables.



D'après la linéarité de la distribution des débits classés (Fig. 3), la source d'Ain Zouaoui n'est pas affectée par le fonctionnement de trop pleins. L'analyse de la récession montre que l'infiltration lente ou retardée est dominante. Ainsi, la recharge de l'aquifère s'effectuerait par les horsts, très fracturés et karstifiés, mais aussi au travers des sédiments mio-pliocènes.

L'hétérogénéité et l'épaisseur notable, autour de 300 m, des sédiments miocènes (COIFFAIT, 1992), confèrent à cette formation une forte perméabilité des niveaux sableux et conglomératiques et une forte capacité de rétention des eaux autour des niveaux argileux. Le tarissement observé à la suite de la récession du 25 Juin 1996 (Fig. 4) est marqué par les fluctuations qui laissent supposer que l'infiltration est complexe. De même, la valeur 0.54 du retard à l'infiltration conduit à faire intervenir un apport d'eau tardif, issu de la formation néogène, et/ou à un écoulement lent dans la zone non saturée.

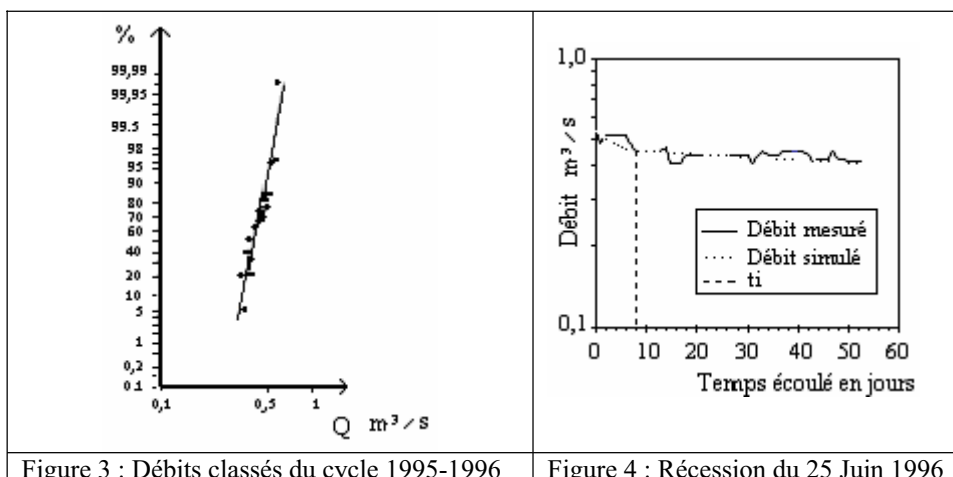


Figure 3 : Débits classés du cycle 1995-1996

Figure 4 : Récession du 25 Juin 1996

Les analyses isotopiques du carbone-13 et de l'oxygène-18 des bicarbonates de l'eau des sources, Salah Bey, Ain Zouaoui et Sidi M'Cid, ont fourni pour l'ensemble des eaux des teneurs très homogènes, en oxygène-18 [$\delta^{18}\text{O}(\text{H}_2\text{O}) = +0.7\text{‰}$ SMOW] comme en carbone-13 [$\delta^{13}\text{C}(\text{CO}_2) = -5.4\text{‰}$ PDB]. La teneur en $^{18}\text{O}(\text{H}_2\text{O})$, très différente de celle des eaux météoriques du Constantinois [environ -4.0‰ SMOW] et la teneur en $^{13}\text{C}(\text{CO}_2)$ très proche du CO_2 d'origine magmatique, signifient que le CO_2 a une origine profonde et que l'eau est soumise à des températures relativement élevées en profondeur.

L'analyse chimique des prélèvements d'eau de sources et de forages (Fig. 1 et Tabl. 1) a révélé un faciès bicarbonaté calcique pour l'ensemble des eaux et de fortes variations de concentrations en $\text{Na}^+ + \text{K}^+$ et en Mg^{2+} .

Tableau 1 : Chimisme des eaux des sources et des forages de la région de Constantine.

Sources et forages	T °C	pH	EC μs/cm	Ca^{2+} mg/l	Mg^{2+} mg/l	$\text{Na}^+ + \text{K}^+$ mg/l	HCO_3^- mg/l	Cl^- mg/l	SO_4^{2-} mg/l
1 Ain Zouaoui	32	7.60	1230	119	26	128	425	135	140
2 Forage	35	7.60	1220	115	32	175	512	131	189
3 Ain Toundji	21	7.50	1360	113	67	171	498	163	200
4 Ain Murat	25	7.70	1120	111	67	76	503	142	106
5 Ain Bou Sba	21	7.80	1370	124	49	124	484	170	143
6 Ain Touta	20	7.90	1720	147	50	202	620	224	182
7 Ain Salah Bey	25	7.80	1080	108	37	189	555	159	150
8 Forage	26	8.00	1130	115	32	199	512	163	195
9 Sidi M'Cid	28	7.50	1200	123	32	138	493	142	129

Ces variations peuvent être expliquées par la dissolution des évaporites des dépôts de couverture et/ou des argiles triasiques jalonnant les accidents majeurs. Dans ce cas l'écoulement général dans l'aquifère carbonaté se ferait depuis le sud vers Hamma Bouziane, le long du fossé Ain Smara – Constantine (Fig. 1). Ceci est très envisageable du fait de la structuration de l'ensemble carbonaté.

Les variables du système $\text{CO}_2\text{-H}_2\text{O-CaCO}_3\text{-MgCO}_3$ sont des informateurs des écoulements dans l'aquifère et de l'évolution de la karstification. Les valeurs atteintes par SIc et Sid, nettement supérieures à 0.5 (Tabl. 2.), caractérisent un dégazage très marqué du CO_2 avant émergence et un début de précipitation de la calcite.

Tableau 2: Caractéristiques du système $\text{CO}_2\text{-H}_2\text{O-CaCO}_3\text{-MgCO}_3$ des eaux de l'aquifère thermal de Constantine.

Sources et forages	SIc	Sid	SIg	pCO_2 10^{-3} bar	$\text{Mg}^{2+}/\text{Ca}^{2+}$
1 Ain Zouaoui	0.90	1.37	-1.34	9.70	0.36
2 Forage	1.27	2.30	-1.25	4.20	0.45
3 Ain Toundji	0.66	1.51	-1.29	1.20	0.98
4 Ain Murat	0.95	2.03	-1.53	8.20	1.00
5 Ain Bou Sba	1.01	2.04	-1.36	5.90	0.66
6 Ain Touta	1.24	2.45	-1.22	5.60	0.56
7 Ain Salah Bey	1.07	2.02	-1.38	7.10	0.57
8 Forage	0.83	1.36	-1.38	1.30	0.45
9 Sidi M'Cid	0.91	1.59	-1.47	1.00	0.42

Par conséquent, les valeurs de $p\text{CO}_2$, existant effectivement dans l'aquifère sont notablement supérieures aux valeurs calculées. Les $p\text{CO}_2$ calculées à l'équilibre sont comprises entre 5 et $10 \cdot 10^{-2}$ bar, ce qui correspond à un apport considérable de CO_2 d'origine profonde. Les rapports molaires $\text{Mg}^{2+}/\text{Ca}^{2+}$ supérieurs à 0.5, parfois proches de 1, témoignent de l'enrichissement relatif en Mg^{2+} par précipitation du CaCO_3 . La nette sous-saturation en gypse, $-1.47 \leq \text{SIg} \leq -1.25$ correspond probablement à une dilution d'eaux ayant traversé les formations évaporotiques (Miocène ou Trias) avec les eaux dépourvues de sulfates de l'aquifère carbonaté.

L'origine profonde du CO_2 , le thermalisme faiblement marqué (32 à 36°C), et les minéralisations inférieures à 2000 mg/l , caractériseraient un mélange d'eaux.

Les données fournies par l'hydrodynamique et la géochimie de l'aquifère ont conduit à un modèle conceptuel (Fig. 5) du fonctionnement du système, faisant intervenir la couverture mio-plio-quaternaire dans la recharge et le mélange des eaux dans la zone noyée où circulent les eaux thermales.

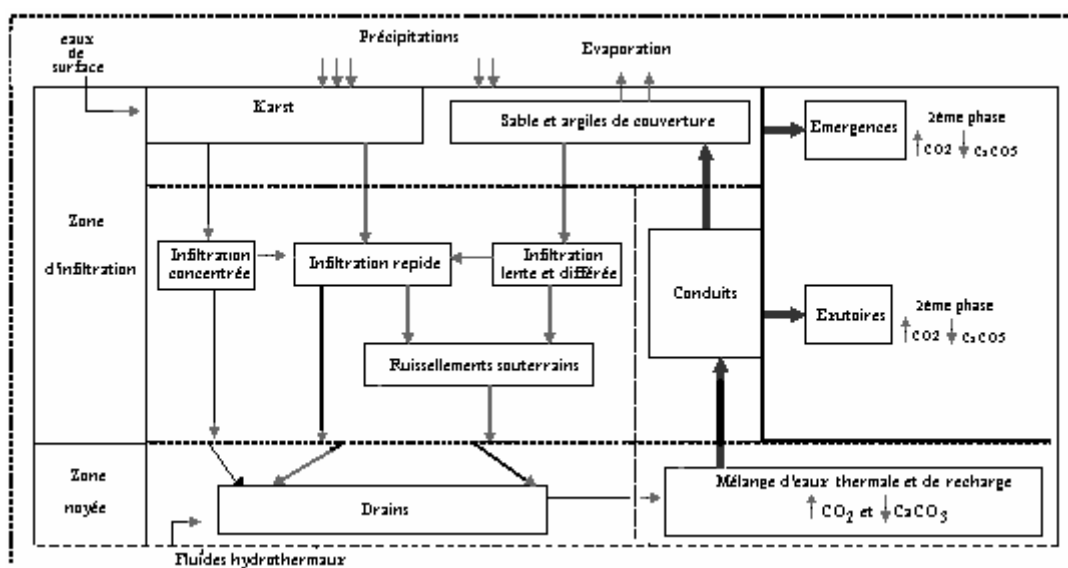


Figure 5 : Modèle conceptuel du système karstique hydrothermal Constantinois.

4. Conclusion

Les conditions d'émergence des sources, les données géologiques récentes, et les résultats fournis par l'hydrodynamique et le chimisme des eaux, ont permis de proposer une nouvelle conception du fonctionnement de l'aquifère thermal constantinois. Les dépôts tertiaires ne sont plus à considérer comme une formation rendant localement captif l'aquifère, mais comme une zone d'infiltration et de recharge du système. Les fortes pressions en CO_2 , rencontrées dans les forages productifs, témoignent particulièrement de la sensibilité de l'aquifère à la surexploitation.

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Nouveaux traçages dans le karst de la Fontaine de Vaucluse

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Abstract

The first two injections in the shaft of Fontaine de Vaucluse characterized the connection between sink holes in the shaft and permanent springs. In February 2004, the tracer injected into sink holes of stream Nesque spent 11.75 days to cover a distance of 21 km to the Fontaine de Vaucluse. Maximum velocity was 74.7 m/h. The following injection took place at Notre-Dame-de-Lure in March 2004. For a distance of 56.6 km, maximum velocity was 74.6 m/h. In March 2005, an injection in sink hole of Saint-Donat showed that the karst located at a few kilometers of the river Durance belonged to the groundwater basin of Fontaine de Vaucluse. The time to first tracer arrival was 70 days after the injection. Maximum velocity was 40 m/h. Modal velocity was 33 m/h. This tracer test of 66.7 km is the longest of Europe. In August 2005, a tracer test showed the connection between the shaft of Fontaine de Vaucluse and the municipal well.

Résumé

Les deux premières injections dans le gouffre de Fontaine de Vaucluse ont permis de caractériser la liaison entre les pertes du gouffre et les résurgences pérennes. En février 2004, le traceur injecté dans les pertes de la Nesque a mis 11,75 jours pour parcourir les 21 km jusqu'à la Fontaine de Vaucluse, soit une vitesse maximale de 74,7 m/h. L'injection suivante a eu lieu à Notre-Dame-de-Lure en mars 2004. Pour une distance de 56,6 km, la vitesse maximale du traceur a été de 74,6 m/h. En mars 2005, une injection dans la perte de la Chapelle Saint-Donat a montré que le karst situé à quelques kilomètres de la Durance appartenait au bassin d'alimentation de la Fontaine de Vaucluse. La première détection du traceur a eu lieu 70 jours après l'injection, soit une vitesse maximale de 40 m/h. La vitesse modale a été de 33 m/h. Ce traçage sur une distance de 66,7 km est le plus long d'Europe. En août 2005, un traçage a caractérisé la relation entre le gouffre de Fontaine de Vaucluse et le puits de captage communal.

1. Introduction

La Fontaine de Vaucluse est la plus importante source karstique de France. Une synthèse sur la géologie et l'hydrogéologie du système karstique de la Fontaine de Vaucluse a été réalisée par PUIG (1990). Le bassin d'alimentation proposé était alors de 1115 km².

D'après la banque de données HYDRO, le débit moyen sur les 40 dernières années est de 18,5 m³/s (Sorgomètre). Les débits sont mesurés par 2 stations, une visuelle datant de 1869 à relevé journalier manuel, le « Sorgomètre », une récente de 2004 avec un enregistreur automatique sur un vrai seuil, le « Moulin ». En basses eaux, le « Moulin » donne des débits 40 à 50 % plus élevés que le « Sorgomètre ».

La méthodologie utilisée pour les traçages est décrite par DUCLUZAU (2001). Le traceur employé est l'ion iodure, qui est mesuré en continu et in situ par des capteurs chimiques. Le seuil de détection d'une variation de la concentration en iodure est inférieur à 0,01 µg/l. La concentration en iode total a été mesurée par 23 analyses par ICP/MS. La moyenne est 2,5 µg/l.

Les deux premières injections dans le gouffre de Fontaine de Vaucluse ont permis de caractériser la liaison entre les pertes du gouffre et les résurgences pérennes, dont la source du Pagodon. Cette source a donc été le point de surveillance des 3 traçages suivants à partir du plateau de Vaucluse. Pour ces traçages, aucun autre point d'eau n'a été surveillé.

Le dernier traçage a caractérisé la relation entre le gouffre de Fontaine de Vaucluse et le puits de captage communal dans la nappe alluviale de la rivière Sorgue.

2. Traçage des pertes de la Nesque

Les caractéristiques et les résultats des traçages sont dans les tableaux Tabl. 1 et Tabl. 2.

Quatre dispositifs indépendants (4 capteurs + 4 enregistreurs) étaient installés sur la source du Pagodon, exutoire le plus bas du système karstique de Vaucluse (coordonnées L2E : x = 824641 m, y = 1883936 m, z = 78,5 m). Le pas de temps était d'une mesure toutes les 20 minutes.

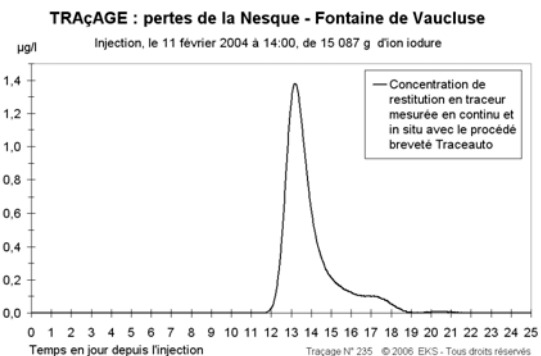


Fig. 1 : Courbe de restitution du traçage des pertes de la Nesque

A partir du 23 février 2004, un pic de restitution très net en traceur a été détecté sur les 4 dispositifs indépendants.

La vitesse moyenne du traceur, en tenant compte des variations de débit (Moulin), est de 63,08 m/h. Les vitesses moyennes des 3 pics visibles sont respectivement 66,0 m/h, 51,2 m/h et 42,8 m/h. Les dispersions sont les mêmes pour les 3 pics : 14 m. Le taux de restitution du dernier pic est faible : 0,07 %.

Le volume écoulé à la Fontaine entre l'injection et le temps moyen est de 32 10⁶ m³ à partir des débits du « Moulin » et 29 10⁶ m³ à partir des débits du « Sorgomètre ». En divisant ce volume par la distance entre les points d'injection et de restitution, on calcule une surface tracée (1600 à 1400 m²) et un diamètre tracé (environ 40 m). Ces définitions sont neutres sans aucune interprétation.

Interprétations :

La courbe de restitution est composée de 3 pics, entre 12 et 15 jours, entre 16 et 19 jours et entre 20 et 21 jours. Ces pics pourraient correspondre à des cheminements différents car les pertes de la Nesque sont graduelles. Le dernier pic bien séparé pourrait aussi avoir pour origine le fractionnement du nuage de traceur par le drain collecteur principal du système karstique. En effet, de fortes précipitations, 91 mm à Sault du 20 au 22 février, ont eu lieu sur le plateau pendant que le traceur avançait dans le karst. Le drain karstique principal provenant de la Montagne de Lure et le drain secondaire provenant des pertes de la Nesque pourraient se rejoindre à 12 km en amont de la Fontaine de Vaucluse.

3. Traçage de la Montagne de Lure

Un seul dispositif de surveillance du traceur (capteur + enregistreur) a fonctionné pendant toute la restitution. Le pas de temps était d'une mesure toutes les 20 minutes.

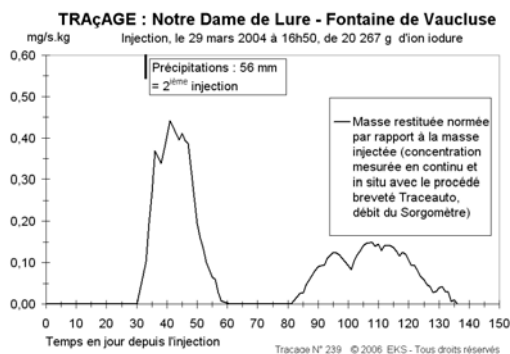


Fig. 2 : Masse restituée du traçage de la montagne de Lure

La courbe de restitution du traçage montre deux restitutions séparées avec des formes de restitution similaires : 4 pics superposés de formes similaires. La dispersion est plus importante pour la deuxième restitution. Le débit moyen du Sorgomètre pendant la première restitution (19 m³/s) est plus important que pendant la deuxième (9 m³/s). Le taux de restitution du premier pic (11,2 % pour 2,3 kg) est donc plus important que celui du deuxième (8 % pour 1,6 kg). La concentration maximale de restitution est faible : 0,09 µg/l.

Interprétations :

L'interprétation du deuxième pic est une restitution tardive du traceur stocké près du lieu d'injection (sol et citerne) remobilisé par les précipitations du 29 et 30 avril (31 jours après l'injection) : 56 mm à St-Etienne-les-Organes.

La première détection du deuxième pic a lieu 53 jours après les précipitations du 29 avril. La vitesse moyenne du deuxième pic, avec les débits du Sorgomètre, est de 30,9 m/h. La vitesse est plus faible car le débit était presque deux fois moins important. Le volume écoulé à la Fontaine entre l'injection différée et le temps moyen est de 89 10⁶ m³, d'où une surface tracée de 1570 m², et un diamètre tracé de 44 m. Ce diamètre est du même ordre de grandeur que le précédent bien que les conditions « d'injection » soient totalement différentes.

4. Traçage de Saint Donat

Cinq dispositifs indépendants (5 capteurs + 5 enregistreurs) étaient installés à la Fontaine de Vaucluse. Les 5 capteurs ont enregistré le même pic de restitution en ion iodure centré fin juin 2005.

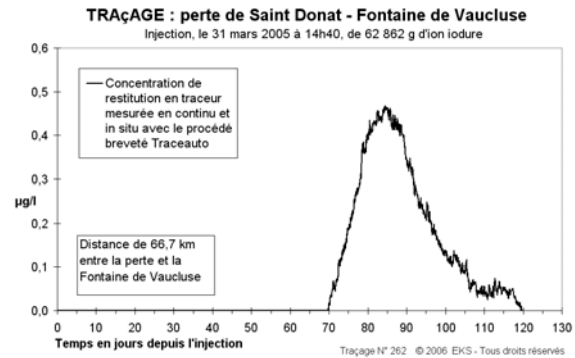


Fig. 3 : Courbe de restitution du traçage de Saint Donat

Interprétations :

Pour les 3 traçages, la surface tracée est toujours de l'ordre de 1500 m² et le diamètre tracé de l'ordre de 40 m. Le diamètre tracé de 40 m semble donc une caractéristique intrinsèque du système karstique de Vaucluse.

Le taux de restitution calculé à partir des débits journaliers du Sorgomètre est 10 %, avec une masse restituée de 6,2 kg. La station du Moulin était en panne pendant la restitution, elle aurait donné un taux de restitution plus élevé. Ce taux de restitution faible vient des conditions défavorables de l'injection (sécheresse) et des caractéristiques de l'aquifère pendant le traçage. La forte crue après l'injection a pu pousser du traceur dans des systèmes annexes aux drains principaux. On ne peut donc pas déduire du faible taux de restitution qu'il existe d'autres exutoires du karst.

Le traçage de Saint Donat permet d'étendre le bassin d'alimentation de la Fontaine de Vaucluse vers l'est, par rapport à celui proposé par PUIG (1990). Les calcaires Barrémo-Bédoulien affleurant en bordure de la Durance, des pertes de la Durance alimentent probablement la Fontaine de Vaucluse. A l'affleurement, ces calcaires ne montrent aucune trace de karstification et toutes les fissures visibles sont colmatées par de l'argile. On peut donc estimer les pertes de la Durance à 50 l/s (perméabilité moyenne 10⁻⁶ m/s et surface 500 * 100 m).

Plus d'une centaine de forages ont été effectués par EDF dans les calcaires sous ou à proximité du lit de la Durance. Deux forages (AC 5 et AC 8) ne montrent aucune trace d'eau alors que leur fond est 20 m sous le niveau de la Durance. Le niveau piézométrique à Saint Donat est à 300 m d'altitude. Avec un gradient de 0,3 %, le niveau de l'eau du karst sous la Durance serait à 335 m, soit 110 m sous le cours d'eau actuel. Le bassin d'alimentation de la Fontaine de Vaucluse pourrait donc s'étendre à l'est de la Durance, c'est-à-dire dans les Alpes, en suivant la structure anticlinale de Beaudouze.

5. Forage et sources temporaires de Chavoul

Le forage de recherche d'eau de Chavoul est situé dans les gorges de Chavoul à 1,5 km de Montbrun-les-Bains (coordonnées L2E : x = 847871 m, y = 1911423 m, z = 576 m). D'une profondeur de 225 m, il est en totalité dans les calcaires Barrémo-Bédoulien. Il n'a recoupé que quelques fissures non karstifiées donnant un débit non exploitable de 1 m³/h. Le niveau statique a été mesuré après la foration et par nous-mêmes en 2005. De fin janvier à fin février 1999, le niveau a baissé de 515 m à 491 m d'altitude. Il était de 479 m le 26/08/05. En très hautes eaux, le forage est artésien et jaillissant avec un faible débit, ARNOUX (2005)

comm. orale, président du Syndicat des Eaux maître d'ouvrage du forage. Les niveaux statiques sont bien corrélés avec les débits de la Fontaine de Vaucluse. Pour obtenir un exutoire à l'étiage dans la rivière Toulourenc, située à peine à 900 m du forage, il faut descendre à 10 km en aval, au lieu dit l'Isle sur la commune de Brantes. Cela est peu probable d'après la géologie. Le forage de Chavoul recouperait donc un aquifère fissuré qui alimente le système karstique de Fontaine de Vaucluse.

Les sources temporaires de Chavoul sont situées au-dessus du forage. La plus importante s'appelle l'Eissuyan Blanc, ARNOUX (2005) comm. orale (coordonnées L2E : x = 848 505 m, y = 1910 130 m, z = 620 m). Les crues, qui pourraient dépasser plusieurs m³/s, sont rares. L'eau sort d'éboulis, d'où la formation de mousse blanche de crue.

En 1996, nous avons repéré une troisième source temporaire possible du karst de Vaucluse (combe de Font Jouvale à St-Saturnin-lès-Apt). Les traces sur le terrain semblaient indiquer que l'eau avait coulé lors de la crue de 1994.

Le système karstique a donc 3 ou 4 exutoires principaux : un permanent à l'ouest (la Fontaine de Vaucluse), deux temporaires à l'est (Saint Donat) et au nord (Chavoul), un temporaire supposé au sud.

6. Synthèse des anciens traçages

L'évaluation de la fiabilité des traçages est basée uniquement sur des éléments techniques liés aux traçages. Aucun argument de cohérence hydrogéologique ne doit être apporté pour valider un traçage.

L'eau de la Fontaine de Vaucluse est d'un beau vert fluorescent à l'étiage, en moyennes eaux et en hautes eaux. Il n'y a qu'après de très fortes crues que la couleur tourne au brun. Dans les eaux karstiques, ce vert est probablement dû à des matières organiques fluorescentes vertes. L'emploi de fluorescéine, colorant vert à fluorescence verte, est donc réservé à des opérateurs expérimentés, qui ont les moyens techniques de distinguer la fluorescence naturelle de celle de la fluorescéine.

Les résultats scientifiques de la première coloration dite positive de 1963 sont donnés par LALLEMAND & PALOC (1964) du BRGM. Le premier fluocapteur « positif », immergé le 6 juillet, a été sorti de l'eau le 10 août 1963, soit 35 jours après l'injection. Ceux du 9 août sont négatifs. La vitesse maximale donnée est donc 26,2 m/h. Or en 1963, le débit moyen de la Fontaine avant le début de la restitution a été de 26 m³/s (Sorgomètre). Le débit moyen avant la restitution du traçage de 2004 a été de 20,3 m³/s (Sorgomètre) pour une vitesse 74,7 m/h. Le débit à l'entrée des Gorges était en 1963 de 400 l/s à 1000 l/s selon les auteurs. En 2004, il était de 120 l/s. Le résultat positif de 1963 est donc totalement impossible. En réalité la restitution aurait dû débuter le 16 juillet 1963, et être totalement terminée début août. Or tous les fluocapteurs journaliers jusqu'au 9 août étaient négatifs.

Cette technique, fluocapteur au charbon actif + fluorimètre, a été testée en 1989 par COUTURAUD & PUIG (1992). En l'absence prouvée de restitution de fluorescéine, ils montrent que la fluorescence verte augmente avec la durée d'immersion du fluocapteur dans la Fontaine de Vaucluse. Les « courbes de restitution aux fluocapteurs » des premières colorations sont dues à un phénomène naturel et non à des restitutions de traceur. COUTURAUD & PUIG (1992) ont également démontré l'absence de relation entre fluocapteurs et prélèvements d'eau. La relation est même inversée. Ils montrent également que les concentrations

inférieures à 1 µg/l pour les fluocapteurs ne peuvent être significatives. Or lors des 5 colorations, les concentrations maximales en fluorescéine seraient restées inférieures à 1 µg/l (estimation à partir des conditions des colorations par rapport aux traçages valides). L'ensemble des résultats positifs annoncés des 5 colorations est donc à rejeter, car la technique de détection employée n'était pas assez sensible.

Les multitraçages de 1974 et 1989 posent des problèmes d'interprétation, à cause de l'emploi simultané de la fluorescéine et d'une rhodamine. IDSTEIN & EWERS (2002) ont montré que les rhodamines étaient peu stables dans le milieu souterrain et que des produits de décomposition (DAR) très stables avaient la même fluorescence que la fluorescéine. Les deux courbes de restitution d'équivalent fluorescéine de 1974 et 1989 pourraient être dues soit à la fluorescéine, soit au produit de dégradation de la rhodamine, soit la somme des deux comme IDSTEIN & EWERS (2002) l'ont souvent montré en réinterprétant d'anciens multitraçages.

En 1974, il n'y a eu aucune détection de rhodamine B. La liaison en 6 jours entre l'Aven du Château et la Fontaine de Vaucluse est valide, mais le fort taux de restitution pourrait être dû à la somme de la fluorescéine et du DAR B. La courbe de restitution de 1989 pourrait aussi être due à l'injection de sulforhodamine dans les pertes de Méthamis. La restitution tardive de 2 à 6 mois après l'injection rendent possible une dégradation totale de la sulforhodamine.

7. Conclusion

L'interrogation posée par le système karstique de Vaucluse est la multitude de failles d'orientation nord-sud ± 30°, alors que la Fontaine est à l'extrémité ouest de l'aquifère et que les circulations principales sont donc de l'est vers l'ouest. Ces failles subméridiennes avaient pendant un siècle limité le bassin d'alimentation vers l'est aux « champs de fracture de Banon ». PUIG (1990) avait émis l'hypothèse avec des arguments géologiques et un bilan d'eau précis que la Montagne de Lure, à l'est des failles de Banon, appartenait au bassin d'alimentation. Les traçages de 2004 et 2005 ont prouvé l'extension du bassin d'alimentation jusqu'à la Durance.

En tant que bureau d'études, nous avons travaillé sur de nombreux forages situés dans les karsts profonds d'origine messinienne. Généralement les forages situés sur les axes des structures anticlinales est-ouest sont exploitables, et ils rencontrent des drains karstiques profonds situés sous le niveau actuel de la mer. GUERIN (1973) a montré en Ardèche une phase de karstification d'axe préférentiellement est-ouest lors de la phase alpine de la fin du miocène (pontien).

Dans le karst de Vaucluse, une seule rivière souterraine est connue, celle du Trou Souffleur. Un collecteur de 100 l/s à l'étiage est suivi sur 1 km dans une galerie horizontale parfaitement est-ouest. De petits décrochements nord-sud à fort dénivelé montrent que le drain principal est au sud du Trou Souffleur. Nous supposons donc que le collecteur principal noyé passe sous l'axe de la structure anticlinale des Monts de Vaucluse.

Les forages EDF au niveau du barrage de Baumas sur la Durance ont montré une épaisseur maximum d'alluvions de 20 m (AUB 14) au-dessus des calcaires Barrémo-Bédoulien. Les essais de perméabilité et les coupes indiquent une probable karstification de ces calcaires. Des pertes importantes de la Durance ont pu avoir lieu au Messinien dans le défilé de Baumas. Le drain principal du karst géant

de Vaucluse pourrait être une paléo Durance creusée au Messinien.

Le drain est-ouest du système karstique de Vaucluse s'est donc formé à la fin du miocène pendant la période messinienne (et probablement la phase alpine pontienne). Le drain principal provenant de la Montagne de Lure passe sous l'anticlinal des Monts du Vaucluse, et non dans les structures synclinales d'Apt ou de Saint Christol.

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En conclusion, la Fig. 4 donne une carte de synthèse montrant le bassin d'alimentation avec le principal drain alimentant la Fontaine de Vaucluse, les 4 relations prouvées par traçage et les 2 sources temporaires du système. Avec les nouvelles données de cet article, le bassin d'alimentation de la Fontaine de Vaucluse atteint 1210 km².

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Nom du point d'injection	Commune	X (m)	Y (m)	Z (m)	Date d'injection	Durée injection (min)	Débit à l'injection (l/s)	Quantité injectée (g)	Distance à la Fontaine de Vaucluse (km)
Pertes de la Nesque	Monieux	840225	1898113	540	11/02/2004 14:00	10	50	15087	21,07
Notre-Dame-de-Lure	St-Etienne-les-Orgues	876961	1905575	1245	29/03/2004 16:50	45	0,05	20267	56,63
Perte de Saint Donat	Châteauneuf-Val-St-Donat	888901	1901891	451	31/03/2005 14:40	70	0,2	62862	66,72

Tabl. 1 : Caractéristiques des traçages

Nom	Temps début	Vitesse maximale (m/h)	Vitesse modale (m/h)	Vitesse moyenne (m/h)	Vitesse minimale (m/h)	Taux de restitution	Nombre Peclet	Dispersion (m)	Volume écoulé (m ³)	Diamètre tracé (m)
Pertes de la Nesque	11 j 18 h	74,7	66,7	63,08	40,5	63 % (Moulin)	1500	14	32 10 ⁶	40
Notre-Dame-de-Lure	31 j 14 h	74,6	52,3			19 % (Sorg.)			62 10 ⁶ (1 ^{er} pic)	37 (1 ^{er} pic)
Perte de Saint Donat	69 j 20 h	39,8	32,9	31,9	23,2	10 % (Sorg.)	225	261	100 10 ⁶	43

Tabl. 2 : Résultats des traçages

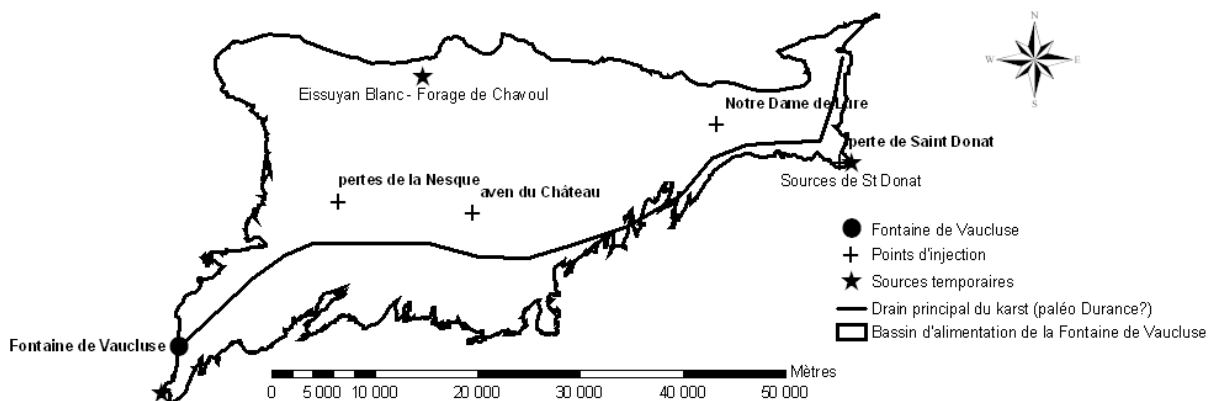


Fig. 4 : Carte du bassin d'alimentation de la Fontaine de Vaucluse

Classification et propriétés des traçages en fonction du temps de première arrivée du traceur

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Abstract

The time to first tracer arrival is an intrinsic and reproducible characteristic of a karstic system in the hydrological conditions of the tracer test. This time practically does not depend on the methodology of tracer injection.

Five classes are defined with the time to first tracer arrival. The limits are established to 1, 3 10 and 30 days. For each of 5 classes, 3 representative tracer tests are selected among the 275 tracer tests already carried out by the author. The parameters of dispersion and velocity are compared and analyzed.

A time (first arrival) lower than one day is described as very short (very easy tracer test). The tracer tests carried out show a Peclet number lower than 50. The apparent dispersion of the curve is strong.

A time higher than 10 days is described as long (difficult tracer test). The Peclet number is generally higher than 200. The apparent dispersion of the curve is weak.

For an increase in the transit time of a few hours to more than 60 days, experiments show increase in the Peclet number and dispersion (in meters). No correlation is obvious between time and velocity.

Résumé

Le temps de première arrivée du traceur est une caractéristique intrinsèque et reproductible d'un système karstique dans les conditions hydrologiques du traçage. Ce temps ne dépend pratiquement pas du déroulement de l'injection du traceur.

Cinq classes sont définies en fonction du temps de première arrivée du traceur. Les limites sont établies à 1, 3 10 et 30 jours. Pour chacune des 5 classes, 3 traçages représentatifs sont choisis parmi les 275 traçages déjà réalisés par l'auteur. Les paramètres de dispersion et de vitesse sont comparés et analysés.

Un temps de transfert (première arrivée) inférieur à un jour est qualifié de très court (traçage très facile). Les traçages réalisés montrent généralement un nombre de Peclet inférieur à 50, c'est à dire que la dispersion apparente de la courbe est forte.

Un temps de transfert supérieur à 10 jours est qualifié de long (traçage difficile). Le nombre de Peclet est généralement supérieur à 200. La dispersion apparente de la courbe est faible.

Pour une augmentation du temps de transfert de quelques heures à plus de 60 jours, on constate expérimentalement des augmentations du nombre de Peclet et de la dispersion (en mètres). Aucune corrélation n'est évidente entre le temps de transfert et les vitesses.

1. Introduction

La caractérisation des traçages est difficile car les résultats dépendent souvent de la méthode de traçage employée. En particulier, le déroulement de l'injection du traceur peut influencer la forme de la courbe de restitution et donc la plupart des paramètres calculés du traçage. Or un traçage hydrogéologique a pour objectif d'obtenir des données sur le milieu souterrain, et non sur la méthodologie de traçage employée.

La réalisation d'une classification des traçages permettra aussi de choisir la bonne technique en fonction du degré de difficulté du traçage. Des techniques, qui fonctionnent parfaitement avec des traçages très faciles où le traceur arrive au bout de quelques heures, ne vont pas fonctionner sur des traçages difficiles où le traceur arrive des mois après l'injection.

2. Importance du temps de première arrivée du traceur

Le temps de première arrivée du traceur est un résultat essentiel d'un traçage. Ce temps ne dépend pratiquement pas de la méthodologie d'injection. En effet, dès que le traceur est injecté, les particules les plus rapides sont entraînées tout de suite par le mouvement de l'eau. Cela est différent du temps modal, du temps de fin de restitution et de la dispersion. Ces résultats peuvent

dépendre fortement du déroulement de l'injection. Pour pratiquement un même trajet souterrain, nous avons montré à la Fontaine de Vaucluse que la dispersion et le nombre de Peclet peuvent varier d'un facteur 10 selon le type d'injection. Une injection directe dans le flux souterrain donne une dispersion, un temps modal et un temps moyen faibles. Une injection dans une grande vasque, ou avec un stockage de traceur près du point d'injection, donne une dispersion, un temps modal et un temps moyen élevés.

Dans les conditions hydrologiques du traçage, le temps de première arrivée du traceur est une caractéristique intrinsèque du système tracé. Si les conditions du traçage sont identiques, c'est une valeur reproductible même avec des opérateurs différents.

De plus ce paramètre est utilisable directement, car la majorité des traçages servent à la protection des eaux souterraines. Ce temps est celui, avec un léger coefficient de sécurité, de l'arrivée d'une pollution dangereuse sur un captage.

3. Définition de la classification

Cinq classes sont définies en fonction du temps de première arrivée du traceur :

- Inférieur à 1 jour : traçage très facile (temps très court)
- Entre 1 et 3 jours : traçage facile (temps court).

- Entre 3 et 10 jours : traçage de difficulté moyenne (temps moyen).
- Entre 10 et 30 jours : traçage difficile (temps long).
- Supérieur à 30 jours : traçage très difficile (temps très long).

4. Propriétés des traçages en fonction des classes définies

Pour chacune des 5 classes, 3 traçages au moins en milieu karstique sont choisis parmi les 275 traçages déjà réalisés par l'auteur. Les moyennes des résultats par classes sont données dans le tableau suivant :

Classe	très court	court	moyen	long	très long
Temps (j)	0,2	1,5	6	13	50
Distance (m)	310	2300	3400	14000	48000
Dispersion (m)	13	15	20	25	300
Peclet	40	200	400	640	250
Diamètre (m)	2	4	6	19	33
Vitesse (m/h)	92	77	25	45	44

La vitesse est la vitesse maximale « à vol d'oiseau » du traceur.

Le diamètre est celui d'un cercle de surface égale au volume écoulé entre l'injection et le temps moyen divisé par la distance. Ce « diamètre tracé » est en relation avec la section des conduits noyés touchés par le traceur.

Dans cette définition du diamètre tracé, on suppose que le temps de parcours dans la zone non saturée est négligeable devant le temps de parcours dans la zone saturée. Cela est en général le cas pour les pertes actives. Dans le cas contraire, par exemple pour une injection dans une tranchée avec rinçage au moyen d'une citerne, il faut mesurer ou estimer le temps de passage dans la zone non saturée et le soustraire au temps moyen du traçage.

Pour les karsts noyés sans rivière souterraine et pour des traçages en conditions normales, hors fortes crues, le diamètre tracé est une caractéristique intrinsèque d'un système karstique. Par exemple, dans le karst de Vaucluse, le diamètre tracé est toujours proche de 40 m.

Pour des temps variant de 0,1 à 70 jours, on constate une augmentation de la dispersion, du nombre de Peclet et du diamètre tracé avec le temps de première arrivée. Le temps est bien sûr lié à la distance.

Pour des temps très courts à courts de 0,1 à 2 jours, l'augmentation de la dispersion avec la distance avait déjà été montrée par JEANNIN & MARECHAL (1997).

Pour un traçage, le nombre de Peclet est obtenu en divisant la distance par la dispersion. L'augmentation du nombre de Peclet avec le temps est nette, sauf pour les temps très longs. Plus les temps sont longs, plus le pic de restitution se resserre et plus la courbe est symétrique sans effet de traine (Fig. 1, Fig. 2 et Fig. 3). Pour des traçages difficiles, nous avons souvent obtenu des courbes qui se calent parfaitement sur le modèle advection dispersion à une dimension, sans aucun effet de traine (Fig. 2). L'explication physique est donnée par JEANNIN & MARECHAL (1997) : plus le temps est long, plus la probabilité pour qu'une particule de traceur passe à la fois par des zones lentes et rapides est grande. Pour des traçages non perturbés et quand il n'y a qu'un seul conduit karstique, la dissymétrie de la courbe diminue, le nombre de Peclet et la dispersion augmente.

Il n'y a pas de corrélation nette entre le temps et la vitesse maximale.

On note également une augmentation du diamètre tracé avec le temps. Plus les karsts sont étendus, plus le diamètre des conduits est grand, ce qui est logique. D'autres facteurs sont également responsables de la taille des conduits : facteurs climatiques, historique du karst.

5. Exemples d'utilisation de la classification

La classification peut aussi être utilisée pour choisir le traceur, les techniques de traçage, les compétences nécessaires et les références de l'organisme qui va réaliser le traçage.

Pour les traçages très faciles, un très grand nombre de produits chimiques ou solides peuvent être utilisés. Les traçages ont pour objet la protection de l'environnement. Etant donné que des dizaines de produits alimentaires et agréés alimentaires peuvent être utilisés comme traceur, nous proposons de ne plus utiliser de produits chimiques artificiels non agréés alimentaires pour les traçages très faciles. Nous proposons que les rhodamines et la fluorescéine au caractère potentiellement cancérigène connu depuis 1978 et 1992 respectivement, IARC (1978) et SAX'S (1992), ne soient plus utilisées pour les traçages très faciles. BEHRENS H. ET AL. (2001), sur la base de 2 tests négatifs de génotoxicité conclut que la fluorescéine est inoffensive. Cependant SAX'S (1992), publication indépendante des utilisateurs passés et actuels des traceurs artificiels fluorescents, donne une description différente pour la fluorescéine : « Experimental reproductive effects. Questionnaire carcinogen with experimental tumorigenic data. Mutation data reported », avec un classement en classe 3, c'est à dire parmi les produits chimiques les plus dangereux. S'il se confirme que le produit est réellement cancérigène, une seule molécule pourrait causer un cancer. Prendre un risque, si faible soit-il, est anormal alors que l'on peut utiliser sans problème des ingrédients alimentaires pour les traçages faciles. Par exemple, le traceur iode est bénéfique pour la santé à cause de la carence actuelle en iode de la population.

Pour les traçages très difficiles, le traceur ion iodeure, naturel, incolore et alimentaire, est bien adapté. Les deux derniers traçages sur la Fontaine de Vaucluse de 56 km (temps de première arrivée 31 j) et 66 km (temps 70 j) le prouvent. Ces traçages sont les plus longs en distance d'Europe. Pour les traçages difficiles, les Fig. 1 et Fig. 2 donnent des exemples de courbes de restitutions obtenues avec le traceur iodeure. SCHUDEL ET AL. (2002) mentionne l'iodeure comme « chimiquement instable, en particulier dans les eaux souterraines chargées en matière organique ». En 275 traçages à l'ion iodeure, le phénomène décrit n'a jamais été observé. Au contraire, plusieurs injections, dans des gouttes à gouttes s'infiltrant dans plusieurs mètres des matières organiques en décomposition, d'humus et d'argile, ont donné des courbes de restitution parfaites et des taux de restitution élevés. Dans des conditions identiques, la fluorescéine, ou tous autres produits fluorescents, aurait été partiellement ou totalement adsorbée par les matières organiques ou l'argile. L'affirmation de SCHUDEL ET AL. (2002) est donc contestable, ainsi que sa conclusion sur le choix préférentiel des produits chimiques artificiels fluorescents par rapport à l'iodeure pour tous les traçages.

En milieu karstique, la fluorescéine est aussi utilisée, mais on démontre souvent que les résultats positifs annoncés sont de fausses restitutions. Par exemple, notre traçage de 2004, à l'ion iodeure sur la Fontaine de Vaucluse,

démontre sans ambiguïté que la coloration de 1963 à la fluorescéine était un échec. En milieu karstique, l'eau est de couleur verte et est souvent chargée en matières organiques fluorescentes ou en argile, le problème vient de l'adsorption dans le milieu souterrain de ce produit chimique artificiel. La seule solution est alors d'augmenter la quantité injectée. Cela est dangereux vu les quantités importantes d'ammoniaque toxique et d'alcool à brûler polluant souvent utilisées pour les injections.

Un autre exemple de l'utilisation de la classification est donné pour les multitraçages avec des colorants fluorescents. IDSTEIN & EWERS (2002) ont montré que les rhodamines étaient peu stables dans le milieu souterrain et qu'il y avait 2 produits de décomposition très stables (détection 12 ans après un traçage). L'un des produits de décomposition a la même fluorescence que la fluorescéine et l'autre la même fluorescence que l'éosine. La formation des produits de décomposition a été observée en un mois lors de traçages pour la rhodamine WT. Un multitraçage de classe moyenne à très difficile avec une rhodamine risque donc de donner des résultats faux.

6. Conclusion

La classification proposée des traçages permet de choisir une technique adaptée à la difficulté du traçage.

Pour une augmentation du temps de transfert de quelques heures à 70 jours, on constate expérimentalement des augmentations du nombre de Peclet, de la dispersion et du diamètre tracé.

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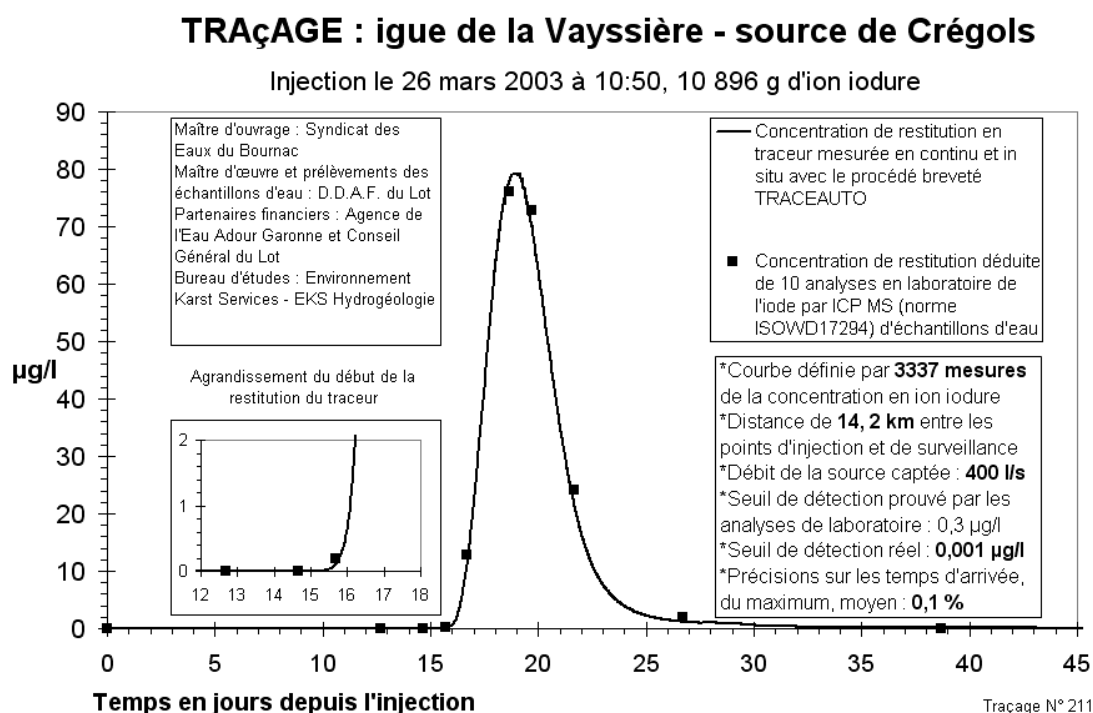


Fig. 1 : Courbe de restitution d'un traçage difficile (temps long) avec un nombre de Peclet de 350 et un pic symétrique

TRAÇAGE : perte du Ravin des Enfers - source d'Aygue Astaud

Injection, le 30 mars 2004 à 16:50, de 2514 g d'ion iodure

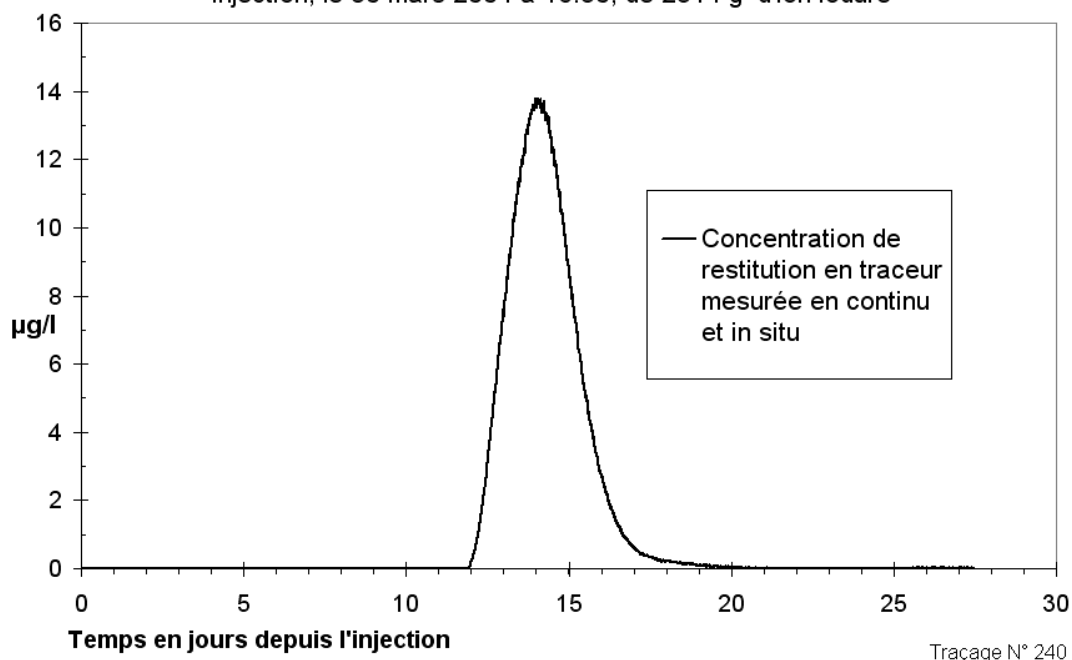


Fig. 2 : Courbe de restitution d'un traçage difficile (temps long) avec un nombre de Peclet de 400 et une courbe sans effet de traine se calant bien sur le modèle advection dispersion à une dimension

TRAÇAGE : gouffre de Fontaine de Vaucluse - sources pérennes

Injection, le 19 octobre 2003 à 11:45, de 838 g d'ion iodure par la SSFV

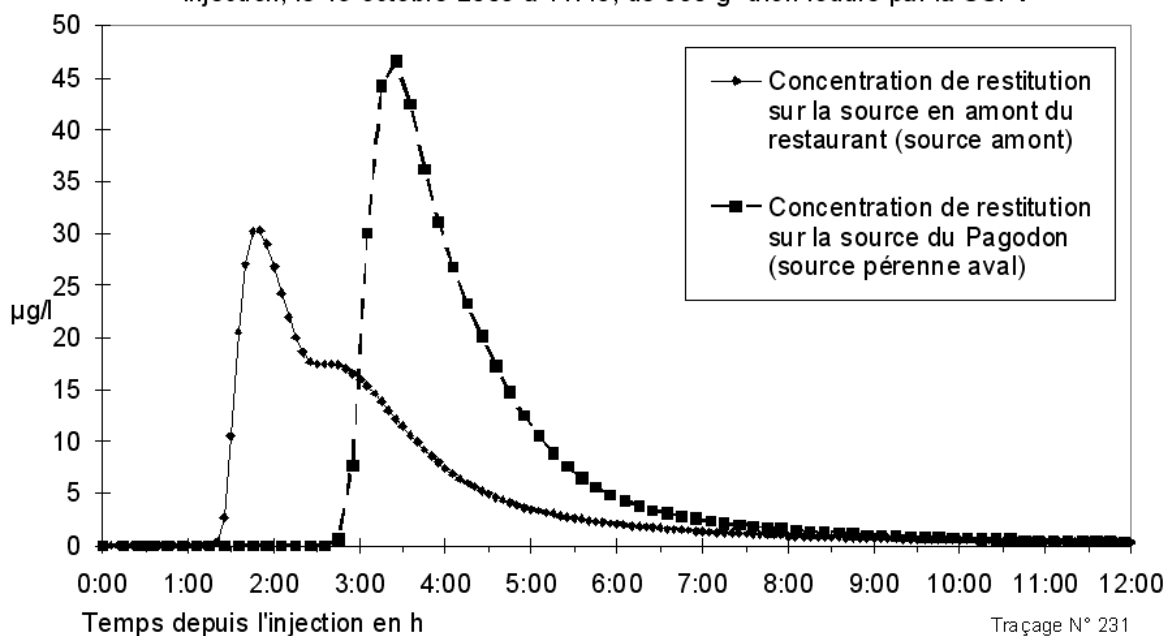


Fig. 3 : Courbes de restitution d'un traçage très facile (temps très court) avec des pics dissymétriques et des effets de traine important

Variabilité du type et de la qualité de l'information issue du traçage naturel en fonction des caractéristiques des systèmes étudiés. Quelques exemples français et espagnols

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Résumé

L'hydrochimie, puissant moyen d'interprétation du fonctionnement de l'hydrosystème karstique, ne saurait être utilisée sans analyse critique des données acquises. Des interprétations hasardeuses pourraient résulter de son application standard. Trois exemples sont donnés.

Deux sources méditerranéennes en crue montrent, l'une, une diminution de la conductivité, l'autre, une augmentation. Le magnésium montre une baisse sur les deux, attestant de l'arrivée d'eaux d'infiltration à faible temps de séjour. Dans ce cas, la conductivité ne saurait être utilisée comme indicateur du temps de séjour.

Sur le second exemple, lors d'une baisse de débit, on observe une baisse des concentrations en nitrate en même temps qu'une augmentation de teneur en COT, deux traceurs pourtant originaires du sol. Cette évolution inversée traduit l'épuisement du stock de nitrates dans le sol en hiver.

Le dernier exemple montre, lors de petits épisodes pluvieux, une plus grande sensibilité du magnésium (décroissance lente) que de la conductivité (palier).

Abstract

Hydrochemistry, a powerful tool to interpret the behaviour of karst hydrosystem, should not be used without a critical analysis of the acquired data. Wrong interpretations could arise from its standard application. Three examples are given.

Two Mediterranean springs in flood conditions display, one of them, an electrical conductivity decrease, the other one, an increase. Magnesium decreases in both, because of the outflow of short residence time infiltration water. In this case, we cannot use electrical conductivity as a tracer of residence time.

In the second example, during a decrease of flow rate, a decreasing trend of nitrate concentrations and an increase of TOC concentrations are observed. Nevertheless, these two tracers are coming from the soil. This antithetic evolution is interpreted as exhaustion of the nitrate soil in winter.

In the case of the last example, magnesium shows a higher sensitivity than electrical conductivity (slow lowering vs stability) during short rainy periods.

1. Introduction

L'hydrochimie, en conjonction avec l'hydro-dynamique, est un moyen puissant permettant de proposer des modèles conceptuels de fonctionnement des systèmes karstiques (Bakalowicz, 1979 ; Blavoux et al. 1992). Cet outil se révèle très sélectif si l'origine du traceur est bien établie et unique. L'utilisation généralisée de l'hydrochimie a fait progresser la connaissance des aquifères karstiques, mais elle peut conduire, faute de réflexion sur les traceurs utilisés, à des interprétations hasardeuses, voire erronées.

A partir de 3 exemples, nous nous proposons d'illustrer et de discuter la diversité de comportement de quelques traceurs communément employés. Puis nous donnerons quelques règles à respecter afin d'éviter une utilisation abusive du traçage naturel pouvant avoir des conséquences fâcheuses, en particulier dans l'exploitation des aquifères karstiques.

2. Exemples de différences dans l'interprétation des traceurs naturels

2.1 Traceurs du temps de séjour – traceurs de l'infiltration (conductivité, magnésium, nitrates).

La figure 1 montre le comportement de la conductivité électrique et du magnésium de deux systèmes karstiques en crue (Fontaine de Vaucluse et Notre-Dame des Angès), contigus et localisés dans la même unité géologique et climatique.

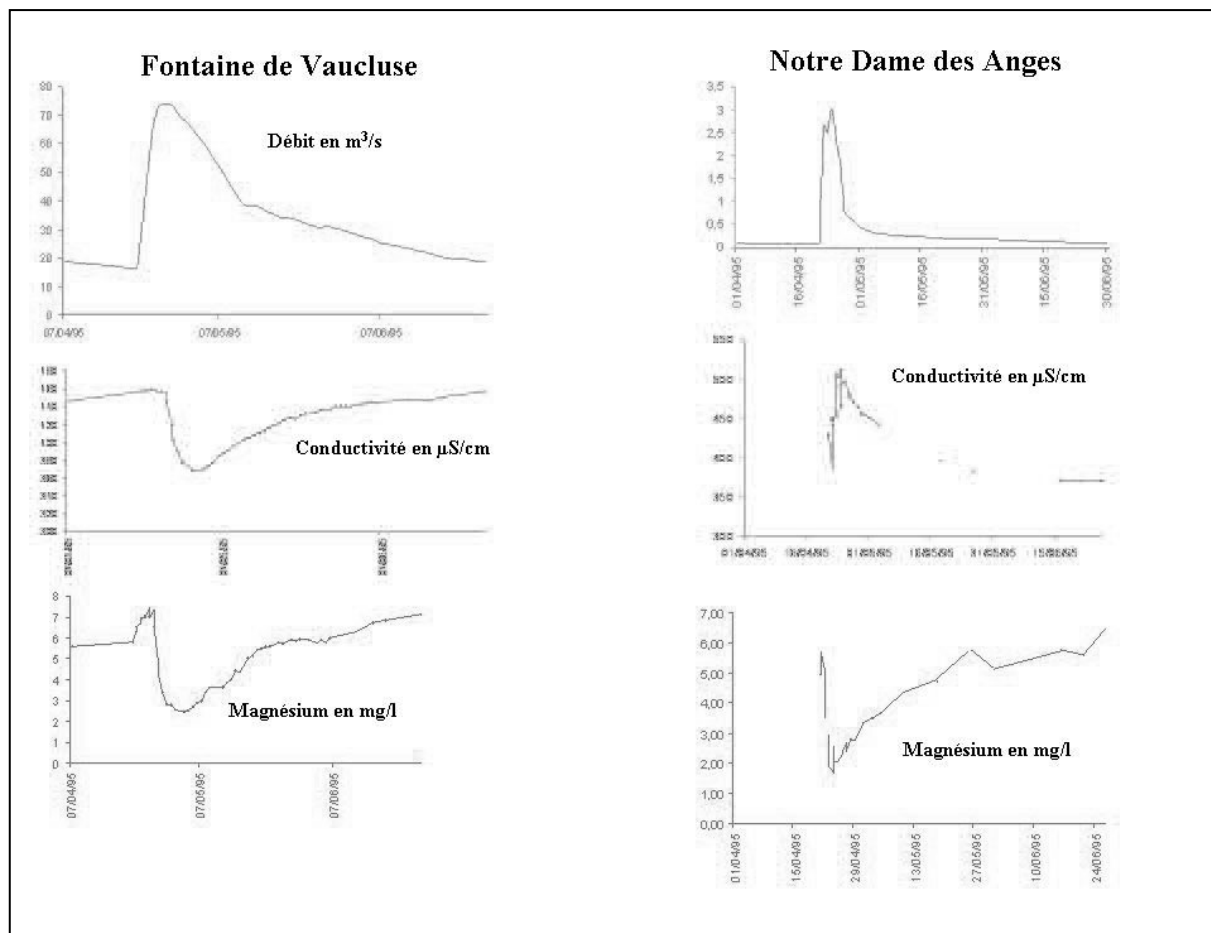


Figure 1 : Evolution de la conductivité électrique et du magnésium à la Fontaine de Vaucluse et à Notre-Dame des Angès (Sud-Est de la France).

L'interprétation communément admise de la conductivité électrique (baisse de conductivité = baisse du temps de séjour, augmentation de conductivité = augmentation du temps de séjour), validée à la Fontaine de Vaucluse par une baisse du magnésium qui peut être utilisé comme un traceur du temps de séjour dans cette entité géologique (Mudry, 1987) est en contradiction avec l'information du magnésium à Notre-Dame des Angès (NDA).

L'augmentation de conductivité enregistré à la source NDA pendant la crue s'explique par des teneurs particulièrement élevées en calcium et bicarbonates (faisant plus que compenser la dilution des traceurs du temps de séjour). Ces eaux à fortes pCO_2 localisées dans l'épikarst, ont eu le temps de se mettre à l'équilibre vis-à-vis de la calcite, mais la dilution du magnésium indique clairement un faible temps de séjour (Lastennet, 1994).

Cet exemple démontre la nécessité de toujours croiser au moins deux traceurs en prenant soin de s'assurer de leur indépendance.

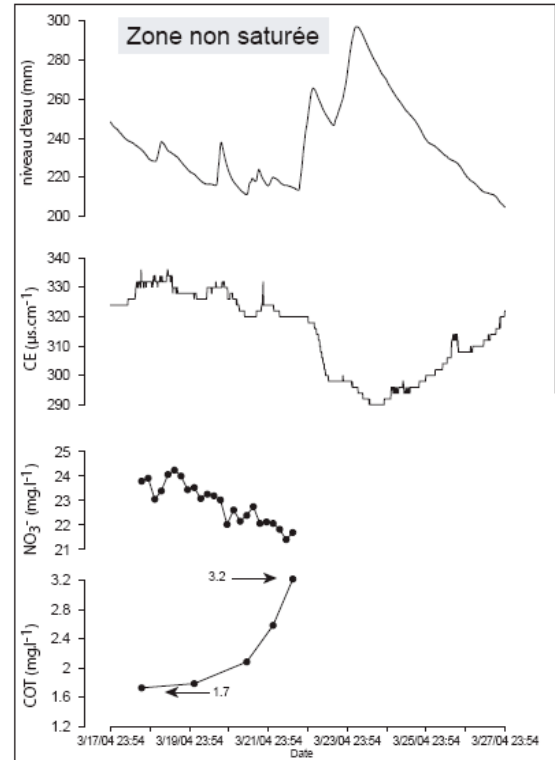
2.2 Traceurs des différents types d'infiltration.

La figure 2 montre l'évolution du carbone organique total (TOC/COT), des nitrates et de la conductivité électrique (CE), durant un épisode de crue dans la zone non saturée du système karstique de Fourbanne (Doubs, France).

Dans ce système, les nitrates dépassent 25 mg/l, indiquant une pollution anthropique attribuable à l'agriculture. Durant la période d'étude (baisse de débit postérieure à une crue non enregistrée), on observe une baisse de concentration en nitrate et une augmentation de teneur en COT. Les deux

traceurs, dont l'origine est le sol, montrent pourtant une évolution inversée. Deux traceurs de l'infiltration peuvent donc donner des informations différentes. Ce phénomène apparemment surprenant s'explique de la manière suivante (Emblanch *et al.*, 2005) : lors de la première crue, les eaux prélevées correspondent à une chasse de l'eau préexistante dans la zone non saturée (forte conductivité électrique et faible COT), les fortes concentrations s'expliquent par le caractère aéré de la zone non saturée, ne permettant pas la dénitrification, l'eau a donc conservé sa charge initiale en nitrates. Puis on observe simultanément une baisse de la conductivité électrique et une augmentation du COT, indiquant une baisse du temps de séjour. Au même moment, les nitrates montrent une diminution, qui s'explique par l'épuisement du stock de nitrates dans le sol à la fin de la période humide. Contrairement à l'interprétation classique, cette diminution de nitrates correspond à une baisse du temps de séjour.

Figure 2 : Evolution de la conductivité électrique CE, du TOC/COT et des NO_3^- dans la zone non saturée (aquifère de Fourbanne, Charmoille, 2005)



2. 3 Diversité de sensibilité des traceurs naturels.

La figure 3 illustre clairement la différence de sensibilité des traceurs naturels utilisés. Alors que la conductivité électrique (CE) reste stable jusqu'au 25/10/2003 (pas de réponse aux petits épisodes pluvieux du mois précédent), le magnésium voit sa concentration diminuer de façon significative dès le 14/10/2003.

La conductivité électrique de l'eau est un paramètre très fréquemment utilisé, du fait de sa simplicité de mesure, de son coût modique et de l'intérêt de l'information délivrée. Ce paramètre apparaît beaucoup moins sensible que le magnésium, ceci peut s'expliquer par le fait que la conductivité intègre l'ensemble de l'information hydrochimique

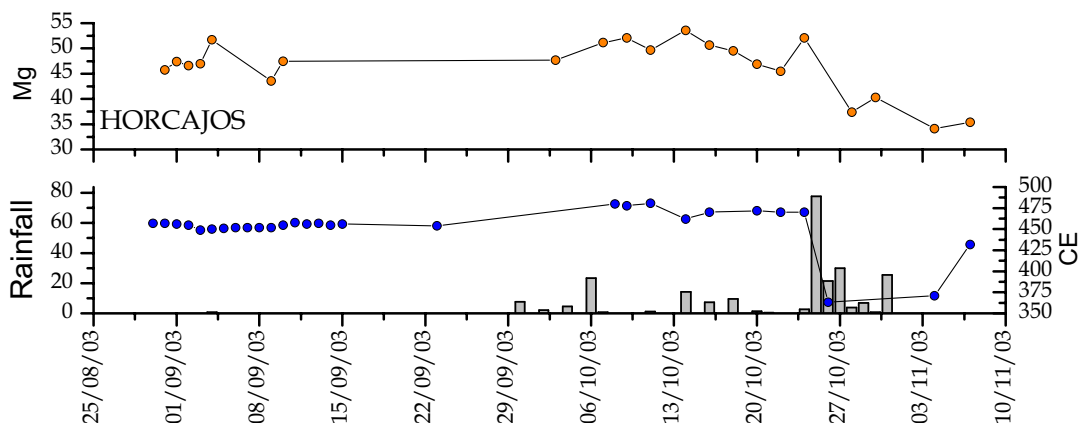


Fig 3 : Evolution de la conductivité (CE) et du magnésium à la source d'Horcajos (Andalousie, Sud de l'Espagne)

3. Conclusions et recommandations

Au travers de ces exemples, apparaît la nécessité de ne pas délivrer une interprétation standard des données hydrochimiques, les traceurs naturels pouvant avoir des origines diverses en fonction des contextes géologique (lithologie et structure), géomorphologique et hydrogéologique (venues profondes, biseau salé...). Il est important de toujours croiser au moins deux traceurs, en prenant soin de s'assurer de leur indépendance statistique. Dans cette optique, il est recommandé de pratiquer des analyses multidimensionnelles exploratoires, destinées à repérer, grâce à la structure des nuages de variables, les traceurs de phénomènes distincts et correctement identifiés : infiltration naturelle (diffuse ou concentrée), interaction eau-roche (au sein de l'hydrosystème karstique), limites géochimiquement contrastées (silicates, évaporites...), apports anthropiques...

Il importe également de considérer comme dynamique le stock de traceurs mobilisables, par exemple dans le sol et dans l'épikarst. Un sel lessivable à la reprise des écoulements d'un cycle hydrologique n'est plus forcément disponible au cœur de la saison pluvieuse : la diminution des nitrates peut donc, de manière atypique, correspondre à une baisse du temps de séjour.

Certains paramètres hydrochimiques, tels la température, la conductivité électrique ou la fluorescence, sont faciles à acquérir en continu, ce qui permet d'en suivre les fines variations temporelles, mais ils peuvent intégrer plusieurs traçages de base : une constance de la conductivité au cours d'un épisode de crue peut en fait être imputable à une diminution des teneurs en ions indicateurs du temps de séjour (bicarbonates, magnésium...) compensée par une augmentation des teneurs en traceurs acquis lors de l'infiltration de l'eau (nitrates, chlorures, sulfates...). Enregistrer la conductivité ne dispense, dès lors, pas de faire des analyses détaillées des ions à un pas de temps adapté à la « nervosité » de la réponse hydrochimique. L'enregistrement en continu permettra d'interpoler les grandes variations individuelles des teneurs entre les instants de prélèvements.

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Hydrogeological investigations in the Rózsadomb area for the protection of the thermal karst system (Budapest, Hungary)

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Abstract

Budapest has a particular hydrogeological setting: lukewarm and hot karst springs arise next to each other along the river Danube. Nevertheless the function of the Buda Thermal Karst System is far from being completely understood, as yet. It is supposed, that a part of the water is being recharged within the urban area of Budapest, including the nearest hill to the Danube, the Rózsadomb area. During the 20th century several caves were discovered beneath the villas of Rózsadomb, 10 to 30 m below the surface. The exquisite mineralization associated with these caves and their situation beneath a densely populated urban area strongly calls for protection. The understanding of the hydraulics of the karst system is crucial here also because of the vulnerability of groundwater. The aims of the present project were: (1) to acquire direct information with the help of discharge measurements and a tracer test on how the Buda Thermal Karst System works between Rózsadomb and its discharge area; (2) to investigate the protective function of the epikarst zone by analysing dripping waters in the Szemlő-hegy Cave and (3) to improve the understanding of discharge characteristics of the Buda Thermal Karst by data acquisition from the the Molnár János underwater cave and by retrospective analysis of natural springs data. The results of these studies can be considered as first steps towards setting up a protection policy for caves and springs in Budapest.

1. Introduction

1.1 Hydrogeology of the Buda Thermal Karst Sytem

The Buda Thermal Karst System belongs to the NE part of the Transdanubian Central Range (TCR) (Fig. 1). The understanding of the operation of the system was hitherto based on conceptual models of SCHAFARZIK (1928), VENDEL & KISHÁZY (1964), ALFÖLDI (1979, 1981) and KOVÁCS & MÜLLER (1980). The aquifer of the TCR is a several thousand metres thick Mesozoic carbonate sequence. The karstic recharge area (about 15% of the whole area) (LORBERER, 1986) feeds a complex geothermal flow system with perennial discharge sites. Infiltrating cold water is heated up because of thermal convection of upward flowing groundwater (ALFÖLDI, 1981). After a deep, regional-scale flow path water returns to the surface as thermal water. The age of this water is about 5000-16000 years based on ^{14}C measurements (DEÁK, 1978). The rest of the infiltrated water discharges as lukewarm springs or seepages via local or intermediate flow-systems.



Fig. 1: Location of the Buda Thermal Karst in the Transdanubian Central Range (1: Subsurface boundary of Mesozoic carbonates, 2: Uncovered Mesozoic carbonates, 3: Buda Thermal Karst System)

Along with Mesozoic carbonates, also Eocene limestones serve as aquifers in the Buda Thermal Karst area. Thermal water discharge is localized here by the step-faulted boundary between the subsided basin to the east and the uplifted hilly range in the west. The course of the

Danube follows this boundary and represents the base level of erosion, which is also the discharge site of the karst waters (Fig. 2).

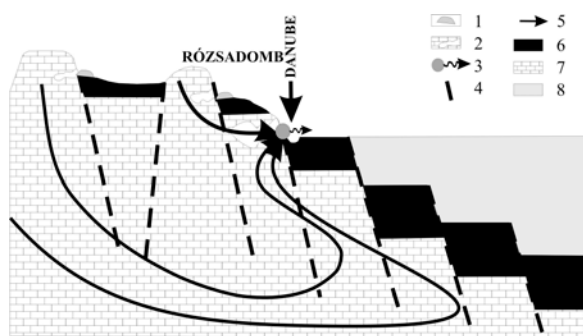


Fig. 2: Hypothetical cross section of the Buda Thermal Karst (modified after KOVÁCS & MÜLLER, 1980): 1: travertine, 2: cave, 3: spring, 4: faults, 5: supposed flow path, 6: aquitard, 7: aquifer, 8: basinal deposits

1.2 Characteristics of the Rózsadomb Area

One of the nearest hills to the Danube is Rózsadomb, which has relatively steep slopes towards the river. Its highest point is 195 m asl. The base level is 104 m asl at the riverside, where the springs of widely different temperatures arise. Surface karst phenomena are rare because of the characteristics of the thermal karst area. Since the 1920s human activities have completely changed this part of Budapest concerning the "natural conditions". The original vegetation has completely been destroyed, and extensive construction works and limestone quarrying have modified the relief. Accelerated urbanization of the capital reached this area in the 1970s, and now more than 80% of the whole 4-km² area has been built over. The infiltration is about 35 mm/y over the marly terrains. The original yield of the springs at the foothills of Rózsadomb is estimated as 34 000 m³/d (HAZSLINSZKY *et al.*, 1993).

During the 20th century several caves were discovered beneath the villas of Rózsadomb, 10 to 30 m below the surface. The presently known five large caves with a total

surveyed length of more than 35 km are hosted in Upper Eocene limestone but the uppermost passages developed also in the overlying marl. The multilevel maze network is arranged in stratigraphically and tectonically controlled levels. These caves are formed by mixing corrosion near the discharge zone of lukewarm and hot karst waters (KOVÁCS & MÜLLER, 1980; TAKÁCS-BOLNER & KRAUS 1989; LEÉL-ÖSSY 1995). Today's springs debouche along the river Danube at 104 m asl while the ancient caves (already dry, like Szemlő-hegy Cave) are situated at an elevation of 262 to 110 m asl. In the Molnár János Cave situated at and below the karst water table (104 m asl), speleogenesis by mixing corrosion is a currently active process.

2. Materials and methods

2.1. Discharge of the springs

The discharge volume of the Rózsadomb' springs represent the overall response of a karstic reservoir to the input precipitation events. The measurement of discharge may help to understand how the karst system works. Such measurements can be utilized also during the planning of tracer tests (injected concentrations, predicted dilution etc.). On the basis of scattered measurements, ALFÖLDI *et al.* (1968) estimated the flow rates for the Boltív Spring at the foothills of Rózsadomb as 3000–6500 m³/d. These values were used for planning the tracer test.

Daily spring discharge measurement was carried out in the foothills of Rózsadomb, on a three-month time-scale.

1.3 Aims

The aims of the present project were: to acquire direct information with the help of discharge measurement and tracer test on how the Buda Thermal Karst System works between the Rózsadomb and its discharge area. The investigation of dripping waters in the Szemlő-hegy Cave – simultaneously recording also the precipitation events – can give information about the unsaturated zone above the cave with special regard to the protective function of the epikarst. By data acquisition from the “natural research laboratory” i.e. the Molnár János underwater cave the understanding of discharge characteristics of the Rózsadomb area is improved. The results of these studies can be considered as first steps towards setting up a protection policy for caves and springs.

Daily precipitation was recorded during the same period. During the discharge measurements extreme difficulties arose because of the urbanized environment of the once natural spring-outlets. Accordingly, measurements had to be focused on the Boltív Spring, the only natural, free outlet feeding the artificial Malom Lake. The discharge volume was estimated from two components. Firstly, with the help of a water-meter on the pipeline it was possible to measure how much water was actually used by the bath associated with this spring (Lukács Bath). The overflow of the Malom Lake was considered as the second component of the total discharge and it was measured by a propeller-type current meter in the canal (Fig. 3).

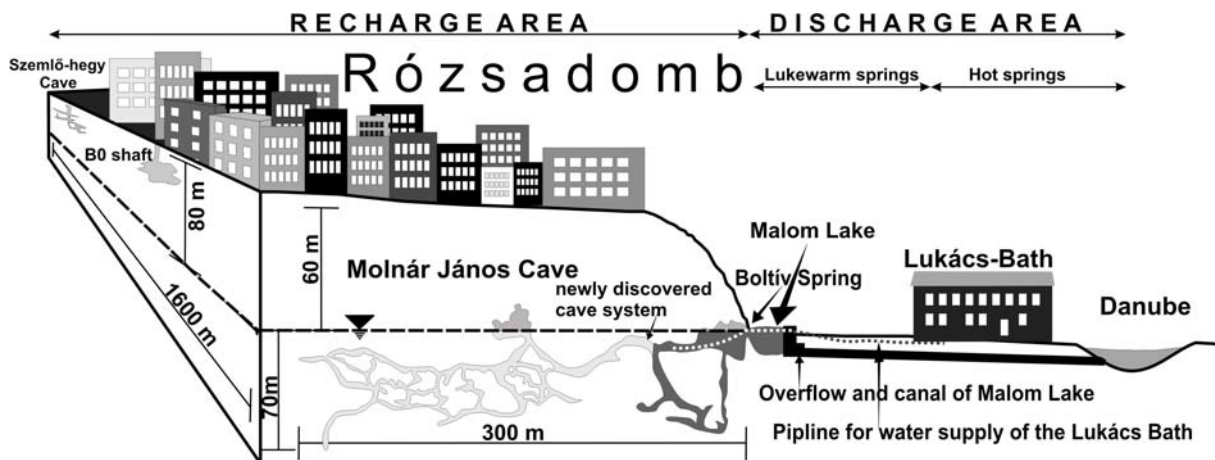


Fig. 3: Sketch of the Rózsadomb area (modified after Wolk, 2003)

2.2. Tracer test between Rózsadomb and Lukács-Bath

According to the results of a previous swallowing test in a cave passage high upon the hill (Fig. 4), it was tempting to suppose direct connection between the Rózsadomb area and the Boltív Spring (SÁRVÁRY *et al.*, 1992). The expected breakthrough-time predicted from this previous study was from 10 hours to 42 days.

Tinopal CBS-X, an optical brightener in a solution concentration about 15 g/l (10 kg dissolved in 600 Litre water) was used as fluorescent tracer for the experiment. Although Tinopal was not the best choice as a tracer due to its relatively high affinity to sorption (SCHUDEL *et al.*, 2003), this was the only allowed tracer which could be used for water tracing around the bath. Shaft B0 in the

Rózsadomb was selected to be the most appropriate injection place (Fig. 4).

Prior to the tracer injection, 500 m³ of water was injected into the shaft to make sure that its swallowing capacity will not limit the operation and that the underground channelways are properly saturated, to avoid obstruction of the descending dye by dry sections.

The sampling period after the tracer injection was two months, with a gradually decreasing sampling frequency. In total, 1100 samples were collected at ten different sites (wells, springs) in the area of the Lukács Bath (Fig. 4). Additionally, at the Molnár János underwater cave the tracer was continuously monitored with a flow-through field fluorometer (SCHNEGG & DOERFLIEGER, 1997) (detection limit for Tinopal: 10⁻⁹ g/ml), connected to a pipeline that

was deeply immersed in the water of the adjoining phreatic Molnár János Cave.

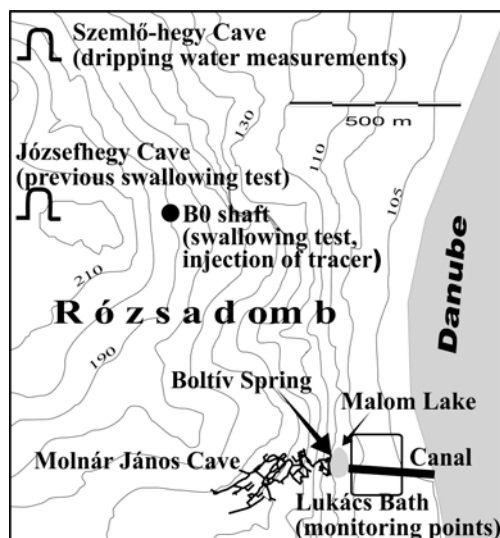


Fig. 4: Location map of the tracer test

2.3. Dripping water analysis in the Szemlő-hegy Cave

In a previous study of MÁDL-SZÖNYI *et al* (2001) it was found that the epikarst of the Rózsadomb area plays a crucial role in the regulation of karstic recharge and hereby influences the protection of caves. In the Rózsadomb area above the karstic carbonate rock an extensive detrital blanket was formed by periglacial processes. Where carbonate debris directly overlies the solid carbonate bedrock without any matrix or with only 20-30 % clay-loess matrix it functions like a porous temporal aquifer. This epikarst has storage capacity and pollutant filtering function depending on the ratio of debris to matrix. Due to sorting in

this clastic cover (TYC 1996, 1997) direct water transport to the carbonate bedrock is also possible at places. Where the carbonate rocks are exposed on the surface, faults, shafts and bedding planes serve as channelways for the infiltrating water. Since the original soil and vegetation acted as natural filters retaining pollution, in the lack of them the epikarst became the only protective layer. The function of the epikarst can be studied in the caves by localizing the dripping points and analyzing the collected dripping water.

In the Szemlő-hegy Cave drip-water measurements have been carried out since 1987, and in October 2005 eight measuring points were reinstalled to represent the whole cave. With weekly sampling frequency in situ measurements (specific electric conductivity, temperature and pH) are carried out. In the laboratory the major cation- and anion contents were measured and simultaneously the precipitation recorded.

2.4. Investigations in the Molnár János cave and Rózsadomb discharge area

A succession of enormous underwater cavities (ca. 5 km total length, deepest point 76 m below the watertable KALINOVITS pers. comm.) is known as Molnár János Cave (Fig. 3) near to the discharge area of Rózsadomb. Retrospective research was accomplished to compile a map showing the position, temperature and chemical characteristics of former lukewarm and hot springs. In the Molnár János cave systematic temperature-measurements are carried out, to understand the underground distribution of the lukewarm and hot waters near the Rózsadomb discharge area. Chemical analysis for major cations and anions were used to evaluate the chemical differences between hot and lukewarm waters of various origins. With stable isotopes measurements the climatic characteristic of the recharge period and mixing rates could be investigated.

3. Results

Discharge measurements in the canal between the Malom Lake and the Danube showed a rather voluminous discharge (~13700–15700 m³/d) much greater than that previous by estimated by ALFÖLDI *et al.* (1968). Based on the simultaneous observation of precipitation and discharge, it was also concluded that there is no significant correlation between them. This suggests that the karst system has high storage capacity but no well developed hierarchical karst network.

Despite the carefully executed tracer measurements no dye breakthrough could be detected during the two-month observation period. One of the possible explanations is that though the tracer has reached the karst water table and descended towards the Lukács Bath, behind the Boltív Spring in the Molnár János Cave, it was met by an enormous water volume that it became overdiluted so that its concentration fell under the detection limit. The size of the newly discovered huge passages of the Molnár János Cave (KALINOVITS pers. comm.) also confirms this idea.

The dripping water studies from the Szemlő-hegy Cave (Fig. 5) show that the rain water doesn't reach immediately the ceiling of the cave rather will be retained by the epikarst. Specific electric conductivity decreases from autumn to spring because epikarstic storage is gradually shortened and fluctuates due to precipitation or snow melting events. The shorter epikarstic storage results lower

HCO₃⁻ content. The human contamination originated from the recharge area can be dangerous for the mineral formations of the cave but due to the adsorption in the 80 m thick unsaturated zone and the attenuation in the saturated zone by the time the water reaches the discharge area it is essentially uncontaminated.

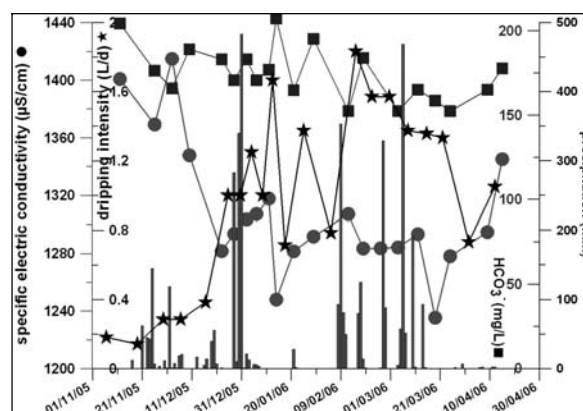


Fig. 5: Dripping water measurements in the Szemlő-hegy Cave: dripping point No. 3. (columns: precipitation; line with squares: HCO₃⁻; line with dots: specific electric conductivity; line with stars: dripping intensity)

From retrospective research at the Rózsadomb discharge area it could be deduced that the hottest springs (55–65 °C) arose near the Danube. The lukewarm springs (20–27 °C) came to the surface between the hot springs and the foothill of Rózsadomb (Fig.3). The hot springs have 1000–1200 mg/l TDS with 140–170 mg/l Cl⁻ and 100–150 mg/l Na⁺ content while lukewarm springs have 500–700 mg/l TDS with 40–60 mg/l Cl⁻ and 20–70 mg/l Na⁺ content (LIEB, 2004).

4. Conclusion

As a consequence of the investigations it turned out that the discharge of the Boltív Spring was 2 to 5 times underestimated by previous authors. No significant correlation was found between the variation of discharge and the recorded precipitation. It means that the direct recharge from Rózsadomb is several orders of magnitude less than the discharge at the Boltív Spring.

The negative tracer test did not provide any direct answer to the original question, namely, that how the Buda Thermal Karst System works between Rózsadomb and its discharge area at the Boltív Spring. However, it greatly improved our understanding for the karst system. It proved that the karst system is of very high storage capacity and that this is the reason for the water table not responding directly to precipitation events.

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L'échantillonnage représentatif en milieu hétérogène et technologie appropriée

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Résumé

En milieu homogène, dans les zones à faibles gradients de vitesses d'écoulement, les charges hydrauliques dans les forages sont considérées comme invariantes suivant la verticale. En milieu hétérogène à porosité d'interstices, en l'absence de différenciations lithologiques perceptibles lors de la foration, il est généralement admis qu'il en est à peu près de même. Par contre, il est possible de montrer que des forages, en milieu à porosité de fissures, hétérogène et discontinu à l'échelle du forage, sont le siège d'écoulements verticaux, qui mettent en relation des fissures présentant des différences de charge hydraulique de quelques centimètres au repos, à plusieurs mètres en régime influencé. Ces différences de charge hydraulique de seulement quelques centimètres, peuvent provoquer des échanges entre fissures de quelques litres à quelques dizaines de l.h⁻¹, il suffit alors de quelques jours pour entraîner le renouvellement de la colonne d'eau d'un forage ainsi que d'une partie de la nappe à proximité, et, éventuellement, un changement complet de chimisme. Des exemples pris dans divers milieux fissurés, où ont été appliquées des méthodes et des techniques appropriées et sélectives, basées sur de longues chroniques et des diagraphies, illustrent ce qui précède. L'ignorance de ce processus peut entraîner de graves erreurs d'interprétation, concernant le fonctionnement et l'organisation hydrochimique des nappes, la cartographie de la distribution d'un élément, l'âge des eaux, la modélisation, etc...

Les phénomènes identifiés sur le terrain ont été reproduits au laboratoire à l'aide d'un modèle physique qui a permis de tester des appareillages existants et d'étudier la mise au point de dispositifs et de méthodes de prélèvement ou d'interprétation bien adaptés à l'étude des hydrogéosystèmes hétérogènes, surtout fissurés.

1. Introduction

L'acquisition de données et d'échantillons représentatifs des conditions de gisements des eaux souterraines, surtout à partir de forages, est sans conteste une opération délicate, dont la difficulté est parfois sous-estimée ou ignorée. Il ne s'agit pas seulement d'obtenir ces données et échantillons à distance dans un espace restreint, à l'aide d'outils plus ou moins bien adaptés, mais aussi, lorsque les problèmes de cet ordre sont résolus, de reconnaître la signification exacte des informations ainsi acquises et leurs limites.

En effet, un forage implanté dans un milieu poreux homogène contiendra une colonne d'eau sensiblement à l'équilibre avec la nappe à son contact, aux faibles mouvements liés à la convection thermique et à la diffusion moléculaire près, hors la zone soumise à l'influence de l'atmosphère. Mesures et échantillons pris dans un forage seront alors assez fidèlement représentatifs de l'eau de la nappe à une profondeur déterminée. Par contre, en milieu hétérogène, que la roche soit carbonatée ou pas, quelle que soit la situation par rapport aux limites et les gradients, surtout si la porosité est de fissures ou de chenaux, il est à craindre qu'une situation aussi simple que celle qui est décrite précédemment ne puisse plus exister.

Nous proposons dans cette note d'illustrer cette éventualité, à l'aide d'ouvrages de reconnaissance (F31 et Pz2) qui présentent un fonctionnement particulièrement révélateur d'un phénomène que l'on rencontre systématiquement dans la zone d'étude, mais avec un caractère démonstratif moindre, et de montrer à quel point elle peut conduire à des interprétations erronées, s'il n'en est pas tenu compte.

2. Evaluation des effets de l'hétérogénéité hydraulique en milieu fissuré

Le périmètre expérimental de Kerveldréac'h (PEK) se prête particulièrement bien à l'illustration des problèmes

d'interprétation de données obtenues en milieu fissuré, à l'aide de forages profonds. Le détail de la description du PEK et des résultats qui y ont été obtenus est détaillé dans SOMLETTE (1998), FAILLAT, SOMLETTE & SICARD (1999) et FAILLAT, CORRE, SICARD & SOMLETTE (1999). Nous ne donnerons ici que les grands traits du contexte hydrogéologique et de l'aménagement du PEK, ainsi que du matériel et des méthodes employés.

2.1. Contexte hydrogéologique et dispositif

L'aquifère fissuré, situé dans le bassin versant du Kerharo (Finistère, région de Châteaulin-Douarnenez), est constitué de schistes briovériens plus ou moins gréseux à faciès flysch, à structure plissée isoclinal très redressée, parcourus par des fractures NW-SE, qui correspondent aux rejeux de failles tardi-hercyniennes. Des cristaux de pyrite isolés ou en filonnets sont fréquents dans les déblais de tous les forages atteignant la roche saine.

Les forages (soigneusement suivis) ont été disposés en trois sites distants de 300 à 400m l'un de l'autre, dans des positions morphologiques contrastées : haut, milieu et bas de versant.

L'ensemble du dispositif est constitué de huit forages de 140mm de diamètre, de 35 à 55m de profondeur, à parois nues, avec des débits instantanés compris entre 4 et 24m³.h⁻¹, isolés des altérites par cimentation, et de dix forages de petit diamètre entièrement tubés (diamètre intérieur de 57mm) dans les altérites, de 10 à 22m de profondeur, isolés du sol par cimentation. Tous ces ouvrages sont distants entre eux de 3 à 60m au plus dans un même site. Ils viennent s'ajouter à deux puits, un forage, 25 drains (0,5 à 1m dans le sol) et une source, préexistant dans le périmètre ou à proximité immédiate, ainsi qu'à 10 sondages à la tarière, de 0,6 à 2,3m de profondeur. Au total, 57 points d'observation sont disponibles sur un tiers de km². Dans la mesure du possible, forages profonds, peu profonds et

sondages à la tarière ont été groupés en triplets. L'équipement du périmètre en deux campagnes de foration (août 1995 et juin 1996) a permis de distribuer au mieux les ouvrages suivant les situations naturelles rencontrées dans chacun des sites.

Sur les 364m cumulés des 8 forages les plus profonds, 38 arrivées d'eau, que l'on peut assimiler à autant de fissures ou groupes de fissures, ont été observées.

Schématiquement, de haut en bas, il est possible de distinguer des formations superficielles (sols, colluvions, alluvions, altérites) plus ou moins argileuses, avec des transmissivités comprises dans un intervalle de 1.10^{-6} à $1.10^{-4} \text{m}^2 \cdot \text{s}^{-1}$ (10 valeurs), d'environ 10 à 15m d'épaisseurs cumulées, qui surmontent la roche saine plus ou moins fissurée. Pour celle-ci, la transmissivité et le coefficient d'emménagement sont respectivement de l'ordre de 1.10^{-5} à $1.10^{-3} \text{m}^2 \cdot \text{s}^{-1}$ et de 1.10^{-4} à 1.10^{-3} (7 valeurs). L'étendue de ces intervalles montre bien l'hétérogénéité de ces milieux. L'ensemble est assimilable à un système bicouche drain-capacité, fortement anisotrope et hétérogène, avec une connectivité très variable. Les débits instantanés maxima des 8 forages profonds sont compris entre 4 et $25 \text{m}^3 \cdot \text{h}^{-1}$. Cinq d'entre eux ont des débits spécifiques de 0,5 à $1,0 \text{m}^3 \cdot \text{h}^{-1} \cdot \text{m}^{-1}$. La surface piézométrique est située entre +1 à -10m par rapport au sol, suivant le lieu et la saison.

Chaque forage est équipé d'une mini-pompe à échantillonner en plastique, à poste fixe, immergée sous 5 à 12m d'eau. Les prélèvements pour analyse sont faits après 0,05 à $0,1 \text{m}^3$ de purge. Ils correspondent donc pour l'essentiel à l'état, à un moment donné, de la colonne d'eau du forage à proximité de la pompe d'échantillonnage. Des prélèvements d'eau et des mesures dans les forages ont

également été faits à l'avancement en foration, lors de pompages globaux de longues durées à forts débits, ou encore, sélectivement à l'aide d'un obturateur double en PVC, à manchettes d'un mètre, léger et peu coûteux, opérationnel jusqu'à une soixantaine de mètres, conçu pour permettre à la fois des prélèvements et des essais hydrauliques. Des diagraphies de température et d'électroconductivité (EC) des colonnes d'eau des forages ont été faites jusqu'à -45m à l'aide du T-L-C meter de YSI, modèle 3000.

2.2. Résultats et interprétation

2.2.1. Chroniques

Dans le PEK, l'eau analysée lors des chroniques est caractérisée principalement par la présence ou l'absence de nitrates, quelle que soit la saison. Un seul forage échappe à cette alternative, le forage F31 (34,5m de profondeur par rapport au sol), situé dans le site en bas de versant, qui présente successivement l'une ou l'autre situation. Cette particularité probablement assez rare (ici, dans un, peut-être deux forages profonds sur huit), a servi de révélateur à un phénomène qui se manifeste ailleurs plus discrètement mais non sans conséquences fortes.

Les réactions de F31 ont été comparées à celles du piézomètre Pz2 (12m de profondeur par rapport au sol), distant de cinq mètres. Par quatre fois en 31 mois, surtout en fin de période estivale, les concentrations en nitrates des prélèvements en F31 ont diminué alors que celles en sulfate augmentaient, tandis que Pz2 ne présentait pas de variations de cette nature (Fig. 1). Systématiquement, la charge hydraulique du forage devenait supérieure à celle du piézomètre.

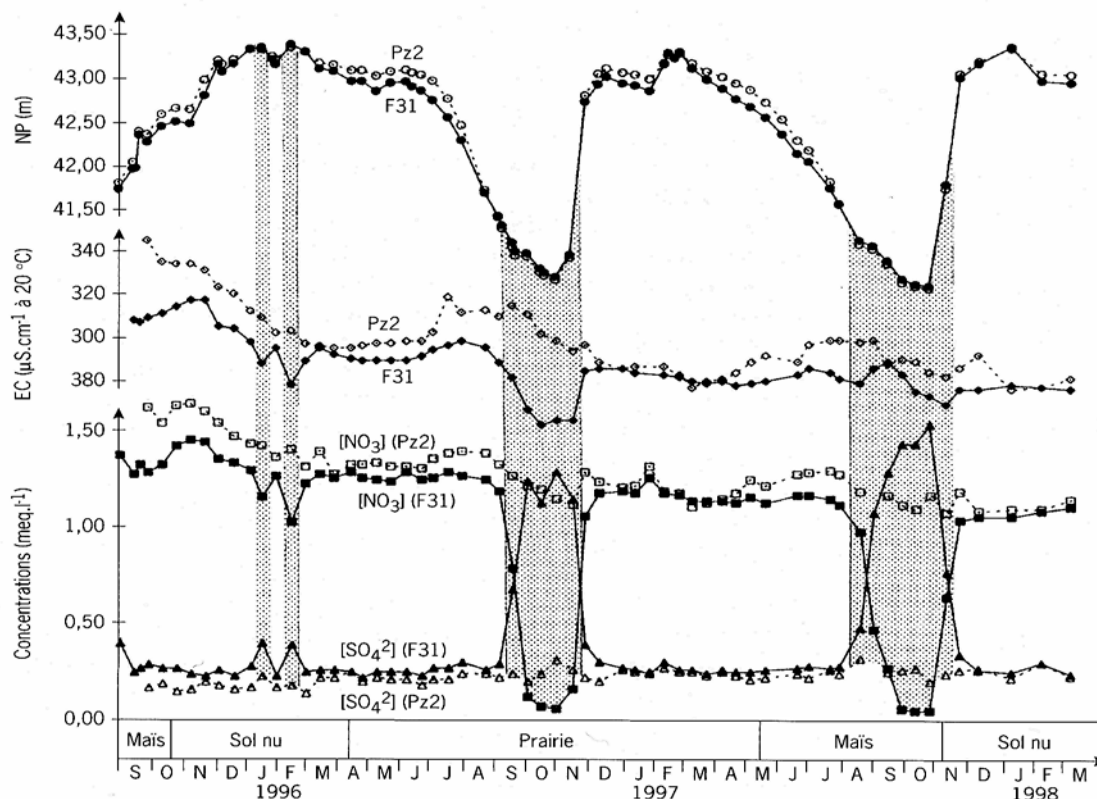


Fig. 1 – Chronique de la piézométrie, de la conductivité et des concentrations en NO_3^- et en SO_4^{2-} du piézomètre Pz2 et du forage F31, de septembre 1996 à mars 1998 (NP : Niveau Piézométrique).

2.2.2. Diagrammes de la colonne d'eau des forages profonds

La réalisation de diagrammes de la colonne d'eau de ce même forage F31 à divers moments permet de compléter ce qui précède concernant le fonctionnement complexe des forages et l'influence que cela peut avoir sur le chimisme des eaux. Ainsi, au repos, la charge hydraulique du piézomètre étant supérieure à celle du forage, F31 est rempli d'eau nitrée d'une conductivité de 280 à 320 $\mu\text{S} \cdot \text{cm}^{-1}$ à 20°C ($\text{NO}_3 = 60$ à 90 $\text{mg} \cdot \text{l}^{-1}$). Par contre, en étiage, hors pompage, lorsque le forage a une charge supérieure à celle du piézomètre, la conductivité de la colonne d'eau est plus faible de 20 à 40 $\mu\text{S} \cdot \text{cm}^{-1}$, correspondant à des prélèvements d'eau contenant moins de 10 $\text{mg} \cdot \text{l}^{-1}$ de NO_3^- .

2.2.3. Hypothèses sur la modification du chimisme de la colonne d'eau des forages profonds

Deux hypothèses peuvent être avancées pour expliquer les observations faites dans le forage F31 :

Dans la première hypothèse, le chimisme de la colonne d'eau de chaque forage serait l'image de celui de la nappe, et pour F31, son évolution traduirait alors celle de l'ensemble de la nappe sur une épaisseur au moins égale à la profondeur du forage, avec des apports successifs d'eaux nitrées ou non nitrées, suivant l'époque de l'année. Cela supposerait le renouvellement de la nappe sur plusieurs dizaines de mètres d'épaisseur plusieurs fois dans l'année, en quelques jours, ce qui est difficilement imaginable dans un milieu uniquement fissuré, quelles que soient les particularités hydrodynamiques ou hydrologiques que l'on pourrait invoquer pour l'aquifère.

Dans la deuxième hypothèse, si l'on assimile le comportement hydrodynamique et hydrochimique du piézomètre Pz2 à celui des fissures les moins profondes du forage F31, il devient alors possible d'expliquer ce qui se passe dans F31 par des échanges d'eau entre des fissures ou des groupes de fissures plus (30m) ou moins (10m) profonds, dans un sens ascendant ou descendant. Cela permet un renouvellement relativement rapide de la colonne d'eau dans une partie du forage, sans qu'il y ait de modifications notables dans la nappe. Dans tous les cas, c'est la variabilité spatiale et temporelle des charges hydrauliques dans les fissures qui contrôle ces mouvements.

2.2.4. Vérification de l'existence d'écoulements internes aux forages profonds

L'isolement successif des fissures du forage F31 à l'aide d'un obturateur double montre que des pompages à faible débit (inférieurs à 2 $\text{m}^3 \cdot \text{h}^{-1}$) permettent d'obtenir de l'eau de plus en plus réductrice avec des concentrations de plus en plus faibles en nitrates à partir des fissures les plus profondes. Mais les faibles volumes d'eau prélevés ne permettent pas une aussi bonne représentativité qu'il serait souhaitable, sans un assez long pompage, le volume de nappe contaminé par un autre niveau pouvant être relativement grand.

Une autre approche a consisté à placer un obturateur à 23-24m de profondeur dans le forage F31 au repos et à prélever des échantillons d'eau tous les deux ou trois jours pendant un mois, dans chacune des deux sections ainsi isolées (Fig. 2). Au départ, la colonne d'eau du forage était entièrement nitrée, tout comme le piézomètre Pz2. La section supérieure présentait une charge hydraulique plus forte de 2 à 10cm que celle de la section inférieure. Si la partie supérieure du forage est restée dans cet état et montrait un chimisme identique à celui du piézomètre Pz2,

par contre, les nitrates et l'oxygène dissous disparaissaient de la partie inférieure en quelques jours, par suite de l'écoulement de la nappe et/ou par dénitrification à proximité du forage des apports nitrés superficiels, mélangés à l'eau réductrice de la partie profonde de la nappe. Simultanément, les concentrations en sulfates et en fer augmentaient.

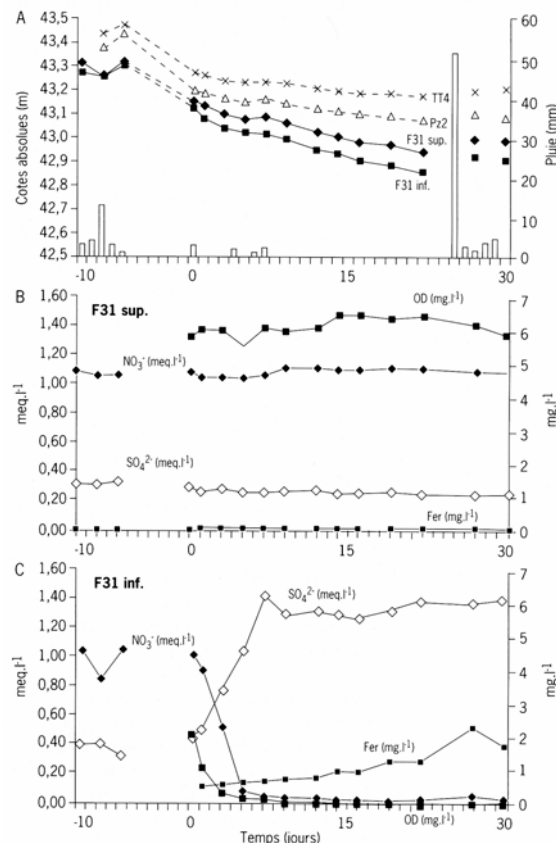


Fig.2 – Suivi de l'évolution du forage F31 après la mise en place d'un obturateur entre le groupe de fissures les moins profondes (10m) et les plus profondes (30m). A : piézométrie, B : Partie supérieure de la colonne d'eau, C : Partie inférieure. Pz : piézomètre, TT : Sondage à la tarière. OD : Oxygène Dissous.

Ainsi, il a été possible de vérifier qu'il y avait bien un écoulement descendant dans ce forage, à cette époque de l'année, d'un débit de 24 $\text{l} \cdot \text{h}^{-1}$ (le déplacement de l'interface entre les deux types d'eau représente le débit d'échange entre les niveaux superficiels et profonds par unité de temps, après l'enlèvement de l'obturateur). Cela confirme bien l'existence de deux catégories d'arrivées d'eau, avec ou sans nitrates et avec des charges hydrauliques différentes. Des mouvements d'eau en conditions non influencées ont également été observés ailleurs en milieu fissuré, à l'aide de courantomètres à haute résolution, par PAILLET (1994 & 1998), et aussi, dans des roches meubles hétérogènes, sables et graviers par CHURCH & GRANATO (1996).

Il est à noter que les colonnes d'eau des trois forages profonds du site à mi-versant et celui du site en haut de versant ne contiennent que des eaux dénitrifiées (mouvement ascendant dans la colonne d'eau), alors que le contraire s'observe dans deux forages sur trois au niveau du site en bas de versant (mouvement descendant), cela sans variations saisonnières perceptibles du sens d'écoulement. Cependant, il s'agit fondamentalement du même phénomène qu'en F31.

3. Contrôles sur modèle semblable (pilote)

3.1. Intérêt du pilote

Un pilote "Milieu Fissuré", installé dans les ateliers d'HydroInvest à Angoulême, a permis de reconstituer des échanges inter-fissures au repos, tels qu'ils sont provoqués par l'implantation d'un forage, et la stratification de la colonne d'eau résultant de cette circulation. Sa conception - éléments en plastique simplement emboîtés dans des manchons de raccordement, différents types de fissures disponibles, rails permettant d'agir sur la charge de 4 fissures (utilisables en même temps) - permet de simuler de nombreuses configurations. L'utilisation d'une centrale d'acquisition DATATAKER et du logiciel DELOGGER PRO autorise l'enregistrement et la visualisation des données en temps réel.

3.2. Résultats déjà obtenus

Une première série de tests a été réalisée sur ce pilote (HYDROINVEST, 2001). Elle concerne principalement les effets que produit sur la distribution des concentrations dans un forage, l'efficacité des préleveurs couramment utilisés et surtout, la mise en échange de 2 fissures. C'est ce dernier cas qui nous intéresse ici, puisque cela correspond à un fonctionnement observé en milieu naturel.

3.3. Comparaison pilote-milieu naturel

Il a été établi un écoulement ascendant entre 2 fissures du pilote séparées d'un mètre, au débit de $9,6 \text{ l} \cdot \text{min}^{-1}$, ce qui donne des vitesses dans le tube et dans les fissures de respectivement $2,1$ et $32 \text{ cm} \cdot \text{s}^{-1}$. La conductivité électrique EC (non standardisée) est de $550 \text{ } \square \text{ S} \cdot \text{cm}^{-1}$ à $22,2^\circ$ dans le tube et de 737 à $22,0^\circ$ pour le traceur. Les profils de conductivité ont la même allure en ce qui concerne les épisodes d'écoulement ascendant dans le forage F31. Cela est d'autant plus remarquable que des différences notables existent entre modèle et forage en ce qui concerne les différences de conductivité entre fissures (172 et 30 à $50 \text{ } \square \text{ S} \cdot \text{cm}^{-1}$ respectivement), les distances entre fissures (1 et 15 m respectivement), et que les sens d'écoulement étaient inversés. En ce qui concerne les débits, les différences sont d'un facteur de 10 et pour les vitesses, de 10 à 100 en "faveur" du pilote.

4. Discussion et conclusion

Des observations et des mesures ont été faites pendant plus de trois années en différents points d'un versant schisto-gréseux, à des profondeurs variables, lors de suivis ou d'opérations ponctuelles. Il a été possible d'y mettre en évidence la superposition d'une zone oxydante superficielle et peu épaisse (moins de $30\text{-}40 \text{ m}$) contenant des nitrates, et d'une zone réductrice, où la matière organique, les sulfures et le fer ferreux sont oxydés par l'oxygène des nitrates, qui sont alors détruits. Un grand nombre d'indices concordants permet de vérifier l'intervention de la dénitrification, principalement : absence de nitrates et d'oxygène dissous, mais abondance de sulfates, de fer et de gaz, de bactéries, Eh faibles, etc...

Cette structure bicouche, identifiable par un suivi soigneux en cours de foration, sera masquée en cours de pompage d'essai par des mélanges d'eau provenant de diverses fissures, et le reste du temps, par des écoulements internes aux forages, responsables d'échanges entre hétérogénéités. Ceci se produit dans tous les forages profonds (8) du PEK, avec des effets spectaculaires sur l'un d'eux, le F31, ce qui explique son utilisation dans cette communication. Ces phénomènes montrent la nécessité

d'une étude sélective des fissures rencontrées par les forages, par exemple à l'aide de dispositifs d'obturation mobiles ou fixes, suivant les problèmes à traiter.

L'utilisation d'un pilote de milieu fissuré a permis la reconstitution de situations observées sur le terrain et des tests de matériels. Il faut souligner que l'utilisation d'un pilote n'a de sens que couplé à des études in situ.

La généralisation de ces biais à l'ensemble des milieux fissurés magmatiques, métamorphiques ou sédimentaires, en particulier carbonatés, dans lesquels la karstification agit comme un amplificateur de l'hétérogénéité et où des échanges entre chenaux sont observables (DROGUE, 1985), semble tout à fait licite. Une approche du même genre que celle qui a été faite dans le périmètre expérimental de Kerveldréac'h, réalisée dans le bassin versant de la Noé Sèche (Côtes d'Armor), permet de voir qu'elle s'applique aussi aux granites (MARJOLET, FAILLAT & SICARD, 2000).

Il faut insister sur le fait que la complexité d'organisation et de fonctionnement des hydrogéosystèmes fissurés et celle des processus hydrogéochimiques mis en jeu, ne permettra de progrès dans leur compréhension et leur modélisation (pérennité du phénomène de dénitrification liée à la présence de corps réduits et aux types de processus mis en jeu, quantification des flux, etc...) que par l'utilisation de dispositifs expérimentaux, de techniques et de méthodes appropriés au milieu fissuré, où l'obtention de la continuité spatiale des observations et de la représentativité des échantillons d'eau constitue une difficulté majeure.

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Study of spring hydrographs of a karst plateau in Southeast Brazil using time series analysis

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Abstract

Lajeado-Bombas is a polygonal karst plateau with 38 km² area, covered by tropical rain forest. Creeks and overland flow coming from the surrounding non-carbonatic rocks promote the allogenic recharge of the karst. Water-tracing tests delineated recharge areas of plateau springs revealing the presence of 3 main subterranean systems. During the period of realization of the tests, hydrometric stations were installed to measure flow rates in springs (Lago and Areias) from two of these systems, and a rain-gauge data logger was positioned near plateau center. Spectral and correlation analyses were performed to obtain global characteristics of the flow signal and rainfall-runoff relationships. Wavelet analysis was employed to understand the runoff variability at different scales resolution. The time series analysis exposed karst springs with contrasting behavior. Areias, a fullflow spring, presents characteristics of a well-drained system, while Lago spring presents an inertial character because of the distributary pattern of its system and possibly by the occurrence of constrictions reservoirs. In Areias system, the value obtained with rainfall-runoff relation is very similar to the value of time to peak concentration obtained in a quantitative tracer test during a rainfall event. This result indicate that in well drained systems the cross correlation function is a good estimator of particle travel times through the aquifer.

Résumé

Lajeado-Bombas est un plateau karstique (un véritable karst polygonal) avec 38 km², couvert par une forêt tropicale humide. Les ruisseaux et les écoulements de surface provenant des roches non-carbonatées de l'entourage promeuvent le rechargement allogénique du karst. Essais avec des traceurs pour identifier les zones de recharge des sources du plateau ont révélé la présence de 3 systèmes souterrains. Pendant la période de réalisation des essais, stations hydrométriques pour mesurer les débits ont été installées dans les sources de deux de ces systèmes (Lago et Areias), et un pluviographe a été positionné proche au centre du plateau. Les analyses spectrale et corrélatrice ont été exécutées pour obtenir les caractéristiques globales du signal du flux et les relations pluie-débit. L'analyse des ondelettes a été utilisée pour la compréhension de la variabilité du débit dans des différentes échelles de résolution. L'analyse des séries temporelles a exposé sources karstiques avec des comportements contrastants. Areias présente caractéristiques d'un système bien drainé, tandis que la source du Lago présente une caractéristique inertielle due au modèle de distribution de son système et possiblement par l'occurrence de réservoirs de constriction. Dans le système Areias, la valeur obtenue pour la relation pluie-débit est très proche de la valeur du temps pour la concentration maximale obtenue dans un essai quantitatif avec traceur lors d'un événement de pluie. Ce résultat indique que, dans le système bien drainé, la fonction de corrélation croisée est un bon estimateur des temps de voyage de particules à travers d'aquifère.

1. Introduction

The main karst landscapes of São Paulo State, Brazil, are in southern Atlantic rain forest mountains. The region is situated into a transitional climatic region between the hot climate of low latitudes and the mesothermic temperate climate of middle latitudes. The climate is classified as warm and wet with no dry season. The mean annual temperature usually lies between 20° C and 22° C and the mean annual rainfall is 1575 mm, which falls mainly between October to March (1062 mm).

Several karst plateaus with polygonal network of interfluvies separating closed depressions are present in the region. These plateaus lie in a Proterozoic Fold Belt divided into several tectonic blocks separated by NE-SW transcurrent shear zones (CAMPANHA 1991). The Lajeado-Bombas (Fig.1) is one of them, with 38 km² of karst surface set between non karstifiable rocks. The tectonic block occupied by the plateau is composed of low metamorphic grade pelitic, psammitic and carbonatic units overlain by a gabbro intrusion.

In Lajeado-Bombas the carbonatic surface is lower (300-500 m) than surrounding crests of psammitic rocks (700-850 m). The contact zone between karst and non carbonatic rocks is characterized by the presence of sinkholes and blind valleys that drain waters from phyllites and metasilstones. Perennial and ephemeral streams and

overland flow feed this allogenic recharge zone; thirty allogenic basins whose areas vary from 0.4 to 5.5 km² direct waters to this place. The autogenic recharge occurs in the polygonal zone, with more than 300 depressions (0.04 to 2 km²) and prevalence of intermittent and ephemeral streams (KARMANN 1994; KARMANN & FERRARI 2002). Two discharge zones drain the plateau, the Bombas and Betari valleys.

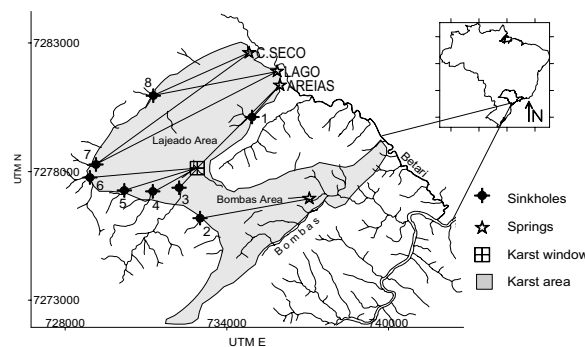


Fig 1. The Lajeado-Bombas karst plateau. The lines are known hydraulic connections identified by water tracing tests. Corrego Seco and Lago Springs drain Corrego Fundo system; Areias spring is the sole spring of Areias System.

Water-tracing tests to delineate the recharge areas of plateau springs, revealed the presence of three main subterranean systems Fig. 1 (GENTHNER et al. 2003). The Areias Cave (5020 m) and Areias spring cave (1168 m) are part of Areias system. Caves with 1357 m and 580 m length, with important vertical development, are inputs for Corrego Fundo system.

A perennial spring with an average flow rate of $0.64 \text{ m}^3/\text{s}$ drains Areias system. The Corrego Fundo system is drained by a perennial and a temporary spring, respectively Lago (average flow rate of $0.28 \text{ m}^3/\text{s}$ - never exceeds $0.45 \text{ m}^3/\text{s}$) and Corrego Seco (discharge rate can increase from 0 to more than $3 \text{ m}^3/\text{s}$ in less than 1 h and back to 0 in few days). The Corrego Seco spring works like a spillway for the system during heavy summer rainfall episodes. Corrego Seco and Areias springs are positioned in opposed flanks of a synclinal structure, near the contact with non-karstifiable beds; the Lago spring is between them, closer to Areias (less than 700 m). This paper focuses on the hydrologic behavior of Areias and Lago springs using monitoring data obtained during tracer tests.

One sector of the karst plateau is inserted in a conservation area (PETAR State Park). The results presented in this paper are part of hydrologic studies carried out to define strategies to protect caves and karst springs.

2. Material and methods

The mechanisms of genesis and development of karstic aquifers lead to a highly discontinuous and heterogeneous medium, not only in its spatial structure but also in its temporal functioning (Mangin 1984, Labat et al. 2000). These characteristics impose limitations to the use of classical hydrogeological techniques because they own a restricted investigation scale. On the other hand, in karst aquifers, the flow of the catchment area is often concentrated in one single spring and this characteristic has stimulated the use of the system analysis to understand their dynamics.

In regions with scarcity or without wells (like Lajeado Bombas), the system analysis based in spring flow time series can be used with data derived from tracer tests to improve the knowledge about aquifer properties. This approach is opportune because uses easily available data which are relatively inexpensive to collect.

The quantitative system approach of karst aquifers based on time series was established by MANGIN (1984) and has been employed in some studies (PADILLA & BOSCH 1995; LAROCQUE et al. 1998; MATHEVET et al. 2004; ZARE et al. 2005). The principle of this technique is to consider the aquifer like a filter which transforms the input signal into an output signal. The degree of transformation of the input signal therefore provides information about the nature of flow in the system. In this approach, it is assumed that spring hydrograph provides an integrated representation of the network of stores and passages delivering water to aquifer outflow point (LAROCQUE et al. 1998).

Based on this idea the spring response has been used to determine if the system is well drained (well karstified) or poorly drained (poorly karstified-fissured systems) MANGIN (1994). However, numerical modeling (Halihan & Wicks 1998) has shown that conduit systems with multiple outlets, or localized aspects of the network of conduits (reservoirs generated by constrictions in conduits), can generate similar responses to those attributed to poorly karstified systems.

In this sense, the time series analyses show how much a system alter an input signal, but cannot explain the exact cause of this alteration. Despite the limitation of the method, its integration with others data leads to better inferences about the system functioning. Thus, results from tracer tests, the knowledge about the existence of active caves passages and the recharge mode of the system, must be considered to interpret spring responses.

The autocorrelation and spectral analysis furnish characteristics about the system hydrodynamics, computing the influence time of input pulses. As autocorrelation and spectral analysis operate over the global signal, it is not possible to observe the temporal variability at different scales resolution and the relations between different scale processes. This limitation can be overcome using the Morlet wavelet spectrum. This tool gives interesting information on the variability of the processes at different scales resolution in a very concise and compact form (LABAT et al. 2000).

The cross-correlation function (CCF) is used to analyze the relations between input and output time series. The maximum CCF value is an estimate of spring response time to input pulses. The delays between inputs and outputs are useful in environmental studies because give an estimation of the particle travel times through the aquifer (LAROCQUE et al. 1998).

These methods were used to explore one-year flow rate time series obtained in hydrometric stations installed at Areias and Lago springs (Fig. 2). The flow rate was obtained with 10 minutes resolution. A rain-gauge data logger positioned near the plateau center measured rain pulses.

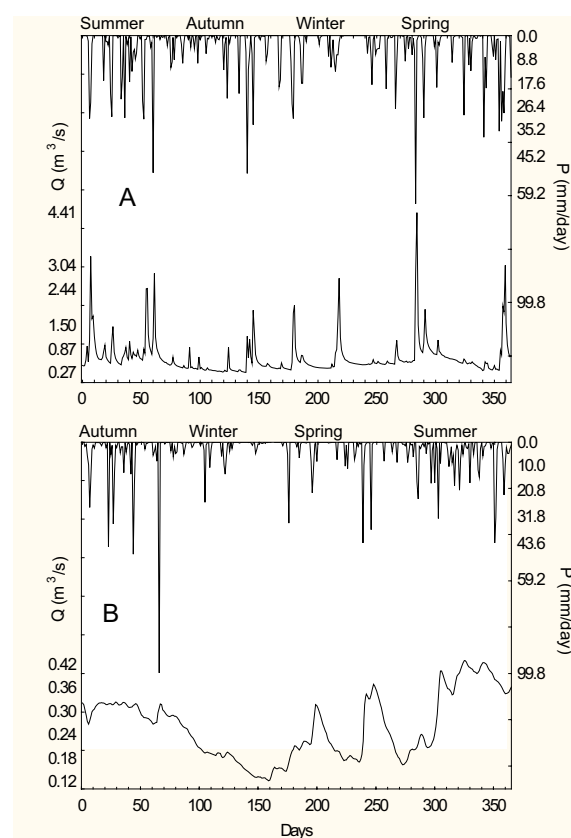


Fig 2. Daily rainfall and outflow rates for Areias (A) and Lago (B) springs.

The Areias spring was monitored during the hydrologic year 2000-2001; the total rain value in this period was 1505 mm. The Lago spring was monitored during hydrologic year 2002-2003, with total rain value of 1535 mm. Spectral, autocorrelation and wavelet analysis were performed using mean daily data, the cross correlation between rainfall and flow rate was based in mean hourly data. More detailed explanations about the used methods can be found in references cited at this item.

3. Results

The autocorrelation and spectral analysis (Fig. 3), represent the behavior of studied springs based in one year flow rate. From the shape of autocorrelations it can be noted that flow rate at Lago diminish very slowly when compared with Areias. The memory effect is 3.5 and 84.94 days for Areias and Lago respectively. From the spectral density graphics it is possible to see that cutoff frequency is very low for Lago when compared to Areias. The regulation time for Lago is 69.96 days and for Areias, 5.3 days.

Although the data have been obtained in distinct hydrologic years, the annual rainfall volumes and distribution patterns are very similar, thus, the contrasting values between the two springs may reflect the systems structures.

Part of the inertial behavior of Lago spring can be explained because Corrego Fundo is a distributary system. However, Corrego Seco is an overflow spring that is not activated with single rain pulses; conversely, it becomes active only after heavy summer storms or after several rainy days (reports from local habitants). This functioning is indicative of the existence of a considerable storage capacity in Corrego Fundo system.

The low values observed for Areias spring, points to a well-drained system that does not offer resistance to the flow. The major part of the concentrated allogenic flow received by the plateau is injected into this system and greatest known cave is part of it.

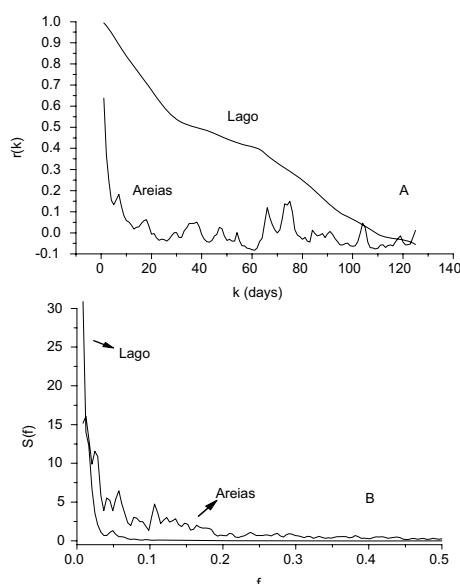


Fig 3. Flow rates time series: (A) auto-correlation functions; and (B) spectral density functions.

The Morlet wavelet spectrums for flow rate data are presented in fig. 4. A grey scale represents wavelet coefficients; the lighter greys indicate lower values. The

wavelet for Areias is characterized by the presence of discontinuous 2-8 days periodic components with variable intensities. When these components own high intensity, energy propagation to greater scales is observed. This behavior is not present at Lago spectrum, where the smallest frequency structures (4-8 days) are homogeneous with respect to intensities and more continuous. In both spectrums can be observed near the 64 days period, the existence of a high intensity component linked to the rainy seasons (spring - summer).

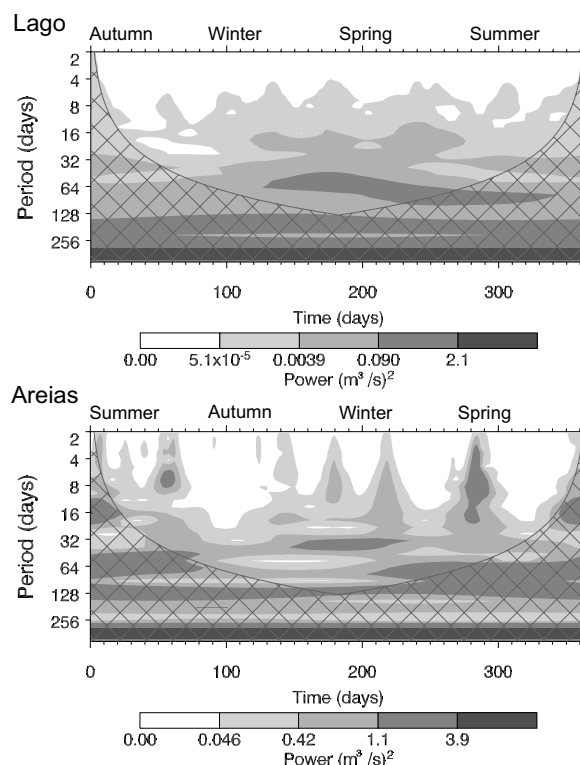


Fig. 4. Flow rates wavelet spectrum.

The cross-correlation function (CCF) analysis used rainfall time series as input data and spring flow rates time series as output data (fig. 5). The maximum CCF occurs 10 and 53 hours after a rain event for the Areias and Lago springs respectively. The decrease of the CCF is more accentuated in Areias than in Lago and the maximum value of the CCF in the last one is very small.

The time for the maximum CCF value obtained in Areias System (10 h), is very near to the value of time to peak concentration obtained in a quantitative tracer test during a rainfall event (8.99 h – GENTHNER 2000). In this test, the tracer (300 g of rhodamine) was injected in a swallow hole located in the SW border of karst plateau (n° 4 – Fig. 1). In an underflow state, the time to peak concentration in this same route takes 18 h. These data indicate that CCF delay can be used to estimate particle travel time through this well-drained system.

The time for the maximum CCF value for Lago spring (53 h) shows some association with tracer tests results. Two quantitative tracer tests were realized in Corrego Fundo system; the first one in an underflow situation (250 g of rhodamine were injected) and the second during an overflow state (Corrego Seco was activated – 300 g of rhodamine were injected). In both tests, the tracer was undetectable after 60 h of monitoring. The tracer passage was proved by charcoal detectors collected three weeks

after the injection. In a straight line, the distances between the injection point (n° 8 – Fig. 1) and the sources (Corrego Seco and Lago), are minors than the one observed in Areias tracer test. From these data it is possible to admit the existence of a feature in the system, which is responsible to dampen discharge peaks and to dilute tracer plumes.

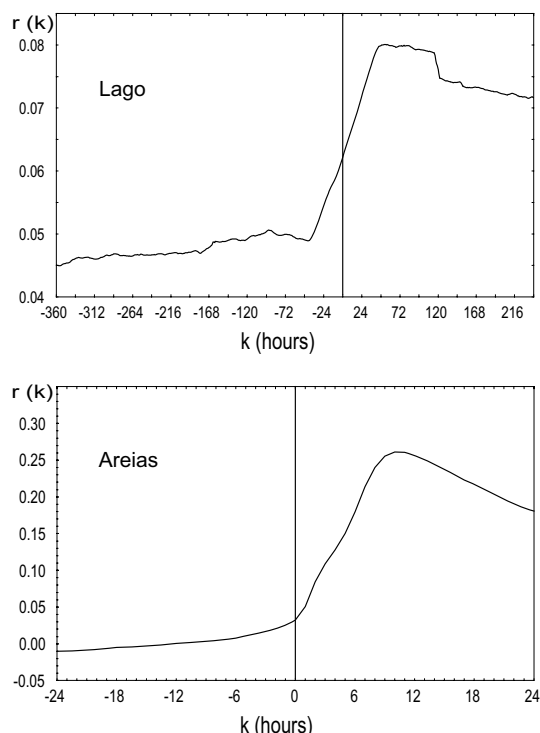


Fig. 5. Cross-correlation functions. Hourly rainfall as input and spring flow rate as output.

4. Conclusion

The time series analysis showed karst springs with contrasting behavior. Areias is a fullflow spring, so, in principle, its time series may translate how the entire system transforms an input signal. In this manner, considering the low values for memory effect and regulation time, and the presence of high frequency structures in the wavelet spectrum, this system does not promote significant transformations in input signals. The presence of important cave passages in this system is indicative of well-developed routes that offer few restrictions to the flow. The proximity between the delay obtained with CCF and the time to peak concentration obtained in a quantitative tracer test, indicates that CCF analysis is a good estimator of travel time of solutes in this well-drained system.

Lago is a baseflow spring, so its time series cannot be used to interpret the whole system. However, the inertial behavior pointed out by the high memory effect and regulation time values may not be explained by this sole characteristic. The dynamics of the overflow spring (Corrego Seco) and the results from quantitative tracer tests indicate the existence of a feature in the system that dampen discharge peaks, and dilute tracer plumes. Taking into account that the caves present in this system indicate the existence of well-developed conduit flow, and that the concentrated recharge is the principal input to the system, part of this inertial characteristic should be attributed to reservoirs generated by constrictions in conduits.

Acknowledgements

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Characterization of karst aquifer vulnerability by means of univariate clustering on electrical conductivity, turbidity and discharge datasets

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Abstract

Time series analyses have proven to be useful in improving understanding of karst systems. However, they cannot compare the variations between electrical conductivity, turbidity and water discharge, which allow us to identify the transport properties within the karst system. This talk presents an univariate clustering applied to these variables to identify the transport properties on the annual hydrologic cycle and their occurrence periods. It was used to a spring of a chalk karst aquifer located in western Paris basin during a part of the 2002 hydrologic cycle (January to September 2002). The results show that the annual volume of spring water is decomposed of 47.7% of ground water, 23.7% of flood water by direct transfer of surface water, 13.2% of spring water flowing during deposition periods of suspended particulate matter and 15.4% of spring water flowing together with resuspension of intrakarstic sediments. The release of ground water occurs during low water and the release of flood water occurs during recharge and storm events in summer time. The release of spring water resulting from deposition occurs during weak rain event intensity from March at June. The release of spring water resulting from resuspension occurs from April to September.

1. Introduction

Measurements of electrical conductivity and discharge have been used extensively to investigate the hydrodynamics of karst systems. Although the conductivity is not directly equivalent to the concentrations of the major ionic species, it has the advantage of being inexpensive and simple to measure, and time series at a very high resolution can be easily obtained. Conductivity is a global parameter that reflects the amount of total dissolved solids TDS in the water, which is mainly controlled by calcium-carbonate equilibrium in karst systems. Consequently, its variation in spring discharge reflects the varying contributions of different masses of water in the system with different types of flow conditions. Moreover, turbidity is used to investigate the transport of particulate matter within karst system and many authors (BAKALOWICZ, 1977, 1979 ; MAQSUD, 1996 ; PLAGNES, 1997 ; MAHLER & LYNCH, 1999 ; MAHLER *et al.* 1999 ; MASSEI, 2001 ; LABAT *et al.*, 2002 ; MASSEI *et al.*, 2002a,b, 2003 ; GRASSO *et al.*, 2003 ; DUSSART-BAPTISTA, 2003) studied the relations between electrical conductivity (EC), turbidity (T) and water discharge (Q) since they may follow the transport properties within karst systems (direct transfer of ground or surface waters, deposition of suspended particulate matter, resuspension of intrakarstic sediments).

The karst hydrologist has at his disposal a set of methods to explore and study karst aquifers. Among these methods, there are the hydrodynamic methods like spring hydrographs, sorted discharge distribution of the annual hydrograph and time series analyses. Hydrodynamic methods are often used on the annual hydrograph of electrical conductivity, water discharge and even turbidity. Time series analyses have proven to be useful in improving understanding of karst systems. However with time series analysis, we lose the time evolution, which have with hydrographs and we cannot compare the variations between some variables.

To answer these problems, we propose the use of univariate clustering with Fisher algorithm (FISHER, 1958) to optimally partition observations in homogeneous clusters based on their description using a single quantitative variable. Applied to electrical conductivity, turbidity and water discharge, this method allows us to define water types (weak, mean or strong values for each one of three variables). The comparison of their time variations allows the identification of the transport properties and the periods where they occur.

2. Materials and methods

2.1 study site and field equipment

The study site is a karst system, located in the Pays de Caux (Haute-Normandie, France) in western Paris Basin, near the small city of Norville, on northern bank of the river Seine, about 40 km upstream from the outlet of the Seine estuary (Figure 1).

The Norville karst system has been widely studied and its hydrological properties became quite well known (MASSEI, 2001 ; MASSEI *et al.*, 2002a,b, 2003 ; DUSSART-BAPTISTA, 2003 ; DUSSART-BAPTISTA *et al.*, 2003a,b). It is composed of a swallow hole and a spring.

The swallow hole is the point of infiltration of Bébec Creek, which runs on the plateau of the Pays de Caux, draining a small catchment area of about 10km². Land use in the catchment area consists of cultivation and grazing. Discharge in Bébec Creek is variable, from 3 l s⁻¹ in summer dry periods until a maximum of 500 l s⁻¹ in winter rainy period. During these periods, the water in Bébec Creek is very turbid, up to 1000 NTU.

The spring, located at the base of the plateau, at the place named 'Le Hannetot', constitutes an overflow of the saturated zone. The Hannetot spring is clearly identified as being the main outlet of waters swallowed in the swallow hole. The water discharge at the Hannetot spring is

composed of the ground water discharge and the surface water swallowed in the swallow hole. After storms, the turbidity of the water discharging from the spring can exceed 600 NTU.

Data from the two sites (swallow hole and spring) were collected continuously on a 15 min time interval and samples were collected during responses to rain events. Field equipment is constituted by multi-parameters 6820 YSI datasonds, each comprising one turbidity probe, one electrical conductivity probe, and one pressure sensor. An ISCO 674 rain gauge is settled on the watershed. Discharge was determined at Hannetot spring using ISCO 4150 Doppler flow meters to perform flow measurements every 15 min.

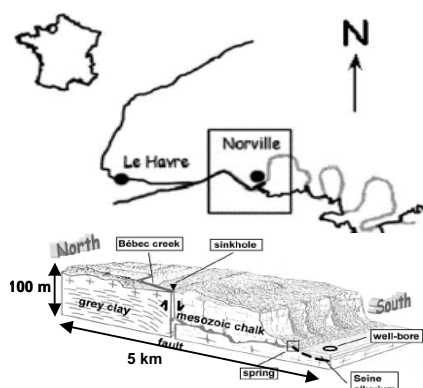


Figure 1 : study site location

2.2 frequency distribution decomposition (FDD)

BAKALOWICZ (1977, 1979) suggested that the different modalities of the conductivity frequency distribution (CFD) of karst spring discharge reflect the existence of geochemically distinct masses of water moving through the aquifer, which do not completely mix together due to the flow in conduits. The assumption is that the average conductivity of an individual water type depends on the spatial distribution of flow conditions. Bakalowicz used CFDs to classify the degree of karstification of different sites, associating the different modalities with the individual water types contributing to the spring, and suggesting that their number and spread represent the intensity of the karstification, or the karst conduit system efficiency. Other authors followed this qualitative approach to describe the karstification of other carbonate systems, for example the chalk of the northwest Paris Basin (MAQSOU, 1996) and the karst of the Grands Causses in central France (PLAGNES, 1997). However, the absence of a quantitative description prevents an effective comparison of different CFDs.

Here, we expand on the ideas proposed by BAKALOWICZ (1979) to show that the frequency distribution of EC and Q can be resolved quantitatively into the sum of two or more normally distributed populations. We propose that these populations represent distinct water types contributing to spring flow over a hydrologic cycle, and we investigate the degree to, which the number of water types.

For each EC and Q, a histogram was constructed with variable divided into 20 to 25 classes. The histograms are smoothed into a curve, and then decomposed into its normally-distributed component populations using a standard peak-fitting software program, such as PeakFit 4.0 (SPSS Inc.). Using the method of residues, all the modes constituting the original frequency distribution curve are

identified, including those not visible as peaks in the original curve.

The frequency distribution decomposition (FDD) defined the number and the modes of peaks. These peaks are interpreted as representing the normally-distributed population of conductivity or discharge of a single water type.

2.3 univariate clustering

We used univariate clustering, using XLStat (Addinsoft), to optimally partition observations in homogeneous clusters by minimization of intra-class inertia, based on their description using a single quantitative variable. The Fisher algorithm (FISHER, 1958) guaranteed that the obtained solution is the best. The results of FDD were used to define the number of classes for the univariate clustering of EC, T and Q datasets. For example, the FDD of water discharge (Fig. 2) and electrical conductivity (Fig. 3) show 6 classes. So, we realized an univariate clustering of Q dataset with 6 classes. Table 1 shows that the results obtained by the FDD are near of these obtained by univariate clustering with a coefficient variation minus to 5%.

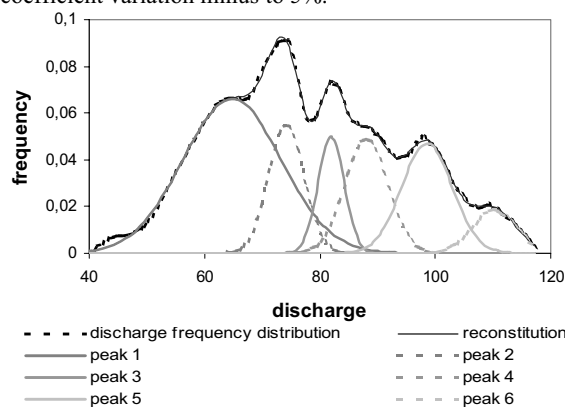


Figure 2 : FDD of water discharge

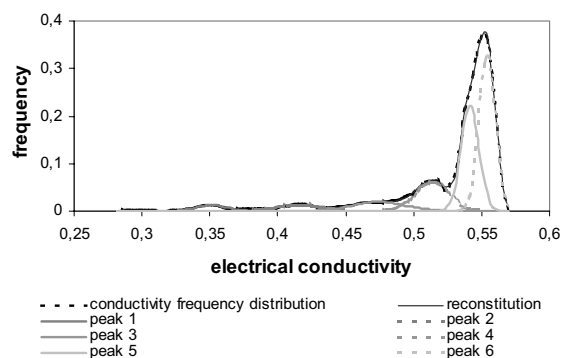


Figure 3 : FDD of electrical conductivity

	class	barycenter from univariate clustering	peak mode from FDD	CV
water discharge l_s^{-1}	1	63,98	64,42	0,7
	2	72,75	73,88	1,5
	3	80,29	81,76	1,8
	4	89,47	88,06	1,6
	5	101,90	98,30	3,7
	6	112,03	110,12	1,7
electrical conductivity $\mu S\ cm^{-1}$	1	345	352	2,0
	2	417	416	0,2
	3	469	474	1,1
	4	513	515	0,4
	5	545	541	0,7
	6	557	556	0,2

Table 1 : comparison of the results obtained by univariate clustering and FDD

3. Results

3.1 typology of water masses

In such binary karst, ground waters are more mineralized than surface waters. An EC decrease indicates the arrival of surface water. An increasing turbidity may indicate the arrival of surface water, but also a resuspension of intrakarstic sediments (MASSEI, 2001 ; MASSEI *et al.*, 2002a,b, 2003) (Tab. 2).

		ELECTRICAL CONDUCTIVITY	
		weak or mean values => surface water	strong values => ground water
T U R B I D I T Y	weak values	deposition of suspended matter	direct transfer of ground water
	mean or strong values	direct transfer of surface water	resuspension of intrakarstic sediments

Table 2 : identification of transport properties according to EC and T variations

The understanding of transport properties are here investigated using the relations between EC, turbidity and water discharge. The hydrologic processes corresponding to each water type are interpreted by comparing the range of EC, T and Q represented by each population. For each variables, three types of water masses are defined:

- weak values, which result from the association of the first and the second peaks obtained by the FDD for each variables,
- mean values, which result from the association of the third and the fourth peaks,
- and strong values, which result from the association of the fifth and the sixth peaks.

By combining these 3 types for the three variables, we obtained 27 theoretical combinations, which only 18 have an hydrological meaning according to their values for the three variables (Tab. 3).

EC	D	T	hydrological meaning
weak	weak	weak	none
weak	weak	mean	none
weak	weak	strong	none
weak	mean	weak	deposition of suspended matter
weak	mean	mean	direct transfer of surface water
weak	mean	strong	direct transfer of surface water
weak	strong	weak	none
weak	strong	mean	direct transfer of surface water
weak	strong	strong	direct transfer of surface water
mean	weak	weak	deposition of suspended matter
mean	weak	mean	direct transfer of surface water
mean	weak	strong	none
mean	mean	weak	direct transfer of surface water
mean	mean	mean	direct transfer of surface water
mean	mean	strong	direct transfer of surface water
mean	strong	weak	direct transfer of surface water
mean	strong	mean	direct transfer of surface water
mean	strong	strong	direct transfer of surface water
strong	weak	weak	direct transfer of ground water
strong	weak	mean	resuspension of intrakarstic sediments
strong	weak	strong	none
strong	mean	weak	direct transfer of ground water
strong	mean	mean	resuspension of intrakarstic sediments
strong	mean	strong	none
strong	strong	weak	direct transfer of ground water
strong	strong	mean	resuspension of intrakarstic sediments
strong	strong	strong	none

Table 3 : theoretical combinations and their hydrological meaning

So at any time, we can see the combination and identify the transport properties during a flood and during the hydrologic cycle (Fig. 4). Also, we can evaluate the proportion of the different water masses (Tab. 4).

transport properties	% of annual volume
deposition of suspended matter	13,2
direct transfer of surface water	23,7
direct transfer of ground water	47,7
resuspension of intrakarstic sediments	15,4
	100,0

Table 4 : % of annual volume of water masses

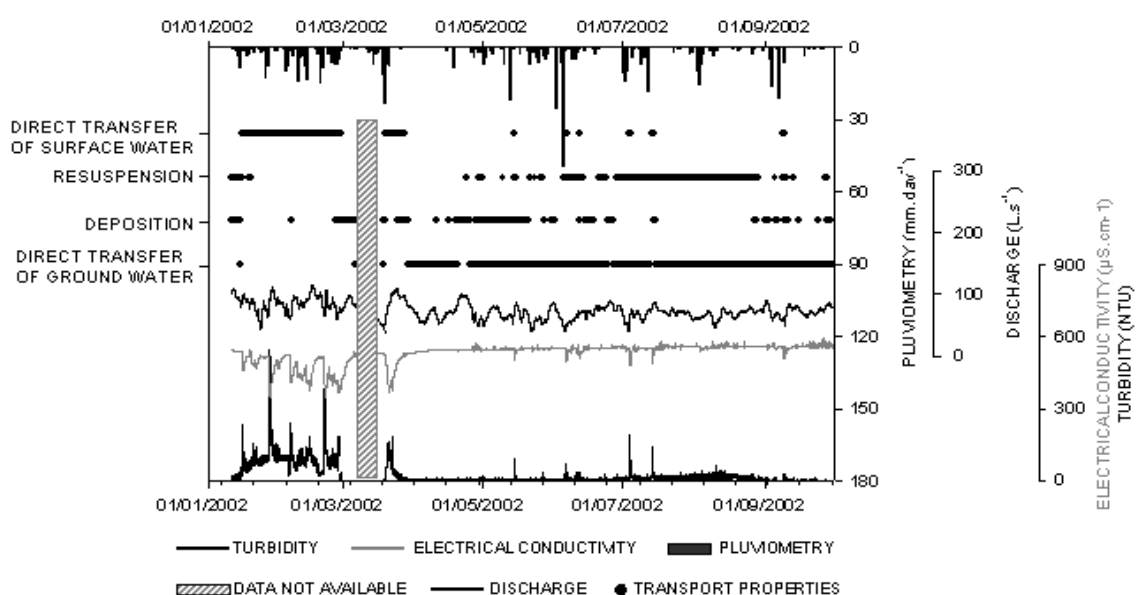


Figure 4 : Transport properties and their occurrence periods at the Hannetot spring during January to September 2002

4. Conclusion

To conclude, this method applied on high frequency time series of electrical conductivity, turbidity and water discharge, allows the identification of transport properties and their occurrence periods on the annual hydrologic cycle. The results show that the annual volume of spring water is decomposed of 47.7% of ground water, 23.7% of flood water by direct transfer of surface water, 13.2% of spring water flowing during deposition periods of suspended matter and 15.4% of spring water flowing together with resuspension of intrakarstic sediments. The release of ground water occurs during low water and the release of flood water occurs during recharge and storm events in summer time. The release of spring water resulting from deposition occurs during weak rain event intensity from March to June. The release of spring water resulting from resuspension occurs from April to September and correspond to vulnerability periods for drinking water supply like the direct transfer of surface water.

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Study of the transport properties of suspended matter by means of multivariate analysis

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Abstract

The aim of this study is to identify the hydrodynamic variables, which control the transport properties of particles within a karst aquifer of western Paris basin. The PSD of water samples swallowed at a swallow hole have been compared with the PSD released at a spring and a well. This comparison allow the identification of particles transport properties within this karst aquifer. Then, the granulometric characteristics of water samples at the swallow hole, spring and well, have been compared with the hydrodynamic variables of karst system (recharge and depletion of ground water, Seine river dynamics, human impact). The variables studied are numerous (46) and mixed (quantitative and qualitative). Under these conditions, it is difficult to identify the parameters, which influence the karst aquifer functioning. Thus, a multivariate analysis (Hill and Smith analysis, Hill and Smith, 1976), which allows the joint analysis of quantitative and qualitative variables, was used to identify the hydrodynamic parameters controlling the transport properties of suspended matter.

1. Introduction

The transport of particles in karstic media is a complex process implying deposition and release phenomena (RODET, 1991, 1993, 1997 ; COQUEREL *et al.*, 1993 ; LACROIX *et al.*, 1998, 2000, 2002 ; MAHLER & LYNCH, 1999 ; MASSEI 2001 ; MASSEI *et al.* 2002 a,b, 2003). Intrakarstic deposition results from a decrease in flow competence. The decrease in velocity allows the larger particles to decant. These particles may later be resuspended and transported by subsequent high competence flow events. Particles transported produce turbidity in the outlet of karst aquifer. So, the observed turbidity would have two potential origins: (i) the direct transfer of particles from the inlet to the outlet of the karstic system (allochthonous origin) and (ii) the resuspension of previously deposited sediments (sub-autochthonous origin).

The particle size distributions (PSD) of suspended particulate matter (SPM) may constitute a tracer of particulate transport, which is related to flow conditions within hydrological systems (ATTEIA & KOZEL, 1997, ATTEIA *et al.*, 1998). The hydrological functioning of aquifers may be complex in particular for the karst aquifers. This complexity is due to the significant number of parameters, which can affect the functioning of aquifers (boundary conditions, base level variation, pluviometry, human impact, interactions with a river or sea, ...). The aim of this paper is to identify the transport properties of SPM by means of study of the relationship between PSD curves and hydrodynamic variables. The PSD comparison of SPM collected at 3 sites (swallow hole, spring and well) of a chalk karst aquifer are compared by means of multivariate analysis allowing the joint analysis of quantitative and qualitative variables: the Hill and Smith Analysis (HILL & SMITH, 1976).

2. Materials and methods

2.1 Study site and field equipment

The karst system studied is located in the Pays de Caux (Haute-Normandie, France) on the right bank of the river

Seine, about 40 km away from the Seine estuary. This kind of system is typical of the karst of the lower Seine valley. The Norville system has been widely studied and its boundaries became quite well known (MASSEI, 2001 ; MASSEI *et al.*, 2002a,b, 2003 ; DUSSART-BAPTISTA, 2003 ; DUSSART-BAPTISTA *et al.*, 2003). The system is composed of a swallow hole, a spring, and a well.

The swallow hole is the point of infiltration of Bébec Creek, on the plateau, and drains a small watershed of about 10 km². Bébec creek discharge is variable, from 3 ls⁻¹ in dry summer periods to more than 500 ls⁻¹ after major winter storms. During these periods, turbidity in Bébec Creek can reach more than 1000 NTU.

The spring, located at the base of the plateau, at the named place 'Le Hannebot', constitutes an overflow of the saturated zone, which is clearly identified as being the main outlet of waters swallowed in the swallow hole. After storms, the turbidity of the water discharging from the spring can exceed 600 NTU. In the past, Hannebot spring water supply, but drinking water is now obtained from a well drilled 130 m away in the alluvial plain. The well is cased through the overlying alluvium and screened in the chalk. Water head in a tank controls the automatic release of pumping in well. During the day (8AM to 10PM), pumping starts automatically when the water level in the tank reaches 2,50m. In the night (10PM to 8AM), pumping occurs when water level in the tank reaches 1,40m. The demand of the population during the day generates a multiplication of the pumping sequences, which last 1 hour (4 to 6 times of 8AM to 10PM). In the night, only one pumping sequence occurs, which lasts 5 hours.

The karst system is located in the Seine river bank at 40km of its estuary. At this point, the tide variations, which are composed of two cycles with periodicity equal to twelve hours (flood and ebb) and fourteen days (falling and rising), defined the water height at Seine river. A tide recorder measures the water height on the Seine river every 5 minutes.

2.2 Sampling periods

ISCO 6700 automatic sampler were installed at the three points of the karst system (swallow hole, spring and well).

Water samples are collected in response to five rain events (December 1999, April 2000, November 2000, February 2002 and October 2002) and three dry events (October 1999, March 2003 and July 2004). December 1999 was an exceptional intense rain event characterized by a cumulative rainfall of 105 mm during one week. April 2000 was a weak rain event after one month without rainfall. November 2000 rain event is characterized by a cumulative rainfall of 10.3 mm in 405 min. February 2002 was an intense rain event with a mean intensity equal to 1.24 mm.h⁻¹ during 4 days. The mean intensity of October 2002 rain event was 0.22 mm.h⁻¹ during 4 days.

Artificial tracer tests were carried out to know hydraulics connections of the swallow hole-spring and swallow hole-well systems (MASSEI, 2001; DUSSART-BAPTISTA, 2003). Knowing the transit times, the granulometric characteristics of each water sample introduced at the swallow hole may be compared with those released at spring and well.

2.3 Particle size distribution of suspended particulate matter

The particle size distributions (PSD) of water samples were determined with a Coulter Multisizer particle counter, using a 100 µm aperture. The PSD were gathered according to grain size (clay<2µm; 2<very fine silt<4µm; 4<fine silt<8µm; 8<medium silt<16µm; 16<coarse silt<31µm; 31<very coarse silt<63µm). From each grain size class (GSC), one index have been computed to identify the transport properties of the karst system:

$$A_{gsc} = P_{gsc} \cdot (S_{gsc} + W_{gsc})$$

with P=swallow hole, S=spring, W=well

these equations have been computed separately for each grain size class.

For each class, the A index indicates the proportion of the particles released at the spring and well compared to the particles introduced at the swallow hole. The A index is a quantitative variable, which allows the identification of the transport properties within the karst system. Starting from the A index value, we define, for each GSC, the T index, which is a qualitative variable with two modes defining the transport properties :

- For each class: if $A_{gsc} < 0$, the PSD released at the spring and well are more important than PSD swallowed at the swallow hole. It puts in evidence the resuspension of intrakarstic sediments, noted by R (in Figure 2, for example Rc indicates the resuspension of clays and Rvcs indicates the resuspension of very coarse silts).

- For each class: if $A_{gsc} > 0$, the PSD released at the spring and well are less important than PSD swallowed at the swallow hole. It puts in evidence the deposition of suspended matter, noted by D (in Figure 2, for example Dc indicates the deposition of clays and Dvcs indicates the deposition of very coarse silts).

2.4 Quantitative and qualitative variables

These granulometric variables (PSD for each GSC at the swallow hole, the spring and the well, A and T indices for each GSC) have been compared with hydrodynamic variables of karst system (piezometry, recharge, depletion, increasing stage, peak, falling stage and recession) and Seine river (tidal range, tide coefficient, flood, ebb, falling, rising). As the variables are numerous (46) and mixed (quantitative and qualitative), the Hill and Smith Analysis

(HSA, HILL & SMITH, 1976), which allows the joint analysis of quantitative and qualitative variables, was used to identify the hydrodynamic parameters controlling the transport properties of SPM. HAS was performed with the ADE4 package (THIOULOUSE *et al.*, 1997) of the R software (THE R DEVELOPMENT CORE TEAM, 2003).

3. Results

3.1 Variable space

The HSA is a multivariate analysis, which allows to identify the correlations between quantitative and qualitative variables. In the variable space, to identify the hydrodynamic variables, which influence the granulometric characteristics of water at the spring and well, we seek the groups of quantitative and qualitative variables, which are plotted i) close one to the other, i.e. which are correlated, or ii) opposed, i.e. which are anti-correlated. The first factorial plane (F1F2) expresses 48% of the total variance, i.e. of all the variations of variables (Figure 1). F1 is related to resuspension of intrakarstic sediments (Rgsc) in the negative part (F1-), opposed to deposition of suspended matter within the karst system characterized by Dgsc and Agsc qualitatively and quantitatively, respectively in the positive part (F1+). F2 is related to PSD of particles observed at the swallow hole (Pgsc), the spring (Sgsc) and the well (Wgsc), then hydrodynamic variables like recharge, piezometric level, falling stage, peak and falling in the negative part (F2-), opposed to tide coefficient, rising, depletion, and recession in the positive part (F2+).

3.2 Sample space

In the variable and sample spaces, principal components (F1 and F2) are the same. The position of the samples on the individual space depends on the values of their variables. Therefore, we can interpret the position of the individuals on F1 and F2 as strong correlations with the variables, which define the principal components. For example, individuals close to the negative part of F1 in the sample space are strongly determined by variables close to the negative part of F1 in the variable space.

Only the individuals of the dry events (October 1999, March 2003 and July 2004) are well separated. The other individuals are dispersed on the positive and negative parts of F1 and F2. Some individuals of various rain events are close to each other and are sometimes difficult to separate. Thus, four groups of individuals can be identified (Fig. 2):

- Group A characterises by strong resuspension of intrakarstic sediments during the rain events of October 1999 (1, 2, 3, 4, 5), December 1999 (11, 14, 17), November 2000 (35) and October 2002 (42, 46, 48, 49).

- Group B characterises by strong particle released from direct transfer during the rain events of April 2000 (19, 20), November 2000 (27, 28, 29, 30, 31, 34, 35) and February 2002 (37, 38, 39, 40).

- Group C characterises by weak particle released resulting from strong deposition of suspended particulate matter during dry events of October 1999 (6, 7, 8, 9), March 2003 (50, 51, 52, 53) and July 2004 (54, 55, 56, 57, 58, 59, 60, 61, 62, 63).

- Group D, very homogeneous, composed by all mean individuals of the rain events (December 1999: 10, 12, 13, 15, 16 ; April 2000: 18, 21, 22, 23, 24, 25, 26 ; November 2000: 32, 36 ; October 2002: 41, 43, 44, 45, 47).

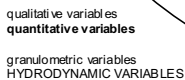


Figure 1 : result on variable space of HSA



Figure2 : result on variable space of HSA

4. Conclusion

Comparison of the results obtained in the variable and sample spaces allows us to identify the variables, which determine the transport properties of SPM. Principal component F1 can be interpreted as a “transport” axis, which opposes the resuspension of intrakarstic sediments characterized by group A (negative part of F1) and the deposition of SPM characterized by group C (in positive part of F1). Principal component F2 can be interpreted as a “SPM release” axis, which opposes the strong particle breakthrough at the spring and well characterized by group B (negative part of F2) to the weak particle breakthrough during dry events characterized by the individuals of October 1999, March 2003 and July 2004 (positive part of F2).

PSD of SPM observed at the swallow hole, the spring and well are close to each other. It reveals a correlation between the particles introduced in the system at the swallow hole and those released at the spring and at the well. Characteristics of particle breakthrough at the outlets of the karst system depend on the granulometric characteristics of particles introduced at the swallow hole and thus they depend on the rain event intensity. However, the events are not classified according to rain event intensity in the sample space. Thus, rainfall characteristics is not the only explanatory variable of the transport properties within the karst system. Indeed, particle released are separated in sample space according to the stage where samples have been taken (increasing stage, peak, falling stage, recession). Samples taken during increasing and falling stages show resuspension of intrakarstic sediments and are in or close to group A. Samples taken during peak show direct transfer and are in or close to group B. Samples taken during recession show deposition of SPM within karst system and are in or close to group A.

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Utilisation du $^{13}\text{C}_{\text{CMTD}}$ dans la discrimination des écoulements de la Zone Non Saturée des systèmes karstiques. Cas du système de Vaucluse.

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Abstract

The karstic system of the 'Fontaine de Vaucluse' has been investigated for many years to get a better understanding of its global functioning. Yet, until now, the Unsaturated Zone (UZ) of this aquifer has been studied indirectly. With the access to the UZ provided by the galleries of the Low-Noise Underground Laboratory of Rustrel in France (LSBB), five flows discharging in the galleries are sampled continuously besides the monitoring at the outlet of the aquifer. First studies allowed getting a first classification of the different identified flows (GARRY *et al.*, accepted), prerequisite for an end-member mixing analysis assessing the contributions from the different sub-systems to the outlet (GARRY *et al.*, submitted). We propose here to use, following the studies of EMBLANCH (1998) and BATIOT (2002), the ^{13}C of Total Dissolved Inorganic Carbon (TDIC) to differentiate the studied flows.

First results (EMBLANCH, 1998) got from the monitoring at the outlet showed that the UZ of this aquifer stayed an opened system towards biogenic carbon dioxide. The results presented here reveal a far more complex scheme: four points (A, B, D and GAS) behave as if they are supplied by a closed system, when the point C is apparently supplied from an opened system during the rising limb of the hydrograph and from a closed system in other cases. These results must lead to the conception of a more elaborated functioning scheme, in order to quantify afterwards the contributions from the different sub-systems to the outlet runoff using the $^{13}\text{C}_{\text{TDIC}}$.

Résumé

L'aquifère karstique de la Fontaine de Vaucluse a fait l'objet de diverses études visant à mieux comprendre son fonctionnement global. Néanmoins, jusqu'à présent, la Zone Non Saturée (ZNS) de cet aquifère a été étudiée d'une manière indirecte. Profitant de l'accès à la ZNS qu'offrent les galeries du Laboratoire Souterrain à Bas Bruit de Rustrel (LSBB), cinq écoulements sont échantillonnés en complément du suivi à l'exutoire de l'aquifère de la Fontaine de Vaucluse. De premières études ont permis d'établir une classification hydrochimique de ces écoulements (GARRY *et al.*, à paraître), préalable à une quantification des contributions des différents pôles de mélange identifiés (GARRY *et al.*, soumis). Nous proposons ici d'utiliser, suite aux travaux de EMBLANCH (1998) et BATIOT (2002), le ^{13}C du Carbone Minéral Total Dissous (CMTD) pour discriminer les écoulements étudiés.

Les premiers résultats (EMBLANCH, 1998), interprétés à partir de prélèvements à la Fontaine, ont montré que la ZNS de cet aquifère restait un système ouvert sur le CO_2 biogénique. Les résultats que nous obtenons dans cette étude mettent en évidence un fonctionnement plus complexe : quatre points (A, B, D et GAS) ont un comportement d'un système évoluant milieu fermé sur le CO_2 biogénique, alors que le point C a un comportement hybride entre un système ouvert et un système fermé. Cette discrimination des écoulements doit conduire à l'élaboration d'un schéma de fonctionnement de la ZNS plus élaboré, dans une perspective de quantification, basée sur le $^{13}\text{C}_{\text{CMTD}}$, des contributions des différents sous-systèmes à l'exutoire.

1. Introduction

Le $^{13}\text{C}_{\text{CMTD}}$, isotope stable commun du carbone, est régulièrement utilisé afin de corriger les teneurs en ^{14}C des eaux souterraines. Cet isotope, déjà utilisé comme traceur dans un contexte cristallin par AMIOTTE-SUCHET *et al.* (1999), n'a été jusqu'à présent que peu utilisé dans un contexte hydrogéologique karstique. Et pourtant, comme le souligne EMBLANCH (1997), les sources potentielles de carbone, *i.e.* le CO_2 et la matrice carbonatée, sont bien identifiées. De récentes études ont ainsi démontré (EMBLANCH, 1998) l'intérêt et la pertinence de l'utilisation du $^{13}\text{C}_{\text{CMTD}}$ comme traceur des écoulements de la zone non saturée des systèmes karstiques. EMBLANCH *et al.* (1998; 2003) ont mis en évidence, à partir de mesures effectuées à l'exutoire du système de Fontaine de Vaucluse (dans le sud-est de la France), l'influence du $\delta^{13}\text{C}_{\text{CO}_2}$ biogénique dans le

marquage de la zone non saturée pourtant épaisse de 800 m. Néanmoins, jusqu'à présent, les études réalisées ne permettent qu'une estimation indirecte de la contribution de la Zone Non Saturée (ZNS) dans le fonctionnement global du système. En effet, la ZNS de ce système a toujours été considéré comme un système ouvert vis-à-vis du CO_2 biogénique, c'est-à-dire que l'on a fait l'hypothèse que l'eau circulant dans cette zone était en équilibre avec le CO_2 biogénique, contrôlant ainsi la signature isotopique du carbone minéral dissous. Pour cette étude, nous avons bénéficié d'un accès direct à l'intérieur de la ZNS, via le Laboratoire Souterrain à Bas Bruit de Rustrel (LSBB), creusé au sein même du massif calcaire. L'objectif principal de notre étude est de mettre à profit cet accès privilégié afin d'estimer le degré d'ouverture des sous-systèmes alimentant

les écoulements identifiés dans les galeries au LSBB et de compléter ainsi la connaissance acquise sur le fonctionnement hydrochimique (GARRY *et al.*, à paraître) de la ZNS de ce système.

2. Présentation du site

Situé dans le Sud-Est de la France, l'aquifère de la Fontaine de Vaucluse (FV), exutoire unique de la grande zone karstique (Fig. 1) comprenant le Plateau d'Albion, le Mont Ventoux et la Montagne de Lure, possède un bassin d'alimentation d'environ 1 100 km² (Puig, 1987). Celui-ci est limité au Nord par la vallée du Toulourenc, à l'Ouest par le Bassin de Carpentras, au Sud par le Bassin d'Apt et à l'Est par le Bassin de Forcalquier. La région étudiée est soumise à un climat méditerranéen de type provençal, caractérisé par de fortes variations interannuelles (CELLE *et al.*, 2000) avec une composante montagnarde pour le Mont Ventoux et la montagne de Lure. L'altitude varie de 84 m pour l'exutoire à 1 912 m pour le Mont Ventoux et une moyenne pondérée pour la zone d'infiltration de 870 m (BLAVOUX *et al.*, 1992).

Ce karst s'est développé dans les calcaires Hauteriviens et Barrémien-Bédouliens (Urgonien) qui peuvent atteindre une épaisseur de 1 500 m entre les marnes du Néocomien et celles de l'Aptien supérieur (MASSE, 1968; MASSE, 1972). La zone d'infiltration de ce système présente la particularité d'être très épaisse avec une moyenne de 800 m (PUIG, 1987).

Le site expérimental du Laboratoire Souterrain à Bas Bruit de Rustrel (LSBB), ex-poste de tir n°1 du système d'armes du plateau d'Albion (Fig. 1) est implanté sur une zone de faible densité de population dans le parc naturel du Lubéron au Nord-Est d'Apt sur la bordure Sud du bassin versant de la Fontaine de Vaucluse (altitude ~ 470 m). La galerie du laboratoire est creusée à 500 m sous le niveau de la surface dans le calcaire Crétacé qui est dépourvu de matière organique. Elle pénètre dans un monoclin de pente sud-ouest et recoupe des failles de direction N-S et NE-SO (THIEBAUD, 2004). La surface d'échantillonnage est de 9560 m² avec une altitude moyenne de recharge estimée à 700 m.

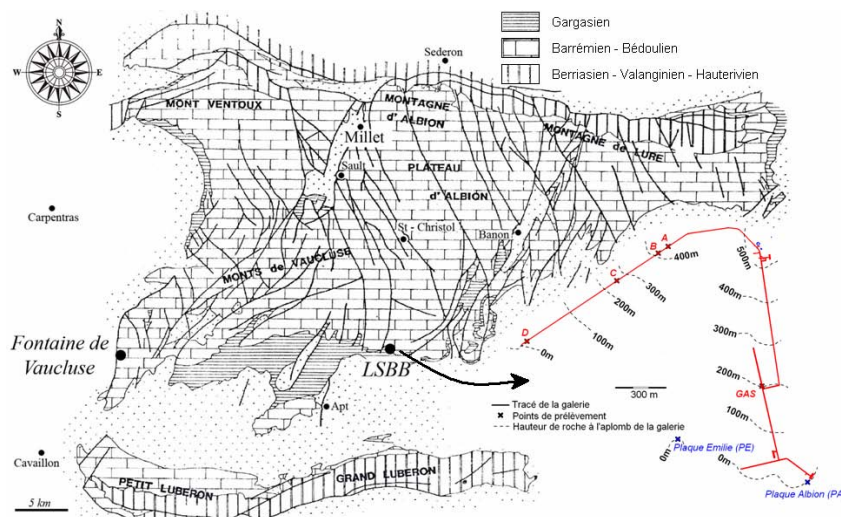


Fig. 1 : Carte géologique du système de Fontaine de Vaucluse et schéma du L.S.B.B. (PUIG, 1987 ; modifié)

3. Matériels et méthodes

3.1 Stratégie d'échantillonnage

Cinq écoulements sont échantillonnés en cinq points des galeries (Fig. 1) depuis février 2002. La fréquence d'échantillonnage est mensuelle de février 2002 à octobre 2003, puis hebdomadaire, voire journalière en périodes de fortes précipitations. Les mesures de débit sont effectuées manuellement. Quatre de ces écoulements (en A, B, C et D) sont pérennes et se trouvent dans la galerie de secours. Le point GAS se situe, quant à lui, dans la galerie anti-souffle et correspond au débouché d'un petit conduit karstique de quelques centimètres. Les points A et B sont recouverts par 450 m de calcaire Urgonien, C et GAS par 350 m et D par 35 m.

Les paramètres température, pH, conductivité et alcalinité totale ont été mesurés sur le terrain. Les échantillons sont ramenés au Laboratoire d'Hydrogéologie de l'Université d'Avignon pour y être analysés. Le Carbone Organique Total (COT) est oxydé puis mesuré par un détecteur infrarouge (TOC Analyser model 700,

BIORITECH). Pour les analyses isotopiques en carbone 13 (¹³C_{CMTD}), les échantillons sont préparés par attaque acide directe sur l'échantillon d'eau afin de transformer le Carbone Minéral Total Dissous (CMTD) en CO₂ gazeux qui est recueilli avant analyse par spectrométrie de masse. La précision analytique est de ± 0.05 mg/l pour le COT (BATIOT, 2002) et de ± 0.2 ‰ pour le ¹³C_{CMTD} (EMBLANCH, 1997).

3.2 Composition isotopique du carbone minéral total dissous (CMTD) en milieu karstique

Le CMTD de l'eau résulte d'une part de la mise en solution du CO₂ provenant du sol, de l'atmosphère, ou de l'oxydation de la matière organique par l'activité bactérienne, et d'autre part de la dissolution des carbonates. La composition isotopique actuelle du CO₂ atmosphérique, dont la pression partielle (pCO₂) est estimée à 3.4 10⁻⁴ atm, est de -8.0 ‰ VPDB (CERLING *et al.*, 1991). Le CO₂ biogénique, produit par la respiration racinaire et par la minéralisation de la matière organique, présente une pression partielle 10 à 100 fois supérieure à celle de l'atmosphère. Contrairement au CO₂ atmosphérique dont la pression partielle est stable, celle des sols varie et dépend

essentiellement de l'humidité et de la température, facteurs régulant l'activité métabolique de la biosphère, qui sont eux même régis par le couvert végétal, l'épaisseur du sol et le climat. EMBLANCH (1997) a réalisé un suivi annuel sur un site à 950 m d'altitude sur le plateau d'Albion et montre une variation saisonnière de la $p\text{CO}_2$ avec un minimum en hiver ($p\text{CO}_2 = 0.2 \cdot 10^{-2}$ atm), période pendant laquelle la végétation est peu active, et un maximum en été ($p\text{CO}_2 = 1.4 \cdot 10^{-2}$ atm), soit une moyenne annuelle de $0.8 \cdot 10^{-2}$ atm (environ 25 fois la $p\text{CO}_2$ atmosphérique). Le $\delta^{13}\text{C}_{\text{CO}_2}$ du sol dépend du type de végétation recouvrant le système étudié, et plus précisément de leur cycle photosynthétique. Dans les régions méridionales, nous assimilons le marquage isotopique en $^{13}\text{C}_{\text{CMTD}}$ du CO_2 du sol à celui des végétaux en C_3 (cycle à trois carbones ou cycle Calvin) qui produisent un CO_2 relativement appauvri en $^{13}\text{C}_{\text{CMTD}}$ avec une signature d'environ -22.0 ‰. EMBLANCH (1997) a réalisé trois mesures du CO_2 du sol sur le bassin d'alimentation de FV donnant des valeurs moyennes de l'ordre de -21.6 ‰, proche de la valeur moyenne considérée. Les carbonates crétacé d'origine marine ont, quant à eux, une signature en $^{13}\text{C}_{\text{CMTD}}$ comprises entre -2.0 et +2.0 ‰ VPDB avec une moyenne calculée de +0.35‰ (EMBLANCH, 1997). Pour simplifier, nous assimilons le marquage isotopique des carbonates à 0.0 ‰.

Ceci permet ensuite de calculer l'évolution théorique du $\delta^{13}\text{C}_{\text{CMTD}}$ des eaux en système fermé et ouvert à température constante (BATIOT, 2002), qui dépend du marquage isotopique du CO_{2g} du sol. En considérant qu'aux pH proches de 7, pH mesurés sur les écoulements dans la galerie du LSBB et à la FV, le CMTD se trouve exclusivement sous la forme de bicarbonates (HCO_3^-) et d'acide carbonique (H_2CO_3), on obtient alors en système fermé :

$$\delta^{13}\text{C}_{(\text{CMTD})} = \frac{[(\text{H}_2\text{CO}_3) + (0.5 \times (\text{HCO}_3^-))][\delta^{13}\text{C}_{(\text{CO}_2)} + \varepsilon_{\text{H}_2\text{CO}_3-\text{CO}_{2(g)}}]}{(\text{CMTD})}$$

où ε_{ij} désigne le facteur de fractionnement entre les espèces i et j.

En système ouvert, on obtient (BATIOT, 2002) :

$$\delta^{13}\text{C}_{(\text{CMTD})} = \frac{[(\delta^{13}\text{C}_{(\text{CO}_2)} + \varepsilon_{\text{H}_2\text{CO}_3-\text{CO}_{2(g)}})(\text{H}_2\text{CO}_3) + (\delta^{13}\text{C}_{(\text{CO}_2)} + \varepsilon_{\text{HCO}_3^--\text{CO}_{2(g)}})(\text{HCO}_3^-)]}{(\text{CMTD})}$$

avec à $T = 12^\circ\text{C}$:

$$\varepsilon_{\text{HCO}_3^--\text{CO}_{2(g)}} = 9.37 \text{ ‰}, \varepsilon_{\text{H}_2\text{CO}_3-\text{CO}_{2(g)}} = -1.12 \text{ ‰}.$$

4. Résultats

Les valeurs moyennes en $^{13}\text{C}_{\text{CMTD}}$ mesurées ($\delta^{13}\text{C}_{\text{mes}}$), sur les 5 écoulements échantillonnés dans le LSBB ainsi qu'à FV sur une période de crue (avril 2005) et sur une période d'étiage (septembre 2005) sont présentées dans le Tabl. 1. Les valeurs mesurées sont ensuite comparées aux valeurs théoriques calculées dans le cas d'un système ouvert et d'un système fermé sur le CO_2 biogénique (respectivement " $\delta^{13}\text{C}_{\text{CMTD}}^{\text{calc}} \text{ ouvert}$ " et " $\delta^{13}\text{C}_{\text{CMTD}}^{\text{calc}} \text{ fermé}$ "). Les valeurs proches de -11.0 ‰ mesurées en A, B et D à l'étiage sont caractéristiques d'une eau chargée en CO_2 biogénique, qui s'est mise à l'équilibre avec la calcite en système fermé. En crue, les valeurs de A et B sont un peu plus enrichies avec -10.2 ‰.

Plus généralement, on remarque que les valeurs pour les points A, B, D, FV et GAS sont systématiquement plus

enrichies que les valeurs théoriques en milieu fermé et en milieu ouvert. Ceci suggère donc que ces écoulements sont alimentés par des réservoirs qui évoluent principalement en milieu fermé. De plus, l'écart observé entre les valeurs théoriques en milieu fermé et les valeurs mesurées montrent que un ou des facteurs externes interviennent dans l'enrichissement en $^{13}\text{C}_{\text{CMTD}}$. Cependant le nombre d'analyses en ^{13}C ne nous permet pas pour le moment d'approcher ces phénomènes. Pour le point GAS, les résultats sont surprenants, car cet écoulement ponctuel est situé au pied d'une faille et ne fonctionne que lors de forts événements pluvieux. Par ailleurs, GARRY *et al.* (à paraître) ont montré que les eaux échantillonnées en GAS ont des teneurs en TOC élevées (~ 2.5 mg/l) accompagnées de faibles teneurs en Mg^{2+} (~ 0.5 mg/l) indiquant un faible temps de résidence des eaux au sein de l'aquifère et un enrichissement rapide en matière organique lors de l'infiltration. Ces éléments permettent d'affirmer que cet écoulement évolue naturellement en milieu ouvert vis-à-vis du CO_2 biogénique. Cette contradiction provient vraisemblablement des conditions de prélèvement qui entraînent un dégazage partiel des eaux, expliquant ainsi leur enrichissement isotopique. Les valeurs en GAS doivent donc être éliminées pour l'analyse.

Les résultats pour A et B en crue sont surprenants puisqu' *a priori* en période de crue, l'arrivée dans le système d'eaux récentes encore riche en CO_2 biogénique dissout, devrait appauvrir le signal en ^{13}C de l'ensemble des réservoirs associés. Ceci révèle un comportement encore mal connu des réserves de la ZNS.

Point	Période	$\delta^{13}\text{C}_{\text{CMTD}}^{\text{mes}}$ (‰)	$\delta^{13}\text{C}_{\text{CMTD}}^{\text{calc}}$ ouvert (‰)	$\delta^{13}\text{C}_{\text{CMTD}}^{\text{calc}}$ fermé (‰)
A	crue	-10.2	-13.3	-12.3
	étiage	-11.1	-13.9	-13.0
B	crue	-10.2	-13.2	-12.2
	étiage	-11.0	-13.8	-12.9
C	crue	-12.0	-13.1	-12.0
	étiage	-11.3	-13.1	-12.1
D	crue	-11.3	-13.1	-12.1
	étiage	-10.9	-13.4	-12.4
FV	crue	-10.9	-14.7	-13.8
	étiage	-10.3	-14.9	-14.0
GAS	crue	-11.3	-12.9	-11.9

Tabl. 1: $\delta^{13}\text{C}_{\text{CMTD}}$ VPDB moyens des écoulements de la ZNS et de FV en période de crue et d'étiage.

Enfin, on peut remarquer que le signal isotopique moyen en C de -12.0 ‰ est plus appauvri que celui des autres points. Pour comprendre cette différence, nous allons étudier ci-après la dynamique de crue en C par le suivi de l'évolution temporelle du signal isotopique.

La Fig. 2 montre les variations au point C du débit enregistré, du COT, du $\delta^{13}\text{C}_{\text{mes}}$, $\delta^{13}\text{C}_{\text{calc}}^{\text{ouvert}}$ et du $\delta^{13}\text{C}_{\text{calc}}^{\text{fermé}}$ sur la période du 13 avril 2005 au 21 juin 2005. Le COT a été choisi pour sa capacité à tracer les eaux à faible de temps de séjour (EMBLANCH, 1998 ; GARRY *et al.*, à paraître). Entre le 20 avril 2005 (pointe de crue) et le 12 mai 2005, on constate, en parallèle de la diminution des débits, une diminution de teneur en matière organique (COT) de 1.8 à 1.1 mg/l et un enrichissement en $^{13}\text{C}_{\text{CMTD}}$ de -13.0 à -10.7 ‰. Ainsi, pour les deux premiers prélèvements, le marquage isotopique est proche du marquage en milieu ouvert (~ -13.0 ‰) puis devient manifestement celui d'un milieu fermé (> -12.0 ‰). Sous l'impulsion des événements pluvieux de la mi-mai, les débits augmentent à nouveau et on observe un appauvrissement en $^{13}\text{C}_{\text{CMTD}}$ et une augmentation de COT.

D'une manière générale, l'augmentation en COT, synonyme d'une diminution du temps de séjours de l'eau dans l'aquifère associée à un appauvrissement en $^{13}\text{C}_{\text{CMDT}}$, est le signe d'un apport d'eau ayant évolué en système ouvert. A l'inverse, lorsque le COT diminue, en raison de l'augmentation du temps de séjour, on observe alors un enrichissement en $^{13}\text{C}_{\text{CMDT}}$ indiquant une diminution, voire un arrêt des apports d'eau ayant évolué en système ouvert. Ce nouveau résultat nous permet ainsi de caractériser le réservoir qui alimente l'écoulement en C. En effet, les variations de signal isotopique de ce sous-système de la ZNS montre une évolution en milieu fermé, mais l'arrivée de nouvelles eaux chargées en CO_2 biogénique vient temporairement appauvrir la signature en $^{13}\text{C}_{\text{CMDT}}$ du réservoir. Cette chute de $^{13}\text{C}_{\text{CMDT}}$ est telle que le réservoir évolue en milieu ouvert sur le CO_2 biogénique révélant sa faible inertie.

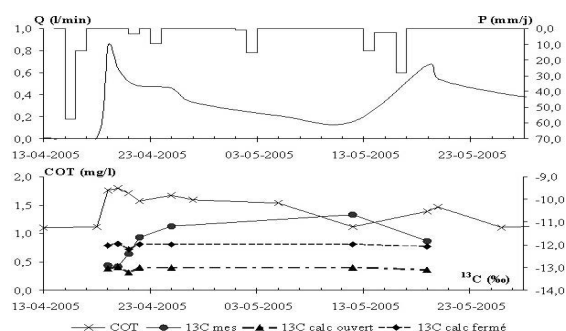


Fig. 2 : Variation du débit, COT et $\delta^{13}\text{C}_{\text{CMDT}}$ de l'écoulement au point C sur la période du 13 avril au 21 juin 2005.

5. Conclusions

A l'instar de EMBLANCH (1998), l'utilisation du $^{13}\text{C}_{\text{CMDT}}$ s'est révélée pertinente comme traceur de la ZNS. Il nous a permis de faire évoluer notre conception de son fonctionnement. En effet, contrairement à l'idée que celle-ci pouvait être considérée comme un système ouvert sur le CO_2 biogénique, nous nous sommes aperçus que les réservoirs alimentant les écoulements de la ZNS peuvent être considéré pour trois d'entre eux (A, B et D) comme des sous-systèmes fermés vis-à-vis du CO_2 biogénique. Néanmoins, même si les valeurs en $^{13}\text{C}_{\text{CMDT}}$ de l'écoulement au point GAS ne peuvent pas être prises en compte dans l'analyse, l'étude au point C a montré un fonctionnement hybride entre un milieu fermé en période de décrue ou d'étiage et un milieu ouvert en début de période de crue. Ceci révèle la complexité de la ZNS qui doit être nécessairement prise en compte dans l'estimation de la ressource en eau. Ainsi, le modèle de mélange à deux réservoirs de EMBLANCH (1997) (le premier caractérisant la ZNS considérée ouverte sur le CO_2 biogénique et le second représentant la zone noyée fermée vis-à-vis de la phase gazeuse) conduit à une sous-estimation puisqu'il donne une participation annuelle de la composante ZNS aux écoulements à l'exutoire de 30 %. Déjà, de nouveaux calculs par l'intermédiaire du COT et du Magnésium, ont permis d'estimer à 52 % (GARRY *et al.*, soumis) la participation de ces écoulements à l'écoulement total à l'exutoire. Il s'agit à présent d'ajuster, en utilisant le $^{13}\text{C}_{\text{CMDT}}$, le modèle à trois pôles proposé par GARRY *et al.*, pour permettre de quantifier et mieux discriminer les arrivées d'eau en provenance de la ZNS des aquifères karstiques.

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Runoff in an Iranian karstic watershed as compared with a neighbor non-karstic watershed

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Abstract

In this paper, we have compared hydrological characteristics for two neighbour catchments. Vehregan watershed as a karstic basin that mostly covered by limestone, and is located in Zagros Mountain Ranges in Iran, has been compared with a non-karstic catchment, namely Golpayegan watershed and located in north of Karstic basin in view of maximum discharge, runoff coefficient, specific discharge, and relation between maximum instantaneous discharge and maximum daily mean discharge having similar climates, catchment areas, slopes and annual mean rainfalls. This research indicates that runoff coefficient of karstic watershed for a concurrent record period is 3.7 times another non-karstic catchment.

Ratio between maximum instantaneous discharge and maximum daily mean discharge during a 20-year period for both karstic and non-karstic watersheds are averagely found 1.45 and 1.38, respectively. The manner of the changes for both instantaneous and maximum daily mean discharges between the watersheds is detected similar but for all of the years, both instantaneous discharge and maximum daily mean discharge in karstic watershed are in a higher level.

Introduction

Technical problems caused that we could not improve and utilize the karstic resources completely, specially in the southern part of the Zagros Mountain Ranges (SALIMKHANIAN, 1990 & 1991). A comparative study about hydrological characteristics of two karstic and non-karstic watersheds, indicated that runoff coefficient for the karstic one was 4 to 5 folds of non-karstic watershed (HAKIMI, 1998). The climate change is one of the processes that having an impact on water resources and runoff and also it can be approved for karstic resources, (BILVARDI, 2000). FELTON (1994) have studied the hydrological parameters of a karstic watershed in Vedfords of America and indicated that the developed formula was not properly worked for karstic watersheds. HAKIMILARIJANI & KHAZALI (2000) have studied two watersheds, one calcareous and another one non-calcareous in the Zagros mountain ranges of Iran and showed that a large volume of water had been entered to karstic watershed by underground canals and they increased both runoff coefficient and specific discharge. ATKINSON (1973) and BROWN (1964) believed that for investigating the karstic watershed, we must have an extensive groundwater field study. DUBLYANSKII *et al.* (1984) and KOVALEVAKY (1993) have studied a karstic area in Krimea and indicated that changing in the groundwater was influenced by changing climate condition.

Iran is one of the arid and semi-arid regions in the world as that water has a very important value. Each year, Iranian water resources losses a large volume of water due to infiltrating in the karstic structures. Surface runoff for all of the watersheds that have not hydrometric stations usually calculated by the methods which based on the physical parameters of watersheds such as area, slope, main stream length, and these methods are not efficient for karstic watershed. The objective of this study is comparing hydrologically, two contiguous watersheds, one karstic and another non-karstic.

Materials and methods

For this research, we have first studied several karstic and non-karstic watersheds for finding two representative watersheds with the most similarities in case of whether, area, and basin mean slope and in the same time one completely calcareous and another one non-calcareous. After that we selected these watersheds, we have studied rainfall mean, changing in the rainfall and other climatic conditions of these watersheds, and also in the same time we analyzed mean of daily discharge, maximum daily discharge, and maximum instantaneous discharge in these watersheds. During the same years of records, from volume of water that produced by rainfall calculated by means of yearly rainfall data and volume of runoff calculated by discharge data, we calculated the runoff coefficient for both watersheds and finally we compared these ratios. We benefited from HYFA (hydrological frequency analysis) software for estimating and improving quintiles of maximum yearly instantaneous discharges and maximum daily mean discharges, and we calculated amount of flows for different return periods. We then compared the ratio between these two discharges for different return periods. Vehregan watershed is a karstic basin that located in central part of Iran. It has a general east-west slope. Golpayegan watershed is a non-karstic watershed in 160 km west of Esfahan, Iran and in neighbourhood of Vehregan basin and it has a general south-north slope. Area of Vehregan watershed is 743.7 km² and this for Golpayegan basin is 816 km² and shape of two basins are protracted (according to Horton method (1932) and

basin circularity method). Mean altitude of Vehregan basin is 2626 and for Golpayegan is 2455 meters from sea surface level and two watersheds have similar conditions that have an impact on runoff coefficient and specific discharge such as protraction, area, slope and precipitation. The average annual precipitation for karstic watershed is about 456.6 mm and its climate is moderate humid according to Demarton method and the average yearly temperature is 9.5 °C. The average yearly precipitation for non-karstic watershed is 469.4 mm and the climate is also moderate humid. Temporal distribution of precipitation is similar for both basins and more rainfall exists in the winter and autumn and dry period is concentrated in the summer.

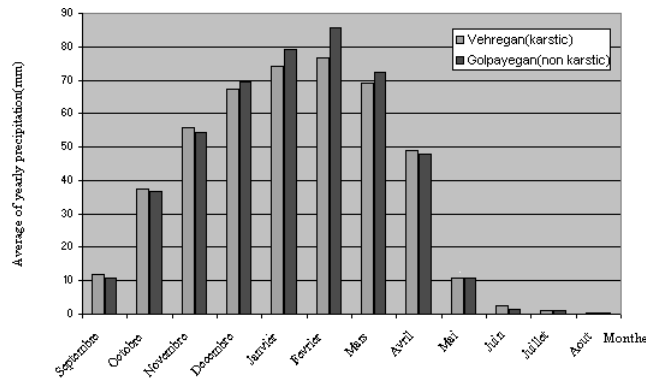


Figure 1: Yearly distribution of precipitation

Yearly distribution of precipitation indicated in Figure (1). This figure showed that the precipitation that is the most important resource of water and runoff for both watersheds is similar. Figure 2 showed geological structure of two watersheds. The geological structure in the karstic watershed of Vehregan covered mostly by calcareous stone and most of precipitation infiltrated in soil and it continues until that water blocked by hard layer. For this reason, there are many springs in this area and also there are different types of karstic topography particularly spring karstic that it transfers water from neighbourhood basin to this watershed and finally increased the runoff coefficient in this basin.

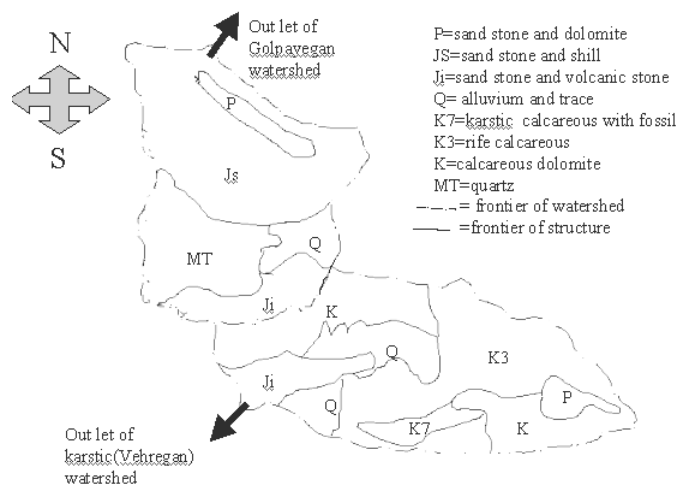


Figure 2: Geological structures for two watersheds

The runoff coefficient (c_r) and specific discharge (q) can be calculated by the following equations:

$$C_r = \frac{V_{ro}}{V_p} \quad (V_{ro}=31.54Q_m, V_p=H_p \cdot A) \quad , \quad q = \frac{Q_p}{A}$$

where V_p is the volume of yearly precipitation (MCM), H_p is the annual precipitation (m), and A is the area of watershed (km^2). V_{ro} is the volume of yearly runoff in million cube meters (MCM) and Q_m is the mean of yearly discharge (m^3/sec). Q_p is the average of yearly maximum instantaneous discharge of basin (m^3/sec).

Results and conclusions

Both watersheds investigated in this research, had hydrometry stations in the outlet with water level recorder. According to data, average yearly discharge for karstic watershed in the outlet point is 12.97 m^3/sec

and with regard to 743.7 km² area, volume of yearly runoff for this basin calculated 409.02 MCM, and volume of yearly runoff for non-calcareous basin (Golpayegan) with 816 km² area calculated 132.14 MCM. Also, both basins have similar efficacious parameter in the runoff, but for comparing amount of runoff we calculated runoff coefficient and specific discharge, because of difference in area. With increasing area, volume of precipitation and runoff will be increased and so, area of watershed has not a specific interference on the runoff coefficient.

Table (1) calculates runoff coefficient, and relation between maximum instantaneous yearly discharge with maximum average daily discharge where S is drainage area, Pm is the average of yearly precipitation, Qm is the average of yearly discharge, Vp is the volume of yearly precipitation, Vro is the volume of yearly runoff, Cr is runoff coefficient, q is specific discharge, Qmiy is the maximum instantaneous yearly discharge, and Qmd is the maximum of average daily discharge.

Name	S	Pm	Qm	Vp	Vro	Cr	q	$\frac{Q_{miy}}{Q_{md}}$
	(km ²)	(mm)	(m ³ /sec)	(MCM)	(MCM)		(m ³ /sec.km ²)	
Vehregan	743.7	456.6	12.97	339.27	409.02	1.206	0.0174	1.145
Golpayegan	816.9	469.4	4.16	383.45	132.135	0.345	0.0051	1.38

Specific discharge is related to area and so two parameters are comparable. According to Table 1, runoff coefficient for non-karstic watershed (Golpayegan) is 0.345 and this coefficient for karstic watershed (Vehregan) is 1.026 and so the runoff coefficient for karstic watershed are 3.7 folds. According to this coefficient, amount of runoff is 1.026 folds of precipitation in karstic watershed but if we have not any losses of rainfall, the runoff coefficient should be 1, and so another factor in addition to precipitation increases discharge for this watershed. Also according to Figure 1, mean and distribution of yearly precipitation is similar for both watersheds, but ratio between amount of precipitation and discharge is completely different for both basins. Correlation coefficient between yearly precipitation and mean discharge was 0.11 for karstic watershed and 0.87 for non-karstic watershed, respectively:

$$Q_m = 0.005 \cdot P_m + 11.06 \quad (R=0.11) \quad , \quad Q_m = 0.02 \cdot P_m - 4.66 \quad (R=0.87)$$

There is no remarkable correlation between precipitation and discharge for karstic watershed and it indicates that a large volume of water in this basin is from karstic resources. This shows that any experimental method based on precipitation is not suitable for the karstic watersheds.

Table 2 estimates the maximum instantaneous yearly discharge and maximum of daily average discharge in two basins for different return periods and their ratios. Where QPv and QDv are respectively maximum instantaneous yearly discharge and maximum of daily average discharge for karstic basin and QPg and QDg are respectively the maximum instantaneous yearly discharge and maximum of daily average discharge for non-karstic basin.

Return Periods	2	5	10	25	50	100	1000
QPv	101.08	149.39	180.84	220.52	249.95	279.18	375.187
QDv	87.25	127.36	153.46	186.40	210.83	235.09	315.53
QPg	87.35	119.89	141.32	168.44	188.65	208.88	277.9
QDg	39.34	73.97	96.90	125.88	147.37	168.71	239.2
QPv/QDv	1.58	1.173	1.178	1.183	1.185	1.187	1.191
Average QPv/QDv	1.24						
QPg/QDg	2.224	1.62	1.458	1.338	1.28	1.238	1.161
Average QPg/QDg	1.41						

Specific discharge for karstic watershed is 17.4 (litre/sec) and for non-karstic is 5.12 (litre/sec). Specific discharge is related to area and for the similar condition of average slope, height, and shape of two basins, we should have more specific discharges in the non-calcareous basin, because it has more precipitations, but specific discharge in the karstic watershed is 3.4 folds of non-karstic. Ratio of maximum instantaneous yearly discharge to maximum average daily discharge between 1975 and 1995 is 1.145 for karstic Vehregan watershed and this is 1.38 for non-karstic Golpayegan basin.

Table 2 shows the ratio between maximum instantaneous yearly discharge and average daily discharge with different return periods. We can utilize this ratio for calculating maximum instantaneous discharge for the years that we have not all of the parameters.

Figure 3 indicates changing in average of yearly discharge and Figure 4 indicates changing in the maximum daily discharge for both basins. According to Figures 3 and 4, trend of changing in discharge is similar for two basins but discharge of karstic Vehregan watershed is greater for all of the hydrological years. Similarly, trend of changing in discharge for both basins indicates that two watersheds have the important and same resources that is precipitation. When discharge in the karstic watershed for all of the years is greater than non-karstic watershed, it shows that this basin has another resource that is known as karst. So in the karstic watershed, we can determine two main resources for runoff, one is the precipitation and another is the karst resources.

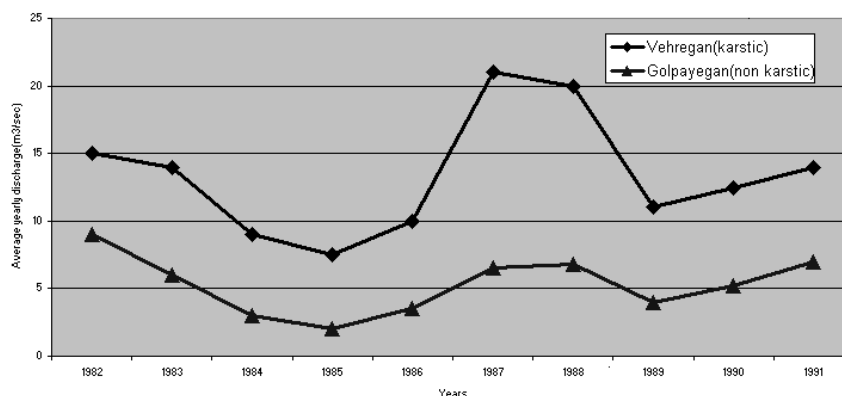


Figure 3: Changing in average of yearly discharge for two karstic and non-karstic watershed

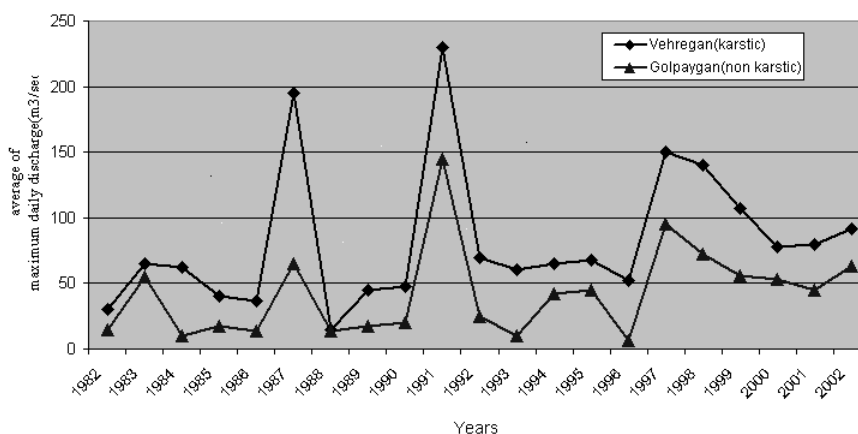


Figure 4: Changing in maximum daily discharge for two karstic and non-karstic watersheds

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Comparison of solute and particle transport in shallow and deep karst aquifer systems

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Abstract

The transport of solute and particulate contaminants, like microbial pathogens, in karst groundwater was simulated by means of comparative tracer tests using fluorescent dyes and fluorescent microspheres with different diameters (1, 2 and 5 µm). The first test site (Hölloch) is a shallow karst system in the Austro-German Alps; the second (Yverdon) is a deeply confined aquifer related to the Swiss Jura Mountains. The tracers were injected in a cave stream and a swallow hole respectively; the samples were taken at springs in several kilometres distance. The microspheres were analysed using two different methods: filtration and subsequent counting under the fluorescent microscope, and with an innovative fluorescence-particle counter. In both test sites, the solute tracers gave regular breakthrough curves with clear maximums and high recovery rates, while the microspheres showed an irregular behaviour, i.e. extremely variable particle contents in subsequent samples, which indicates heterogeneous transport processes. The differences between the shallow and deep system are quite low. The results of this study demonstrate a high groundwater vulnerability to microbial and other contamination.

1. Introduction

In many karst areas, microbial contamination resulting from agriculture and untreated domestic wastewaters is the most important groundwater quality problem. Contaminants, including microbial pathogens, can easily enter the subsurface through thin soils or via swallow holes. Inside the aquifer, they are often rapidly transported towards drinking water springs and wells (DREW & HÖTZL 1999).

Tracing techniques allow groundwater flow and contaminant transport to be studied. Solute tracers make it possible to simulate the migration of solute contaminants, while particulate tracers can be used to investigate the transport of microbial pathogens and other colloidal contaminants in groundwater (AUCKENTHALER *et al.* 2002).

Particle tracers can be classified into colloids (< 1 µm) and larger particles and include both biological and artificial substances. Bacteriophages are biocolloids that provide analogues for pathogenic viruses. Specific types of harmless bacteria can be used as tracers to simulate the transport of pathogenic bacteria (HARVEY 1997). However, some species that have often been used as tracers, like *Serratia marcescens*, are now classified as pathogens and thus require special precaution. Furthermore, bacteria that are introduced into groundwater may be involved in biological processes that are difficult to control, like inactivation, active attachment, and predation by protozoans.

Therefore, fluorescent microspheres are increasingly used as particle tracers. They are available in different colours (excitation/emission wavelengths), diameters (0.05–10 µm), and electrical surface charges. Microspheres are invisible in the water, chemically and biologically stable, not harmful, and show little interference with other tracers. Spheres < 1 µm provide analogues for viruses and small bacteria. Larger spheres can be used as surrogates for larger bacteria and protozoans, e.g. *Cryptosporidium* oocysts (3–7 µm) (EMELKO & HUCK 2004).

Two test sites have been selected to study the transport of particulate and solute contaminants by means of comparative tracer tests: the Hölloch cave in the Austro-German Alps, and the Yverdon test site in the southern foreland of the Swiss Jura Mountains. The first site is a shallow flow-system, i.e. the water flows in turbulent cave

streams on top of the underlying marl aquiclude. The second site is a confined karst system, where groundwater flows in permanently saturated conduits below the hydrologic base level, before it re-emerges to the spring. The detailed monitoring of turbidity, organic carbon (OC), bacteria, discharge, temperature and electrical conductivity in this test site makes it possible to compare the tracer test results with the observed natural variations (see PRONK *et al.* in this volume).

2. Materials and methods

Overview and sampling

The field experiments included comparative tracer tests with fluorescent dyes and different types of microspheres, and flow measurements. Water samples were taken at the springs for subsequent tracer analyses. In the Yverdon site, uranine was also measured continuously, and a variety of natural parameters have been monitored.

The samples for the tracer analyses were collected with automatic samplers (ISCO). For the Hölloch tracer test, due to extremely low winter temperatures of -20°C, the sampler was placed into a purpose-built heated isolate box. The samples were transfilled daily and split into two fractions. One fraction was decanted into 200 mL brown plastic bottles and analysed for fluorescent dyes and 5-µm spheres. The other fraction was transfilled in 250 mL brown glass bottles and analysed for 1-µm and 2-µm spheres.

Particle tracers and analyses

Fluorescent polystyrene microspheres with 3 different excitation/emission wavelengths and 3 different diameters were used as particle tracers: Polychromatic Red Microspheres 1.00 µm (PC Red, Polysciences Inc.), Yellow-Green Microspheres 1.00 and 2.00 µm (Fluoresbrite YG, Polysciences Inc.), and Scarlet Microspheres 5.00 µm (FluoSpheres BB, Molecular Probes). Two methods were used for the analyses: the 1-µm and 2-µm spheres were counted under the fluorescence microscope; the 5-µm spheres were measured with a fluorescence-particle counter.

The microscopic analyses were done in the AGK lab using a well-approved method (KÄSS 1998). The water samples were homogenised with a sonicator, filtered using

0.8 μm nitrate paper filters with an imprinted reticule (Sartorius), and finally counted under an Epifluorescence Microscope (Zeiss Standard FL) at different wavelengths.

The fluorescence-particle counter was developed by the Markus KLOTZ Company. After initial tests in surface streams, it was first used for karst groundwater tracing within the frame of this study. At the same time, the new technique was tested in fractured and porous media (GÖPPERT, unpublished data). The instrument consists of a particle counter coupled with a fluorescence detector (Fig. 1).

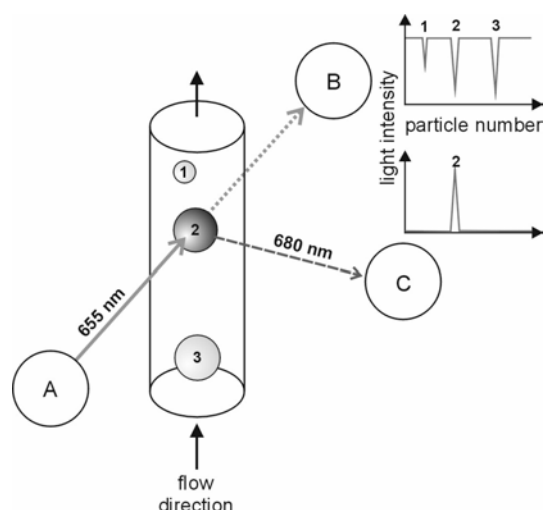


Fig. 1: Illustration of the fluorescence-particle counter. The water passes through a fine tube illuminated by the light source A. The equivalent diameter of the particles n° 1, 2 and 3 is determined as a function of the intensity decline measured at the light detector B. Only the fluorescent particle n° 2 create a signal at higher wavelengths that is measured at the detector C.

The water sample, usually 10 mL, passes through a fine tube (0.3 mm) illuminated by a laser ray ($\lambda = 655 \text{ nm}$) perpendicular to its axis. A light detector measures the intensity decline caused by the individual particles, i.e. the “shadow” of each particle. This makes it possible to count the particles and measure their equivalent diameters. A light detector perpendicular to the laser ray receives the fluorescence light emitted by each particle ($\lambda = 670\text{--}750 \text{ nm}$). The excitation/emission wavelengths of the 5- μm spheres used for this study are 645/680 nm. Natural organic particles show fluorescence at shorter wavelengths, so that the scarlet microspheres can be identified unambiguously. Even single scarlet microspheres can be detected in samples that contain large quantities (e.g. $> 10^5/\text{mL}$) of natural particles.

Solute tracers and analyses

The fluorescent dyes uranine and sulforhodamine G (sulfoG) were used as solute tracers. The monitoring included automatic water sampling and continuous online measurement. The samples were analysed in the CHYN lab using a spectral fluorimeter (PERKIN ELMER LS 50 B). The continuous uranine measurement in the Yverdon site was done with a flow-through fluorimeter (GGUN-FL30) that also records turbidity, TOC and temperature.

Natural parameters

In the Hölloch test site, the discharge (Q) at the main spring was measured two times during the tracer test using

the salt-dilution method. In the Yverdon site, various natural parameters, including Q, T, EC, turbidity and TOC, have been monitored continuously at a swallow hole and at two karst springs (see below). The employed techniques and the results are described by PRONK *et al.* in this volume.

3. Shallow karst system: Hölloch

Description of the test site

The 9343-m-long Hölloch, Germany’s largest cave, is situated in the Hochifen-Gottesacker area in the Austro-German Alps. Geologically, it belongs to the Helvetic Säntis nappe and consists of Cretaceous sedimentary rocks. The Schrattekalk limestone (100 m thick) forms a karst aquifer overlying a marl aquiclude. In most parts of the area, the aquifer basis is above the hydrologic base level so that the underground drainage generally occurs near the top of the underlying aquiclude (shallow karst). Previous tracer tests with a total of 16 injection points demonstrated that the plunging synclines form the main flow paths while anticlines act as groundwater divides (GOLDSCHIEDER 2005).

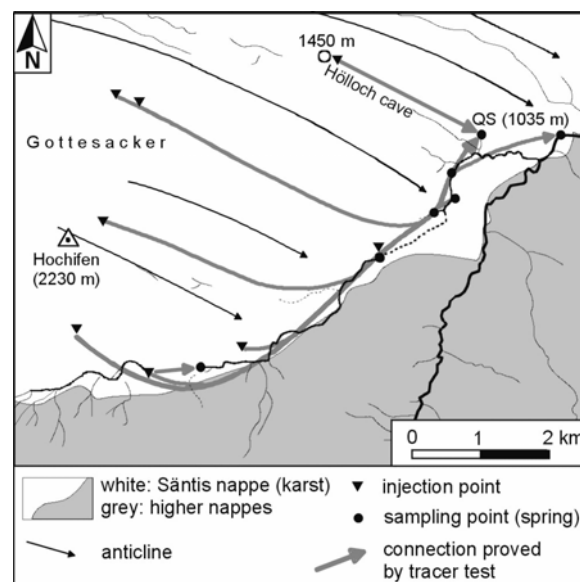


Fig. 2: Simplified hydrogeological map of the eastern Hochifen-Gottesacker area with the Hölloch cave entrance and the Sägebach spring (QS). Only the tracer test between these two points is described here (modified after GOLDSCHIEDER 2005).

At an altitude of 1450 m, an 80-m deep shaft provides the entrance to the cave, which follows an ESE-wards plunging syncline (Fig. 2). Turbulent open-channel flow can be observed in the cave, but there are also small lakes and siphons. Permanent saturation has to be expected in the lowermost, inaccessible part of the cave. The cave system is drained by a permanent spring (Sägebach) at 1035 m, with a discharge of about 150–2000 L/s. Previous tracer tests showed that this spring receives additional inflow from the entire eastern Hochifen-Gottesacker area.

Comparative tracer test

On the 26.02.05, a tracer test was carried out in the Hölloch-Sägebach system during low-flow conditions, in order to obtain better information on the cave hydrology, and to compare solute and particle transport in a shallow

karst aquifer. 4 tracers were injected at the nearly same location in the cave stream, 2.5 km upstream from the spring: 200 g uranine, 200 g sulfoG, $9.1 \cdot 10^{10}$ YG 1- μ m spheres and $3.2 \cdot 10^9$ scarlet 5- μ m spheres. An auto-sampler was installed at the spring; several other sites were sampled manually.

The main tracer breakthrough occurred on the 01.03.05. On this day, a spring discharge of 172 L/s was measured. This value was taken to calculate the recovery rates.

The fluorescent dyes were first detected 66.4 h (uranine) and 69 h (sulfoG) after the injection, and peaked after 98.3 and 93 h respectively. The breakthrough curves have a regular shape, a single maximum and a short tailing, which indicates advective-dispersive solute transport (Fig. 2). The recovery rates are 99 % for uranine and 51 % for SulfoG. The quasi-total uranine recovery suggests that the Sägebach spring is the only relevant outlet of the cave stream. The lower recovery for sulfoG suggests tracer loss due to adsorption on cave walls and sediments and/or suspended clay minerals. Rhodamine dyes are known to be more prone to adsorption than uranine (Käss 1998).

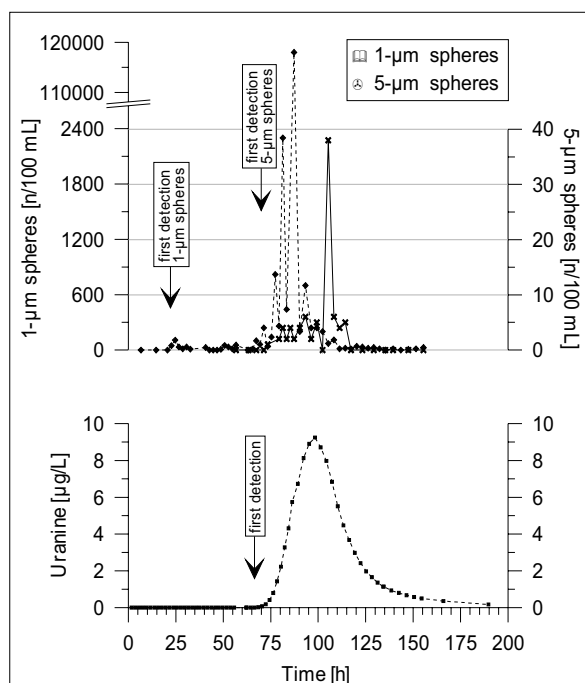


Fig. 2: Results of the comparative tracer test in the Hölloch-Sägebach shallow flow system (without sulfoG).

The curves of the microspheres display an extremely irregular shape, including several sharp peaks. The first 5- μ m sphere was detected 73 h after the injection, which coincides well with the first detection of the solute tracers. In the following samples, the contents varied between 1 and 6 particles/100 mL. An aberrant value of 38/100 mL was detected in a sample taken 105 h after injection. Due to the highly irregular curve, the determination of the recovery rate is ambiguous. The calculated value is 27 % when the aberrant value is removed, and 48 % when it is included.

The 1- μ m spheres were first detected after 23 h, long before the first arrival of the solute tracers. In the subsequent samples, the contents vary between 0 and 107 spheres per 100 mL. A continuous but highly irregular breakthrough started 65 h after injection. An aberrantly high content (118.000/100 mL) was found in a sample taken 87 h

after injection; only 440 and 200 spheres per 100 mL were detected in the previous and subsequent samples. The recovery rate is 118 % when this aberrant value is removed but unrealistically high when it is included in the calculation.

The maximum of the 1- μ m spheres occurs before the maximums of the solute tracers, which suggests that the particles travel at a higher mean velocity. The solute tracers show a short tailing effect, while the particle contents rapidly decrease after the maximum has been reached. The extreme irregularity of the microspheres curves makes it difficult to determine transport parameters, like the Peclet number or dispersivity, and to compare the curves more quantitatively with those of the fluorescent dyes.

4. Deep karst system: Yverdon

Description of the test site

The second test site is a deep karst aquifer system consisting of Jurassic and Cretaceous carbonate rocks, located near Yverdon-les-Bains at the southern slope of the Swiss Jura Mountains. A swallow hole (Feurtille) draining an agricultural surface frequently causes microbial contamination at the Moulinet spring in a distance of 4.8 km, and at another spring (Cossaux) (Fig. 3). Sandstones and marls confine the karst aquifer between the swallow hole and the springs. Fully saturated and pressurised conduit flow in a depth of several hundred metres has to be expected along the flow paths. For details see PRONK *et al.* in this volume, and PRONK *et al.* (2006).

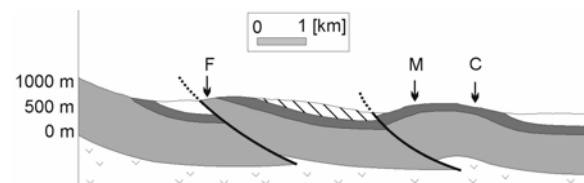


Fig. 3: Hydrogeological cross section of the Yverdon test site. Jurassic: light grey, Cretaceous: dark grey, F: Feurtille swallow hole, M: Moulinet spring, C: Cossaux spring.

Comparative tracer test

Three tracer tests with uranine during low, medium and high-flow conditions proved connection between the swallow hole and the springs and made it possible to better characterise flow and solute transport in the system. The times of first tracer detection at the Moulinet spring range between 40 h (high-flow) and 292 h (low-flow).

The tracer test during average-flow conditions additionally aimed at comparing particle and solute transport. On the 02.06.05, 1 kg of uranine, $2.28 \cdot 10^{11}$ red 1- μ m spheres (5 mL) and $2.84 \cdot 10^{10}$ yellow-green 2- μ m spheres (5 mL) were injected into the swallow hole. All tracers were detected at the two springs; only the results for the Moulinet spring are presented here. Uranine was first detected after 86.4 h and peaked after 103.8 h. The curve shows a regular shape with a single maximum and a long tailing (Fig. 4). Contrastingly, but similar to the findings in the Hölloch, the microspheres display irregular breakthrough curves. The 1- μ m spheres were first detected 85.4 h after the injection, i.e. 1 h before the first detection of uranine. The concentrations during the breakthrough mostly varied between 0.5 and 40/100 mL; an aberrant value (1200/100 mL) was found in a single water sample.

The maximum occurred before the uranine maximum, and the breakthrough essentially ended after 105 h, approximately at the time when uranine reached its maximum. The 2- μm spheres were first detected after 94.4 h. The breakthrough curve is irregular and discontinuous with concentrations of only 0.5 to 3.8/100 mL.

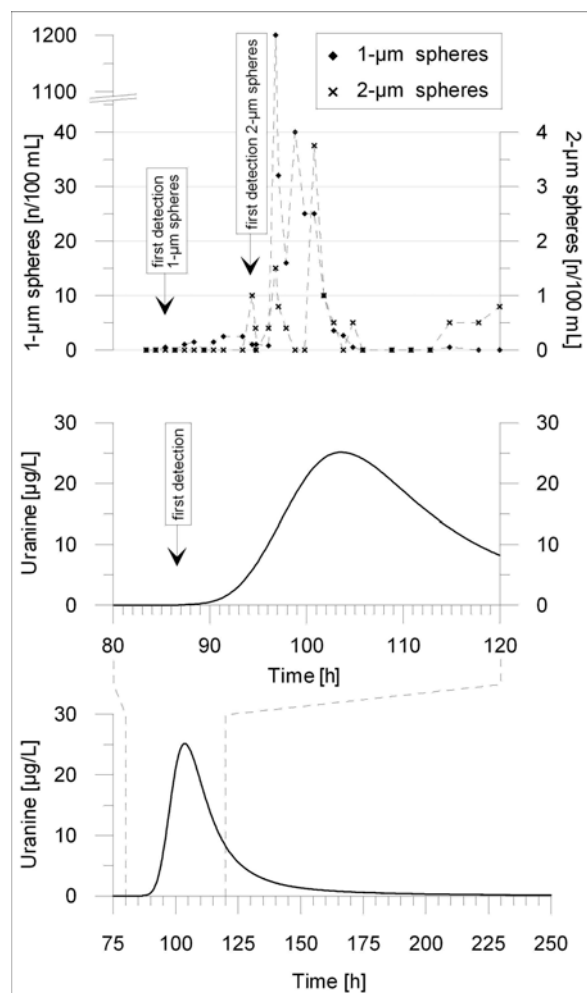


Fig. 4: Results of the comparative tracer test during medium-flow conditions in the Yverdon karst aquifer system.

4. Discussion and Conclusions

Comparative tracer tests with fluorescent dyes and fluorescent microspheres with different diameters have been carried out in a shallow and in a deep karst aquifer system in order to compare the transport of solute and particulate/colloidal contaminants.

In both tests sites, the solute tracers gave regular breakthrough curves with a single maximum, followed by a more or less pronounced tailing. Amongst the particle tracers, the clearest results were obtained with the 1- μm spheres (comparable to the size of pathogenic bacteria). They arrived before the first detection of uranine, and peaked before the uranine maximum; their breakthrough curves are highly irregular and show no significant tailing.

The results of the spring water monitoring in the Yverdon test site are consistent with this finding: The maximum concentrations of particles and faecal bacteria often precede those of TOC and nitrate (PRONK *et al.* in this volume).

The 2- μm and 5- μm spheres also arrived at the springs, but their concentrations were generally very low, which is a result of lower injection quantities and, mainly for the 5- μm spheres, sedimentation and/or filtration processes. Due to the very low concentrations, the curves are influenced by statistical effects and cannot be interpreted quantitatively. However, the results clearly show that particles in the size of *Cryptosporidium* cysts (5 μm) can be transported in the groundwater over distances of several km.

The tracer tests in the two test sites gave quite similar results, although their hydrogeological setting is very different: The Hölloch belongs to a shallow alpine karst area with turbulent open-channel flow, while Yverdon is a deep karst system with pressurised flow in saturated conduits.

A similar behaviour of particle vs. solute tracers has often been observed in porous media, mainly on a lab scale, and is explained by different types of exclusion processes, i.e. the particles are mainly transported in the pore centres where flow velocities are higher. In karst aquifers and on a field scale, these processes have been studied much less.

The experiments described in this paper also included the use of an innovative fluorescence-particle counter, which allows single microspheres to be identified and counted in water samples that contain high numbers of suspended particles. To date, only microspheres $\geq 5 \mu\text{m}$ with a specific fluorescence (645/680 nm) can be detected unambiguously. For particle tracer tests on a field scale, smaller microspheres are preferred because their transport properties are more favourable, and a much larger number of particles can be injected for the same price.

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Fonctionnement et classification des systèmes aquifères karstiques du synclinal du Dyr – Tébessa (Algérie) par l'approche fonctionnelle.

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Résumé

Dans cet article, nous présentons les synthèses d'application des méthodes d'approche fonctionnelle dans le traitement et l'analyse des chroniques des débits d'écoulement à l'exutoire des systèmes aquifères karstiques du synclinal du Dyr qui se situe au nord est du territoire algérien et faisant partie du bassin d'effondrement de Tébessa. L'étude descriptive des débits classés de l'écoulement des principales exurgences, suivant des représentations graphiques qui répondent à un ajustement conformément à la loi normale de Gauss, permet l'identification du fonctionnement de l'exutoire de l'aquifère "épikarst" de l'Eocène et celui de l'aquifère "holokarst" du Maestrichtien. L'analyse des courbes de récession permet, selon le modèle à deux réservoirs (MANGIN A. 1970), la caractérisation du fonctionnement de la zone d'infiltration par la détermination des coefficients d'hétérogénéité d'écoulement (ϵ) et de vitesse d'infiltration (η), elle permet ainsi de définir le comportement de la zone noyée en évaluant le coefficient de tarissement (α).

Le report des valeurs des paramètres hydrodynamiques i et k qui représentent respectivement les retards à l'infiltration (i) et le rapport entre le volume dynamique maximum obtenu sur la période d'étude et le volume de transit interannuel sur la même période (k), sur le diagramme de classification de Mangin (1996) permet de définir le modèle karstique analogue.

Abstract

In this article, we present the application syntheses of the methods of functional approach in the treatment and the analysis of the mass flow chronicles at the discharge of the karstic aquifer systems of the Dyr synclinal located at the north east of the Algerian territory and forming part of the collapse basin of Tébessa. The descriptive study of the classified mass flows of the principal exurgences, following graphical chart which agrees with an adjustment according to the normal law of Gauss, allows the identification of the operation discharge system of the aquifer "épikarst" of the Eocene as well as Maestrichtien aquifer "holokarst". The analysis of the curves of recession allows, according to two tank MANGIN model (1970), the characterization of the operation of the infiltration zone by the determination of the heterogeneity flow (ϵ) coefficients and infiltration speed (η), it thus makes it possible to define the behaviour of the drowned zone by evaluating the running dry coefficient (α). Reporting the values of the hydrodynamic parameters I and K which respectively represent the delays to the infiltration (i) and the ratio between the maximum dynamic volume obtained on the study period and the inter annual volume transit over the same period (K), on the classification Mangin diagram (1996), makes it possible to define the similar karstic model.

1. Introduction

Parmi les pays karstiques de l'Algérie est le synclinal du Dyr. Ce dernier, couvrant une superficie de 162 km², est situé à vingt kilomètres des frontières Algéro – Tunisiennes. Il constitue la bordure nord-est du bassin d'effondrement de Tébessa. Les travaux d'étude réalisés dans cette région ont permis de déceler deux systèmes aquifères : Le premier, constitué par des calcaires fissurés d'éocène, est de type épi karst; Le second, constitué par des calcaires fissurés du Maestrichtien est de type holokarst. Ces études ont révélé une complexité du mécanisme hydrodynamique des deux réservoirs. La nature lithologique, la géométrie et la variabilité hydroclimatique sont les principaux indices de détermination de l'ampleur de cette complexité qui restait indéfinie. Comme démarche, il y'a lieu de confirmer que ces deux systèmes sont karstiques. Ensuite, on essayera de faire déceler les principaux points recherchés : La présence d'impluvium fluctuant aux travers des fuites ou apports éventuels, pouvant se produire entre différents systèmes ; Les conditions de mise en fonctionnement de trop-plein ; L'estimation des volumes de transit (volume dynamique) et aboutir à une classification des dits systèmes selon une typologie mondiale.

2. Matériels et méthodes

L'application des méthodes de l'approche fonctionnelle est un moyen d'étude des réservoirs karstiques (CROCHET & MARSAUD 1996). On se limitera aux présentations des résultats de l'analyse descriptive des débits classés pour la compréhension du fonctionnement de l'exutoire et ceux de l'analyse des courbes de récession

pour la classification des systèmes. Ces méthodes s'appuient sur le traitement des données de vidange des aquifères à leurs exurgences. On dispose de données de deux stations hydrométriques, sur une période de vingt six cycles hydrologiques (1974/1975 à 1999/2000), d'écoulement de deux sources Ain Erkel et Ain Zerga qui représentent respectivement les vidanges des systèmes aquifères karstiques de l'Eocène et du Maestrichtien. Les débits journaliers ont été déduits à partir de l'établissement de la courbe de tarage relative à chaque cycle hydrologique, liant les niveaux journaliers de la lame d'eau écoulée et les débits instantanés jaugés.

3. Résultats

3.1. Identification du fonctionnement de l'exutoire : Le traitement par voie automatique, à l'aide du logiciel HYDROLAB qui permet de tracer le graphe de la distribution d'une population selon la loi normale de Gauss, des valeurs de débits journaliers relatives à cette période d'observation, a permis d'obtenir, pour les deux systèmes, des représentations graphiques qui répondent à un ajustement gaussio-arithmétique (Fig. 1). Cet ajustement traduit une infiltration lente. En se référant aux cinq cas les plus fréquents de rupture de pente de la droite représentative des débits classés, l'analyse descriptive de l'ensemble des courbes de débits classés des deux exutoires montre que :

- Le système aquifère éocène se caractérise, par une décrue dont la représentation graphique est une droite à simple rupture répondant à l'expression : $\alpha_2 < \alpha_1$ où le modèle dominant est celui de type 2 (Tabl. 1) ;
- Le système aquifère maestrichtien se caractérise par des décrues de vidange dont les courbes représentatives sont aussi des droites : - A simples ruptures, obéissant aux expressions : $\alpha_2 > \alpha_1$, $\alpha_2 < \alpha_1$ et à doubles ruptures, conformément aux l'expressions: $\alpha_2 > \alpha_1$ et $\alpha_3 < \alpha_1$ (Tabl. 2).

Selon les modèles de référence, le fonctionnement de l'exutoire du système aquifère maestrichtien est très variable dans le temps où l'on assiste à une intermittence d'apports en provenance d'un autre système, apport d'une réserve issue d'un cycle antérieur, mise en fonctionnement de trop-plein, stockage momentané ainsi qu'aux phénomènes de piégeage d'une réserve lors de la décrue et les processus de constitution de réserves. La complexité du fonctionnement de l'exutoire du réservoir Maestrichtien est donc plus palpable par rapport à celui de l'Eocène plus simple.

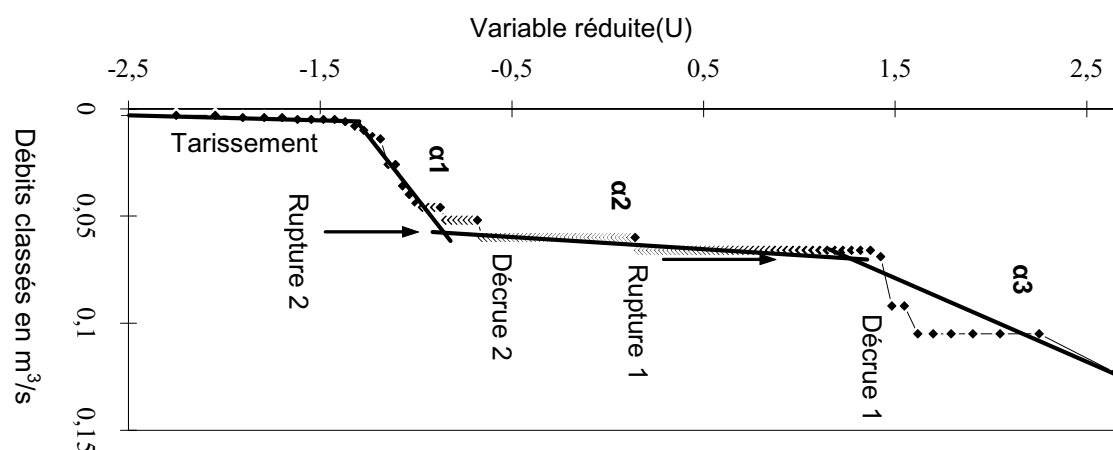


Fig. 1 Exemple de courbes des débits classés de l'écoulement de la source Ain Zerga

Nature de La pente	Expression	Modèle de rupture	Nombre de cycles	Interprétation hydrogéologique
Pente à simple rupture	$\alpha_2 < \alpha_1$	Type 2	15	- Apport en provenance d'un autre système - Jaugeage d'écoulement n'appartenant pas au système
		Type 4	03	- Apport d'une réserve issue d'un cycle antérieur
	$\alpha_2 > \alpha_1$	Type 3	02	- Constitution d'une réserve

Tabl. 1 Cas de rupture de pente de la droite représentative de débits classés de la Source Ain Erkel

Nature de la pente	Expression	Modèle de rupture	Nombre de cycles	Interprétation hydrogéologique
Pente à simple rupture	$\alpha_2 > \alpha_1$	Type 1	03	- Mise en fonctionnement de trop-plein - Fuites vers un autre système - Stockage momentané - Fuites ou débordement de la station de Jaugeage en hautes eaux
		Type 3	02	- Constitution des réserves
	$\alpha_2 < \alpha_1$	Type2	05	- Apports en provenance d'un autre système - Jaugeage d'écoulement n'appartenant pas au système
		Type4	05	- Apports d'une réserve issue d'un cycle antérieur
Pente à double rupture	$\alpha_2 > \alpha_1$ $\alpha_3 < \alpha_1$	Type5	02	- Piégeage d'une réserve lors de la décrue et restitution au cours du tarissement

Tabl. 2 Cas de rupture de pente de la droite représentative de débits classés de la source Ain Zerga

3.2. Fonctionnement de la zone d'infiltration et celle noyée:

3.2.1. Principe de la méthode : L'aquifère est assimilé à un réservoir se vidangeant au travers d'un bouchon poreux. On ne s'intéresse qu'à la partie décroissante de l'hydrogramme de crue d'une source. Le modèle utilisé possède deux réservoirs : L'un d'infiltration se vidange avec un débit d'infiltration (q_i), l'autre de type exponentiel correspond à la zone noyée et se vidange avec un débit Q . De ces courbes de récession, on peut distinguer : - La décrue, durant laquelle l'infiltration (q_i) influence la vidange du réservoir ; - Le tarissement, qui débute à l'instant où le débit d'infiltration (q_i) devient nul. La vidange du réservoir principal obéit alors uniquement à la loi de MAILLET. L'expression globale utilisée est de la forme : $Q(t) = \varphi(t) + \psi(t) \dots (1)$, avec $\varphi(t) = Q_{R0} \cdot e^{-\alpha t} \dots (2)$ décrivant la vidange de la zone noyée (MAILLET). $\psi(t) = q_0 (1 - \eta t) / (1 + \varepsilon t) \dots (3)$ décrit une fonction homographique se rapportant au fonctionnement de la zone d'infiltration. Avec q_0 : débit d'infiltration à l'instant t marquant la pointe de crue ($q_0 = Q_{MAX} - Q_{R0}$) ; ε : coefficient d'hétérogénéité d'écoulement, décrivant la concavité de la courbe et variant entre deux fourchettes, la première (0.001 à 0.01) indiquant une infiltration lente, et la seconde (1 à 10) traduisant une décrue très rapide ($\varepsilon = Q_{R0} - Q_i$) ; Q_{R0} : débit à l'instant t_0 ; η : coefficient de vitesse d'infiltration est égale à $1/t_i$ correspondant à l'instant où l'infiltration cesse, il est assimilé à la vitesse moyenne d'infiltration, et varie entre 0 et 1 ; quand η tend vers 1, il traduit une infiltration rapide. Le coefficient de tarissement (α) permet de calculer le volume dynamique correspondant au volume d'eau en mouvement au moment du tarissement, aussi bien au dessus qu'en dessous de l'exutoire (MANGIN 1970). Ce paramètre (V_{dyn}) est une estimation par défaut des réserves. Il est donné par l'expression : $V_{dyn} = \int Q_i \cdot e^{-\alpha t} = Q_i \cdot c / \alpha \dots (4)$ Q_i : débit où le tarissement devient effectif ; t_i : l'instant où l'infiltration cesse ; c : constante, prise égale 86 400.

3.2.2. Fonctionnement des zones d'infiltration et celles noyées : Conformément à l'expression (1) et à l'aide d'un traitement des données, on a pu obtenir des représentations graphiques dont les essais d'ajustement selon la fonction exponentielle n'ont permis de retenir que dix huit courbes de récession pour Ain Erkel et onze courbes de récession pour Ain Zerga (Fig. 2).

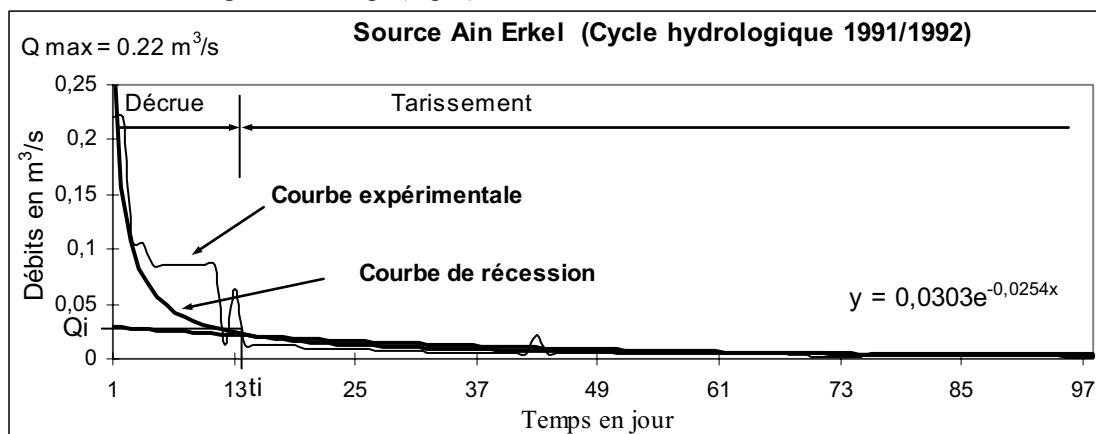


Fig. 2 Exemple d'ajustement des courbes de récession de la source Ain Erkel

Le calcul des paramètres hydrogéologiques a donné des éléments caractéristiques du fonctionnement des zones d'infiltration et celles noyées des deux réservoirs (Tabl. 3). Les très faibles valeurs de (ε) repérées à Ain Erkel ($\varepsilon_{moy} = 0.0095$) sont synonymes d'une infiltration épi karst lente au sein du système éocène. Pour le Maestrichtien, une variabilité de vitesse d'infiltration est notée où ε oscille entre deux fourchettes : L'une de très faible valeur $0.001 < \varepsilon < 0.01$ indiquant une infiltration lente; La seconde, de valeur plus importante $0.01 < \varepsilon < 1$

signe d'une infiltration moyennement lente. D'autre part, les valeurs de (η) convergent ou tendent vers zéro, ce qui confirme la persistance de l'infiltration lente. Les variations interannuelles des valeurs de tarissement (α), ainsi que celles du volume dynamique, s'expliquent par des régimes d'écoulement influencés au sein des zones et sont certainement dues aux augmentations des pertes de charge à l'amont de chaque exutoire. Les valeurs du volume dynamique varient de 0.056 à 1.728 Mm³ et les paramètres d'ajustement sont dispersés ce qui peut constituer un indice de la présence de réserves importantes. Les zones noyées sont alors en régime influencé durant une grande période.

Paramètres	Source Ain Erkel			Sources Ain Zerga		
	Min	Moy	Max	Min	Moy	Max
Coefficient de tarissement (α)	0.0033	0.0109	0.0263	0.0009	0.0275	0.076
Volume dynamique en 10 ⁶ m ³	0.0748	0.3614	0.9850	0.0560	0.3784	1.728
Coefficient d'hétérogénéité (ε)	0.0011	0.0095	0.0716	0.0002	0.0283	0.175
Coefficient de vitesse d'infiltration (η)	0.0357	0.0568	0.1100	0.0290	0.0854	0.143

Tabl. 3 Paramètres statistiques de l'ajustement des courbes de récession des sources Ain Erkel et Ain Zerga

3.3 Classification des systèmes aquifères : Elle passe par le calcul des retards à l'infiltration "i", par application de l'expression $i = (1 - \eta t)/(1 + \eta t)$ et par le calcul de k, qui représente l'aptitude de la zone noyée du karst à stocker les précipitations et les restituer progressivement dans le temps, selon le rapport entre le volume dynamique maximum obtenu sur la période d'étude et le volume de transit interannuel sur la même période ($k = V_{\text{dyn max}}/\sum V_{\text{dyn}}$). Le report des valeurs de ces deux paramètres sur le diagramme de classification (MANGIN, 1970), ont révélé que les deux systèmes aquifères carbonatés sont karstiques ($k < 0.5$) et complexes. Ils sont caractérisés par une géométrie de grande taille avec de nombreux sous-systèmes, malgré la nette diversité dans leurs comportements. Ils possèdent un fort pouvoir régulateur et des retards à l'alimentation importants traduisant la présence de réserves. Ils sont alors analogues au modèle de la fontaine de Vaucluse (fig. 3).

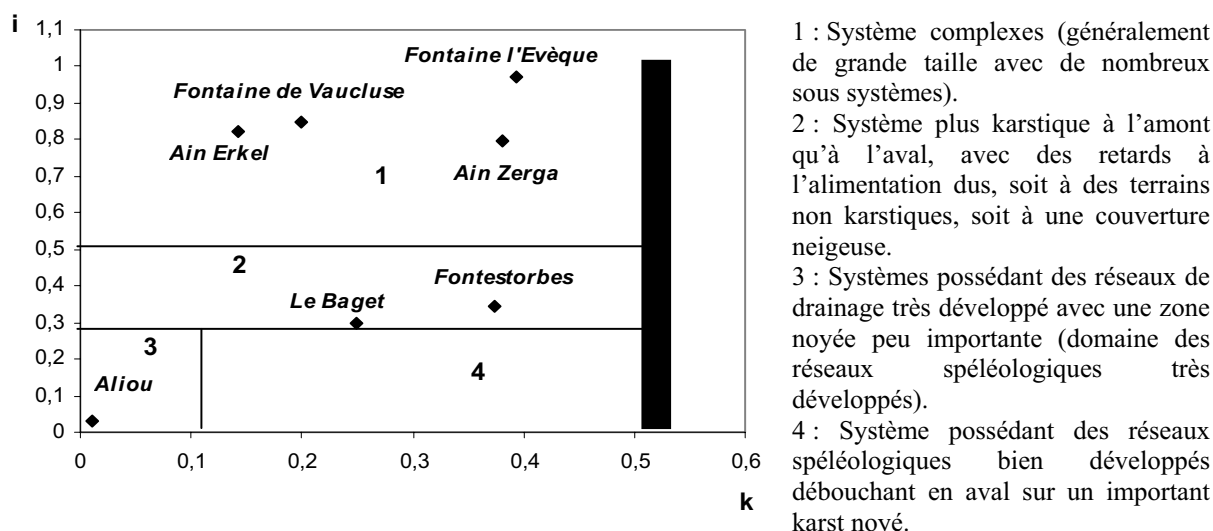


Fig. 3 : Classification des systèmes karstiques éocène et maestrichtien du synclinal du Dyr (Tébessa – Algérie).

4. Conclusion

L'essai d'application des méthodes de l'approche fonctionnelle aux systèmes karstiques du synclinal du Dyr, a permis de montrer une complexité très significative du comportement de l'exutoire du réservoir maestrichtien, tandis que celui de l'éocène est plus simple. Une persévérance des apports d'alimentation en provenance d'autres systèmes annexes a été mise en évidence. Selon les valeurs calculées des paramètres caractéristiques du milieu, une infiltration lente à moyennement lente au sein du système maestrichtien a été visualisée. Un régime d'écoulement influencé dû aux augmentations des pertes de charge à l'amont de chaque source est perceptible. Les deux réservoirs en question sont analogues au modèle de la fontaine de Vaucluse.

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A template for academic/NGO partnership in the evaluation of cave and karst resources

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Abstract

Western Kentucky University's Hoffman Environmental Research Institute works with both US government agencies and non-governmental organizations (NGO's) in the US and China to assist with karst resource evaluation. This paper presents a case study focusing on collaboration with The Nature Conservancy to evaluate the extent, resource, and environmental quality of caves on three tracts of land in Metcalfe, Green, and Adair counties, Kentucky, USA. These caves have formerly received little scientific attention, being overshadowed by the world-class systems of the Mammoth Cave Plateau just to the west. There is synergy in collaborations between such NGOs who provide the legal and financial means to protect such resources, and academic agencies that provide technical expertise to evaluate them. Simultaneously, students gain technical experience in the environmental sciences.

Field methods involved in the project included cave survey and inventory, biologic specimen collection, fluorescent dye tracing, water quality analysis, and photo documentation. Geographic Information Systems (GIS) tools were used for spatial and attribute data organization and visualization. In one case, these activities led to a large grant to set aside a 40-hectare tract containing a picturesque gorge and several key caves, aiding in water quality protection for the Little Barren River.

1. Introduction

Western Kentucky University's (WKU) Hoffman Environmental Research Institute, housed within WKU's Applied Research and Technology Program works with both US government agencies and non-governmental organizations (NGO's) in the US and China to assist with karst resource evaluation. Within the context of this work the Hoffman Institute has been steadily building a relationship with the Kentucky chapter of The Nature Conservancy. The Hoffman Institute studies best management practices and other resource protection methodologies in karst and other rural areas in order to enhance environmental quality. The Institute is also involved in developing specialized GIS tools to support the projects involved with resource management, particularly with regard to karst. The Nature Conservancy is a leading international, nonprofit organization dedicated to preserving the diversity of life on Earth. They strive to preserve plants, animals and natural communities by protecting the lands and waters they need to survive. The organization is sectioned into chapters, representing the state they are located in; the specific section for this project area is the Green River (Kentucky) Basin.

The Institute originally interacted with The Nature Conservancy through cooperative involvement in a grant to purchase about 60 ha of rural land for Western Kentucky University to protect an entrance to the Mammoth cave System that lies outside of Mammoth cave National Park. The relationship was built through subsequent work done in Metcalf County, Kentucky on the caves of the Dry Fork area.

In 2005 the Nature Conservancy instigated a new collaboration by inviting Institute personnel to pay an initial visit to two sites to evaluate the potential for two caves that TNC was considering protection strategies for, one located in Green County and the other in Adair County, Kentucky.

Monin cave is located in Green County, Kentucky, near the town of Crailhope. Garnett cave is located outside of Columbia, Kentucky in Adair County. Monin cave is located in the lower Carboniferous aged St. Louis limestone, and Garnett Cave is formed in the older, but still lower Carboniferous, Fort Payne Formation.

Methods

A work plan was developed with The Nature Conservancy to evaluate the two cave systems. A cave survey was conducted following standards set by the Hoffman Institute. The GPS data were taken by a Garmin Legend with accuracy of up to 10 m. Along with the cave survey, a resource inventory was completed. Water chemistry measurements including temperature, pH and conductivity were taken at both locations. Photo documentation was performed on all trips. A dye trace was performed at Monin Cave to connect the main cave system with a spring that appeared to be draining the system. TNC also asked the Institute to perform a biologic collection of stygobites from both caves. The samples were then forwarded to TNC contract scientists for identification

Results

After the completion of the cave survey, the information was compiled into the cave survey program Compass. It was georeferenced using the GPS locations acquired in

the field. This data were then used to complete the cartography of each of the caves. Due to the size of Monin cave, a quadrant map format was used. The Compass data were also exported as a shapefile. One shapefile was used to import the line plot onto a topographic map of the region in ArcGIS. Other shapefiles were modified to be used as a catalog for the resource inventory. These files were then added as layers in ArcGIS. An interactive map was created in ArcGIS displaying different inventory layers, such as locations of different speleothems in the cave. The results from the dye trace at Monin Cave provided a positive conclusion that the stream in Monin Cave was the stream that came out at Monin Spring. This was added as another layer to the interactive GIS map.

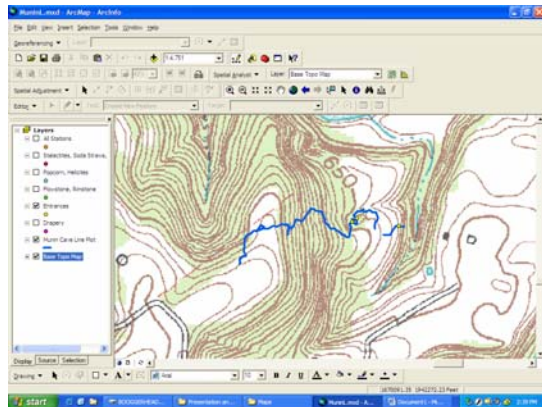


Figure 1: Inventory line plot in ArcGIS.

Future Work

The above details the preliminary work done at Garnett and Monin caves, and future work will directly correlate with the needs of The Nature Conservancy. It is planned to attempt a direct physical connection and survey between Monin Cave and Monin Spring. When the dye trace was performed, a visual confirmation was expected, due to the locations on the topographic map. A positive result was obtained, but not in the time frame expected. Investigation of the passage in-between might answer the time question. An overland survey will also be conducted to determine the accuracy of the GPS placed points.

More information about Garnett Cave also needs to be obtained. There is a local rumor that there is another entrance to the cave, which was not found during survey. The topographic map shows an old homestead with a possible spring source in the direction of the survey. A clean-up project for this cave will be coordinated with the local cave club to take out glass and other debris that has washed into the cave.

Conclusions

The Hoffman Institute worked with The Nature Conservancy to develop and conduct a work plan for the evaluation of Monin Cave and Garnett Cave. The collaboration in this project is beneficial to both parties. Field methods and technical skills are built by students and associates of the Hoffman Institute that are transferable to other projects. It also builds a relationship with other agencies and the Institute is involved in providing services to the Commonwealth of

Kentucky. The Nature Conservancy benefits by gaining knowledge about the systems they are seeing to protect. It helps drive the conservation-decision process for where they should work and what is required to conserve certain systems.

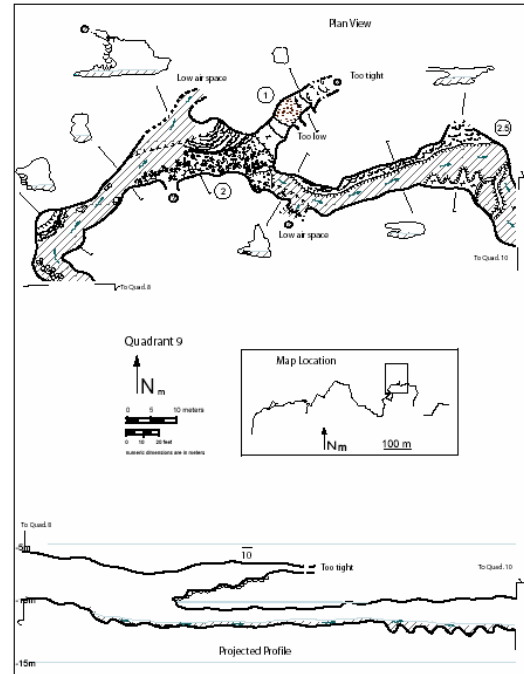


Figure 2: Finished Quadrant 9.

There is synergy in collaborations between such NGOs who provide the legal and financial means to protect such resources, and academic agencies that provide technical expertise to evaluate them. Simultaneously, students gain technical experience in the environmental sciences.

Acknowledgements

We acknowledge the Center for Cave and Karst Studies for assistance in the dye trace. We also thank Joel Despaigne, Brad Hacker, Carrie Crockett, Pete Reutter, Andrea Croskrey, Angél Cortes, Wendy Breatheren, Mark Phillips, Todd Armstrong, Jeremy Tallent, and Rick Olson for assistance with the cave survey.

Analysing karst spring response to microbiological contamination in the framework of a national groundwater monitoring network

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Abstract

The National Groundwater Quality Monitoring Network of Switzerland (NAQUA) comprises various types of sampling sites in different aquifers, including some typical karst springs. Base parameters, such as discharge, temperature and electrical conductivity form the basis for the interpretation of the chemical and isotopic composition of the periodically sampled groundwater. Furthermore, the continuous data form the essential base for the interpretation of special studies, such as investigating the occurrence of bacteria, protozoa and viruses in groundwater. The present paper illustrates the occurrence of viruses and protozoa in response to discharge of a karst system.

1. Introduction

In Switzerland, a national network for the monitoring of groundwater quality has existed since 2002 (NAQUA). The network includes approximately 500 observation sites (SAEFL/FOWG, 2004). They were chosen by criteria such as type of aquifer or land use in the recharge area. Groundwater sampling at the observation sites is conducted regularly throughout the year. The water is analysed for various organic and inorganic compounds.

NAQUA_{TREND} (NT), the core of the monitoring network includes 50 observation sites (GREBER *et al.* 2002). Here, in addition to contaminants, major elements are analyzed in the water samples, and other parameters such as discharge, electrical conductivity, temperature or turbidity are continuously recorded by data loggers. For every observation site of NT, a detailed documentation on the local geological and hydrogeological situation as well as the land use characteristics is available.

2. Karst springs in the National Groundwater Quality Monitoring Network

The land use in the recharge area of the karst springs of NAQUA_{TREND} in the Jura and Alps is dominated by forest, grassland and livestock farming, and locally by tillage. Hence, typical chemical contaminants are nitrate and pesticides. However, the major quality problems of karst springs result from bacteriological contamination resulting from the combination of specific land use and the hydrogeological characteristics of karst aquifers. The classical analyses for microbiological water quality do not include viruses and protozoa. Until now, little has been known about the occurrence of viruses and protozoa in the groundwater of Switzerland (GREBER *et al.* 2005).

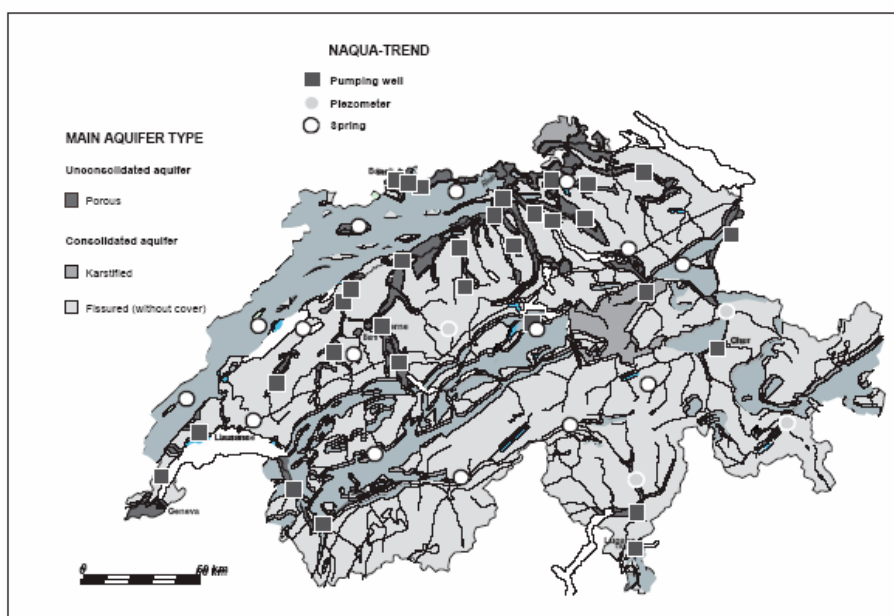


Fig. 1: Location of the 50 observation sites of NT.

In the framework of NAQUA_{TREND}, the occurrence of viruses and protozoa was investigated in various kinds of aquifers. In this pilot study, four karst springs (Jura and Alps), two fracture springs (Molasse) and two pumping wells in porous media were sampled four times within a year (GREBER *et al.* 2005). The selection of sampling sites focused on conditions where the characteristics of land use and hydrogeology indicate a high probability of microbiological contamination, i.e. livestock farming in the recharge area, known waste water component or elevated turbidity during high water events.

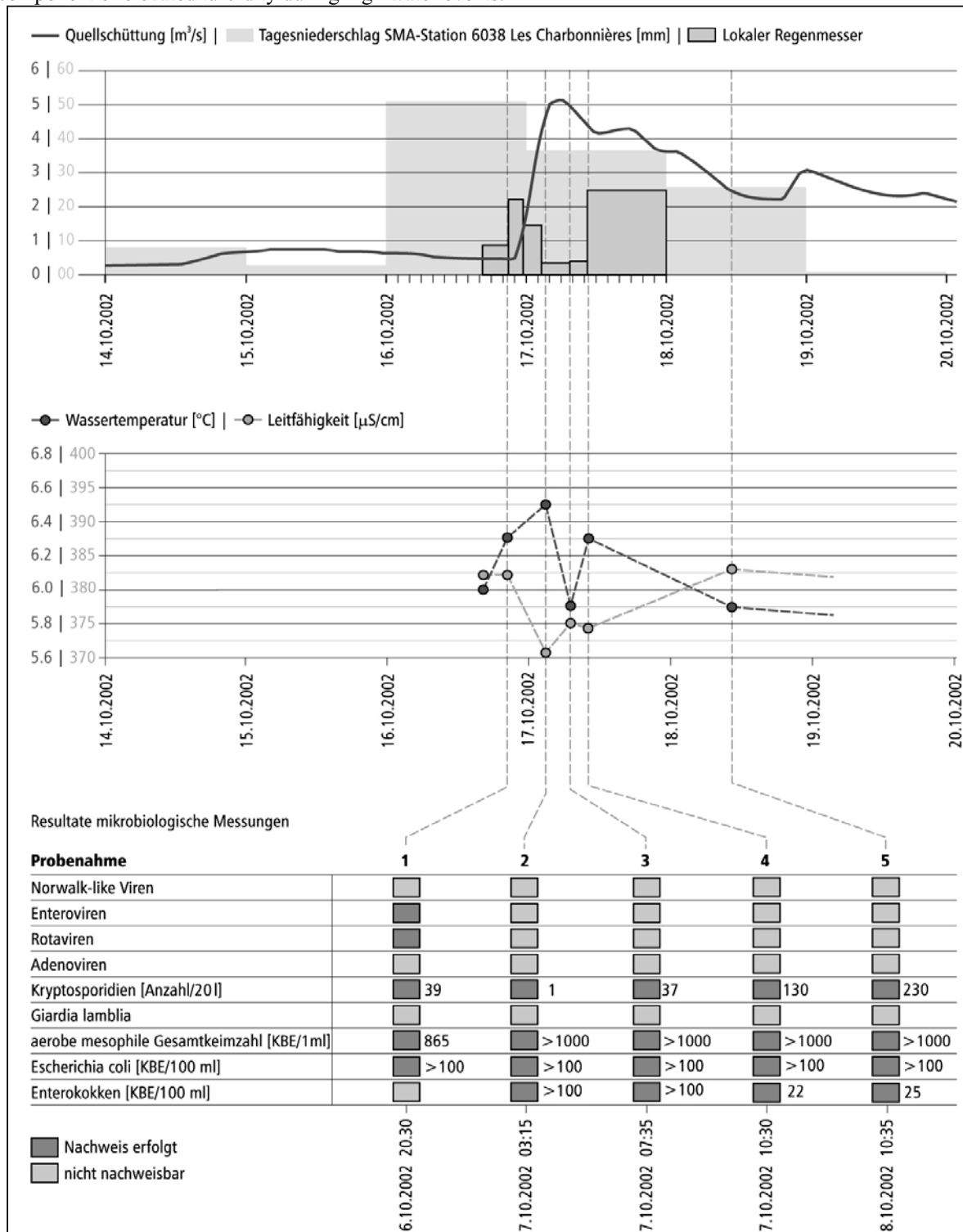


Fig. 2: Significant rise of microbiological contamination during a high water event (GREBER *et al.* 2005).

3. Results

At all of the four karst springs, significant bacterial contamination was observed. Additionally, *Cryptosporidium* as well as entero- and rotaviruses were detected at some of the springs. The analytical methods were used to identify viruses and protozoa but do not indicate viability as infectious agents.

In order to study the occurrence of viruses and protozoa with respect to the spring discharge, one of the karst springs was sampled during a high water event (Fig. 2). The results show that the type and amount of contamination highly depends on the sampling time. *Cryptosporidium* was detected in all of the samples, with the lowest count at peak discharge and the highest one day later. In contrast, viruses were only detected before maximum discharge.

The sampling sites at springs in fissured rocks and pumping wells in unconsolidated sediments show no contamination of viruses and protozoa, with the exception of one sample containing *Cryptosporidium*.

4. Conclusion

Cryptosporidium was regularly detected in karst springs, followed by entero- and rotaviruses. The results show that the type of contamination and the concentration highly depend on the sampling time during a high water event. The results also suggest that the protection areas of the springs are not adequately established or that the land use restrictions in the protection areas are not respected.

In comparison with other hydrogeological settings (GREBER *et al.* 2005), Viruses and protozoa were almost exclusively detected in water from karst springs. Typically, they are observed in those groundwater samples, which already show a generally high bacteriological contamination. But they have also been detected in water samples where the classical bacteriological indicators suggest no relevant quality problems.

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Differentiation of the role of the unsaturated and the saturated zones in the hydrogeological behaviour of carbonate aquifers from Southern Spain, by means of the use of $\delta^{13}\text{C}_{\text{TDIC}}$, Total Organic Carbon (TOC) and several hydrodynamical parameters

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Abstract

This work focuses on the use of Total Organic Carbon (TOC) and Carbon-13 combined with the classical chemical and physico-chemical tracers to study several karst aquifers from S Spain with different lithologic and hydrodynamic contexts. The TOC content originates from the soil and it is an interesting tracer of fast infiltration. TOC content varies antipathetic to those of magnesium, usual indicator of the residence time of water. However, the TOC seems much more sensitive than chemical and isotopic tracers usually used in the study of karstic flows. The complementarity of magnesium and TOC has been used as a relevant tool in order to characterize the behaviour of the aquifers.

In addition, the Carbon-13 of Total Dissolved Inorganic Carbon ($\delta^{13}\text{C}_{\text{TDIC}}$) was a good tracer to identify processes that take place: slow versus fast infiltration, role of the epikarst and unsaturated and saturated zones. It can be considered that water from the unsaturated zone evolve as an open system to the CO_2 , (more depleted $\delta^{13}\text{C}_{\text{TDIC}}$) whilst water from saturated zone evolve in a closed conditions with respect to the biogenic CO_2 (less depleted $\delta^{13}\text{C}_{\text{TDIC}}$).

This research is interesting for a better understanding of the behaviour of these aquifers, but also to characterize karstic flows and to evaluate the vulnerability of the systems.

1. Introduction

Total Organic Carbon (TOC) is commonly used to study local pollutions and to evaluate anthropogenic influences within karstic aquifers, but in the case of unpolluted aquifers, which are not supplied by surface waters, the whole TOC is originated from the soil, so that it is an interesting tracer of fast infiltration (EMBLANCH *et al.*, 1998). If magnesium, usual indicator of the residence time of water (MUDRY, 1987), is homogeneous in the carbonate rock then TOC most of the time follows an opposite variation with it (BATIOU *et al.*, 2003). Consequently, the TOC may be used as a much more sensitive tracer than chemical and isotopic ones (Mg^{2+} , NO_3^- , $\delta^{13}\text{C}$), which are usually used in the study of karst systems. The combined use of all of them allows the characterization of the aquifer behaviour (BATIOU *et al.*, 2003).

EMBLANCH *et al.* (2003) studied different karst systems which belong to the experimental site of Vaucluse and showed that ^{13}C of Total Dissolved Inorganic Carbon ($\delta^{13}\text{C}_{\text{TDIC}}$) was a good tracer to differentiate the types of water that participate in the karstic flow (fast infiltration, unsaturated zone, saturated zone). On the other hand, JIMÉNEZ *et al.* (2005) studied the hydrogeological behaviour of the Sierra de las Nieves (S Spain) by means of TOC and $\delta^{13}\text{C}_{\text{TDIC}}$ and compared this results with those obtained at the Fontaine de Vaucluse (SE France), demonstrating that the use of these parameters appears as a relevant tool to study the hydrogeological response of carbonate aquifers located under different climatic, lithological and hydrodynamical contexts.

This work focuses on the use of TOC and $\delta^{13}\text{C}_{\text{TDIC}}$, combined with the conventional chemical and physical-chemical tracers, to study the main springs of three karst system in Southern Spain with different climatic and geological conditions (lithology, geological structure, karstic development). The aims are to show the importance of these tools to know the aquifer behaviour which could be interesting for groundwater management and contamination vulnerability.

2. Hydrogeological setting and methods

The Yunquera-Nieves is a karst massif located near the Atlantic-Mediterranean convergence (southern Spain), in the Alpine context of the Betic Cordillera (Fig. 1). This hydrogeological unit is made up of dolomitic and calcareous rocks of Mesozoic-Tertiary age (over 1,500 m thick), folded and fractured, which have been strongly karstified (karren, sinkholes and even a polje with swallow holes as well as abundant karstic cavities). The unsaturated zone is more than 300 m thick. In the Sierra de las Nieves aquifer, the average annual precipitation is approximately 1200 mm in the western area and 800 mm in the eastern area. The spatial distribution of precipitation is strongly influenced by the relief in Sierra de las Nieves (ANDREO *et al.*, 2004; LIÑÁN, 2003). The seasonal distribution of rainfall is similar during all years: with local stormy precipitations in autumn, moderate to heavy continuous precipitations during winter and scarce or nearly absent ones during spring and summer. The recharge of this aquifer takes place by the infiltration of rainwater and snowmelt. The existence of large N130E fractures cutting across the unit divides it into distinct

systems (PISTRE *et al.*, 2002; LIÑÁN, 2003). Groundwater discharge occurs basically through springs on the southern edge of the unit (Fig. 1) and through a few wells used for water supply and irrigation. The most important springs are Grande, with an average flow of 625 L/sec, Genal 75 L/sec, Verde 350 L/sec and Jorox 110 L/sec. These springs increase rapidly in flow, from zero to several m³/sec after rainfall (LIÑÁN *et al.*, 1999).

The western sector of Sierra de las Nieves system is the recharge area of Genal spring and the limestone outcrops of Sierra Prieta system are the main area of recharge of Jorox spring. The hydrodynamic and hydrochemical analysis of these springs (LIÑÁN *et al.*, 1999) demonstrate that residence times are short because the calcareous rocks have a developed karst drainage pattern (conduit flow system).

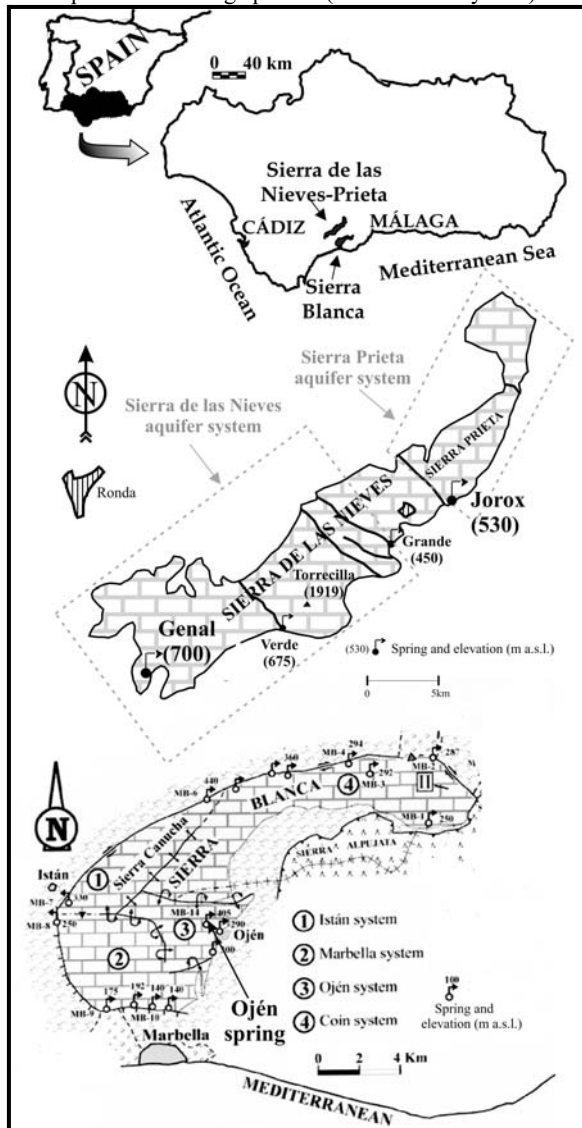


Fig. 1. Location of the Sierra de las Nieves, Sierra Prieta and Ojén aquifer systems.

Sierra Blanca is a carbonate massif in Southern Spain, located at the North of the city of Marbella (Fig. 1) which comprises a Palaeozoic lower metapelitic formation overlying carbonates. These carbonates include lower white dolomitic marbles, 300 m thick, and upper blue calcareous marbles, 300 m thick, with thin and discontinuous metapelitic intercalations. The white marbles are fractured

and outcrop mainly in the eastern Sierra Blanca and the blue marbles, which include karstic voids, are present mainly in western Sierra Blanca. The geological structure reveals an interference of N–S and E–W folds in the western Sierra Blanca, while the eastern Sierra Blanca displays a tabular structure (ANDREO *et al.*, 1997).

ANDREO *et al.* (2002) distinguished four aquifer systems in Sierra Blanca (Fig. 1): three in western Sierra Blanca (Istán, Marbella and Ojén) and Coin aquifer system in eastern Sierra Blanca. The springs of western Sierra Blanca increase rapidly in flow (from 0 to several hundreds of L/s) after rainfalls (ANDREO *et al.*, 2002).

In this preliminary research, the chemical composition of the groundwater samples has been monitored once fortnight approximately. The electrical conductivity (EC) was measured in situ and major ions were analysed by means of standard methods in the laboratory of the University of Málaga. The analyses of TOC and $\delta^{13}\text{C}_{\text{TDIC}}$ were performed in the University of Avignon using samples previously treated with HgCl_2 . Modern methods exist for continuous online measurement of TOC using field fluorometer (Pronk *et al.*, in press) but they have been not used in this work.

3. Results

Genal spring

During the study period, the Genal spring shows rapid increases in flow (from 100 to several hundreds even thousands of L/sec) after rainfalls (Fig. 2). At the same time, decreases in the electrical conductivity of water occur (Fig. 2), the magnitude of which varies from 10 to 50 $\mu\text{S}/\text{cm}$. This variation in the electrical conductivity also appears in the evolution of $\delta^{13}\text{C}_{\text{TDIC}}$, with a maximum value of -9.5‰ and a minimum value of -12.3‰ (Fig. 2). However, the variations of TOC are very small, from 0.4 to 0.9 mg/L, what indicates that the residence-time of groundwater within the aquifer is long enough for bacterial degradation of dissolved organic carbon. Thus, these two parameters display clear global seasonal variations and the evolution of TOC and $\delta^{13}\text{C}_{\text{TDIC}}$, (except $\delta^{13}\text{C}$ march 03) together with electrical conductivity, Mg^{2+} , Cl^- , NO_3^- and Si_{CAL} have permitted to differentiate the participation of the saturated or unsaturated zones in the hydrogeological behaviour of the system (Fig. 2), in spite of the sampling periodicity.

During the flood event of December 2002, the waters drained during this period show low electrical conductivity (Fig. 2), related with short residence-time water (high NO_3^- and TOC). In addition, the Mg^{2+} content and $\delta^{13}\text{C}_{\text{TDIC}}$ decrease. Therefore, $^{13}\text{C}_{\text{CMTD}}$ flowing waters towards the outlet correspond of waters coming predominantly from the unsaturated zone. On the other hand, during the flood event of the beginning of March, electrical conductivity and $\delta^{13}\text{C}_{\text{TDIC}}$ increase, whilst NO_3^- and TOC contents decrease because long residence-time water from saturated zone is pushed out by recently infiltrated water. Then, after the flood event, the electrical conductivity and $\delta^{13}\text{C}_{\text{TDIC}}$ falls, whereas the TOC content increase (Fig. 2) because the fraction of water from the unsaturated zone increases.

In the flood event produced at end of March 2003 another type of response has been observed. In this case, Mg^{2+} and $\delta^{13}\text{C}_{\text{TDIC}}$ diminish, whereas TOC contents increase (Fig. 2). This can be explained by a piston-effect during the recharge of the system and it shows the participation of the unsaturated zone during this period. Finally, in the depletion (summer 2003), the waters drained show an increase of EC

related with high values of parameters characterizing the saturated zone (Mg^{2+} and $\delta^{13}C_{TDIC}$). Nevertheless, it is possible to detect the influence of the unsaturated zone because Cl^- contents increase and TOC remains constant. Therefore, this evolution confirms that a fraction of water coming from the unsaturated zone exists during depletion.

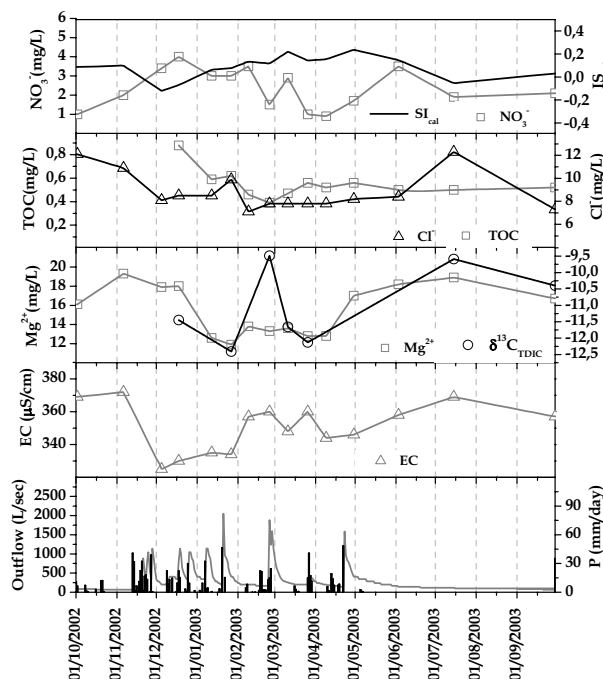


Fig. 2. Temporal evolution of chemical and isotopic parameters at Genal spring.

Jorox spring

During the hydrological year 2002/2003, the precipitation in the recharge area of Jorox spring was 540 mm, whilst in the western Sierra de las Nieves, the recharge area of Genal spring was 920 mm. Thus, during the first five months of the hydrologic year, the Jorox spring displays slight flow variations and a progressively decreasing trend of the electrical conductivity (Fig. 3).

The evolution of the $\delta^{13}C_{TDIC}$ shows a decreasing trend with the minimum value during the flood event of April 2003, whilst the TOC evolution is more significant, but this parameter shows a scarce range of TOC variations (0.4 to 0.8 mg/L). Therefore, the general evolution shows certain inertia, interrupted by important precipitations that produce rapid increases of the flow rate and reductions of the water mineralization.

Jorox spring displayed a low increase of the flow rate, due to the rainfall quantity and intensity between October and February (Fig. 3). During this period, the waters drained show a decreasing trend of the electrical conductivity and a progressive reduction of Mg^{2+} and $\delta^{13}C_{TDIC}$. The significant recharge events of February and March 2003 seem to provoke increase of TOC and Cl^- , and decrease of $\delta^{13}C_{TDIC}$. Therefore, the scarce recharge that takes place during the five first months of the hydrologic year permits a progressive dilution, because of water coming from the unsaturated zone. During the high water conditions, the water stored in the unsaturated zone is rapidly pushed towards the outlet. This response reflects again the influence of the unsaturated zone in the discharge of the karst system. Nevertheless, it is necessary to consider that the number of samples of TOC and $\delta^{13}C_{TDIC}$ in Jorox and Ojén springs

(Figs. 3 & 4) are lower than in the case of Genal spring (Fig. 2), so that the data interpretation is more difficult, in terms of hydrogeological behaviour.

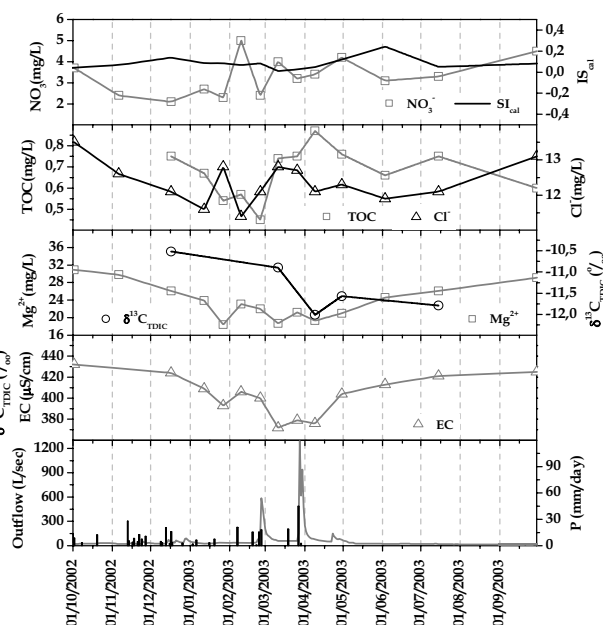


Fig. 3. Temporal evolution of chemical and isotopic parameters at Jorox spring.

Ojén Spring

The hydrodynamic and hydrochemical response of the Ojén spring seems a mixture of the behaviours observed in the former springs (Fig. 4). Thus, Ojén spring, like Genal spring, shows rapid flow increases (from 0 to several tens of L/sec) after rainfalls (Figs. 2 & 4). However, the evolution of the electrical conductivity shows a relatively lower variation (Fig. 4).

During the studied period, Ojén spring shows a relatively low variation range in $\delta^{13}C_{TDIC}$ (Fig. 4), from -10.2 to -12.2 ‰. In addition, the variations of TOC are very small, from 0.3 to 0.9 mg/L and its general evolution shows higher values during the flood events that occurs in the first part of hydrologic year and, later, decreases progressively until at main flood periods when reach the lower values.

In the most important flood period (March 2003), electrical conductivity, Mg^{2+} content and $\delta^{13}C_{TDIC}$ decrease (Fig. 4). In addition, Cl^- content also diminish whilst NO_3^- remains constant and TOC content increase slightly. The isotopic composition and Mg^{2+} value of this water indicate scarce time into the aquifer. The rest of parameters indicates that this water comes mainly from the unsaturated zone. Therefore, a mixture of water coming from the saturated and unsaturated zones exists during high water conditions.

Finally, during the depletion curve, the waters drained show an increase of mineralization (electrical conductivity) related with an increase of Mg^{2+} content, but also high values of TOC (Fig. 4). Therefore, this evolution shows again a joint response of the saturated and unsaturated zone even during low water conditions.

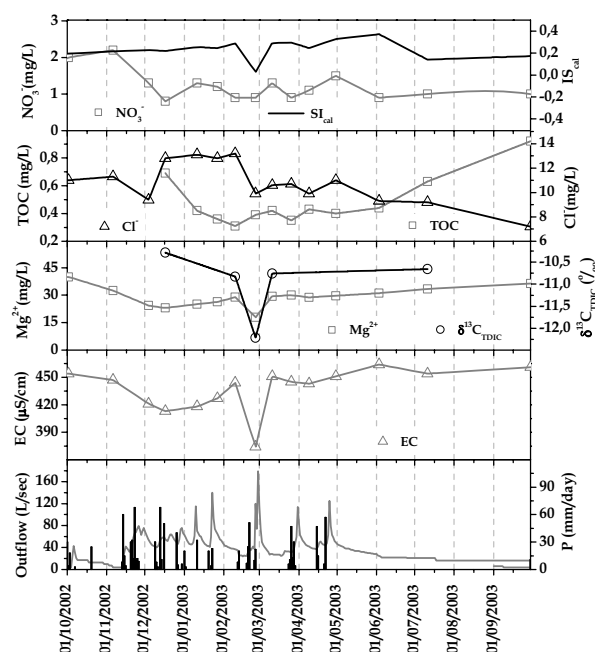


Fig. 4. Temporal evolution of chemical and isotopic parameters at Ojén spring.

4. Conclusions

Three aquifer systems from South Spain have been investigated by means of conventional physical-chemical parameters, TOC and $\delta^{13}\text{C}_{\text{TDIC}}$ to know the hydrogeological behaviour of these aquifers and to differentiate the origin of water that contributes to karst flow (unsaturated zone, saturated zone). The groundwaters drained by the main springs of the systems show relatively similar evolutions of electrical conductivity, TOC and $\delta^{13}\text{C}_{\text{TDIC}}$, during the hydrological year 2002/2003, although their geological (lithology, geological structure, karst development) and climatic (rainfall) contexts are different.

The first rainfall events of the hydrologic year provokes slight or important flow variations in the springs, but always show a progressively decreasing trend of the electrical conductivity and reductions in the parameters characterizing the saturated zone (Mg^{2+} and $\delta^{13}\text{C}_{\text{TDIC}}$) because of the water coming from the unsaturated zone. During high water conditions both saturated and unsaturated zones participate in the flow of the springs, although may be more the unsaturated one. And, during the depletion episodes the saturated zone has a prevailing role, but the importance of the unsaturated zone can not be neglected in aquifers such as those that have been studied here, with unsaturated zones several hundreds of meter thick.

Acknowledgements

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Karst landscape and aquifer evolution: The numerical perspective

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Abstract

The evolution and flow in a karst aquifer is studied with numerical simulations, based on the KARST model (Karst Aquifer Simulation Tool). The aquifer consists of a three-dimensional interconnected network of conduits representing fractures in the rock, and a porous rock matrix representing the finer fissured system in the rock. Flow through the aquifer can be driven by both diffuse recharge from precipitation and localised sinking streams, and the aquifer drains towards a large karst resurgence representing the base level. Superimposed onto the karst aquifer is a landscape, which can evolve with time by small-scale diffusive processes, large-scale river erosion, and karst denudation.

Fractures in the aquifer are enlarged with time by chemical dissolution, enhancing the secondary porosity of the karst aquifer. The enlargement of fractures results in a dramatic increase of the aquifer conductivity over several orders of magnitude, and a change of flow pattern from an initially pore-controlled to a heterogeneous fracture-controlled aquifer. During the evolution, the water table is falling from an initially high position close to the land surface to a lower level coinciding with the actual base level.

A typical model scenario is studied to elucidate the long-term karst aquifer evolution in three dimensions. The evolution model is then complemented by event-type spring discharge modelling, which can be used as a predictive tool for karst spring discharge and contaminant transport.

1. Introduction

Groundwater flow in a karst aquifer is an important subject, as rapid flow of water through highly permeable karst aquifers is prone to contamination. However, large-scale heterogeneities in the karst aquifer make it difficult to predict flow and transport in a karst catchment. While the primary conductivity of limestone is fairly low ($K \sim 10^{-8}$ m/s), fractures in the rock enlarged by chemical dissolution increase their conductivity by orders of magnitude ($K > 10^{-1}$ m/s). The enlargement of initially small fractures into larger passages then forms a pattern of cave passages, which carry the majority of water flowing through the karst aquifer. The *short-term response* of a karst aquifer observed today is guided by the *long-term evolution* of secondary porosity within the karst aquifer. Hence we need to discuss both long-term evolution and short-term response to understand the complicated dynamics of a karst aquifer.

Aquifer geometry:

PALMER (1991) has analysed several thousand of cave passage patterns and found that 57% of the total cave passages follow bedding planes and 42% are fracture-oriented. Only 1% of the total passages are related to intergranular porosity in the rock. In his classification, PALMER (1991) introduced four distinct patterns of cave passages:

(i) *Branchwork caves* (65% of total passage length), resembling single passages joining larger passages downstream as tributaries and thus similar to surficial dendritic river patterns.

(ii) *Network caves* (17%), with passage patterns of angular grids formed by dissolutional enlargement of most of the available fractures.

(iii) *Anastomotic caves* (10%), with patterns of intersecting, curvilinear tubes, often superimposed on branchwork caves.

(iv) *Ramiform and spongework caves* (8%), with random-like patterns of small cavities and larger, irregular rooms, mainly resulting from hydrothermally controlled dissolution and therefore not directly related to surface recharge.

These widely different cave passage patterns pose the important question of identifying the characteristic

processes responsible for the evolution. We need to answer questions such as 'how important is the recharge condition', 'is the structural setting guiding the evolution', 'is water chemistry decisive in the long term evolution'. Here, numerical simulations of long term karst aquifer evolution both in terms of flow and geometry have been used intensely during recent years (e.g. DREYBRODT, 1990; PALMER, 1991; GROVES et al, 1994a,b; HOWARD et al, 1995; CLEMENS et al, 1996, 1997; HANNA et al, 1998; SIEMERS et al, 1998; KAUFMANN et al, 1999, 2000; GABROVSEK et al, 2000, 2001; ROMANOV et al, 2002, 2003; KAUFMANN, 2002a,b,2003a,b), and a state-of-the-art summary of modelling in two dimensions can be found in DREYBRODT et al (2005). Common to all of these numerical models is the change from an initially relatively homogeneous aquifer with small fractures, where flowrates are low and flow is diffuse, to a strongly heterogeneous aquifer with flow being fast and concentrated along enlarged fractures. Often, a preferential flowpath is established in the models, which guides recharge through cave passages towards a karst spring.

Aquifer response: The large variability in karst aquifer properties cannot be observed properly by field measurements, as typical methods such as injection, packer, and slug tests are more suitable to porous groundwater flow. Flow in the enlarged fracture system is, however, more difficult to estimate. Only in few cases it is possible to explore the enlarged fracture system directly. Hence, the prediction of flow and contamination needs to be studied by other means.

A global observation of the response of a karst aquifer are hydrographs taken from large karst springs. These resurgences often collect water from the entire catchment area and show a rapid response to sudden recharge events, followed by an exponential decrease of discharge seen along the recession limb of the hydrograph. Analysis of

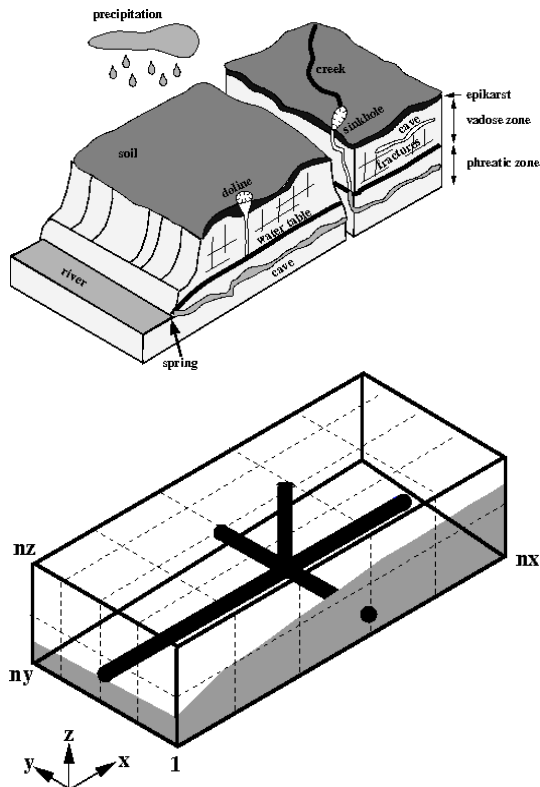


Fig. 1: Top: Conceptual model of the fractured, porous aquifer. The surface along the top boundary receives a spatially constant recharge, the base level is given as a resurgence in the valley, and above the resurgence a seepage face might be present. Phreatic fractures are shown as Thick gray lines, vadose fractures as black lines. The thick black line depicts the water table. Bottom: Realisation of the aquifer. The model domain is discretised into rectangular blocks, representing the porous-flow component. Fractures are implemented as circular conduits, which can grow with time.

karst spring hydrographs is a common tool to study the response of a karst aquifer to groundwater flow (e.g. ATKINSON et al, 1973; SAUTER, 1992; GRASSO et al, 1998; BAEDKE et al, 2001; JEANNIN, 2001). However, the hydrograph analyses are often limited by a number of factors, e.g. the large catchment of karst aquifers, the strong heterogeneities in conductivity, and the scarcity of data (see JEANNIN et al, 1998, for a review).

Besides analysis of a measured spring discharge hydrograph, hydrograph time-series can be generated by numerical flow models, which describe a karst aquifer geometry in a simplified fashion. This numerical approach excludes per definition uncertainties in the hydraulic properties, and thus can be used to study the response of a karst aquifer to recharge events. For example, EISENLOHR et al (1997a,b) have used the numerical model of transient groundwater flow described in KIRALY (1998), with flow driven by the rainfall time-series from the Areuse and Serriere catchments in the Swiss Jura. The finite-element model includes both low-flow, high-storage porous matrix, and high-flow, low-storage karst conduits into the approach.

While EISENLOHR et al (1997a,b) have assigned a constant conductivity to karst conduits, which were placed in the

numerical model domain in an 'ad-hoc' fashion, we discuss another possibility to access the karst aquifer properties: We make use of numerically generated long-term karst aquifer models and simulate short-term transient flow behaviour driven by a realistic recharge times series. Our aim is to test synthetic karst aquifer models, which are calculated as evolution models under the assumption of steady-state recharge, under transient event-based recharge conditions. We predict spring discharge time-series and show that a long-term numerical karst evolution model is capable of describing the flashy response typical to mature karst aquifers.

2. Theory

Dissolution chemistry: In karst landscapes, soluble bedrock such as limestone or dolostone is dissolved by water enriched with carbon dioxide. On the surface, material is dissolved under conditions open to the atmosphere, and removed with the aqueous solution. This process, termed karst denudation, lowers the bedrock fairly uniformly, and a karst plateau will be lowered over time. Within the karst aquifer, material is removed from narrow fissures and bedding partings, in which water enriched by carbon dioxide circulates. The voids are enlarged over time, altering the permeability of the bedrock, and also changing the flow patterns through the aquifer. Depending on the situation within the aquifer, dissolution proceeds either under open-system conditions, where the solution is in contact with the atmosphere (soil, epikarst, vadose caves), or under closed-system conditions with solution completely filling the voids (phreatic caves, voids below the water-table), thus being decoupled from the atmosphere.

The dissolution process continues until the solution is in equilibrium with respect to calcite. The time needed to attain equilibrium is controlled by various rate-limiting reactions. Hence the need to understand both the equilibrium chemistry within the ternary system water (H_2O), carbon dioxide (CO_2), and calcite ($CaCO_3$), and the reaction kinetics controlling the speed of dissolution.

Experimental data on natural limestone under closed-system conditions (EISENLOHR et al, 1999) can be described by a linear calcium fluxrate for large undersaturation, and a non-linear calcium fluxrate F close to equilibrium:

$$F_{Ca^{2+}} = k_i \left(1 - \frac{c}{c_{eq}} \right)^{n_i}, i = 1, 2 \quad (1)$$

Here, $F_{Ca^{2+}}$ [mol/m²/s] is the fluxrate, k_1 and k_2 [mol/m²/s] are rate coefficients, c and c_{eq} [mol/m³] the input and the equilibrium calcium concentration, and n_i a power-law exponent (see KAUFMANN, 2005, for details). Eq (1) accounts for the *slow linear* fluxrate with $n_1=1$ for $c < c_1$, and the several orders of magnitude lower *high-order* fluxrate with $n_2=4$ for $c > c_1$. Here, c_1 is the calcium threshold concentration, at which the fluxrate changes from slow to high-order kinetics; it is generally between 70 and 90% of c_{eq} .

Single fracture: Several numerical models on single fracture growth have been published (e.g. DREYBRODT, 1990; PALMER, 1991; GROVES et al, 1994a; DREYBRODT, 1996; KAUFMANN et al, 1999; DREYBRODT et al, 2000; PALMER, 2000). These models simulate the evolution of a single fracture as a simple analog of flow along a fracture or a bedding plane, and the evolution of the fracture width is controlled by the dissolution rate. At the exit of the fracture, slow dissolution rates are responsible for the slow growth of

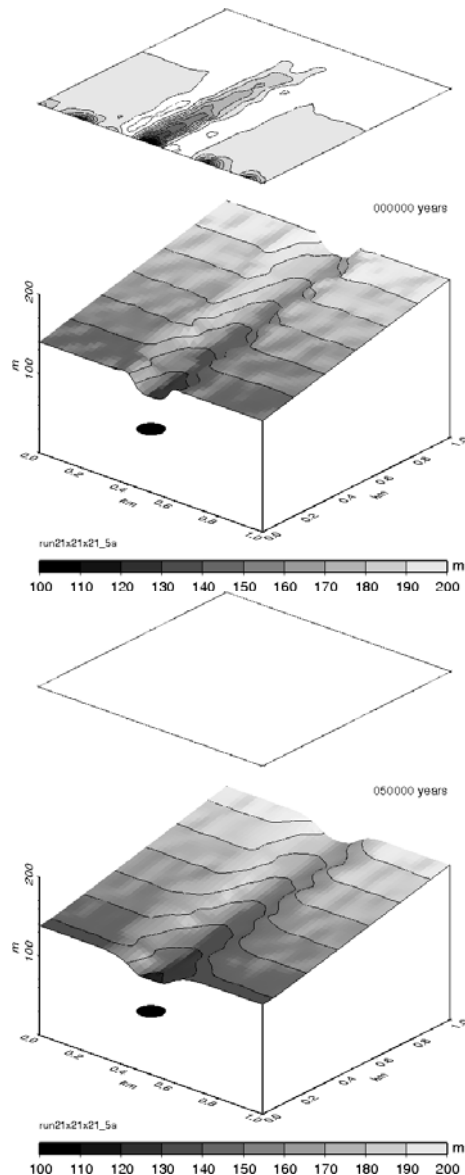


Fig. 2 : Topography of karst plateau, shown at the start of the model run (top) and after 50,000 years (bottom).

Heights are colour-coded. The map on top of the topography shows surface runoff, with black-ish colours indicating excess water running off the surface, and white colours no surface runoff.

the exit diameter, while the entrance of the fracture grows faster during this initial phase. As time proceeds, the solution at the exit becomes sufficiently undersaturated to increase the dissolution rate by several orders of magnitude, and a breakthrough is achieved. From then on, the entire fracture grows at a constant pace. Details on the implementation of flow and dissolution can be found in KAUFMANN (2005).

Surface evolution: Large-scale landscape evolution on tectonic timescales is controlled by a number of processes. *Short-range* hillslope processes, such as weathering, slope wash, mass wasting, and soil creep, redistribute mass over short distances, while *long-range* flow processes, such as fluvial erosion and sedimentation, and karst denudation, control landscape evolution over long distances. The mathematical approach for both short- and long-range

processes used in model simulations can be found in KAUFMANN et al (1999b).

Subsurface flow: The concepts on limestone dissolution, surface evolution, and flow through fractures developed in the previous sections are used to assemble a three-dimensional numerical model representing a karst aquifer, called KARST (Karst Aquifer Simulation Tool, Fig. 1). In the scenarios discussed, the aquifer is assumed to comprise a flat-lying layer of limestone, which is underlain by a layer of impermeable material such as sandstones or schists. Hence, all water entering the limestone layer must leave the model through its defined output.

Flow in a karst aquifer is modelled using the discrete fracture approach (e.g. GUREGHIAN, 1975 ; HUYAKORN et al, 1983; KAUFMANN et al, 2000), which incorporates fractures directly into the medium. The advantage of the discrete fracture approach is the combined modelling of matrix and fracture elements and the consistent modelling of flow in the matrix-fracture medium. Flow in the fractured, porous karst aquifer is modelled by the transient continuity equation:

$$\frac{\partial}{\partial x} \left(K(t) \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(K(t) \frac{\partial h}{\partial y} \right) + \frac{\partial}{\partial z} \left(K(t) \frac{\partial h}{\partial z} \right) = S_s \frac{\partial h}{\partial t}. \quad (2)$$

Here, x [m] and y [m] are the northward and eastward coordinate directions, z [m] the vertical direction, t [s] is time, h [m] is the hydraulic head in the model domain, K [m/s] is the conductivity, S_s [1/m] is specific storage. For a total of N nodes in the model domain, (2) represents a set of N equations for the N unknown heads h_i at nodes i . Equation (2) is solved with a Galerkin finite element method, with parallelepipedal elements for the porous matrix and linear elements for fractures and fissures (Fig. 1).

The conductivity K_m in the matrix representing micro-fissures in the rock remains constant throughout the evolution, $K(t)=K_m$, while the conductivities in the fractures, $K(t)=K_c(t)$, increases with time due to the enlargement of the elements by chemical dissolution. Water entering the porous matrix can be aggressive ($c < c_{eq}$), water leaving the porous matrix is saturated ($c=c_{eq}$). At the intersections of fractures and fissures, instantaneous and complete mixing of incoming water is assumed.

3. Results

Long-term evolution: The 3D model simulates a 1 km² karst aquifer (1,000 m x 1,000 m), with a plateau dropping from 200 m in the north to 140 m in the south. A small central valley channels surficial flow towards the base level (Fig. 2a). The karst aquifer is bounded by impermeable rocks to the sides and below, hence all water flows out to the defined base level. The domain is discretised into 21x21x21 nodal points, resulting in 9261 nodes, 8000 parallelepipedal elements as porous matrix, and 25620 circular conduits as fractures, oriented either horizontally or vertically. The initial fracture diameter is $d_i=0.1$ mm, and matrix conductivity is $K_m=10^{-7}$ m/s. Water reemerges to the base level in the south located at 80 m elevation. Groundwater recharge is 600 mm/yr and spatially constant, and water runs onto the plateau surface during the first couple of thousand years. Water disappearing underground has a temperature of 10 °C and a carbon-dioxide pressure of 0.5 atm. The water sinking into the aquifer is already at 90% calcium concentration.

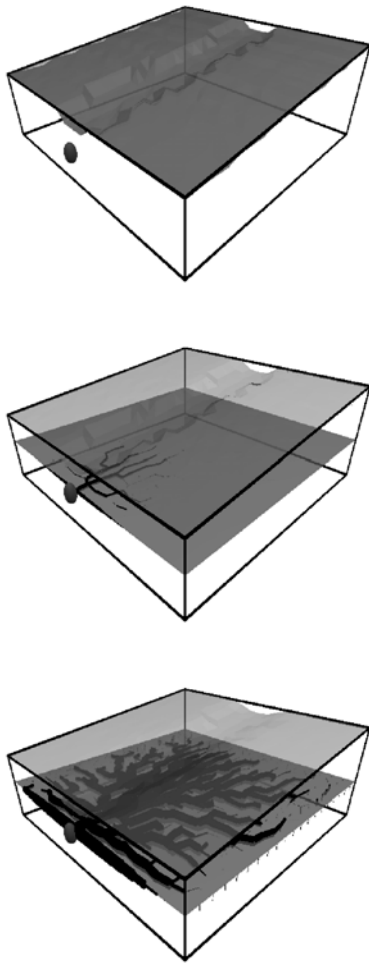


Fig. 3 : Three time slices of fracture-diameter evolution in the karst aquifer. Light gray surface is the topography, darker gray surface is the water table. Gray lines indicate fractures in the unsaturated (vadose) and black lines fractures in the saturated (phreatic) zone. The line thickness corresponds to the width of the fractures.

The long-term model is modelled over 50,00 years, using a steady-state approximation of (2), thus $S_s=0$.

After 50,000 years of evolution (Fig. 2b), the landscape has become smoother, and the entire surface has been lowered. However, as surface runoff has been switched off early due to the increase of subsurface flow (see below), fluvial erosion has become very low.

The water table (Fig. 3, top, gray surface) has reached the surface at the initiation of the model run, enabling surface runoff. Then, dissolution in the fractures has increased the permeability, and the water table has dropped quickly during the first 10,000 years (Fig. 3, middle), and has reached a steady-state niveau along the base level, as it can be seen after 50,000 years (Fig. 3, bottom). Between the surface and the final water table, a zone of enlarged fractures is present. Here, some fractures above the

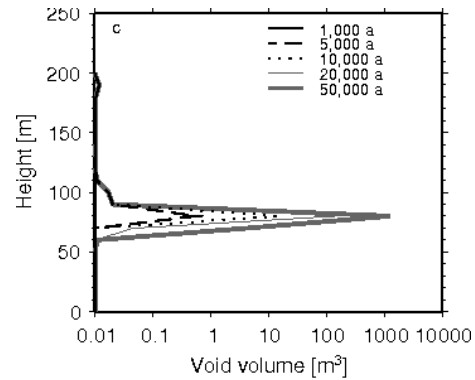


Fig. 4 : Void volume as a function of depth for different times.

resurgence have been enlarged first, but since then they have fallen dry (gray conduits). After the water table dropped to the base level, a dendritic drainage network of enlarged fractures has evolved (black conduits), which carried water along the water table towards the resurgence. The enlargement, however, has been selective: The quick drop in water table allowed only for significant fracture enlargement close to the resurgence area, and along the final water table, only the most direct flow routes have been enlarged. It should be noted that the 3D model essentially evolves as a 2D plan model, once the water table has reached its final steady-state position.

In Fig. 4, the void volume within the aquifer is shown as a function of depth. The void volume is calculated by summing up the volume of all conduits within a given depth. During the initial phase, no significant enlargement has taken place, and thus only a small hump in the curve for 1,000 years close to the surface is visible. However, after already 5,000 years, the water table has reached its steady-state niveau, and significant enlargement along the water table taken place. The void volume along the steady-state water table then increases by orders of magnitude, until the experiment is terminated at 50,000 years. At that time, the largest fractures have reached meter-sized dimensions.

Short-term spring response : We now use selected times of the long-term evolution model with a time-dependent recharge event and solve the full transient form of (2). Specific storage for the porous matrix and the fractures are 10^{-5} m^{-1} and 10^{-8} m^{-1} , respectively. The recharge time series, shown as gray bars on top of Fig. 5, is taken from the Gallusquelle in southwest Germany (SAUTER, 1992), and covers 134 days. It is characterised by a sequence of recharge events, each lasting two to three days, with a maximum of 21 mm/day.

At the initiation of the karst evolution (0 yrs), the hydrograph response (gray line) is characterised by a moderate recession limb resulting from the combination of permeable matrix and small fractures. Then fractures start evolving around the karst resurgence, and 5,000 years into the evolution, the southern part of the aquifer is efficiently drained through the enlarged fractures, while in the northern part flow is still slow and pore-controlled. The discharge series (dotted line) reflects this bimodal behaviour, small peaks from the rapid transmission of water through the enlarged fractures, followed by a long recession limb from the slow drainage of the northern part.

After 10,000 years, the fractures within the entire karst plateau are enlarged, and drainage is fast and effective (dashed line). Discharge is characterised by large peaks and

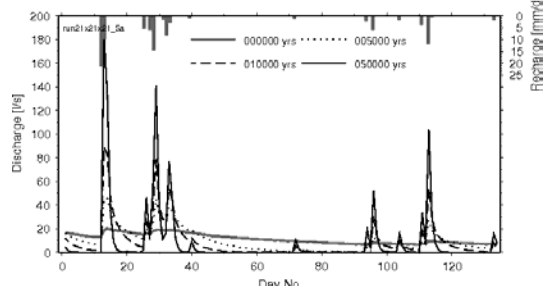


Fig. 5 : Recharge and discharge time series for different evolution times.

a very short recession limb. Hence, within 10,000 years, flow through the evolving karst aquifer has changed significantly. Even further into the evolution (50,000 years), the transmission of flow is almost instantaneous, as every part of the karst plateau is directly drained through enlarged fissures.

Following HOBBS et al (1986), the evolution can be classified as follows: Early discharge is characterised by high storage in the matrix and flow transmission controlled by both diffuse flow through the porous matrix and concentrated flow through the small initial fractures. Later, discharge is almost entirely concentrated to the enlarged fissures, and storage is insignificant.

4. Conclusion

We have developed a 3D-numerical model for the evolution of a small karst aquifer. The model includes an evolving landscape controlled by diffusion, fluvial erosion, and karst denudation processes, and the increase of secondary porosity within the aquifer by chemical dissolution. The main control on the evolution is the position of the water table: Initially the water table is close to the surface, allowing for surface runoff and dissolution within the entire aquifer. When fractures are enlarged by dissolution, the water table drops, the surface falls dry, and enlargement of fractures moves deeper into the aquifer. Finally the water table reaches its steady-state position along the base level niveau, and fractures are enlarged along the stagnant water table.

Driving the evolution model with a time-dependent recharge series, the response of the karst resurgence strongly depends on the secondary porosity, thus on the evolution time. Early into the model run, discharge is controlled by the slow-flow, high-storage regime in the porous matrix and the not yet enlarged fissures. Once the fissures are sufficiently enlarged, discharge becomes very flashy, as the fast-flow, low-storage fracture system controls the flow.

Acknowledgements

The figures in this paper are drawn using the GMT graphics package (Wessel & Smith, 1991,1998).

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Integrated technologies for a better understanding and management of karst aquifer systems

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Abstract

A number of articles have been published related to the study of karst aquifers that highlight the need for an integrated approach using methods to explore and study karst systems as defined by BAKALOWICZ (2005).

The study of paleokarst reservoirs at the wellbore and system scales has allowed the construction of an improved static model with the application of geostatistical methods.

Facies models have been derived from the analysis of the response of the various karst zones to microresistivity imaging of wellbores and other larger scale geophysical, e.g. seismic, measurements.

In this approach, the static hydrogeologic model is further validated and calibrated using dynamic simulation studies. The installation of multi-port wells and the analysis of test data can help derive an estimate of the areal and vertical variations of the distribution of hydraulic conductivity.

A case study attempts to briefly evaluate the various modeling approaches and demonstrate the means of measuring properties and calibrating models. Lastly, the issue of how those new developments in oilfield technology could improve the assessment of water resources from karst systems is addressed.

Résumé

De nombreux articles publiés sur l'étude des aquifères karstiques ont mis en lumière le besoin d'une approche intégrée utilisant différentes méthodes d'exploration et d'étude des systèmes karstiques (BAKALOWICZ 2005).

L'étude des gisements pétroliers situés dans des paléokarsts, tant à l'échelle du puits qu'à celle du gisement, a permis la construction de meilleurs modèles statiques de ces gisements, en utilisant des méthodes géostatistiques.

Différents faciès karstiques ont ainsi pu être identifiés, à partir de l'analyse des réponses des diverses zones du karst à l'imagerie électrique de la paroi des puits ainsi qu'à d'autres mesures géophysiques à plus grande échelle, comme la sismique de surface.

Dans une telle approche, le modèle hydrogéologique statique est ensuite validé et calé à l'aide de simulations dynamiques. L'installation de puits permettant des prélèvements et des mesures à différents niveaux et l'analyse des données d'essais hydrauliques peuvent aider à estimer les variations tridimensionnelles de la distribution de la conductivité hydraulique.

Une étude de cas essaie de brièvement évaluer les différentes approches de modélisation, tant statique que dynamique, et de démontrer les différents moyens qui existent pour mesurer les propriétés et caler les modèles. Enfin est abordé la question de savoir si, et comment, ces nouveaux développements de la technologie pétrolière peuvent servir à améliorer la caractérisation des ressources en eau dans les systèmes aquifères karstiques.

1. Introduction

Carbonate reservoirs contain a little more than half of the world's remaining oil in place, and the giant carbonate oilfields of the Middle East are still expected to play an important role in the world's oil production in the 21st century.

Since the 1970s, the oil industry has gradually paid more and more attention to the characterization of carbonate rocks. With time, the approach to studying carbonates in the oil industry evolved, and various technologies were developed to obtain the results commented in the case study of East Asia, presented herein, contributing to a better understanding of oil-filled paleokarsts.

The heterogeneities at all scales of carbonates make them difficult to study and make flow prediction incredibly difficult. Therefore, a particular method might not be suitable at all relevant scales of investigation and a combination of techniques helps span all scales. At the intermediate scale, from a few millimeters to tens of meters, geological features such as bedding, vugs, stylolites, fractures or karst features are readily observable through

wellbore technology. At the system scale, from hundreds of meters to kilometers, for insights into the inner geometry and outer boundaries, seismic reflection data, log correlation, and history matching of dynamic measurements are used. The data provided by those various technologies are integrated to build and constraint the model.

The question now is: could all these oilfield technologies allow a better understanding and management of modern karst systems?

The present paper describes the methods used for a case study in East Asia. It is submitted to the conference so as to allow the discussion of this approach for karst aquifer studies and thereby to help define the way forward.

2. Materials and Methods

Borehole microresistivity imaging

Wireline borehole imaging was the first real breakthrough that was made by the oil industry for the characterization of carbonate formations. It suddenly enabled geologists to see the formations crossed by a borehole, and thus to better understand the reservoir properties.

Most carbonate reservoirs display higher permeability zones that are due to carbonate dissolution features, even if not all of them are identified as karstic zones. Permeability in non-karstic carbonates is claimed to be predominantly a function of the presence of fractures and of the rock porosity types.

Borehole imaging tools, especially the electrical micro-scanners that were developed in the 1980s, have been successful in achieving both fracture identification and rock type classification. This is due to the high resolution imaging capability of those scanners. As electrical micro-scanners are pulled up a borehole, formation conductivity measurements are made every 2.5 mm, both horizontally and vertically. Small-scale conductivity variations in the electrical images can thus be detected and permit to identify the presence of thin fractures or of vuggy porosity in carbonate rocks (AKBAR *et al.* 2000).

A turning point, though, was the discovery in 1975, in the vertical well Rospo Mare-1, about 1300 m below the bottom of the Adriatic Sea, of a heavy oil accumulation in the Albian to Cenomanian carbonate formations. The oil reservoir was thought to be fractured so a development with horizontal wells was decided upon in order to increase the drainage area by intercepting as many fractures as possible with a limited number of wells.

A first horizontal well, Rospo Mare-6, was drilled in 1982 and was an immediate success, producing 20 times more oil than a neighboring vertical well. Borehole images were acquired with both borehole televiewers and electrical imagers, and it became apparent that the oil accumulation was in fact located in a paleokarst capped with impervious formations of Miocene age (GAUCHET & CORRE 1996).

The ten appraisal wells and twenty eight production wells (two vertical and twenty six horizontal) that have been drilled in the field at various times, confirmed the presence of karstic features (either enlarged fractures, conduits, or small caves) over the entire reservoir. Nuclear logs, giving significant radioactivity levels related to thorium, pinpointed the presence of residual clay (terra rossa) plugs.

A methodology was determined for characterizing karstic features from borehole images. Vertically, four different zones could be distinguished from top to bottom by the oil company (Fig. 1):

- the epikarst where fractures had been enlarged by dissolution but partially infilled by clay,
- a percolation zone where large fractures, vertically enlarged by dissolution, spaced by tens of meters, have been intersected by the horizontal wells,
- the so-called underground rivers zone characterized by numerous horizontal conduits, formed along the natural fracture system relative to the paleohydrodynamic gradient and where the wells sometimes crossed breccia zone resulting from cave top collapse,
- a deep zone with tight rocks where dissolution was limited to non-connected vugs (examples in Fig. 2).

Surface seismic imaging

Another measurement method came afterwards: seismic surveying, where interpretation with new mapping technologies revealed not only the surficial geomorphologic features of Paleozoic paleokarst surfaces with exceptional clarity but also mapped features within the carbonates.

An example of this approach, a 3D seismic survey covering the large structural high, identified as the Obelix structure, in the central area of the Loppa High, Barents Sea, was acquired in 1998. Seismic attributes (i.e.

	Zonation	Average porosity (after editing)	Karstic void organization	Water circulation
	Epikarst (0 to 35 m)	1.2 %		Vertical drainage (locally horizontal on screens)
	Percolation zone (15 to 45 m)	0.6 %		
	Underground rivers zone (35 to 70 m)	1.9%		Horizontal drainage (high speed)
	Deep phreatic zone with limited circulation	0.4%		Horizontal drainage (low speed)

Fig. 1: Vertical Zonation of the Rospo Mare paleokarst (from GAUCHET & CORRE, 1996)

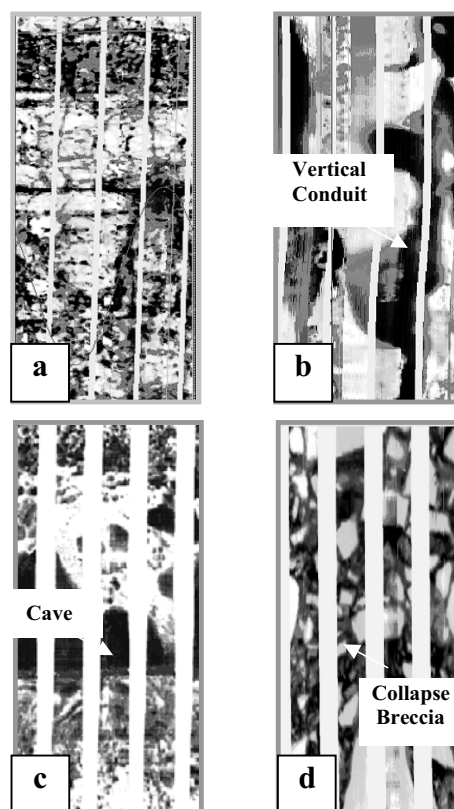


Fig. 2: Examples of karst features in borehole electrical images: a) epikarst; b) vertical percolation conduit; c) cave; d) collapse breccia.

parameters such as amplitude, frequency, phase, dip, azimuth, continuity, derived from the analysis of the seismic traces) were analyzed to determine the spatial variation in rock properties (HUNT *et al.* 2003).

The methodology conceived by that time for highlighting karstic features at the reservoir scale combines dip maps from PSDM (Pre-Stack Depth Migration) with a 3D seismostratigraphic mapping approach. The SeisClass mapping technique was used for seismic facies

classification, combining geometric and signal attributes into a 3D seismic facies cube. It compares seismic data (velocity tomography) from multiple subsets of the 3D seismic data cube. When compared, data with common features are isolated, using statistical or neural network classifiers for inferring the (statistical) relationships between patterns of the seismic data, or between attribute responses corresponding to the original data.

texture attributes data, as described in RANDEN *et al.* (2003), (i.e. reflection strength, variance, flatness, fault edge,

The 3D attributes mapping revealed structural and stratigraphic information, and notably showed the presence of karstic features and large scale karstified zones in the case of the Barents Sea data volume, by producing maps and 3D images of seismic facies (amplitude-derived seismic facies), as illustrated in Fig. 3 and Fig. 4.

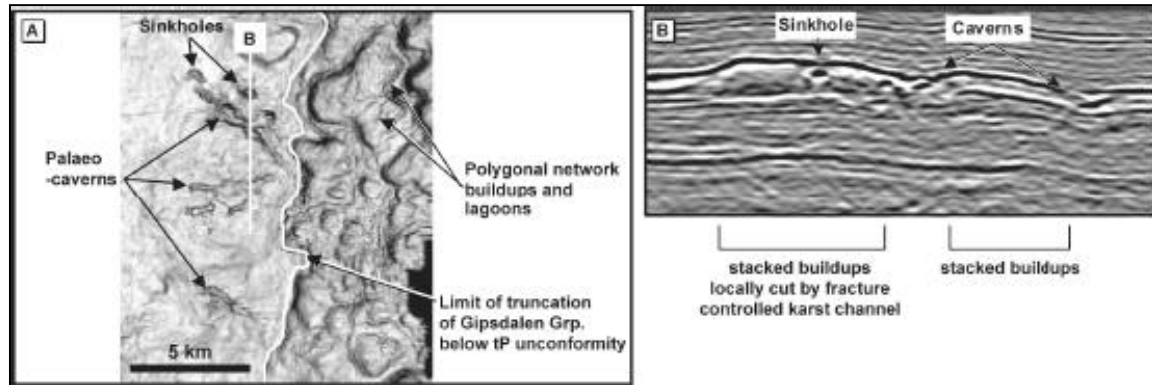


Fig. 3: A) Dip map showing the polygonal arrangement of fault-controlled carbonate buildups. The fault and fracture control on karst distribution is demonstrated by comparison of velocity tomography, variance and seismic reflection data; channels locally cut straight through the carbonates. B) Seismic line through the fracture/fault controlled karst channels. Sinkholes apparent from dip maps often appear at the intersection between faults, they are coincident with zones of poor reflector continuity and anomalously low velocity (from HUNT *et al.* 2003).

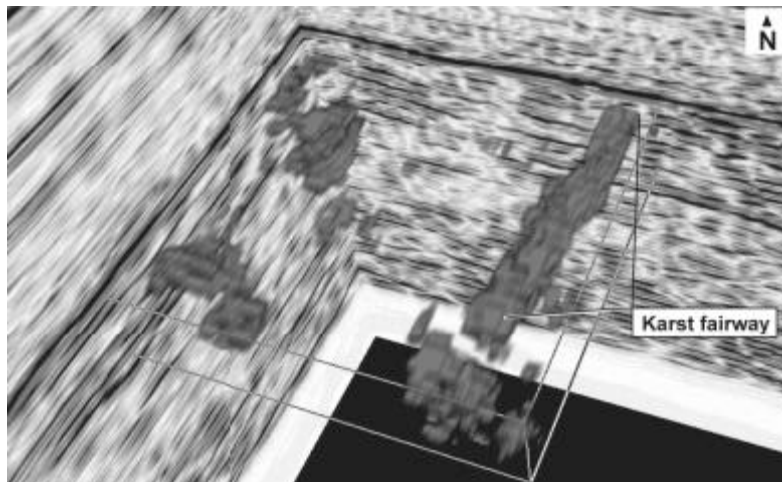


Fig. 4: 3D image representing the 'karst' subvolume from a Seisclass 3D classification, mapping within the carbonate volume the karst features that were visible on the dip maps of the top surface. Note the linear fairway of karst features mapped-out (from RANDEN *et al.* 2003).

3. Results

The aforementioned methods were integrated for a case of oil accumulation in paleokarst that has recently been described (WANG 2005) in Paleozoic formations of East Asia. Locating and managing water to control its production while producing the oil was a key objective of this project. A combination of seismic surveying and borehole imaging has permitted the unravelling of karst structure and geometry.

Borehole electrical images taken in three horizontal wells and five vertical wells have allowed the characterization of the density and orientation of the natural fracture network, and the recognition of different karst zones, quite similar to

what could be observed in Rospo Mare, but herein identified as surface, vadose, phreatic, and deep zones. In the images it is possible to identify, the fractures in the surface zone (epikarst), the vertical conduits in the vadose zone, the wide horizontal conduits (and roof-collapse breccia) in the phreatic zone and the tight rock in the deep zone.

Guided by a classification of 3D seismic attributes, such as variance and flatness, and based on the precise information brought by the electrical images, a 3D stochastic reconstruction of the karst zonation was performed. It is possible to observe that as a result of

paleoclimatic changes and of tectonics, several cycles of karstification occurred and are now present on top of each other. Production and well test data showed that the so-called surface and phreatic zones were providing better storage and productivity. These results were used to construct a fluid flow simulation model and to design horizontal well trajectories so as to maximize oil production, while minimizing water breakthrough.

To obtain more realistic estimations of the volume geometries and flow properties having an impact on oil production and water breakthrough, an uncertainty analysis was performed using a geostatistical approach to take into account uncertainties related to the geological interpretation, seismic depth conversion, facies modeling or petrophysical modeling. In the flow simulator that was used for the East Asian case study, commercial software using a dual porosity and dual permeability approach, and based on the WARREN & ROOT approximation that is normally geared to fracture systems, was employed despite the karstic nature of the various reservoir zones.

A new fluid flow simulation module dedicated to karst formations is under development in the ECLIPSE simulator. It combines dual porosity / dual permeability media with conduit networks forming a set of multi-segmented trees. It also includes the modeling of carbonate conduit network genesis and enlargement by dissolution, based on the processes described by BAUER *et al.* (2003) and LIEDL *et al.* (2003). Flow, transport, and rock/fluid interaction are solved in a fully coupled manner over time, in order to address the impact of saltwater and freshwater mixing on the dissolution of carbonates.

4. Conclusion

Models depend on an understanding of the processes that created or altered the formations, and on measuring rock properties at adequate resolutions (LUCIA 1995 & 2000).

The integration of data at various scales and stochastic modeling did contribute to the construction of representative models for paleokarsts. At the wellbore scale, the integration of borehole logging and imaging data - that are normally used to characterize the natural fractures and quantify the vug fraction in carbonate rocks - helped in the identification of karst zones. At the system scale, seismic facies mapping revealed unexpected complexity of the carbonates and contributed to the identification of karst features.

The question is what can those methods bring to improve the characterization, modeling and exploitation of the water contained in present day karsts and, if so, what has to be done?

As mentioned by BAKALOWICZ (2005), the key difficulty in karst aquifers is dealing with the karst network itself. The main problem to be solved in carbonate aquifer formations is how to determine if such a karst network exists, where it is located, and what role it plays in the functioning of the hydrodynamic system. Fluid flow simulation also remains a challenge and adequate simulation methods are yet to be developed for karst. In addition, a methodology for model calibration is required. Head and water quality measurements from multiport wells can be instrumental here, as demonstrated in the karst limestones of Puerto Rico (SEPÚLVEDA 1999).

The validity, for the study of karst aquifers, of combining methods developed for oil reservoir characterization has

now to be tested, in order to answer questions such as:

- on the measurement side, what will be the impact of the presence, in actual karst, of dry and wet zones, compared to the same measurements in a paleokarst containing only saturated oil-bearing and water-bearing zones?

- on the methodology side, will this type of detailed characterization bring representative pieces of information for the study of modern karsts, as it does for paleokarsts, and what will this new information bring to the systemic approach commonly used for present day karsts?

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The important role of Nappe Tectonics in the coastal aquifers. The case study in the island of Kythira (Greece)

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Abstract

The main problem of the littoral aquifers of islands and coastal areas is the seawater intrusion, which takes place when there is direct hydraulic communication and low hydraulic head. This happens when groundwater exists in very permeable rocks, the recharge is not satisfactory and/or the pumping is intensive. This study notes the importance of nappe tectonics in hydrogeological conditions. Such a case is the Kythira Island, where the tectonic structure favours the storage of satisfactory exploitable quantities of good-quality groundwater without the danger of groundwater salinity.

1. Introduction

Saline intrusion into the coastal aquifers constitutes the main problem of the groundwater in Greece resulting in groundwater salinity often in distances of 10km and 15km from the shore. This phenomenon has affected a great number of Greek islands, as well as the coastal zones of Greek mainland, in places where the aquifers hydraulic heads are low and the hydraulic communication with the sea is open.

Such conditions are typical of areas with karstic carbonate rocks, which cover about 1/3 of the country and are usually of very high permeability, not permitting the development of hydraulic head as they allow quick discharge of the groundwater into the sea. Similarly, very low hydraulic heads are observed in the coastal low altitude flat zones, in which the over-exploitation of water, in order to cover the water demand, often intensifies the saline intrusion. Under these negative hydrological conditions it is necessary to look for favourable geological structures, which prevent the occurrence of saline intrusion. Such a characteristic case is the case of the island of Kythira, which is studied in this paper.

2. Description of Kythira Island

Kythira and Antikythira islands are the southernmost islands of the Ionian Sea between Peloponnese and Crete. Geographically, they are extensions of the southeastern Peloponnese. Kythira has a surface of 281 km² and a coastline of 91.5 km.

Two elongated mountain ranges with North-South direction dominate on the island, one on the east side and the other on the west side. They both consist of carbonate rocks of Tripoli's geotectonic zone. The highest altitude is 490 m on the east side and 507 m on the west side. Because of this morphology, Kythira has a significant hydrographic network of torrents and several streams, where few springs of small discharges can be found. Between the two mountain ranges there is a plateau with altitudes ranging from 300 to 350 m. This plateau consists of limestones of the overlying Pindos geotectonic zone (THEODOROPOULOS 1973).

By processing the available data at the Kythira meteorological station for the period 1931-2001, the following conclusions are: a) The average annual rainfall is 576 mm b) The average annual temperature is 17.7 °C c) The average annual relative humidity is 66% d) The prevailing wind has north eastern direction

3. Geological and hydrogeological characteristics of the area

The geological formations building up the island belong to the following units (DANAMOS 1992) (Fig. 1):

a) Metamorphic rocks: The northern part of the island consists of Neopaleozoic gneisses and schist, which cover approximately 1/6 of its total surface. They constitute the deeper geological basement of the island.

b) Tripoli's geotectonic zone: Limestones and dolomitic limestones of Jurassic – Cretaceous form the carbonate complexes of the two mountain ranges on the east and west sides of the island, as well as the central part of the island between the two mountain ranges. They constitute the bedrock of the island under the overthrusting formations of the Olonos – Pindos geotectonic zone. Flysch sediments constitute the upward continuation of the carbonate upper horizons of the Tripoli's zone.

c) Olonos –Pindos geotectonic zone: The formations of Olonos-Pindos zone cover a large part of the island, occupying its central and southern part between the two big mountain ranges of Tripolis zone limestones. They consist of various rock types, mainly thin-bedded limestones and flysch clay schists of upper -Cretaceous in age.

They are overthrust on Tripoli's zone formations.

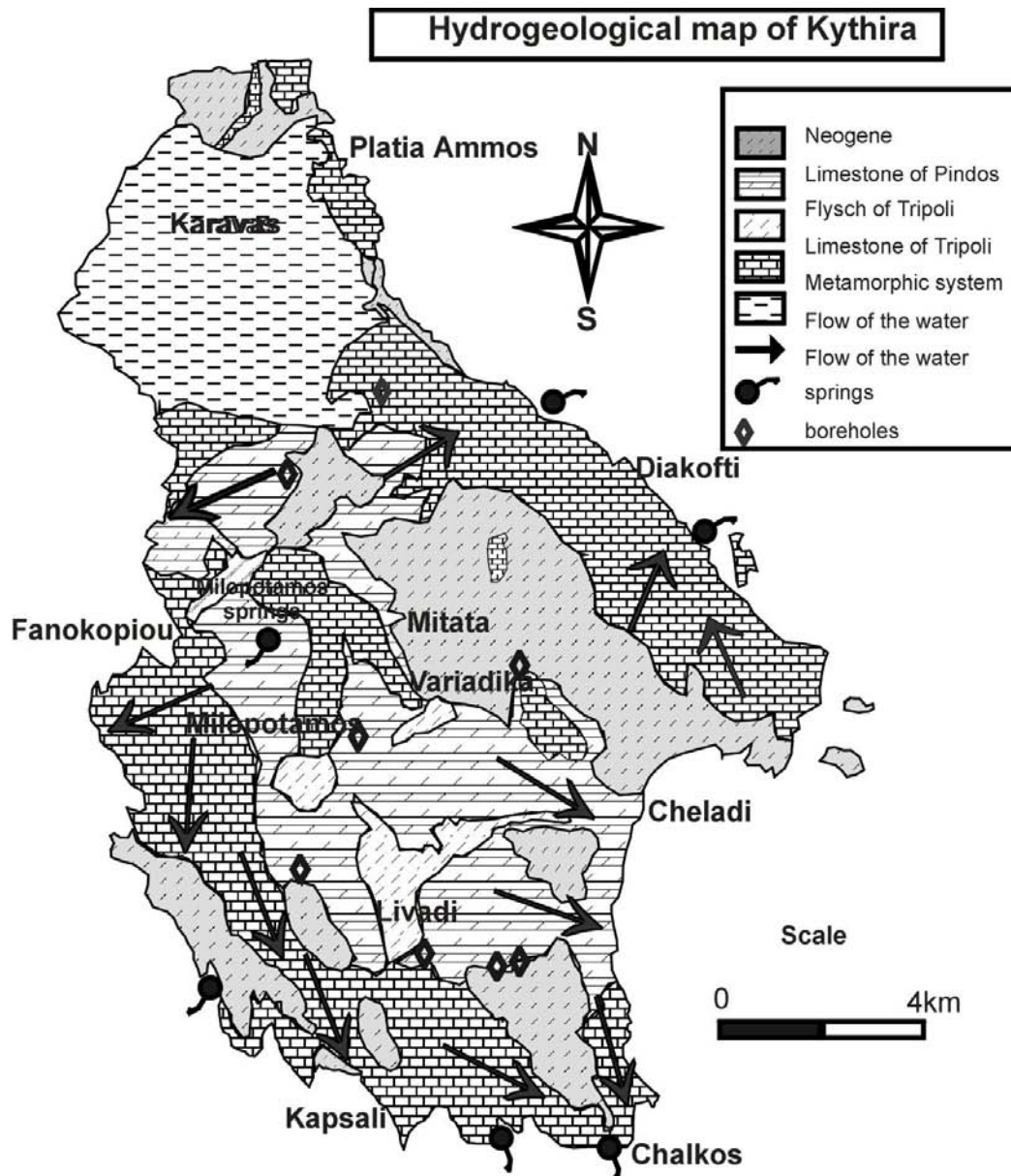


Figure 1: Hydrogeological map of Kythira (IGME 1966)

d) Neogene deposits: *Neogene lacustrine and saline* sediments consist mainly of conglomerates, sandstones, marls, marly limestones and sands with lignite intercalations.

e) Quaternary deposits: They include coastal formations, alluvial and continental depositions, as well as talus formations. Their thickness exceeds 65 m in some places.

Permeable (55% carbonate rocks), semi-permeable and impermeable formations are found in the Kythira Island:

(a) Permeable formations

❖ *Limestones of Tripolis zone*

The limestones and the dolomitic limestones of Tripolis zone are developed in a karstic network with high permeability. They are overthrust on the Paleozoic metamorphic basement of the island. These rocks are of high permeability allowing large quantities of groundwater to flow through them; however, the direct contact with the sea and the low water table of the aquifers that is near to the altitude of the sea, cause groundwater salination. The discharge of the karstic aquifers takes place from both submarine and littoral springs, which are located in various sites of the island. Submarine karstic springs are located in the areas of Vani and Skydias bays at the southwest coasts and Kakis Lagadas and Diakofti at the eastern coast. Coastal springs are located in the area of Fanokopiou, at the bay of Felotis, Kapsaliou, Chalkou, Diakofti and Cheladi cape. (PERISORATIS & MITROPOULOS 1983) (Fig.1).

❖ *Limestones of Olonos-Pindos zone*

Clay schist intercalations within the limestones of the Olonos-Pindos unit prevented limestone karstification. The hydrogeological interest of these rocks is in the contact with the underlying impermeable flysch of Tripolis zone where aquifers are developed because of the impermeable basement and the morphology of the local area. Such hydrogeological conditions were found at Paleokastro, Livadi and other areas in the center of the island.

❖ *Neogene marly limestones*

These rocks have small surface development in the west of Viaradika area and northeast of Mitata area. Impermeable clays and marls form their basement. Their thickness does not exceed 50 m.

❖ *Quaternary deposits*

These deposits are located in the valleys and in the Livadi lowland. A small aquifer is developed in favorable geological conditions.

(b) Semi permeable -and impermeable formations

a) Metamorphic rocks b) Flysch formations of c) Neogene sediments d) Quaternary deposits

The hydrolithological behaviour of the various formations, as well as the morphology of the contact between permeable-impermeable formations determines the shape, extent and importance of the forming hydrogeological units. The most interesting ones are:

(a) Hydrogeological units created in the contact of metamorphic bedrock and limestones of Tripolis zone.

The geometry of the tectonic contact between limestones and metamorphic rocks, which inclines towards the sea at the North part of the island or is entirely at negative altitudes, makes the creation of exploitable underground hydrogeological units impossible, since the groundwater is channelled quickly towards the sea and is discharged at coastal or submarine springs. The main parts of Tripolis limestones that are found on the island are in open contact with the sea and discharge quickly to the sea all the water quantities they receive (PAGOUNIS & GERTZOS, 1984).

(b) Hydrogeological units of the Pindos nappe.

The upper Cretaceous limestones of Pindos zone present a satisfactory percolation factor since they are highly fractured and faulted. The hydrogeological units of the Pindos nappe are favoured by the wide surface extend of the limestones. The single hydrogeological units inside the limestones are demonstrated by springs and existent wells. The decrease in the supply of these wells with time shows that these single hydrogeological units are small and therefore, only the renewable groundwater quantities that are dependent on the annual rainfall must be pumped out. In the case of over-pumping a negative balance with a series of negative consequences will be observed. The tectonic contact of the Pindos limestones with the Tripolis flysch, the latter being an impermeable formation, plays an important role in the isolation of the single hydrogeological units within the limestone nappe.

The % surface area extends of the permeable and impermeable formations were determined at 55% (carbonate rocks) and 45%, respectively. For the carbonate rocks the percolation factor is equal to 30% of the total rainfall and the runoff factor of 3%. These factors have been estimated by the available relative data (rainfall, drainage, etc) and by comparison to similar conditions in other Greek areas. Similarly, for the

impermeable formations, the percolation factor has been estimated to 3% of the total rainfall and the runoff factor to 30%. Consequently, the evapotranspiration is 67% of the total rainfall in average (NOTARAS 1993).

	Rainfall	Percolation	Drainage	Evapotranspiration
Water volume 10^6 m^3	161.8	28.9	24.6	108.3

4. Water quality in Kythira

The purpose of the hydrochemical investigation was to determine the qualitative characteristics of karstic aquifers of the overlying limestones, in order to determine their possible communication with seawater. Twenty-three samples were collected (December 2005) and then chemical analyses were performed at the Laboratory of Hydrogeology of the School of Mining Engineering and Metallurgy of the National Technical University of Athens, Greece (PSYCHOGIOS 2006). We conclude the following:

- The values of electrical conductivity range from 450 to 550 $\mu\text{S}/\text{cm}$
- The Total Dissolved Solids (T.D.S.) range between 283 and 350 mg/l.
- The pH values range from 6.92 to 7.92 and they indicate neutral to slightly basic character of the waters. The pH at the Saint Mamas spring, which is supplied by the metamorphic rocks, is an exception and it has a value of 6.04.
- The calcium values range from 12.8 mg/l to 129.8 mg/l with a mean value of 93.02 mg/l. The movement and the possibly large time interval of the underground contact of waters with carbonate rocks, causing an increase of the calcium concentrations, can explain all the concentrations.
- The magnesium values range from 0.7 mg/l to 32.5 mg/l with a mean of 13.9 mg/l.
- The concentration of Cl^- is very low suggesting very well ‘‘protected’’ aquifers from the seawater intrusion.
- The NO_3^- concentrations are low indicating very low influence from the human activities.
- Thus, in general, Kythira waters in the overthrust limestones are protected against salinity and they range from very good to medium quality for irrigation purposes.

5. Conclusions and discussion

As it is concluded from the above, in some cases the nappe tectonics provides favourable hydrogeological conditions, as it ensures the storage of satisfactory exploitable quantities of groundwater of good quality as in the case of Kythira, without the risk of salination. These conditions dominate in a large area of the central island, where the Tripolis impermeable flysch is overthrust by slab bedded, fractured limestones of Pindos zone. The top of the flysch is at a positive altitude favoring the creation of many small single karstic aquifers in the overthrusting carbonate zone (upper karstic aquifers). These aquifers have good-quality water and are developed in order to cover the island water needs without the danger of salination even if they are over-exploited, because of the positive altitude of the impermeable flysch floor. In the contrary to the above geological structure, the carbonate rocks of the underlying Tripolis zone, found at large scale and constitutes the bedrock of the island, discharge very rapidly their waters at coastal and submarine springs. Their piezometric surface is almost at sea level and their salination is extended.

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Spring hydrograph analysis as means for determining karst aquifer parameters

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Abstract

A method for determining geometric and hydraulic parameters of karst aquifers from spring hydrograph analysis has been developed. Analytical formulae deduced from simple conceptual models describe the connection between aquifer parameters and hydrograph recession coefficient. Two significantly different flow domains exist, depending on the overall configuration of aquifer parameters. During the baseflow recession of karst systems, the drainage process is influenced by the size and hydraulic parameters of the low-permeability blocks alone. During the baseflow recession of fissured systems and the flood recession of karst systems, the recession process is dependent not only on the hydraulic parameters and the size of the low-permeability blocks, but also on conduit conductancy and the total extent of the aquifer. Analytical formulae demonstrated the drawbacks of equivalent models. While equivalent discrete-continuum models of fissured systems may reflect their real hydraulic response, there is only one adequate parameter configuration for karst systems that yields appropriate recession coefficient. Consequently, equivalent discrete-continuum models are inadequate for simulating the global response of karst systems. The global response of equivalent porous medium models corresponds to the transition between karstic and fissured type flow regimes, and thus cannot be directly applied for modeling either karst or fissured aquifers.

1. Introduction

In order to be able to construct coherent distributive groundwater flow models of karst systems, the definition of realistic hydraulic and geometric parameters is essential. The lack of such data gives rise to significant problems in modeling karst hydrogeological systems.

A simple conceptual model of karst catchments consists of a rectangular aquifer shape, a regular network of high-conductancy karst conduits embedded in the low-permeability fissured rock matrix, and a single karst spring that drains the conduit network. This model can be characterized by the hydraulic parameters of the low-permeability matrix and the conduit system, conduit spacing, and the spatial extent of the aquifer (Fig. 1.).

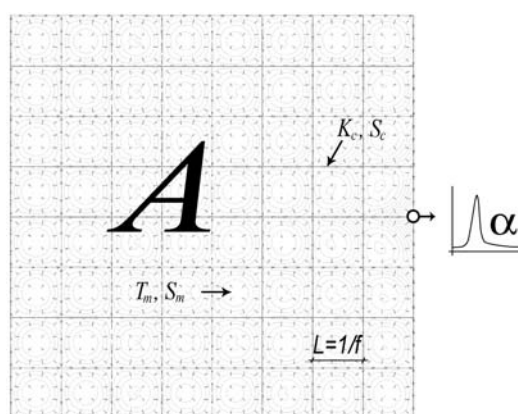


Figure 1. Simplified conceptual model of karst catchments. T_m [L^2T^{-1}] transmissivity of the low-permeability matrix, S_m [-] storativity of the low-permeability matrix, K_c [L^3T^{-1}] conduit conductancy, S_c [L] conduit storativity, A [L^2] spatial extent of the aquifer, f [L^{-1}] frequency of karst conduits.

Classical geological, geophysical and hydrogeological investigation techniques can provide only very limited information on the geometry and hydraulic properties of a conduit system. However, in most cases spring discharge time series data, coupled with information on the hydraulic properties of the low-permeability rock matrix are available.

Every hydraulic process taking place in a karst aquifer manifests in the temporal variations in spring discharge. The plots of spring discharge versus time are referred to as spring hydrographs. Hydrographs consist of a succession of individual peaks, each of which represents the global response of the aquifer given to a precipitation event (Fig. 2). Hydrograph peaks consist of a rising and a falling limb. The falling limb comprises a steep and a slightly sloped segment. The former is called flood recession, while the latter is referred to as baseflow recession, which is the most stable section of any hydrograph.

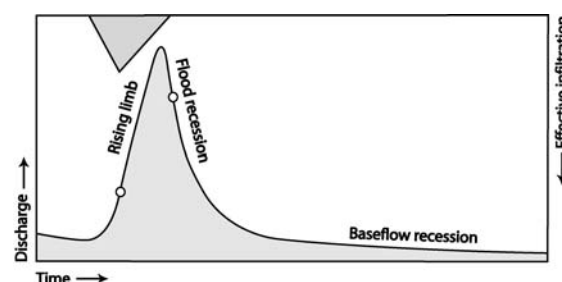


Figure 2. Typical features of a spring hydrograph.

The first mathematical characterization of the baseflow recession was provided by Maillet (1905). This interpretation is based on the drainage of a simple reservoir, and presumes that the spring discharge is a function of the volume of water held in storage. This behavior is described by an exponential equation as follows:

$$Q_{(t)} = Q_0 e^{-\alpha t} \quad (1)$$

where Q_t is the discharge [$L^3 T^{-1}$] at time t , and Q_0 is the initial discharge [$L^3 T^{-1}$], α is the recession coefficient [T^{-1}] usually expressed in days. Plotted on a semi-logarithmic graph, this function is represented as a straight line with the slope $-\alpha$. This equation is usually adequate for describing baseflow recession of karst systems, and is believed to reflect the drainage of the saturated low-permeability fissured matrix.

The aim of this paper is to quantitatively characterize the connection between the hydraulic and geometric properties of karst aquifers and their global response, in order to facilitate distributive groundwater flow modeling of karst systems. This has been achieved by deducing analytical solutions for the global hydraulic response of simple two-dimensional (2D) domains. The analytical formulae were then tested by numerical models. Resulting formulae express the connection between the hydraulic and geometric properties of a karst system and spring hydrograph recession coefficient, which is believed to be a characteristic parameter of the global response of a karst system.

3. Methodology

In order to provide a mathematical characterization of diffusive flux from a square conductive block, a 2D analytical solution has been derived from the heat flow equation solution of Carslaw & Jaeger (1959), using the following formula:

$$Q_{(t)} = \frac{128}{\pi^2} T \sum_{n=0}^{\infty} e^{-\frac{(2n+1)^2 \pi^2 T t}{SL^2}} \sum_{n=0}^{\infty} \frac{e^{-\frac{(2n+1)^2 \pi^2 T t}{SL^2}}}{(2n+1)^2} \quad (2)$$

where uniform hydraulic heads are assumed as boundary conditions along the sides of the square block, and initial conditions comprise uniform hydraulic heads over the block surface. Neglecting the higher order terms of the series, the discharge can be approximated as

$$Q_{(t)} = \frac{128}{\pi^2} T (e^{-\frac{2\pi^2 T t}{SL^2}}) \quad (3)$$

Comparison of this solution with the classical formula of Maillet (1905) (**Eq. 1**) shows that the recession coefficient of a 2D homogeneous block may be expressed as

$$\alpha_b = \frac{2\pi^2 T}{SL^2} \quad (4)$$

The recession of an entire karst system can be determined using this approach if the conductive capacity of the water drainage system is assumed to be sufficiently high and the storage in the conduit network is neglected. Based on this approach, the discharging water from low-permeability blocks is assumed to reach the outlet of a system instantaneously.

In order to check the domain of validity of the above formula and the influence of changing conduit

characteristics on the recession coefficient, numerical sensitivity analyses have been performed, using the combined discrete-continuum method (Király 1985, 1998; Király & Morel 1976). The numerical codes FEN1 and FEN2 were applied throughout the simulations. With an initial set of selected simulations, the sensitivity of the recession coefficient to the conduit conductance was investigated. A second series of simulations was performed in order to investigate the sensitivity of the recession coefficient to conduit frequency. Conduit storage coefficients S_c [L] were calculated from relevant conduit apertures assuming water compression alone.

Numerical simulations show that an increase in conduit conductance results in a rise in recession coefficient, until reaching the value of the analytical solution for a single homogeneous block. By exceeding a threshold value, the increase of the conduit conductance has no further influence on the baseflow recession coefficient, the recession process is controlled by the hydraulic parameters of the low-permeability blocks alone. Conduits act as fix-head boundary conditions as assumed by the above analytical model. This flow condition has been defined as *matrix-restrained flow regime (MRFR)*.

If the conduit conductance is lower than the threshold value, the recession coefficient is strongly dependent on this parameter, and the analytical **Equation (4)** is no longer valid. This flow condition has been defined as *conduit-influenced flow regime (CIFR)*. The influence of changing conduit frequencies on the recession coefficient follows similar principles as the alteration of conduit conductance. A threshold value of conduit frequency exists, above which the change of conduit frequency entails the change of the recession coefficient.

By arranging the hydraulic parameters of a porous medium in a manner that an elementary volume of the aquifer transmits the same specific discharge and releases the same amount of water from storage as that of a heterogeneous domain, an equivalent porous medium can be made.

Supposing that the recession of a conduit influenced heterogeneous domain can be approached by the drainage of an equivalent porous domain of the same dimensions and same boundary conditions, the equivalent transmissivity parallel to karst conduits and equivalent storage (S_{eq}) may be expressed as follows:

$$T_{eq} = K_c f + T_m \quad (5)$$

$$S_{eq} = S_m + 2S_c f \quad (6)$$

where T_{eq} is equivalent transmissivity [$L^2 T^{-1}$], K_c is conduit conductance [$L^3 T^{-1}$], S_m is matrix storativity [-], S_c is conduit storage coefficient [L], f is conduit frequency (number of conduits per unit length) [L^{-1}], and T_m is the transmissivity of the matrix medium [$L^2 T^{-1}$].

We assume that the formulation of recession coefficient of diffusive flux from a homogeneous domain of arbitrary shape takes the following general form:

$$\alpha = \frac{\gamma T}{SA} \quad (7)$$

where γ is a geometric factor [-] dependent on domain shape and boundary conditions, and A is the domain area [L^2].

Substituting the equivalent parameters into this general formulation, the recession coefficient becomes

$$\alpha_h = \frac{\gamma}{A} \frac{(K_c f + T_m)}{(S_m + 2S_c f)} \approx \gamma \frac{(K_c f + T_m)}{S_m A} \approx \gamma \frac{K_c f}{S_m A} \quad (8)$$

where the formula can be further simplified by neglecting matrix permeability and conduit storage. The geometric parameter γ in Equation (8) was determined and the equation tested by a large variety of discrete-continuum models (Fig. 3).

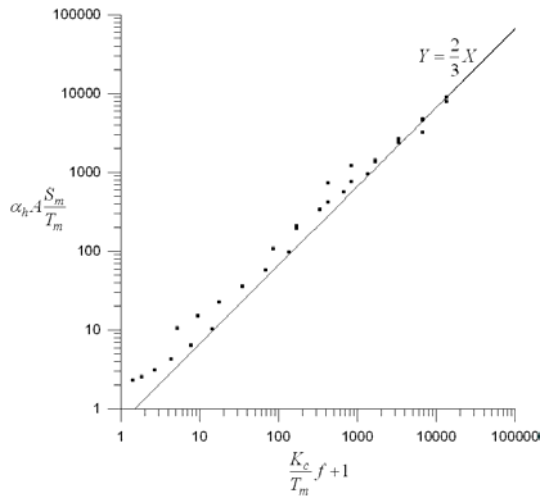


Figure 3. Evaluation of the geometric parameter (γ) of conduit-influenced regular karst systems by the means of curve fitting to numerical model results.

Curve fitting on a dimensionless graph of numerical model results yields the value of $\gamma=2/3$. The final formula for the recession coefficient in the conduit-influenced baseflow domain can be expressed as follows:

$$\alpha_h = \frac{2}{3} \frac{(K_c f + T_m)}{A(S_m + 2S_c f)} \approx \frac{2}{3} \frac{(K_c f + T_m)}{S_m A} \approx \frac{2}{3} \frac{K_c f}{S_m A} \quad (9)$$

3. Results

The alteration of aquifer hydraulic and geometric properties results in changing recession coefficients. The dependence of recession coefficient on aquifer properties follows two fundamentally different principles:

The MRFR flow regime is controlled by the hydraulic parameters of the low-permeability medium. This case can be mathematically characterized by the drainage of a homogeneous block (Eq. 4).

The CIFR flow regime is mainly controlled by the conductive capacity of the conduit system. This case can be mathematically characterized by Eq. (9).

The two principal baseflow domains are linked by a transition zone. The threshold separating the two principal flow domains can be mathematically expressed by equating the matrix-restrained and conduit-influenced recession coefficients:

$$K_c^* = 3\pi^2 T_m A f - \frac{T_m}{f} \approx 3\pi^2 T_m A f \quad (10)$$

For smaller values of K_c , flux is mainly restrained by the high-permeability conduit network, while for higher values of K_c , flux is restrained by the low-permeability matrix (Fig. 4)

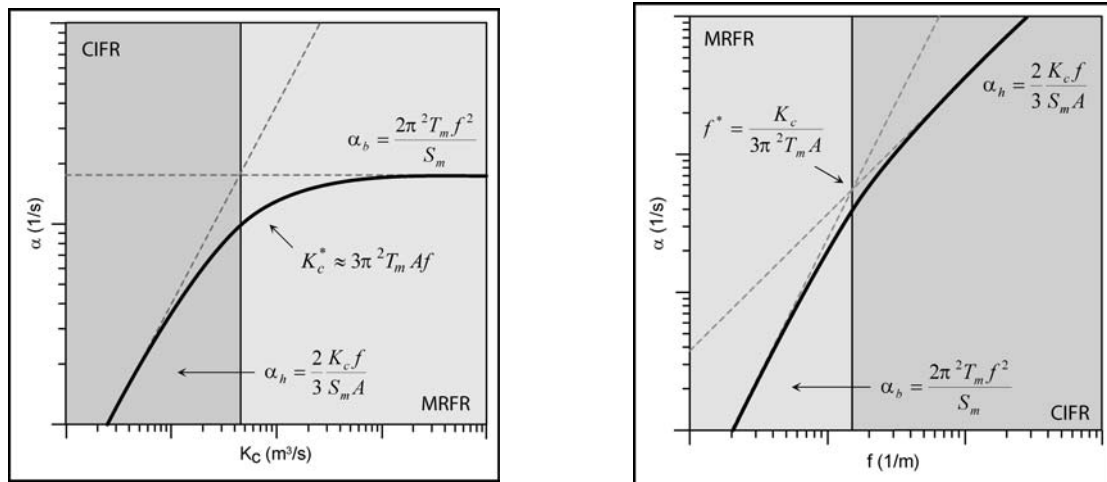


Figure 4. Relationships between recession coefficient, conduit conductance and conduit frequency on log-log graph. K_c^* and f^* represent the threshold values of conduit conductance and conduit frequency. Modified after Kovács (2003) and Kovács et al. (2005).

The evaluation of threshold parameters facilitates the quantitative classification of karst systems: while

MRFR flow regime is a characteristic of mature karst

systems under baseflow conditions, early karst systems and fissured systems exhibit CIFR baseflow.

The validity and applicability of equivalent models is a crucial question in modeling fissured and karst systems. Numerical tests demonstrated that in the MRFR baseflow domain, equivalent discrete-continuum models yield systematically different recession coefficients, and thus are invalid (**Fig. 5**). Only the real parameter configuration gives appropriate results. In the CIFR baseflow domain, equivalent discrete-continuum models are appropriate as they yield identical recession coefficients. In the centre of the transition zone, threshold parameters designate an inflection point that corresponds to the recession coefficient of an equivalent porous medium. This is the only point of validity of the equivalent porous medium approach when simulating strongly heterogeneous systems.

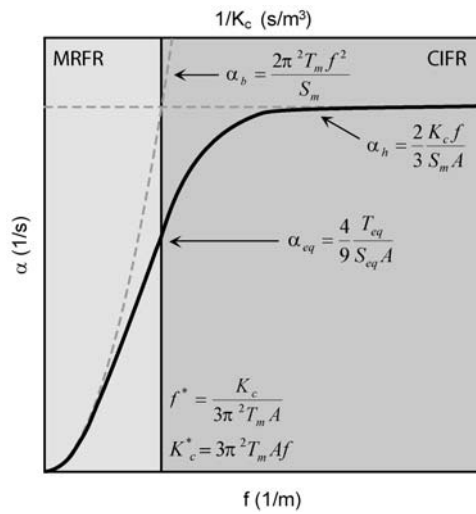


Figure 5. Variations of the recession coefficient for equivalent discrete-continuum domains.

4. CONCLUSIONS

Discharge time series represent a very important means of determining of information about the overall structure of a karst system. The drainage of an arbitrary porous domain can be described by a general exponential model, with the following recession coefficient:

$$\alpha = \frac{\gamma T}{SA}$$

where γ is a geometric factor that depends on boundary conditions and aquifer shape. Numerical sensitivity analyses identified two fundamentally different flow domains, depending on the overall configuration of aquifer parameters. The baseflow recession coefficient of fissured systems or weakly karstified systems can be expressed as follows:

$$\alpha \approx \frac{2 K_c f}{3 S_m A}$$

while the baseflow recession coefficient of karst systems is:

$$\alpha = \frac{2 \pi^2 T_m f^2}{S_m}$$

The validity of equivalent models is strongly restricted. While equivalent discrete-continuum models may lead to adequate recession coefficients in the conduit-influenced baseflow domain (fissured systems and early karst systems), there is only one parameter configuration that yields appropriate recession coefficient in the matrix-restrained baseflow domain (mature karst systems). The direct application of equivalent porous medium models is inadequate both in the case of karst and fissured systems.

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Tetxa inversac (Baia river): its influence on the confined zone of the Subijana Carbonate Unit (Basque Country)

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Abstract

Tetxa inversac is located on the Baia river channel along the Subijana Carbonate Unit (Basque Country), where there exists an important aquifer-river relationship. This structure, in low-water period, works as a sink (shallow hole) infiltrating part of the superficial water in some moments, and even the whole of it in the end of the period. In high waters, on the contrary, the structure works as a discharge point of the aquifer. Since river water comes after the contact with the Murgia diapire (gypsum), its sulphate signal is extremely different from the groundwater in the aquifer; so the spatial-temporal influence of the infiltrated water on the aquifer has been investigated by controlling the hydrochemical signature in the wells down-gradient from Tetxa, where the aquifer becomes confined.

We used sulphate as natural tracer to have evidence of this influence. Along the beginning of the 2002 summer, the exhaustive hydrochemical and local piezometric control allowed us to determinate the flow pattern in the confined part of the aquifer, giving evidence of a north-south component of flow with some lateral dispersion. In conclusion, while in some wells superficial water has a direct and immediate influence, in other wells it is weak and almost non-existent.

Although just right now these wells are not exploited, in the near future they will be, and so the understanding of this phenomenon is relevant from the point of view of its vulnerability.

Key words: inversac, aquifer-river relationship, hydrochemical signature, piezometer, vulnerability

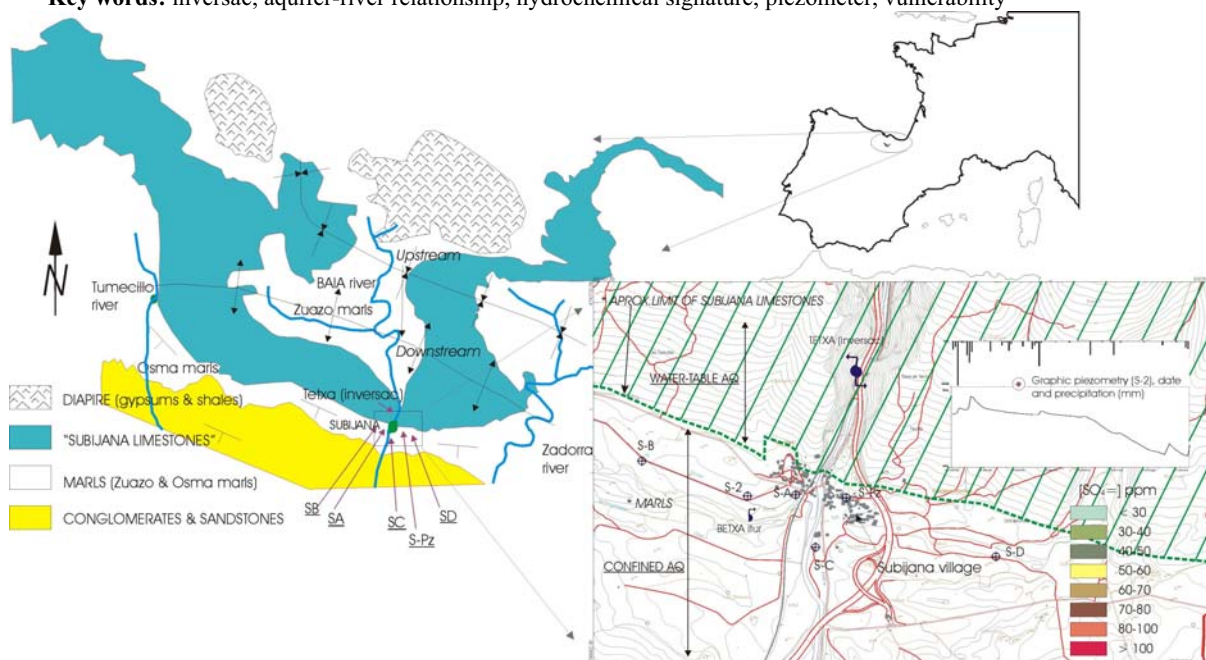


Figure 1. Schematic geology of the Subijana Hydrogeological Unit and the closer and main lithologies. Location of Tetxa inversac, wells and piezometers in the Subijana sector.

1. Introduction

The Subijana Hydrogeological Unit (170 km², 700-900 mm/year of rainfall), fractured and partly karstified Upper Cretaceous limestones of about 300 m thickness, is divided in 5 sectors by previous studies (EVE, 1996; LARRUZZEA & ANTIGÜEDAD 2003). This paper is focused on the, maybe most important of them, the sector around the village of Subijana (Fig. 1). The limestones in this sector have a monoclinial disposition deeping toward the south where are

covered by marls; so, the carbonate aquifer «goes» from a water-table to a confined aquifer.

The drainage of the aquifer in the Subijana sector is controlled in high water periods by the Baia river, a river which crosses the limestones. In this narrow pass exists an inversac structure, working as a spring in high-water conditions and as a swallet (shallow hole) in low-waters when river sinks into the aquifer.

This aquifer, scarcely exploited, is really interesting because of its strategic location, near the city of Vitoria-Gasteiz (250.000 inhabitants), and because it has important

water resources (estimation of 72 hm³/year for the whole unit). Around 8 wells (up to 300 m deep) are located in the sector under consideration, with high productivity and big water level fluctuations (up to 40 m). Groundwater levels (GWL) are continuously measured in two piezometers in the Subijana sector, S2 (www.entevascodelaenergia.com/redbas) and S5, and fortnightly in 5 wells.

Figure 1 shows the schematic geology of the Unit and the closer and main lithologies. It is remarkable the marls around the limestones, Zuazo marls upstream and Osma marls downstream. This last lithology is the one that confines the aquifer toward the south, down gradient. The Murgia Diapire is also important, with gypsum and shales, where superficial water (Baia river) gets its sulphate mark. This superficial water is monitored in Tetxa inversac before it sinks (Fig. 2). Figure 1 also shows the location of the inversac, wells and piezometers in the Subijana sector.

To know the flow pattern of groundwater in the aquifer and its relationship with the river in the Subijana sector, we carried out a strong hydrochemical control (Fig. 2) in superficial water in Tetxa inversac, before it sinks, and in the wells (SA, SB, SC, SD and Subijana PZ, in Fig. 1) on summer 2002. All the wells are located in the confined zone of the aquifer, and almost parallel to the border of the limestones.

The data obtained in this control are shown in Figure 2. As the infiltration in the inversac is progressive, our intention was to begin with the chemical control before the river sank completely in summertime. When it happens, it is also possible to check that piezometry has a clear break point around 513.5 m, with an acceleration in piezometric decrease. This break point is possibly linked to the «regional flows» (LARRUZZA & ANTIGÜEDAD 2003).

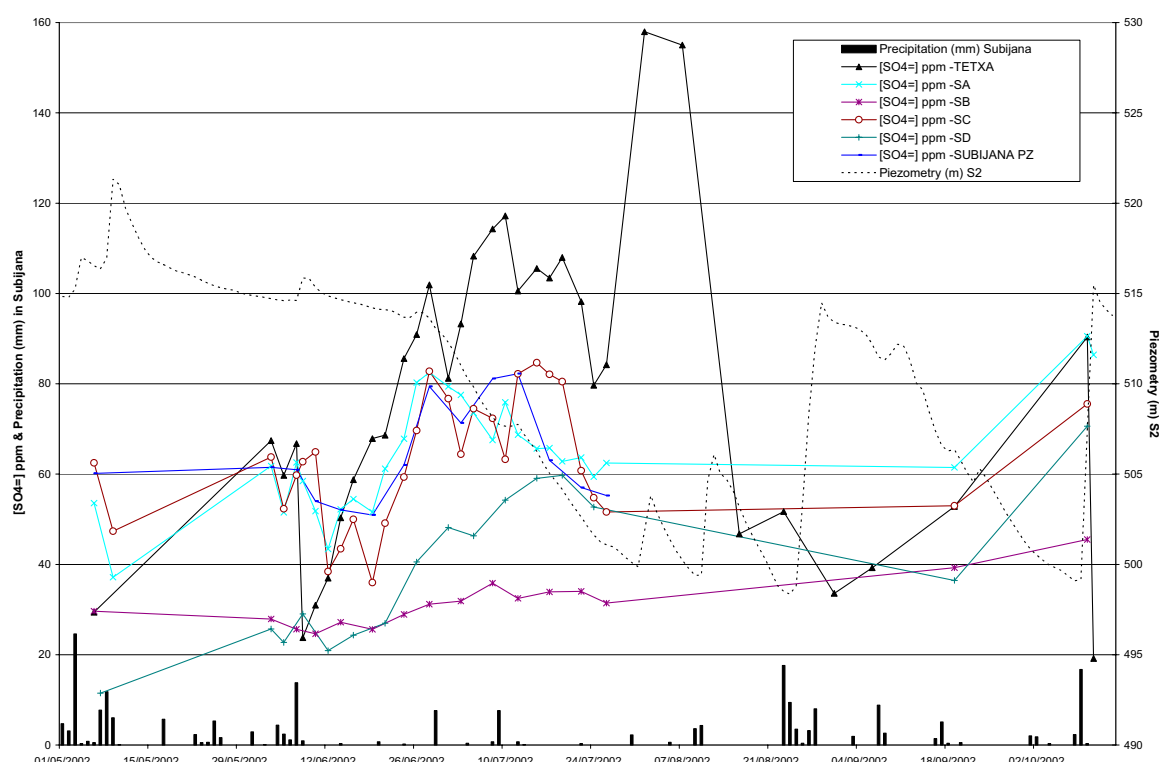


Figure 2. Hydrochemical data in Tetxa inversac (superficial water) vs. wells data (SA, SB, SC, SD and Subijana PZ). Precipitation (mm) and piezometry (S-2).

2. Materials and methods

We have chosen sulphate as natural tracer because in the studied sector it can be considered as an external lithologic marker, and at the same time sulphate concentration has the highest differentiation with the background of the aquifer (20-30 ppm, LARRUZZA, in preparation) and the highest quantitative concentrations.

To know the direction of water after it sinks into the aquifer, we used the sulphate mark of the superficial water for the 2002 summer. As we have mentioned before, the superficial water in Baia river goes through Murgia Diapire ; at this point there is a dissolution of gypsums that gives a mark of sulphate up to 160 ppm (Fig. 2).

From another point of view we consider sulphate to be conservative, at least in the zone under investigation; in fact, no reduction ambient has been detected at this scale of time and space.

The water samples have been taken at different depths in each well by using a sampler, that allows us to accede to the productivity zones of each well. These productive zones themselves and to which of them fresh water arrives have been described before by using a multi-parametric probe (LARRUZZA & ANTIGÜEDAD 2002; LARRUZZA, in preparation).

In Figure 2 there is not any hydrograph of Baia river because there is not any gauging station near this sector. This test has been, mostly, focused on the beginning of 2002 summer, from 2002/06/03 to 2002/07/26.

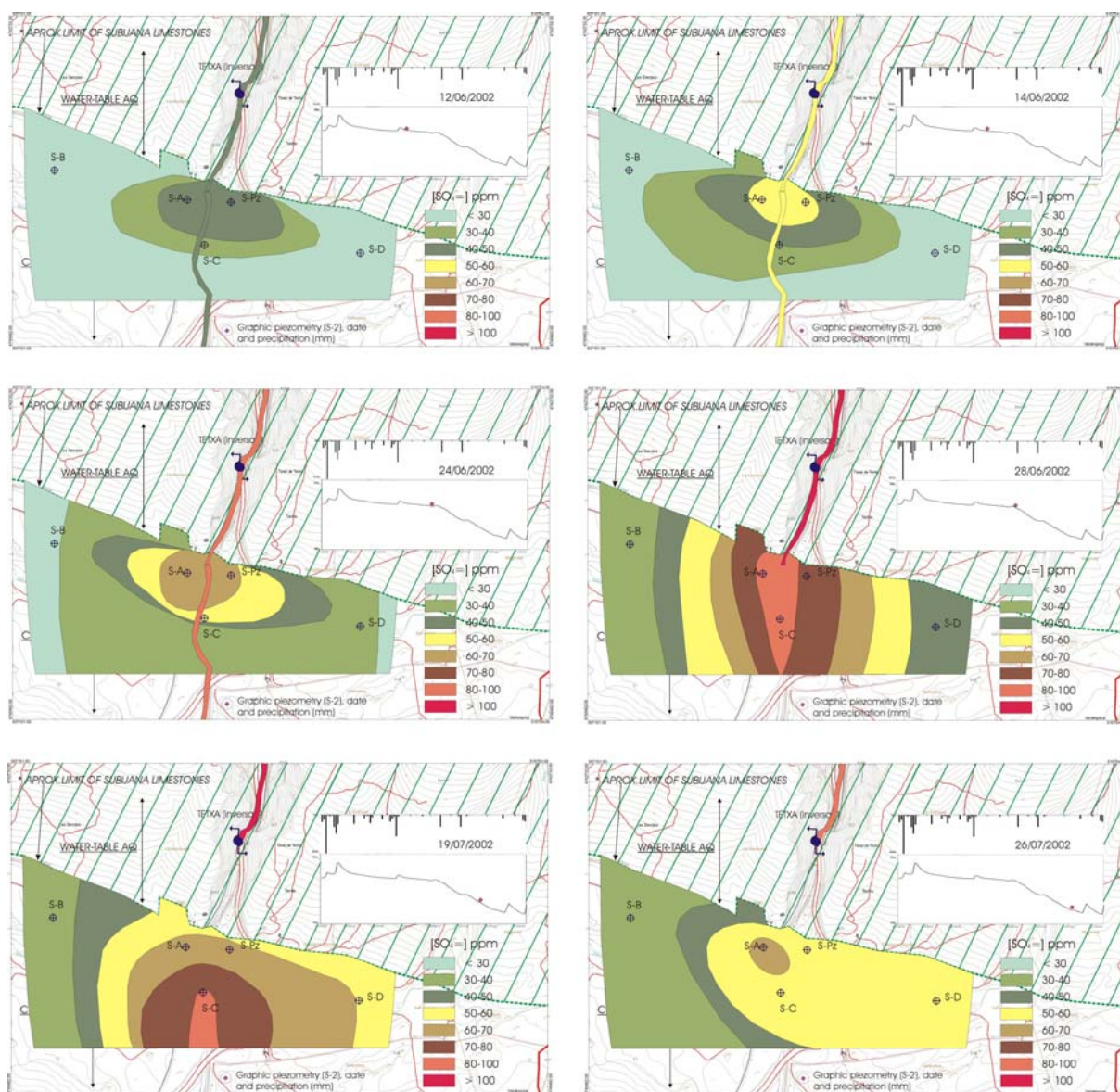


Figure 3. Sulphate iso-concentrations maps in Subijana Sector for 2002 summertime.

3. Results

In an initial interpretation of the data from the wells (Fig. 2), there exists a clear sulphate concentration increase along the Baia river (Tetxa) when discharge decreases (due to a greater influence of sulphates from the diapire), from a minimum around the background of the aquifer, up to a maximum of 160 ppm. Evidently rainfall events cause a decrease in the sulphate concentration. It is also clear a great influence of the superficial water on SA, SC and Subijana PZ wells, lower on SD well (east side of the sector) and much more lower on SB well (west side).

But in this figure it is hard to see any spatial influence on the confined part of the aquifer, any preferential corridor and any flow direction. In the same way, it is difficult to see a time-spatial reaction with the input signal (superficial water). In order to avoid all these difficulties, sulphate iso-concentrations maps have been represented by using as model the one represented in Figure 1.

It is evident that reality has to be more complex than these simple maps, and even much more complex for this karstified sector of the aquifer.

The infiltration in the inversac is progressive, so the different stages in Figure 3 have been drawn with three different coloured stretches: three drawn/ coloured stretches when river does not sink completely, this is, when river flows after the limestones; two stretches when it sinks completely but arrives up to the limestones limit downstream; and, finally, just one stretch when it sinks completely in the principal location for the inversac structure, this is Tetxa inversac.

Since it is impossible to show in this paper all the sulphate concentration maps along the duration of the hydrochemical control, we have chosen the most representative of them all.

In Figure 3 in order to make easier its interpretation, it is also possible to check the piezometric evolution for piezometer S-2 (m) and precipitations (mm/day) in Subijana meteorological station, signaled them both by a rhombus.

The 2002/06/12, after some precipitations, was the beginning of a new period. For this day most of the Sector had a sulphate concentration close to the aquifer background, this is, a concentration under or close to 30 ppm. Just sulphate concentration in SA and Subijana-Pz wells is similar to that of the superficial water, between 40 and 50 ppm.

On the 2002/06/14, the similitude between SA and Subijana-Pz wells and Baia river continues, with an increase of sulphate all around the sector. It seems that for this time there is no influence on SB and SD wells.

On the 2002/06/24, ten days later without rainfalls, a greater influence of the superficial water on the aquifer is evident, mostly for SA and Subijana-Pz wells. At this point concentration in the river water is higher than 80 ppm.

Baia river water's influence arrives for the first time to SD well (2002/06/24). For SB well it will be later (on the 2002/07/08 -not shown-).

On the 2002/06/28, it is remarkable that Baia river does not continue flowing after the limestones, sinking the whole of the water just when the piezometry gets its «break point» (513.5 m). In this figure it is also evident the great influence of the superficial water on the aquifer with a clear corridor towards the south. Water in the central part of this corridor (more than 80 ppm) is similar enough to that in the river (close to 100 ppm).

On the 2002/07/19, three weeks later and just in the recession of the piezograph, the concentration in superficial water is around 100 ppm. There is lower and lower flow in the river and now it sinks the whole of the water in the main inversac and no, like previously, all along the limestones. So, in this situation with a lower «pressure» from the superficial water, it is evident a displacement of the highest concentrations of the sulphate following the central corridor southward and a clear non-connection between the more influenced well, SA, and the superficial waters.

This displacement toward the south could be interpreted as a consequence of the «regional flows». In fact, according to the geological setting, groundwaters are to be driven beneath the marls of Osma in a regional pathflow (discharge point of this flow is not clear up to now).

Baia river water's influence on SD well (east side of the sector) is really high at this moment (near 60 ppm).

On the 2002/07/26, after a lower sulphate «pressure» from the superficial water due to a lower inflow of the river, it is evident a cleaning up of the aquifer. This event could be the result of the defined lower «pressure», a dilution along the groundwater flow, but, mostly, due to the mentioned regional flow toward the south or, may be, south-east.

4. Conclusion

It is possible to investigate the spatial and temporal influence of the infiltrated river water on the aquifer by hydrochemical signature in the wells down-gradient from Tetxa (inversac), where the aquifer becomes confined.

As we have shown in the tracing test based on the natural sulphate signal, there is an important spatial influence from the superficial water on the Subijana aquifer. This influence is maximum along a corridor of north-south direction, that includes SA, SC and Subijana Pz wells, with a south component (down gradient).

The temporal influence could be described as almost immediate for SA well (from some hours to few days) and bigger for SC well, approximately, between three and four days. Subijana Pz well would had a value close to the last

one. For SD and SB wells influence is smaller and with a greater delay. SD is better connected than SB. So, after horizontal distances approximately speed (m/day) could be calculated: SA, 770; SC, 250; Subijana Pz, 150; SD, 105 and SB, 50.

This influence pattern would be also shown by the electrical conductivity. This parameter median value for the test in Baia river (373 $\mu\text{S}/\text{cm}$) is closer to the measured in the central corridor (SA, 375; SC, 389; Subijana Pz, 427). SD has a higher value (439) and SB (579) has the biggest difference.

So, while in some wells superficial water has a direct and immediate influence, in others it has an influence weak and almost non-existent.

Finally, we can conclude that Baia river, with its differential hydrochemical signal, is a pressure element that has the maximum influence on the aquifer when the infiltration and superficial water's sulphate signal is also maximum. This occurs when Baia river flows up to the limestone contact (piezometry around 513.5 m) and after some days without rain (maximum concentration). On the other hand, when river sinks completely due to a lower flow, Baia river's pressure is lower, allowing a cleaning up of the aquifer.

The understanding of this phenomenon is relevant from the point of view of the vulnerability of the aquifer in this sector. So, it is important to know when would be the maximum influence from the superficial water, mostly if the wells of this sector begin to be exploited.

In this way, if the sector is exploited, and groundwater levels depressed, Tetxa inversac will work as swallet for a longer season during the year. So, it is important a good management of the superficial basin upstream from this structure, this is, Baia river itself (SCI- *site of Community importance*) and all around it, especially industry and population areas.

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Vulnerability mapping in the Cradle of Humankind World Heritage Site, South Africa

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Abstract

Dolomitic aquifers are vital sources of water in South Africa and under increasing stress due to expanding urban, mining and industrial developments. The karst aquifers of the Cradle of Humankind World Heritage Site (COHWHS) are an example of an area where the suitability of the water resource is threatened. The COHWHS is predominantly underlain by strata of the Chuniespoort and Pretoria groups. The dolomites of the Malmani Subgroup act as the main aquifer and are an example of a karst developed on a very old dolomite that has been subjected to deep burial, tectonization, folding, uplifting, and prolonged episodes of natural erosion. The continuity of the aquifer is interrupted by various, low permeability, igneous intrusions (vertical and sub-vertical dykes) resulting in groundwater compartments within the dolomite. Although numerous groundwater vulnerability studies have been carried out in South Africa, no karst specific aquifer vulnerability determination method exists. For this reason the COP vulnerability mapping method will be adapted to compensate for the unique properties of South African karst terrains like the COHWHS as well as the climatic conditions in semi-arid environments. Climatic data analyses illustrate the highly seasonal and erratic nature of the rainfall in the study area. Precipitation typically occurs during intense thunderstorm events, exceeding the infiltration capacity of the overburden and generating large amounts of surface runoff.

1. Introduction

The karstified dolomites of the Chuniespoort Group are capable of sustaining high-yielding boreholes and are the only readily available water resource for many towns, rural areas and farms. They form also a vital component of the water resources needed for the expanding urban and industrial complexes in Gauteng and Rustenburg; hence they are considered as one of the most important aquifers in South Africa (BARNARD 2000). The present pressure for urban development onto dolomite areas is a major concern regarding the safety of residents (sinkholes) and will also create an enormous potential source of pollution to the dolomitic aquifers underlying these areas. Despite its importance no scientific based outline of areas that need protection from potential harmful activities, and areas where such activities would constitute a minor risk to the natural resources and sources exists.

The Cradle of Humankind World Heritage Site (COHWHS) is one such dolomitic terrain that is under threat. The gold mines in the Western Basin stopped dewatering operations, which caused the water table to rebound, and in 2002 acid mine water started to decant from the so-called Chinese shaft near Krugersdorp, south of the COHWHS. This water is extremely acidic and highly mineralised and flows into the COHWHS catchment via the Bloubankspruit, which later enters the aquifer via a swallow hole. Another significant source of pollution for the water resources of the COHWHS is the large volume of sewage effluent return flow municipal sewage works that enter the catchment and therefore threaten the suitability of the karst aquifer system as a water resource. The concept of vulnerability mapping will be an important decision tool when expansion of urban areas onto dolomitic terrains must be evaluated and the concept should be expanded to include the surface stability (potential for sinkholes and dolines) if the resource is used as a regular or emergency water supply during future scenarios. A karst terrain vulnerability mapping procedure that takes into account the international knowledge gained in karst vulnerability mapping as well as the specific conditions of the South African dolomites, soils and climatic conditions should be developed.

2. Geological setting of the study area

The COHWHS is situated approximately 40km northwest of Johannesburg, South Africa and covers 47 000 hectares of land that is mostly privately owned (Fig. 1). The site was declared a UNESCO World Heritage Site in 1999 and comprises an exceptionally large, unique and scientifically significant band of palaeo-anthropological sites which have yielded valuable insight in regard to the origin of modern humans. These include a dozen dolomitic limestone caves containing the fossilised remains of ancient forms of animals, plants and most importantly, hominids.

The site is underlain predominantly by strata of the Chuniespoort and Pretoria groups of the Transvaal Supergroup. Rocks of the Halfway House Granites, Ventersdorp Supergroup and Witwatersrand Supergroup underlie minor sections of the area as shown in (Fig. 2). The Halfway House Granites are Swazian age intrusives that are exposed as a domical “window” of granitic basement rocks (SACS 1980). They consist of tonalitic gneisses, migmatites, gneisses and porphyritic granodiorites, dated at 3203 ± 65 Ma (ALLSOPP 1961 in SACS 1980:87).

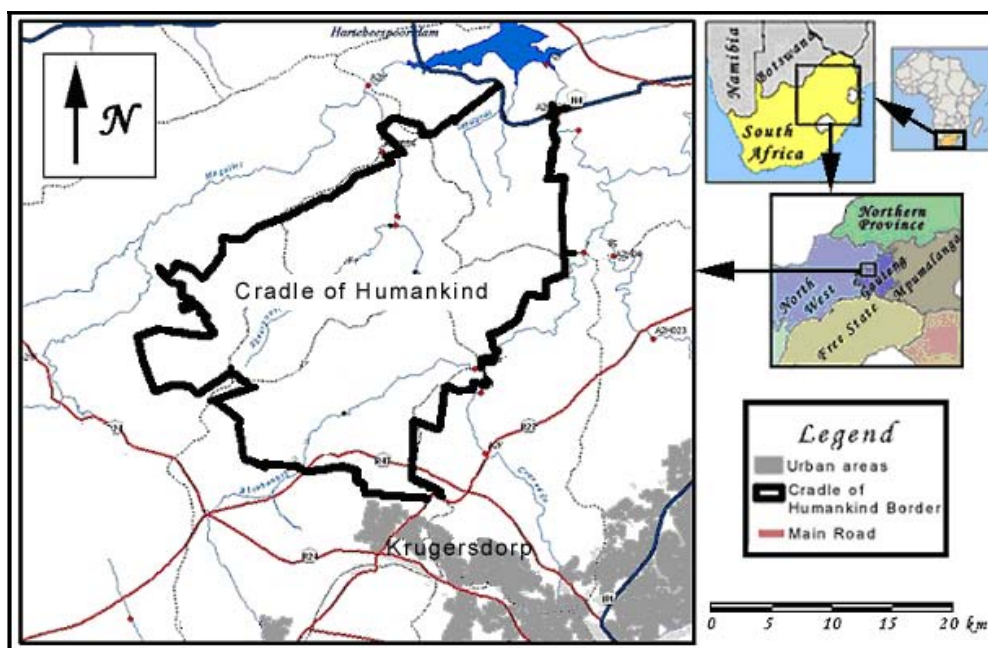


Fig. 1. Study area locality map (adapted from HOLLAND *et al.* 2005).

The Witwatersrand Supergroup is Archaean in age (3074 to 2714 Ma) and consists of a lower West Rand Group (maximum thickness 5150m) and an upper Central Rand Group (maximum thickness 2880m). Sandstone and shale dominate within the West Rand Group whereas the Central Rand Group consists predominantly of sandstone and conglomerate with only minor shale (ratio in order of 1:12) (ROBB & ROBB 1998). The Ventersdorp Supergroup (2714-2709 Ga) is made up of a lower Klipriviersberg Group, middle Platberg Group and the Bothaville (sedimentary) and Allanridge (volcanic) Formations, which are all unconformable (BRINK 1979). The Ventersdorp Supergroup is an accumulation of andesitic to basaltic lavas with related pyroclastic rocks (agglomerates and tuffs) and a number of sedimentary intercalations and has a maximum thickness of 8000m (BRINK 1979).

The Neoarchaeo-Palaeoproterozoic Transvaal Supergroup is preserved in three separate structural basins (the Kanye (Botswana), Griqualand West and Transvaal basins) and comprises five major lithostratigraphical subdivisions: the so-called protobasinal rocks (purely a descriptive and not a genetic connotation), the Black Reef Sandstone Formation, the Chuniespoort Group carbonate-banded iron formation (BIF) platform succession, the volcano-sedimentary Pretoria Group and the uppermost Bushveld-related Rooiberg Felsite Group (ERIKSSON & RECZKO 1995). The Transvaal basin (present in study area) forms the floor rocks to, and outcrops in a limited band around, the 2.05 Ga old Bushveld Complex intrusion.

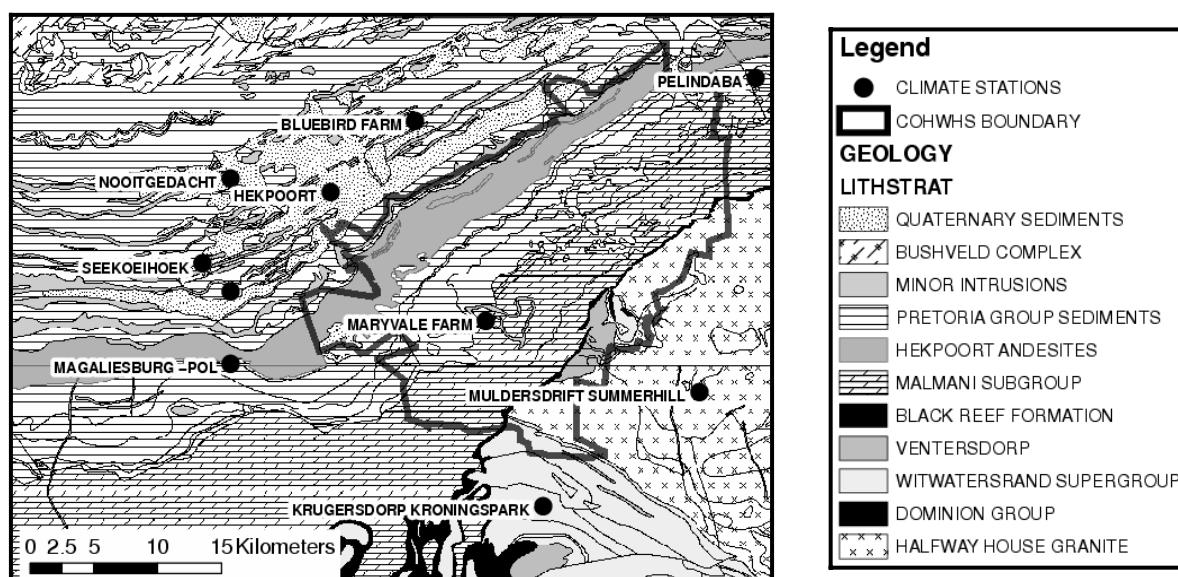


Fig. 2. Geological setting of study area.

The protobasinal rocks are preserved in discrete units around the margins of the preserved basin but are not present within the study area. The Black Reef Formation succeeds the protobasinal successions unconformably and ranges in thickness from a few meters to thirty meters. Basal conglomerates are overlain by interbedded sandstones and mudrocks and the arenites increase in textural and mineralogical maturity with increasing stratigraphic height (ERIKSON & RECZKO 1995). The Black Reef Formation is conformably overlain by the Chuniespoort Group consisting of the lower Malmani subgroup (stromatolitic dolomite with chert interbeds) and the upper Penge (banded iron formations) and Duitschland (mixed clastic and carbonate rocks) Formations. The later two units are absent in the study area. The Malmani Subgroup is subdivided into the Oaktree, Monte Christo, Lyttleton, Eccles and Frisco Formations based on the interbedded cherts and shales, the variety of stromatolite structures present and the low-angle unconformities (BUTTON 1973; SACS 1980).

The Chuniespoort Group is unconformably overlain by the Pretoria Group and this succession buries a palaeokarst landscape developed on exposed dolomites during a depositional hiatus of at least 80 Ma. The Pretoria Group is predominantly an alternating sequence of mudrock and sandstone formations with subordinate conglomerates, diamictites and carbonate rocks and significant interbedded basaltic-andesitic lavas (ERIKSSON & RECZKO 1995).

The entire Transvaal Supergroup originated between 2658 ± 1 Ma and 2224 ± 21 Ma (ERIKSSON *et al.* 2001) with the carbonate sequence being deposited over a period of at least 120 m.y. between 2643 and 2520 Ma (OBBES 2000). Deformation of the Transvaal Basin sedimentary strata is limited to open interference folding, faulting and syn-Bushveld dykes and sills (ERIKSSON *et al.* 2001) and the Transvaal strata dip at angles up to 20° toward the centrally located Bushveld intrusives (ERIKSSON *et al.* 1995).

3. Hydrogeological description

The generally high to very high storage capacity (storativity) and high permeability of the Chuniespoort dolomites make it one of the most important aquifers in South Africa (BARNARD 2000). The Karst of the Malmani Subgroup Dolomite, Chuniespoort Group is an example of a karst developed on a very old dolomite that has been subjected to deep burial, tectonization, folding, uplifting, and prolonged episodes of natural loss of soil and rock debris in the interior of the Kaapvaal Craton. The karstified dolomite acts as the main aquifer, though fractures extending to considerable depths within the non-karstified dolomite can also support high yields. However the continuity of the dolomite aquifers of the Chuniespoort Group is interrupted by various igneous intrusions (vertical and sub-vertical dykes) that serve as low permeability or impermeable barriers to groundwater movements (BARNARD 2000). These dykes therefore result in the formation of groundwater compartments within the dolomites (Fig. 3). Groundwater is transferred from one compartment to the next either via leakage in near surface weathered zones of dykes or via overflow, i.e. high yielding springs at their lowest surface elevations near an impermeable boundary.

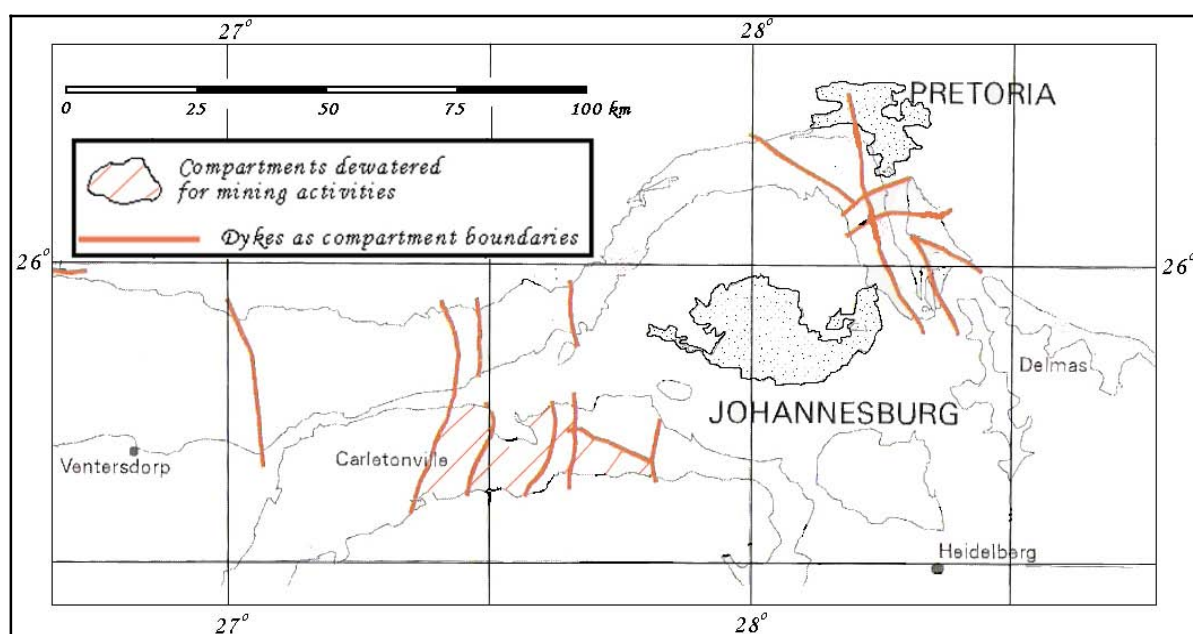


Fig. 3. Distribution of individual dolomite compartments (adapted from BARNARD 2000)

4. Vulnerability mapping in South Africa

Scientific investigations of the karst regions in South Africa started in the early sixties (DE KOCK 1964; BRINK & PARTRIDGE 1965). Since then many investigations have been carried out by geologists, hydrologists and geomorphologists. The cycle of karst development associated with the regional dolomite have long attracted attention from South African geologists (MARKER 1972; MARTINI & KAVALLIERIS 1976; MARKER 1980; WOLMARANS 1986; MARTINI *et al.* 2003).

Water has always been needed for the growing population and this has led to numerous large-scale and widespread groundwater investigations on the groundwater potential of the regional dolomite (e.g. ENSLIN & KRIEL 1967; FLEISHER 1981; FOSTER 1984; BREDEKAMP *et al.* 1986; KUHN 1986; BREDEKAMP 1995; AND BARNARD 1997). Nonetheless past groundwater mismanagement has resulted in the degradation of groundwater resources, for example impact of the gold mining has become most evident. In an attempt to redress such past mismanagement the National Water Act (Act 36 of 1998) (GOVERNMENT GAZETTE 1998) and the emerging Integrated Water Resource Management (IWRM) concept have been created, in order to place emphasis on groundwater recharge as one of the key factors in determining the sustainable management of groundwater resources.

LYNCH *et al.* (1994) produced a national-scale groundwater vulnerability map of southern Africa using the DRASTIC methodology but omitted the hydraulic conductivity (CR) parameter in their analyses due to insufficient data availability. Because of the small map scale and the assumptions made in the DRASTIC evaluation process, the produced map shows large areas of equal vulnerability and therefore can't be used for practical applications, which illustrates the need for local scale aquifer vulnerability mapping. SILILO *et al.* (2001) presented a provisional groundwater vulnerability classification system for South Africa based on two criteria. The first of these criteria is the "hydraulic attenuation" which is given a rating from 1 (maximum hydraulic attenuation) to 5 (minimum hydraulic attenuation) based on the physical properties of the soil in the area. The second criterion is the "chemical attenuation", which consists of three ratings for cationic, anionic and organic contaminants, each ranging from 1 (maximum chemical attenuation) to 5 (minimum chemical attenuation) based on the soil type according to the South African soil classification system (SILILO *et al.* 2001). However, the proposed groundwater assessment strategy does not address the unique properties of karst terrains.

A Water Research Commission (WRC) project, "Improved methods for aquifer vulnerability assessments and protocols (AVAP) for producing vulnerability maps, taking into account information on soils" (WRC Project K5/1432), is currently underway in South Africa. However, no karst specific division existed within the AVAP program and as such the need for the development of a local karst aquifer vulnerability mapping method was identified. For this reason the project entitled "Vulnerability mapping in Karst terrains (VUKA) (WRC Project No. K8/669), exemplified in the wider Cradle of Humankind World Heritage Site" was initiated in 2006 and associated to the AVAP research group.

5. The COP method

The COP method, developed by the Hydrogeology Group of the University of Malaga (GHUMA), assumes that contaminant transport depends predominately on the characteristics of the water to move through the aquifer, and that the contaminant infiltrates from the surface by means of rainfall. COP is an acronym derived from the initials of the vulnerability factors considered. The O factor indicates the capability of the unsaturated zone, by means of various processes, to filter out or attenuate contamination and thus reduce its adverse effects. The C and P factors are used as modifiers that correct the degree of protection provided by the overlying layers (O factor). The C factor takes into account the surface conditions that control water flowing towards zones of rapid infiltration and can alter the protection capacity of groundwater described by the O Factor to any degree between nullifying (0) or not altering it (1) at all. The P factor considers the characteristics of the transport agent in the unsaturated zone, i.e. precipitation. Since the influence of precipitation on vulnerability is not as great as that of the flow concentration, the value of the P factor ranges only from 0.4 to 1. While the P and O factors can be used to evaluate the vulnerability of any type of aquifer, the C factor is specific to karst aquifers (VÍAS *et al.* 2003). The final vulnerability indexes are classified into five classes, ranging from "Very Low" to "Very High". The higher classes are assigned mainly depending on the influence of the C factor. The Very Low class refers to zones in which the C and P factors have little influence on protection.

6. Outlook

The Cradle of Humankind World Heritage Site (COHWHS) will act as the study area in which the guidelines for aquifer vulnerability mapping in South African karst terrains will be developed. The semiarid to subtropical climate of South Africa has more common characteristics with that of Mediterranean climate of southern Spain (where the COP method was developed) than with the central European climate of e.g. Germany, (where the PI method was developed) and as such the COP method is more likely to be applicable for South African aquifer vulnerability mapping. The COP method includes a specific factor that takes into account the intensity of rainfall

events in an area and the effects thereof on the aquifer vulnerability. Hence the COP method is again more suitable for the study area's environment, where precipitation is highly seasonal and predominantly occurs as high intensity thunderstorms during the summer months (November to March).

The COP aquifer vulnerability mapping method will thus be used as a starting point for the aquifer vulnerability mapping to be performed in the VUKA project. The COP method has also been selected because it appears to be a more detailed mapping method (and therefore can provide more information to decision makers) in comparison to other European vulnerability mapping methods.

The COP method will however be modified to compensate for the unique properties of South African karst terrains such as the distinctive regional karst of the Malmani Dolomite subgroup of the Chuniespoort Group (Neoproterozoic-Palaeoproterozoic sequence that has been exposed to deep burial, uplifting, tectonization, folding, weathering and erosion cycles) as well as the climatic conditions in semi-arid environments.

Preliminary analyses of available data from weather stations shown in Fig. 2 reveal the precipitation data record to be largely incomplete and not ideal for the proposed analysis. However some important trends are identified. Firstly the highly seasonal nature of the rainfall in the study area is confirmed as most rainfall occurs in the summer months with the winter months being exceptionally dry. Secondly a significant variance exists in the yearly mean rainfall for each of stations considered. This is shown in Table 1 where it can be seen that although the average yearly rainfall for the study area is between 550- 795 mm, single years can have a yearly rainfall of up to 1265 mm or as little as 268 mm.

Table 1. Average yearly rainfall and range for selected climate stations in the COHWHS area.

Station number	Station name	Mean annual rainfall [mm]	Max. annual rainfall [mm]	Min. annual rainfall [mm]	No. of years considered
0512056 7	Seekoeikhoek	659	1059	416	20
0512738 3	Pelindaba	677	1231	392	19
0475661 2	Muldersdrift Summerhill	690	1080	419	33
0512418 4	Maryvale Farm	674	998	513	9
0512090 5	Magaliesburg - Police	643	1037	318	34
0475456 8	Krugersdorp Kroningspark	739	1224	427	44
0512204 3	Hekpoort	657	995	355	22
0512082 1	Hekpoort Nooitgedacht	795	1266	268	32
0512320 9	Bluebird Farm	557	686	385	5
0512087 0	Bekker Hoer Landbouskool	626	823	483	15

While the dry years may pose water shortage problems they are favorable from an aquifer vulnerability point of view whereas the intense, above-average, years will have a significant effect on the aquifer vulnerability of the area. Since most of the rainfall is in the form of intense thunderstorm events the infiltration capacity of the overlying layers will be exceeded and large amounts of surface runoff will occur. This runoff may then be concentrated and enter the aquifer via karst features such as swallow holes and sinkholes.

The infiltration capacity of the overlying layers in the study area need to be determined and based on this, and further investigation on the actual precipitation rates during thunderstorms, the P factor ratings of the COP method will be modified in order to accommodate for the intense precipitation events that occur in the study area.

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Utilisation du signal en Carbone 13 dans le traçage des eaux épikarstiques. Cas de la grotte de Lascaux (Dordogne)

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Résumé

Le paléokarst de la colline de Lascaux est un lieu privilégié pour l'étude des traceurs naturels géochimiques. Ceux-ci sont utilisés pour l'étude du transit des eaux, l'identification des types d'eau aux exutoires des émergences karstiques, ainsi que dans la caractérisation des conditions d'écoulement. Le suivi du Carbone 13 et des équilibres calco-carboniques réalisés en 2003-2004 dans les eaux à l'exutoire du système situé dans la grotte de Lascaux, a apporté des informations nouvelles sur l'origine des écoulements et l'histoire de l'eau dans les formations géologiques.

A partir de la chimie des eaux et du calcul des concentrations des formes du CMTD, il est possible de simuler un $\delta^{13}\text{C}$ théorique correspondant à une stricte évolution de l'eau soit en milieu fermé soit en milieu ouvert sur le CO_2 du sol. Les teneurs en $\delta^{13}\text{C}$ mesurées pendant un cycle dans les eaux de l'épikarst de Lascaux, comparées aux valeurs théoriques, ont permis de préciser les conditions du milieu traversé au cours du transit et d'identifier des mécanismes de dégazage et de précipitation.

Le marquage des eaux en carbone 13 associé à l'étude des équilibres calco-carboniques révèle une évolution en milieu fermé qui permet de préciser le fonctionnement du système épikarstique. Les comblements sablo-limoneux parfois argileux du paléokarst pourrait expliquer les particularités observées dans les eaux.

Mots clef : Lascaux, épikarst, zone d'infiltration, transit, carbone 13.

1. Introduction

L'étude menée sur le site de Lascaux durant le cycle hydrogéologique 2003-2004 s'intègre dans une problématique globale de compréhension du marquage des eaux d'infiltration dans le karst. L'épikarst joue en effet un rôle remarquable dans le fonctionnement de l'aquifère karstique ainsi que dans la mise en place de son réseau de drainage (BAKALOWICZ, 1981). Cette partie du système karstique est le siège d'un ruissellement souterrain mais aussi d'un écoulement retardé, de type matriciel ou venant d'aquifères perchés, constituant « la réserve suspendue » du système (LASTENNET *et al.*, 1995). Elle peut atteindre une épaisseur considérable et jouer un rôle important dans la dynamique et le soutien à l'étiage de certains karst (LASTENNET, 1994 ; EMBLANCH, 1997). Le suivi du chimisme des eaux épikarstiques de Lascaux réalisé durant le même cycle montre de plus que l'essentiel de la minéralisation est acquise en zone insaturée, dans les premiers mètres de l'infiltration.

Les résultats et les interprétations proposées ici concernent le marquage des eaux en carbone 13, et l'utilisation de ce marqueur isotopique comme un indicateur des conditions que rencontrent les eaux lors de leur infiltration à travers l'épikarst.

2. Le site étudié

La colline dans laquelle se développe la grotte de Lascaux est une butte témoin formée de calcaires bioclastiques plus ou moins gréseux, karstifiés puis comblés, datés du Coniacien au Santonien inférieur. Les calcaires marneux du Coniacien inférieur constituent à la fois la base de la colline et le mur de l'aquifère étudié. Même si leur qualité d'imperméable semble médiocre (présence de nombreuses fissures et fractures dans la série), le contraste de perméabilité avec les formations

sus-jacentes est suffisant pour permettre la création d'un aquifère karstique perché. Celui-ci est drainé par des émergences pérennes (sources) parfois diffuses le long de cette discontinuité hydraulique d'altitude 120 m NGF (Nivellement Général Français).

La grotte ornée s'inscrit dans le réseau fossile de ce massif karstique en grande partie comblé par un remplissage sablo-argileux (VOUVE, 1968). Le paléoconduit est inclus dans un promontoire calcaire, entouré par deux paléo canyons (R1 et R2) comblés de sables et d'argiles d'âge encore indéterminé (Tertiaire ?, Fig. 1.). La géomorphologie générale est de type « grand lapiaz ».

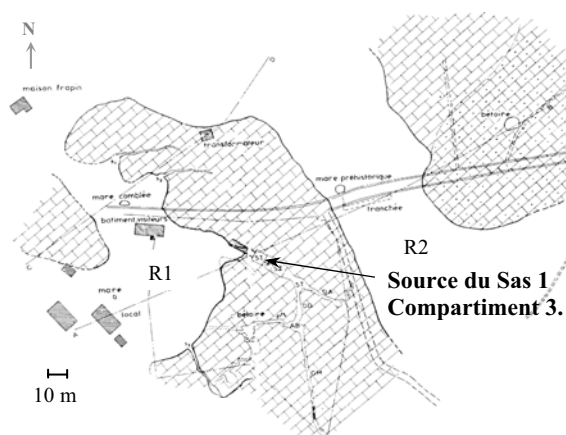


Fig. 1. Carte géomorphologique du site selon VOUVE (1968)

L'émergence située dans la grotte de Lascaux, au toit du Sas 1 compartiment 3 (alt : 185 m NGF, Fig. 1), est l'exutoire d'un système épikarstique. Cet écoulement non pérenne s'établit au droit d'un horizon décimétrique plus marneux dans la série calcaire du Coniacien sommital. Cet

horizon apparaît continu même s'il est actuellement impossible d'exclure la présence de fissures plus ou moins ouvertes qui viendraient l'entailler. La puissance des terrains carbonatés présents au-dessus de cette discontinuité hydraulique évolue entre 5 et 7 mètres. Ces calcaires sont fissurés, mais les fractures semblent peu nombreuses et majoritairement comblées par des formations de remplissage.

3. Matériel et méthode

Parallèlement à l'étude de la dynamique des écoulements et au suivi du chimisme des eaux qui transitent par l'épikarst (LASTENNET *et al.* à paraître), les teneurs en carbone 13 des eaux du Sas 1 - compartiment 3 (Fig. 1.) ont été mesurées au pas bimensuel durant une partie du cycle 2003-2004. Notre connaissance préliminaire du fonctionnement de l'aquifère semble montrer que les vitesses de transferts varient énormément en fonction de l'état de recharge de l'épikarst (LASTENNET *et al.* 1999). Il est probable qu'il en soit de même pour le transit des eaux et des prélèvements supplémentaires à pas de temps rapproché ont donc été effectués lors d'événements pluvieux intenses.

La méthode d'interprétation repose sur l'analyse croisée des teneurs en isotope stable du carbone dissous et des équilibres calco-carboniques. La mise en solution du carbone minéral total dans l'eau (CMTD : H_2CO_3 , HCO_3^- , CO_3^{2-}) résulte de divers phénomènes : la mise en solution du CO_2 du sol ou de l'atmosphère et la dissolution des carbonates constitutifs de la matrice. Chaque espèce dissoute possède un marquage propre en carbone 13 (Tabl. 1.).

Source de Carbone		$\delta^{13}\text{C}$ VPDB	Références
CO ₂ atmosphérique		- 6,4 ‰	FRIEDLI <i>et al.</i> (1986)
CO ₂ biogénique	Plantes C3	- 27 ‰	VOGEL (1993)
	Plantes C4	- 12,5 ‰	VOGEL (1993)
	Plantes CAM	- 27 ‰ jour - 12,5 ‰ nuit	VOGEL (1993)
Sols	Remplissages	- 21,5 ‰	Echantillonnage terrain juin 2003
	Calcaires	- 21,6 ‰	Echantillonnage terrain octobre 1998
Roches	Mondmilch	- 8,5 ‰	
	Carbonates marins du Coniacien	- 0,4 ‰	

Tabl. 1. Marquage isotopique en ^{13}C de quelques réservoirs

La valeur résultante du $\delta^{13}\text{C}_{\text{CMTD}}$ de la solution est exprimée en part pour mille par rapport à un standard (VPDB): un rostre de *Belemnitella americana* de la formation Pee Dee du Crétacé supérieur en Caroline du sud (USA). Elle est régie par les deux pôles d'alimentation en Carbone (CO_2 gazeux et dissolution des carbonates) et par les équilibres chimiques réversibles et thermodépendants du système calco-carbonique. Chaque changement de phase et de forme de carbone minéral dissous entraîne un fractionnement différent qui peut être calculé grâce aux équations présentées dans le Tabl. 2. :

Facteur de fractionnement	Equation
$\varepsilon^{13}\text{C}_{\text{CO}_2(\text{aq})-\text{CO}_2(\text{g})}$	$10^3 \ln \alpha^{13}\text{C}_{\text{CO}_2(\text{aq})-\text{CO}_2(\text{g})} = -0,373 (10^3 \text{ T}^{-1}) + 0,19$
$\varepsilon^{13}\text{C}_{\text{HCO}_3(\text{aq})-\text{CO}_2(\text{g})}$	$10^3 \ln \alpha^{13}\text{C}_{\text{HCO}_3(\text{aq})-\text{CO}_2(\text{g})} = 9,552 (10^3 \text{ T}^{-1}) - 24,10$
$\varepsilon^{13}\text{C}_{\text{CO}_3(\text{aq})-\text{CO}_2(\text{g})}$	$10^3 \ln \alpha^{13}\text{C}_{\text{CO}_3(\text{aq})-\text{CO}_2(\text{g})} = 0,87 (10^6 \text{ T}^{-2}) - 3,4$
$\varepsilon^{13}\text{C}_{\text{CO}_2(\text{g})-\text{CaCO}_3}$	$10^3 \ln \alpha^{13}\text{C}_{\text{CO}_2(\text{g})-\text{CaCO}_3} = -2,9888 (10^6 \text{ T}^{-2}) + 7,6663 (10^3 \text{ T}^{-1}) - 2,4642$

Tabl. 2. Facteurs de fractionnement du $\delta^{13}\text{C}$ pour les différentes formes de carbone en fonction de la température T en °Kelvin.

Avec ε (enrichissement) = $10^3 \ln \alpha$ (coef. de fractionnement)

Et $\varepsilon^{13}\text{C}_{\text{b-a}}$ du composé (a) vers le composé (b).

Le marquage de l'eau correspond au mélange des formes du CMTD et dépend alors directement du pH de la solution (CLARK & FRITZ, 1997).

Par ailleurs, il est nécessaire de faire une distinction entre l'évolution des teneurs en carbone 13 dans les eaux qui circulent en milieu ouvert ou fermé sur le CO_2 gazeux du sol. En système ouvert, même si du carbone est emprunté aux carbonates solides par dissolution, le rééquilibrage en ^{13}C avec l'atmosphère du sol considérée comme infinie et continue, est permanent. La signature isotopique de la solution sera donc imposée par la phase gazeuse. Lorsque la solution entre en système fermé, sa composition isotopique dépend pour moitié du marquage du CO_2 biogénique et pour moitié de celui de la matrice carbonatée. Lors de la dissolution de calcite (à saturation) il y a répartition stœchiométrique entre les bicarbonates HCO_3^- provenant de la mise en solution du CO_2 du sol et ceux de l'attaque acide de la calcite

Connaissant le marquage moyen en ^{13}C du $\text{CO}_2(\text{g})$ des milieux où transitent des eaux ($\delta^{13}\text{C}_{\text{CO}_2(\text{g})} = -21,5\text{‰}$) ; il est possible d'estimer, grâce à l'analyse chimique et aux calculs des concentrations des formes du CMTD, un $\delta^{13}\text{C}$ théorique correspondant à une stricte évolution de l'eau en milieu ouvert sur le CO_2 gazeux, sans dégazage ni précipitation.

$\text{CMTD} = \text{H}_2\text{CO}_3 + \text{HCO}_3^- + \text{CO}_3^{2-}$, d'où :

$$\delta^{13}\text{C}_{\text{CMTD}} = [(\delta^{13}\text{C}_{\text{H}_2\text{CO}_3} * (\text{H}_2\text{CO}_3) / (\text{CMTD})) + [(\delta^{13}\text{C}_{\text{HCO}_3^-} * (\text{HCO}_3^-) / (\text{CMTD})) + [(\delta^{13}\text{C}_{\text{CO}_3^{2-}} * (\text{CO}_3^{2-}) / (\text{CMTD}))]$$

Aux pH rencontrés dans les eaux épikarstiques de Lascaux (pH < 8) la forme CO_3^{2-} du CMTD devient négligeable ; l'équation se simplifie alors :

$$\delta^{13}\text{C}_{\text{CMTD}} = [(\delta^{13}\text{C}_{\text{CO}_2(\text{g})} + \varepsilon_{\text{CO}_2(\text{aq})-\text{CO}_2(\text{g})}) * (\text{H}_2\text{CO}_3) / (\text{CMTD})) + [(\delta^{13}\text{C}_{\text{CO}_2(\text{g})} + \varepsilon_{\text{HCO}_3-\text{CO}_2(\text{g})}) * (\text{HCO}_3^-) / (\text{CMTD}))]$$

avec $\delta^{13}\text{C}_{\text{CMTD}}$: teneur en carbone 13 attendue pour l'aquifère étudié.

$\delta^{13}\text{C}_{\text{CO}_2(\text{g})}$: teneur en carbone 13 du CO_2 équilibrant propre au sol considéré.

ε_{i-j} : facteur de fractionnement entre les espèces i et j

De la même manière, en système fermé, nous pouvons écrire :

$$\delta^{13}\text{C}_{\text{CMTD}} = \delta^{13}\text{C}_{\text{H}_2\text{CO}_3} * [(\text{H}_2\text{CO}_3) / (\text{CMTD})) + \delta^{13}\text{C}_{\text{HCO}_3^-} * [(\text{HCO}_3^-) / 2(\text{CMTD})) + \delta^{13}\text{C}_{\text{CaCO}_3} * [(\text{HCO}_3^-) / 2(\text{CMTD}))]$$

Or, si nous considérons que $\delta^{13}\text{C}_{\text{CaCO}_3}$ est approximativement de 0 ‰, alors :

$$\delta^{13}\text{C}_{\text{CMTD}} = [\delta^{13}\text{C}_{\text{CO}_2(\text{g})} + \varepsilon_{\text{H}_2\text{CO}_3-\text{CO}_2(\text{g})}] * [(\text{H}_2\text{CO}_3) / (\text{CMTD}) + [\delta^{13}\text{C}_{\text{CO}_2(\text{g})} + \varepsilon_{\text{HCO}_3^--\text{CO}_2(\text{g})}] * (\text{HCO}_3^-) / 2(\text{CMTD})]$$

Enfin, l'évolution du marquage en ^{13}C est le résultat de certains états transitoires que rencontre l'eau au cours de son transit dans le système :

Lorsqu'une eau transite d'un milieu à forte pCO_2 vers un milieu moins carboné, des dégazages permettent une remise à l'équilibre du CMTD de la solution avec son nouvel environnement. La solution fille, sursaturée, s'enrichit en carbone 13 lourd par la perte des formes H_2CO_3 très négatives. Cette remarque prend de l'importance car, quand une eau ayant évolué en milieu fermé dégaze, le $\delta^{13}\text{C}_{\text{CMTD}}$ mesuré de la solution peut être plus enrichie que le $\delta^{13}\text{C}_{\text{CMTD}}$ théorique de système fermé ; les deux phénomènes d'enrichissement se superposant.

Le rééquilibrage d'une eau sursaturée en calcite entraîne des précipitations de carbonates enrichis en ^{13}C par rapport aux formes dissoutes. L'estimation de ce fractionnement isotopique est difficile car les effets cinétiques dus à l'état transitoire dans lequel se trouve l'eau sont mal connus aujourd'hui. Il dépend à la fois des vitesses et des variations du taux de précipitation. L'appauvrissement de la solution fille peut tout de même être estimé de -1 à -2 ‰ (BATIOU, 2002). Comme précédemment, une eau ayant évolué dans un milieu ouvert sur la pCO_2 du sol et qui serait amenée à précipiter pourrait voir son $\delta^{13}\text{C}_{\text{CMTD}}$ mesuré plus appauvrie que celui calculé en système ouvert.

Ces remarques montrent déjà l'importance de l'interprétation croisée des différents paramètres hydrogéochimiques mesurés ou calculés, notamment la comparaison des données en $\delta^{13}\text{C}_{\text{CMTD}}$ avec l'indice de saturation vis à vis de la calcite (IS_c), marqueur de l'état d'équilibre des eaux au moment du prélèvement. Nous rappellerons à cet effet la formulation de l' IS_c tirée de la réaction de dissolution / précipitation de la calcite :

$$\text{IS}_c = \log [\text{Ca}^{2+}] + \log (\text{K}_2) + \log [\text{HCO}_3^-] - \log \text{K}_c + \text{pH}_{\text{mes}} \approx \text{dpH} = \text{pH}_{\text{mes}} - \text{pH}_{\text{sat}}$$

4. Résultats

La Fig. 2.b montre l'évolution du marquage en carbone 13 des eaux du Sas1 au cours du cycle 2003-2004. Contrairement aux éléments majeurs qui restent assez stables durant cette période, le $\delta^{13}\text{C}$ montre des variations très significatives de l'ordre de 2,3 ‰. Les eaux sont enrichies en ^{13}C en début de cycle ($\delta^{13}\text{C} = -12,5$ ‰ le 1-déc-03), et évoluent ensuite vers un appauvrissement continu pour atteindre -14,8 ‰ le 25-mars-04. Les crues C1 et C2 (Fig. 2.a) semblent favoriser la vitesse d'appauvrissement des eaux en ^{13}C mais cet effet paraît secondaire par rapport à l'évolution générale du signal. Les évolutions théoriques des $\delta^{13}\text{C}$ calculées en système ouvert ou fermé sur le CO_2 sont représentées sur la Fig. 2.b.

EMBLANCH *et al.* (1998) ont constaté une nette corrélation positive entre les évolutions en $\delta^{13}\text{C}$ et en Mg^{2+} utilisé dans leur étude comme marqueur du temps de séjour. Ils

ont montré que le ^{13}C était un traceur sensible du temps de séjour de l'eau dans la zone non saturée des aquifères épikarstiques. Or, le système épikarstique de Lascaux ne répond pas de la même manière que ceux du Vaucluse : les valeurs les plus appauvries en $\delta^{13}\text{C}$ (fin mars 2004) correspondant aux concentrations en magnésium les plus élevées (Fig. 2.b). Nous interpréterons alors le signal en ^{13}C non pas comme un marqueur du temps de séjour mais comme un indicateur des conditions physico-chimiques qu'a rencontrées l'eau au cours de son transit dans la zone non saturée.

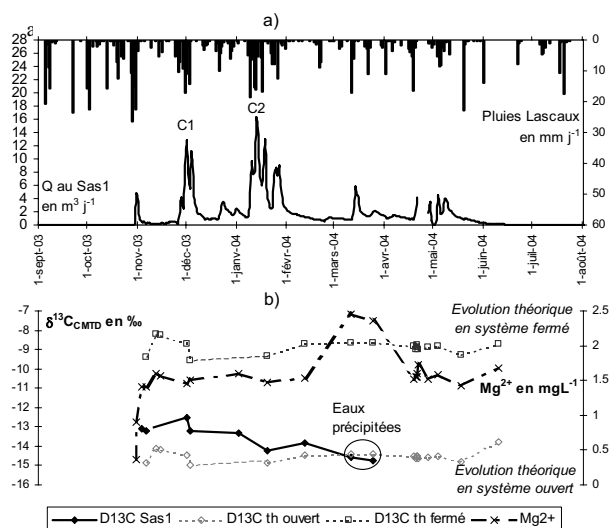


Fig. 2. a) Hydrogramme au Sas1 pour le cycle 2003-2004
b) Evolution des teneurs en ^{13}C (‰) et en Mg^{2+} (mgL^{-1}) des eaux du Sas 1.

Les eaux de novembre 2003, les plus enrichies du cycle, subissent l'influence du marquage en ^{13}C de la matrice carbonatée. Ces eaux issues de la recharge automnale et stockées dans un milieu capacitif (fissures colmatées, calcaires de sub-surface délités) sont restées au contact de la matrice en milieu relativement confiné. Leur mise en mouvement est effective lorsque le profil hydrique du sol atteint des teneurs en eau suffisantes pour un déplacement gravitaire. On observe alors un effet piston dans la zone non saturée drainant ces eaux plus profondément vers l'émergence. La connectivité hydraulique étant réalisée, les eaux suivantes transitent plus rapidement dans le système en suivant des axes de drainage privilégiés et aérés. Ces eaux, marquées par l'atmosphère du sol, sont appauvries en ^{13}C . A la fin du cycle, les flux se ralentissent pour devenir très faibles. Nous supposons alors des précipitations possible dans la zone de drainage aérée expliquant ainsi des teneur en ^{13}C plus appauvries que le calcul théorique en système ouvert. Les eaux collectées à l'émergence du Sas1 sont donc probablement le résultat d'un mélange d'eaux stockées dans des zones capacitives confinées, prépondérantes en début de cycle, et d'eaux à circulation rapide dans les zones transmissives et aérées du système.

De plus, comme le signal $\delta^{13}\text{C}$ dépasse les limites théoriques calculées pour une évolution de l'eau en système ouvert, d'autres phénomènes sont susceptibles d'affecter ce signal. L'interprétation du marquage des eaux du Sas en ^{13}C ne peut alors se faire qu'en croisant les informations afférentes aux conditions d'écoulement dans le système.

La Fig. 3 montre ainsi l'évolution du $\delta^{13}\text{C}$ des eaux du Sas1 en fonction de leur état de saturation vis à vis de la calcite. Globalement, en dehors des crues C1 et C2, les eaux enrichies en ^{13}C sont celles qui sont le moins sursaturées vis à vis de la calcite. Ceci confirme l'hypothèse de l'apport à l'écoulement d'eaux ayant évoluées en milieu confiné où la matrice carbonatée imprime son marquage isotopique à l'eau qui ne peut dégazer. A l'inverse, les eaux appauvries en isotopes lourds sont nettement sursaturées vis-à-vis de la calcite. Ce sont des eaux marquées isotopiquement par l'atmosphère du sol et dont le dégazage est possible.

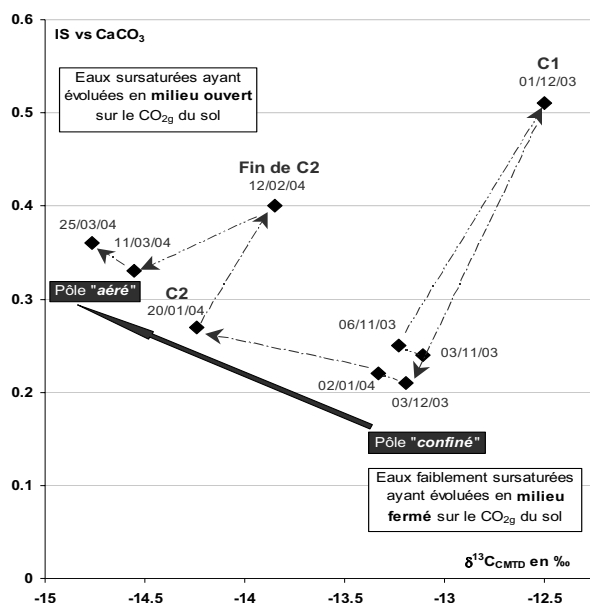


Fig. 3. Evolution du marquage en ^{13}C des eaux du Sas 1 en fonction de leur état de saturation vis-à-vis de la calcite.

Enfin, un phénomène particulier se réalise à deux reprises pendant le cycle étudié :

Durant la crue C1, intervenant en début de la période de recharge, les eaux montrent une nette augmentation de leur saturation corrélée avec un enrichissement de la solution en carbone 13. Ceci peut s'expliquer par un changement des conditions d'écoulement du à la mise en charge du système provoquant un dégazage des eaux en milieu diphasique. Après la crue, on retrouve le marquage initial : eaux appauvries en ^{13}C et moins sursaturées vis à vis de la calcite. Les eaux sont alors de nouveau alignées sur la droite fictive $\text{ISc} = f(\delta^{13}\text{C})$ de mélange du pôle «confiné» vers le pôle «aéré».

Ce phénomène se reproduit mais après l'épisode de crue C2. Un état de la réserve différent du fait de la recharge hivernale peut expliquer ce décalage. C'est en effet au cours de la décrue (février) survenant après la période de hautes eaux que le mécanisme de dégazage se produit (vidange des fractures ouvertes ?), d'où cette eau plus enrichie en ^{13}C et fortement sursaturée.

Le signal continue à s'appauvrir pendant le drainage de la réserve au printemps du fait de l'aération croissante du milieu qui favorise le dégazage des eaux et même leur précipitation.

Ces mécanismes ne sont pas révélés par le suivi de la chimie des majeurs. Le marquage en ^{13}C constitue donc un complément intéressant pour affiner la caractérisation des conditions de transit des eaux épikarstiques.

5. Conclusion

Les données en $\delta^{13}\text{C}$ récoltées durant le cycle d'étude semblent caractériser les conditions physico-chimiques des milieux que rencontrent les eaux au cours de leur transit dans l'hydrosystème. Elles confirment les différentes origines possibles des eaux du Sas et la tendance générale des écoulements à être de plus en plus marqués par une évolution en système ouvert.

De plus, l'augmentation de pression pendant les périodes de crues perturberait les conditions d'écoulement des eaux dans l'épikarst. Elles subiraient des phénomènes de dégazage mis en évidence par l'évolution du signal $\text{ISc} = f(\delta^{13}\text{C})$. Nous pouvons penser que pendant ces événements, les eaux sont « chassées » de milieux capacitifs et confinés vers des secteurs plus transmissifs et aérés des calcaires. Le transit des eaux dans des milieux aux conditions physico-chimiques internes contrastés, peut donc être caractérisé de manière pertinente par l'utilisation du marquage en carbone 13.

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Les mises en charge du réseau de la Luire (Vercors, France) : enregistrements et implications géomorphologiques

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Abstract

Flood of the Luire karst system (Vercors, France): records and geomorphological implications

Flood in epiphreatic zone of the mountain karsts generates intense erosion in relation with the flood frequency and the flood process. The Luire network (40 km of development), in the south of Vercors knows elevation of groundwater level more than 480 m height, most important known in the world. The autonomous recording heights of water with regular step (luiuographe) in several points of the network, morphological observations of some sectors and the analysis of the geometry of the system make it possible to show the role of flood events in the organization of the conduits and the role of the network in the structuring of the flood hydrograph (discharge of the subsystems and “crevaisons” of the temporary spring). This temporary spring of the system would be a paleo well recut by the lowering of topographic surface. The temporary discharge of water (called “crevaisons” here) generates a considerable acceleration of flow and erosion. The geomorphological analysis of the network and the fine recordings water level variation show the space and temporal disparity of the erosion in the epiphreatic zone of this karst.

Résumé

La zone épinoyée des karsts de montagne est le siège d’une morphogénèse intense liée autant à la fréquence des crues qu’aux conditions de déroulement de ces crues. Le réseau de la Luire (40 km de développement), dans le sud du Vercors connaît des mises en charges de plus de 480 m de hauteur, les plus importantes connues au monde. L’enregistrement autonome des hauteurs d’eau à pas régulier (luiuographe) en plusieurs points du réseau, le relevé morphologique de quelques secteurs clefs et l’analyse de la géométrie du système permettent de montrer ici le rôle des crues dans l’organisation des conduits (étagement des drains) et à l’inverse, le rôle de ces conduits dans la structuration de l’onde de crue (déversement des sous-systèmes et crevaisons de l’exutoire). Cet exutoire temporaire du système serait une ancienne cheminée d’équilibre recoupée par l’abaissement de la surface topographique. Le déversement temporaire des eaux (appelé ici les crevaisons) conduit à une accélération considérable des vitesses d’écoulement et génère une érosion décuplée. L’analyse géomorphologique du réseau et les enregistrements fins des hauteurs d’eau montrent ainsi la disparité spatiale et temporelle de la morphogénèse dans la zone épinoyée de ce karst de montagne.

1. Introduction

Les mises en charge dans la zone épinoyée des karsts de montagne ont un rôle spéléogénétique indéniable lié à la récurrence des crues et au processus liés à ces crues (variabilité chimique, détente mécanique post ennoiment...) (JAILLET, 1999). Le réseau karstique de la Luire (GARNIER, 1997) (Fig. 1) dans le sud du Vercors (France) connaît des mises en charge sur plus de 450 m de dénivellation ce qui constitue le record mondial connu à ce jour (AUDRA, 1997). Il s’agit d’un ensemble de drains de plus de 40 km de développement de forme plus ou moins circulaire, horizontaux, pentés ou verticaux. L’ensemble de ces drains s’ennoe jusqu’à l’exutoire temporaire du système : la Luire. On parle alors de « crevaisons » de la Luire.

Le propos du présent article est de montrer en quoi la spéléogénèse des conduits endokarstiques est liée à ces mises en charge (échelle des drains) et en quoi, les crevaisons de la Luire implique une morphogénèse accrue dans la zone épinoyée (échelle du système karstique). Pour cela des travaux de terrains croisant une approche hydrométrique et des observations morphologiques ont été mis en place.

2. Matériels et méthodes

L’analyse hydrologique est réalisée à partir de l’enregistrement des hauteurs d’eau dans le réseau en utilisant

un Luiuographe. Cet appareil a été conçu spécifiquement pour ce réseau (MOREL, 1996) et équipe aujourd’hui plus d’une dizaine de grands réseaux karstiques français. Il permet de mieux connaître et de comprendre le fonctionnement hydrologique des cavités. L’étude est menée sur plusieurs années mais seuls quelques épisodes sont présentés ici (Fig. 2). On peut alors connaître la fréquence des crues et la hauteur d’ennoeiment, les temps de réaction, la vitesse de l’onde de crues dans la cavité, les temps de montée et de descente du niveau de l’eau.

Depuis 1995, le Luiuographe fonctionne de manière entièrement autonome, sur plusieurs années. Il regroupe les capteurs, l’enregistreur et l’alimentation dans un boîtier étanche. Les centrales sont placées dans plusieurs endroits stratégiques du réseau (Fig. 1) et relevée régulièrement (généralement une fois l’an).

Parallèlement à ces analyses hydrologiques, des observations géomorphologiques ont été menés dans le réseau souterrain. Il s’agit essentiellement de relevés parietaux permettant d’identifier, à partir des coups de gouges (CURL, 1966, 1974) les sens et vitesses de l’écoulement en période de crue. Une telle approche permet de proposer à l’échelle d’un drain une vitesse et un débit pour la pointe de crue. On considère en effet que pour chaque conduit existe un débit borné de crue (lié aux contraintes géomorphologiques du site) et que c’est ce (ou ces) débit (s) connaissant une fréquence de retour

importante qui sont responsables de l'édification des coups de gouges.

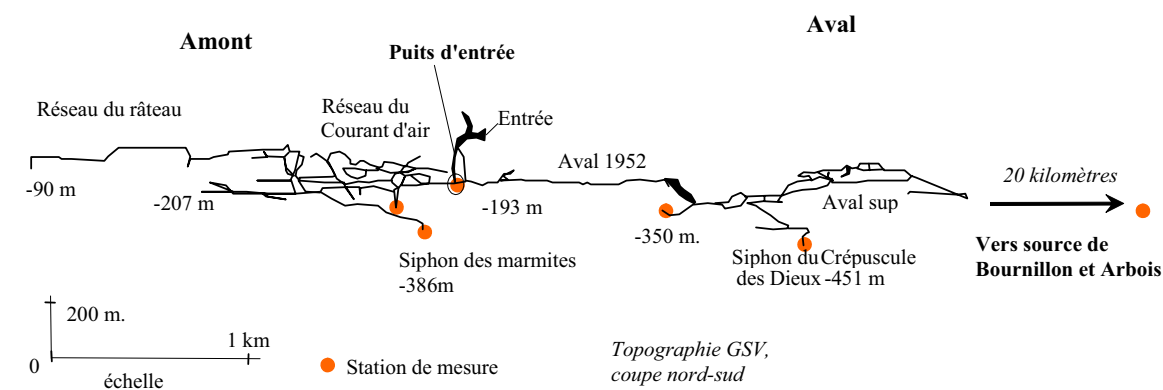


Figure 1 : Coupe schématique Nord/Sud et localisation des stations de mesures dans le réseau.

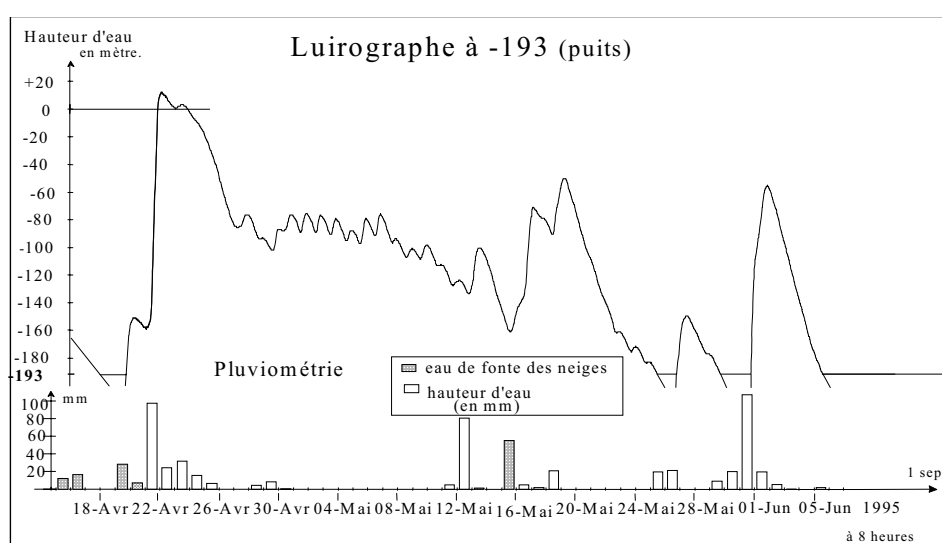


Figure 2 : Enregistrement des hauteurs d'eau dans le réseau (station luirographe à -193 m). Période Avril – Juin 1995.

Le réseau de la Luire fait partie du vaste système de Bournillon (227 km²). Il présente un étagement de conduits avec des niveaux identifiés à -450 et -300 (actif pérenne) et à -200 (conduites forcées temporaires). Trois sites ont fait l'objet d'investigation particulière. Ils ont été choisis en fonction de cet étagement des conduits dans le réseau. Il s'agit du fond du puits du crève-cœur à -300 m. C'est un regard sur l'écoulement pérenne du système. Plus haut, un site dans les conduits forcés vers -200 m a été relevé et quelques observations complémentaires ont été réalisées sur la zone d'entrée.

puis, depuis 2 ans, à -450 au dessus du crépuscule des Dieux. Au total 5 sites ont été étudiés dans la Luire (Fig. 1).

Vitesse de montée de l'eau :

Pendant l'année 2004-2005 la vitesse de montée de l'eau atteint 100m/h lors de la crue du 11 aout 2004. Plusieurs crues par an dépassent les 60m/h (Fig. 3). De manière générale, les crues sont rapides à la fin du mois d'aout, septembre et à l'automne. En hiver le manteau neigeux ralentit fortement les vitesses de montées. Les vitesses de montées aux marmites et à -450 ne sont pas identiques sauf à partir de -193 dans la zone des puits. La zone de -450 à -193 est noyée 14% du temps.

3. Résultats

Fréquence des crues :

Un Luirographe enregistre depuis 1995 les variations de hauteur d'eau dans les puits. Il est situé à -193 m (Fig. 2). Depuis 1999, un autre Luirographe enregistre les hauteurs d'eau dans la galerie des marmites à -275, où un écoulement temporaire (plusieurs mois par an) est observé lors de pluies modérées (Fig. 1). On observe une inversion du sens l'écoulement, par une mise en charge lors de fortes précipitations. La résurgence de Bournillon est équipée d'un Luirographe afin de connaître son débit. En 2000 un autre a été posé dans le crève cœur et en 2001 à -350, dans l'aval,

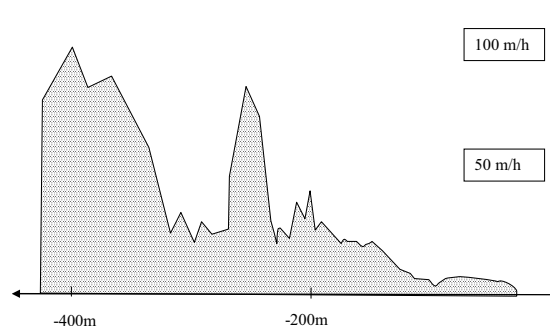


Figure 3 : Vitesse de montée maxi enregistrée en fonction de la profondeur.

On peut observer des montées en moyenne de plus de 40m/h c'est-à-dire que la Luire peut se remplir à partir de -450 en une douzaine d'heures.

Le puits du Crève-cœur :

Le fond du puits du Crève-cœur à -300 m est un éboulis donnant accès à une rivière souterraine (Fig. 4) pérenne dont il a été montré par traçage qu'il s'agit du « cours actif ». Dans ce tronçon pérenne, on note la présence de coups de gouges de petites dimensions (2 à 4 cm) dans la partie chenalisée du conduit (dimension moyenne l: 1m x h: 0,5m). Ce surcreusement correspondant à de petites crues (écoulement à surface libre). La vitesse déduite des coups de gouges (1 m/s) implique un débit de l'ordre de 400 à 600 l/s pour ce chenal. Au plafond du même conduit, des vagues d'érosion moins marquées indiquent un sens d'écoulement remontant vers le puits du Crève-cœur. La vitesse déduite serait de l'ordre de 10 à 15 cm/s pour une section mouillée totale de l'ordre de 3 à 4 m². Le débit (cette fois ascendant) serait donc de l'ordre de 300 à 600 l/s.

Il est intéressant de constater que dans un sens ou dans l'autre on retrouve des débits maximums similaires pour ce conduit actif se mettant en charge et connaissant des inversions de sens du courant. Une telle observation milite pour la présence d'un diaphragme à l'aval du site de mesure qui borne le débit précisément autour de 500 l/s dans un sens ou dans l'autre.

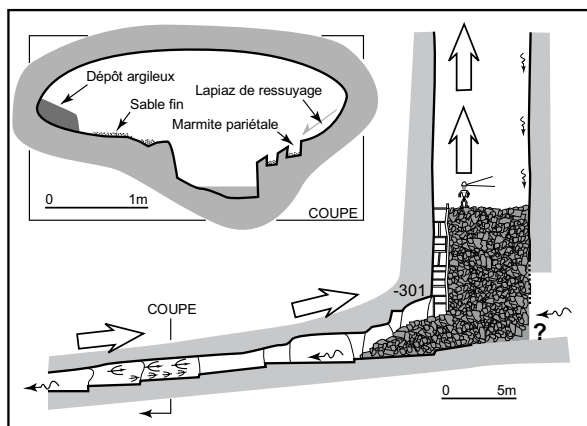


Figure 4 : A la base du puits du crève-cœur (- 300 m), on note une inversion des sens de courant selon l'importance de la crue. Dans un sens comme dans l'autre, le débit est borné à 400 l/s ce qui montre l'existence d'un diaphragme limitant précisément l'écoulement des eaux dans ce secteur de la cavité.

Les conduites forcées à - 200 m :

Le même phénomène est observé dans le secteur des conduites forcées vers -200m. Cette zone (« l'aval 52 ») offre des conduits de dimensions plus importantes (jusque 10 ou 20 m² de section). Dans ce secteur, une conduite forcée de plus d'une trentaine de mètres de longueur présente des coupes de gouges petits dans un chenal surcreusé indiquant un sens d'écoulement sud – nord (vers l'aval). Le débit a pu être évalué supérieur à 1 m³/s. Au plafond on retrouve des coups de gouges moins marqués mais indiquant un écoulement nord – sud (vers l'amont) avec des vitesses moindres, mais une section totalement noyée. Ces observations montrent bien que c'est bien l'aval du système qui envoie la totalité des drains de la Luire et non uniquement l'amont du système (alimentation par les hauts-plateaux) qui sature en crue et provoque les mises en charge.

Cette observation a été faite par la mesure simultanée de deux Luitrographe, un au crève cœur et l'autre à l'aval à -350. Cet

écoulement de l'amont vers l'aval se produit une dizaine de jours par an en moyenne.

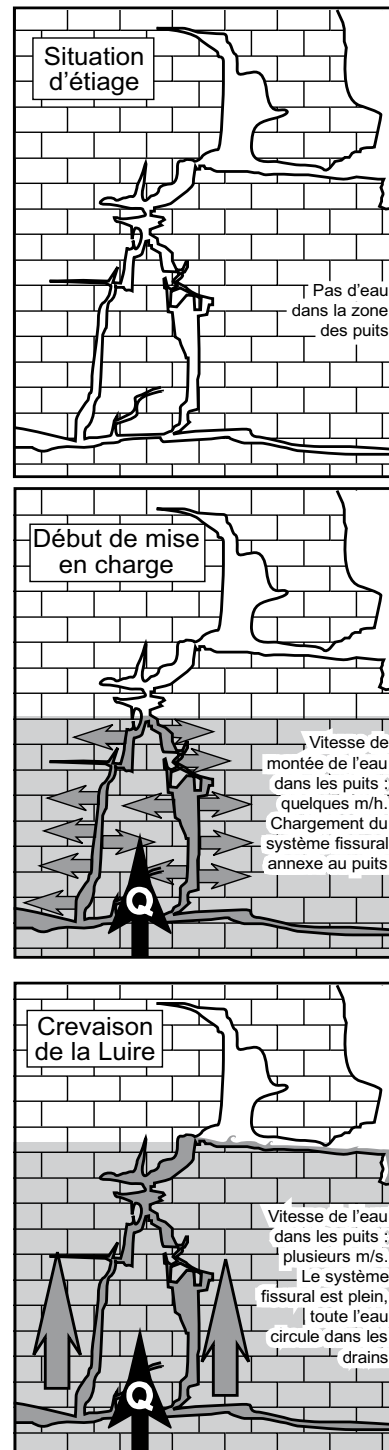


Figure 5 : Mise en charge dans la zone des puits entre 0 et - 200 m. Noter les variations des vitesses de l'eau dans les puits au cours de la mise en charge et au moment du débordement. Avec un même débit, la crevaision génère une accélération considérable des vitesses de circulation dans les puits impliquant une morphogenèse décuplée.

Le fonctionnement dans la zone des puits :

La remontée des eaux dans la zone des puits (200 m de dénivellation) est un exemple particulièrement éloquent de mise en charge d'une zone épinoyée associée à un déversement temporaire de l'écoulement. En effet, la zone des puits n'est pas une zone de transit vertical du karst (zone non

saturée). On ne note pas d'écoulement à l'étiage. En début de mise en charge (Fig. 5), avec un débit nominal Q les puits se remplissent comme le système fissural annexe à ces puits. La vitesse d'écoulement de l'eau est minimale (quelques dizaines de mètres par heures). Parvenu au seuil de déversement, le trop plein se déverse à l'extérieur, la Luire crève (Fig. 5). Cette fois le système fissural est plein et la totalité du débit nominal (plusieurs dizaines de m³/s) passe par les puits. Ceux-

ci présentent des sections minimales pouvant être de 2 à 3 m² impliquant des vitesses de circulation de l'ordre d'une dizaine de mètres par seconde. Ces vitesses considérables génèrent une accélération exceptionnelle de la morphogénèse. A ces vitesses, aucune cupule ne se met en place. Des galets coincés dans des fissures accréditent ces vitesses considérables mais ponctuelles dans l'espace de la cavité.

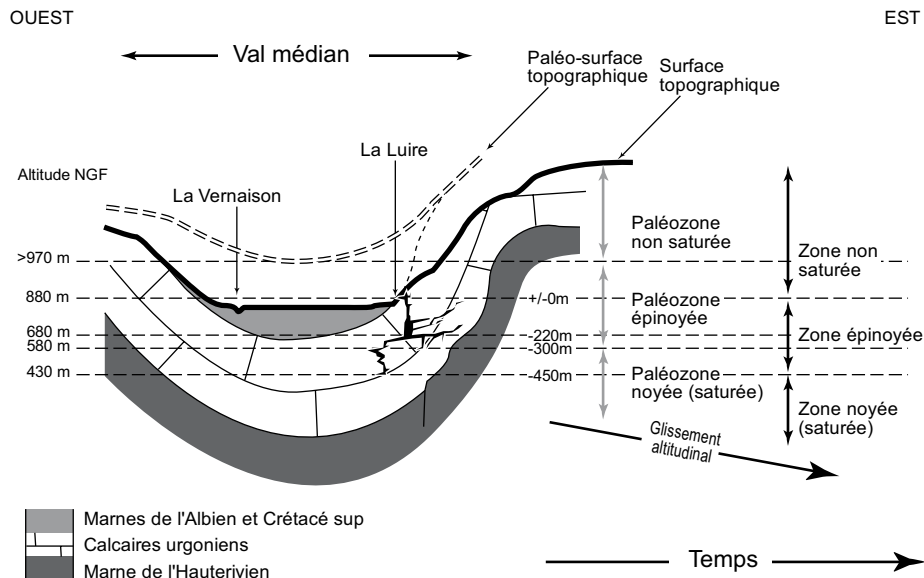


Figure 6 : Coupe synthétique Est – Ouest du synclinal médian présentant le réseau de la Luire dans son contexte morpho-structural. Noter l'étagement des zones noyées, épinoyées et non saturées du système karstique et le glissement de ces zones associé à l'abaissement du niveau de base et à l'incision de la vallée de la Vernaison. Coupe d'après Delannoy, 1997, complétée.

4. Discussion et Conclusion

Si une telle morphogénèse est possible dans le réseau de la Luire, c'est bien parce que les crevaisons ont lieu de temps en temps. En effet, dans ces périodes brèves, l'érosion des drains (verticaux et horizontaux) est maximale car les vitesses de circulation dans les conduits sont décuplées.

Or, il existait une période où la vallée de la Vernaison (cours d'eau dans lequel se jette la Luire) devait se situer plus haut (Fig. 6). En effet, le fond imperméable du val médian, structure synclinale étroite, subméridienne du Vercors, draine des écoulements de surface vers les gorges des Grands Goulets. Des travaux géomorphologiques menés sur les gorges de la Bourne et des Grands Goulets montrent que l'incision de la Vernaison semble tardive par rapport à celle de la Bourne.

Les observations présentées ici accréditent ce fait. En effet, alors que la vallée de la Vernaison était moins incisée, les mises en charges de la Luire ne devaient pas générer de déversement des eaux. La vitesse de circulation de l'eau restait modeste. Le réseau de la Luire se comportait comme une vaste cheminée d'équilibre du système Bournillon. Aujourd'hui, l'incision de la Vernaison autorise les crevaisons de la Luire et génère une érosion décuplée dans le réseau souterrain.

Parallèlement à cette érosion accrue du réseau souterrain, c'est toutes les zones noyées, épinoyées et non saturées du système karstique qui migrent vers le bas d'une centaine de mètres environ. Les volumes souterrains situés au dessus de la grotte de la Luire (salle Decombaz) restent les témoins d'une zone épinoyée aujourd'hui déconnectées par l'incision de la Vernaison. Cette dernière bien que possédant un bassin

modeste et se situant en contre-haut, en prélevant une partie des écoulements dévolus à la Bourne, capture une partie du bassin de cette dernière (DELANNOY, 1997).

On l'aura compris, l'analyse croisée de données hydrométriques (ici des fréquences de mises en charges) et des observations morphologiques sous terre permettent, dans un cadre géomorphologique connu, de préciser les modalités de la spéléogénèse de la zone épinoyée à l'échelle du drain comme à l'échelle du système tout entier.

Références

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Easytopo : appareil de topographie portatif.

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Abstract

La mesure de distance d'inclinaison et d'azimut, permet de replacer un point dans un espace, ainsi de reconstituer la cartographie 3D de n'importe quel lieu. Un appareil automatique portatif, nommé Easytopo, vient d'être mis au point. En un clic, il relève avec précision la position de points, les affiche, les mémorise et les exporte vers un ordinateur ou un PDA grâce à la technologie Bluetooth. L'acquisition se fait alors en quelques secondes et permet d'envisager de nouvelles méthodes de topographie.

1. Introduction

Soutenu par la Région Rhône Alpes via son l'incubation au sein de Créalys (Incubateur Rhône Alpes Ouest), le projet Easytopo est avant tout un projet de création d'entreprise innovante. Ce projet vise au développement d'un produit portatif et simple d'utilisation, pouvant être utilisé dans des environnements difficiles (atmosphère humide, terrain accidenté...) et ouvert en terme de fonctionnalités. Un premier prototype a été mis au point avec le concours scientifique du Centre de Génie Electrique de Lyon (Cegely).

Le Cegely a pour objectif la maîtrise de l'énergie électrique. Ses domaines de compétence concernent les grands domaines du Génie Electrique et de l'électromagnétisme : la compatibilité électromagnétique (CEM), la haute tension, le contrôle-commande de systèmes, l'électronique de puissance, les matériaux, ainsi que la modélisation des phénomènes électromagnétiques. Les compétences de ce laboratoire de l'Université Claude Bernard Lyon 1, dans les domaines des champs magnétiques, ont permis de développer principalement la mesure d'azimut.

EasyTopo est avant tout la réponse à un besoin, par sa capacité à intégrer les trois instruments de mesure dans un système électronique automatisé, permettant, en une manipulation, de relever les valeurs, les afficher, les mémoriser et les exporter sous différents formats vers ordinateurs et assistants personnels.



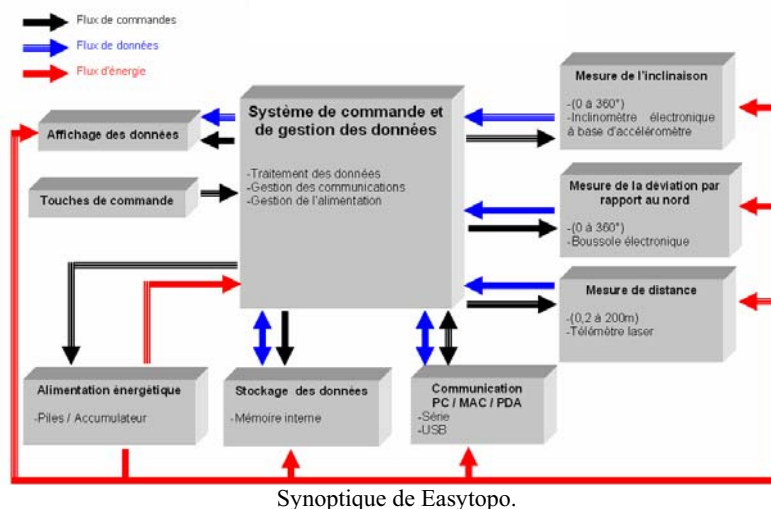
Prototype

Technologie employée :

- Laser (distance)
- Accéléromètre 3D (inclinaison)
- Magnéto inductive 3D (azimut)

Précision attendue :

±3 mm pour la distance
0.3° pour les angles.
Temps pour une mesure <1s

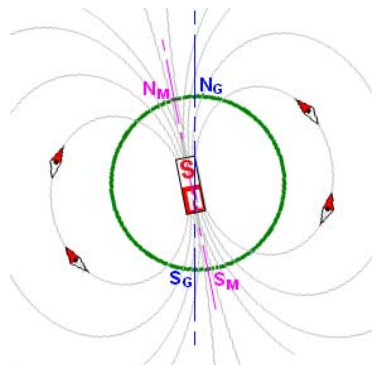


Synoptique de Easytopo.

2. Champ magnétique

La Terre possède un champ magnétique produit par les déplacements de son noyau - composé essentiellement de fer et de nickel en fusion conducteurs - qui se comporte comme une gigantesque dynamo.

Le champ magnétique peut être comparé, en première approximation, à un aimant droit (dipôle) ou à une bobine parcourue par un courant.



Ligne de champ magnétique terrestre

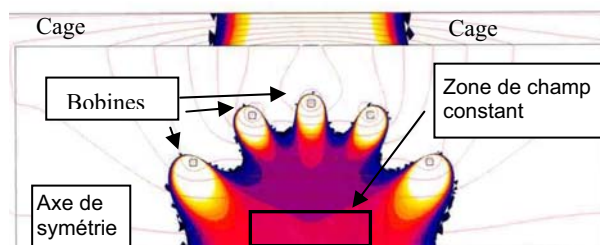
En un point donné du champ magnétique terrestre, le vecteur d'induction magnétique B possède une composante verticale B_v (dirigée vers le centre de la Terre) et une composante horizontale B_o . Aux pôles magnétiques la composante horizontale a une valeur nulle, à l'équateur les lignes de champ sont parallèles à la surface du globe. L'angle formé par B et B_o est appelé angle de pénétration. Il augmente lorsque l'on se rapproche des pôles en tendant vers 90° . La valeur de l'induction magnétique est exprimée en tesla. Actuellement, elle est de l'ordre de $47 \mu\text{T}$ (10^{-6} T) au centre de la France. Pour se repérer on utilise la composante horizontale B_o .

Le champ magnétique est très faible $47\text{e-}6\text{T}$, un aimant même de qualité médiocre peut fournir un champ autour de 1T (dans un environnement proche) soit vingt mille fois plus ! D'autre part la moindre structure métallique, ligne électrique va perturber la distribution des lignes de champ magnétique. De plus notre environnement est baigné par un champ électromagnétique variable sur une bande de fréquence allant du hertz à quelques gigahertz. Mesurer le champ magnétique terrestre avec précision dans un environnement perturbé est délicat.

Pour cette raison nous avons réalisé une chambre amagnétique permettant de s'affranchir de toute perturbation électromagnétique et magnétique. Le champ magnétique résiduel dans la cage est de l'ordre de la centaine de nanotesla, bien inférieur aux grandeurs mesurées. La cage se compose de trois couches de mumétal, alliage de fer et de nickel. Ce matériau à très forte perméabilité magnétique vient canaliser les lignes de champ à sa surface. A l'intérieur le champ est donc quasi nul et une bobine recrée un champ magnétique homogène de la valeur du champ terrestre. Une étude en simulation en éléments finis (Flux 2D) a permis de trouver le courant nécessaire traversant la bobine et d'optimiser la forme de la bobine pour obtenir une zone homogène où le champ est constant en amplitude et en direction. Une plateforme amagnétique actionnée de l'extérieur permet de faire tourner les capteurs dans les trois dimensions.



Cage amagnétique Ø0.7m L1m.



Simulation en éléments finis de la cage amagnétique.

3. Azimut.

Les magnétomètres permettent de mesurer en amplitude et en direction le champ magnétique, les plus performants sont capables de mesurer le nanotesla, soit environ 50.000 fois le champ terrestre et encore mille fois moins pour les appareils très spécifiques. Ces appareils coûtent cher et sont destinés principalement aux laboratoires. Actuellement plusieurs entreprises se

lancent dans la fabrication de capteur magnétique à faible coût et surtout à grandes précisions. Dans cette gamme on trouve principalement trois technologies de capteurs, magnétorésistif, fluxgate et magnétoinductif.

Cependant il n'existe pas de capteurs permettant la mesure directe de l'azimut. De plus l'appareil peut être positionné dans n'importe quelle direction. Il suffit simplement de viser avec le laser le point à mesurer, ce qui fait un point fort de l'appareil. On utilise alors 3 capteurs montés à 90° formant un trièdre le plus parfait possible, couplé aux capteurs d'inclinaison.

4. Inclinaison.

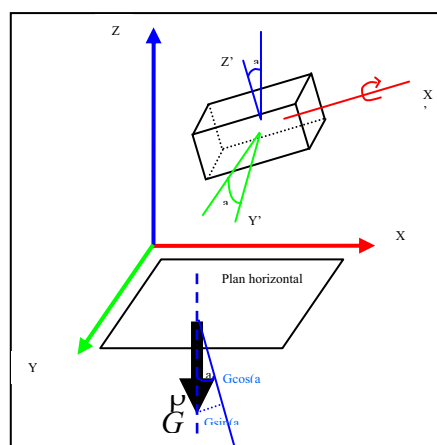
Caractériser l'inclinaison nécessite aussi une mesure en trois dimensions, mais de l'accélération de pesanteur. Plusieurs technologies existent, les plus courantes sont basées sur la technologie piézoélectrique, capteur capacitif et CMOS MEMS.

Les accéléromètres MEMS, micro-usinés, disposent d'une bonne précision pour un coût réduit. Leur grande sensibilité et la précision qu'ils offrent les rendent aujourd'hui incontournable, le tout bénéficiant de tous les avantages (fiabilité, reproductibilité...) de la forte intégration des fonctions sensibles dans un circuit imprimé.

Les mouvements ou inclinaisons sur trois axes entraînent de légers déplacements de structures de silicium mobiles dans le capteur MEMS, modifiant la capacité entre pièces fixes et pièces mobiles. Le circuit d'interface, intégré dans le même boîtier, traduit ces minuscules changements de capacité en tensions analogiques calibrées proportionnelles au mouvement. Le circuit d'interface est " bridé " en usine afin d'assurer des performances répétables sans ajustement du produit.

5. Matrice de rotation.

Easytopo peut être placé dans n'importe quelle direction, n'importe quelle inclinaison. On peut également faire tourner le module sur lui-même. Pour pouvoir afficher la valeur de l'azimut dans n'importe quelle position, il va falloir prendre en compte les valeurs des trois accéléromètres. Ceci est réalisé par une double matrice de rotation. Cependant une grande maîtrise des capteurs est nécessaire, même une petite erreur sur un des capteurs peut être vite répercutée sur les valeurs finales.



Repère du module dans un repère fixe.

6. Auriga



Auriga sur palm, fenêtre d'édition à gauche, fenêtre graphique à droite

Easytopo envoie les données par la liaison bluetooth ou par une liaison sériel RS232. Tout ordinateur équipé d'une de ces technologies permet de recevoir les trois grandeurs.

Auriga, développé par Luc Le Blanc, est un logiciel de topographie souterraine tournant sous Palm OS et destiné à être utilisé sous terre comme carnet topo intelligent (<http://www.speleo.qc.ca/auriga>). Au fil des mesures, Auriga affiche le

cheminement de la grotte sous forme graphique, produit des statistiques, facilite la détection et la correction des erreurs et aide au dessin à l'échelle.

Easytopo connecté à Auriga permet de voir en temps réel la topographie se dessiner.

Auriga peut être utilisé tout à la fois comme :

- un carnet électronique pour la prise de notes sous terre
- une calculatrice scientifique pour la conversion des visées topo en coordonnées cartésiennes
- un dispositif de sauvegarde électronique des données topo en attendant de retourner à son PC/Mac
- un outil d'affichage du cheminement topographique de la grotte (vue en plan ou en coupe avec zoom et déplacement)

7. Conclusion et perspective

Pour les environnements lointains, exigus ou difficiles Easytopo permet l'acquisition de plusieurs centaines de points sans difficulté et avec une grande précision. Il permet par exemple de topographier le filaire ainsi que les sections des galeries à chaque station et donc de pouvoir les représenter en 3D.

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Procédures de protection des captages d'alimentation en eau potable en milieu karstique en France : Bilan et préconisations

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Abstract

The establishment of the mandatory protection perimeters for drinking water resources in France is far behind schedule. Especially where karst aquifers are concerned, it seems to have come up against many difficulties, whether they be technical, economic, legal, administrative, organisational or human.

As groundwater protection in karst environment is a very important stake for the Adour-Garonne and Rhône-Méditerranée basins, both water agencies jointly launched a survey in 3 phases:

- First overall view of the subject: specificities of the procedures in karst terranes, main strategies used these last ten years for the 3 stages of the procedure, elements of failure and success. This step has been carried out using experience feedbacks available in the bibliography and with advice given by experts;
- About fifteen case studies, mainly based on discussion with the people involved in the control of the procedures, in the south of France, the Alps, the Jura, and Normandy;
- Recommendations for improvement in all the stages of the procedures and all their aspects: organisational, administrative, methodological, legal and technical. They are addressed to the facility owners, to the administration and to the approved hydrogeologists.

Résumé

La mise en place des périmètres réglementaires de protection des captages d'eaux destinées à la consommation humaine accuse en France un retard important ; elle semble se heurter dans le cas des aquifères karstiques à de nombreuses difficultés, d'ordre technique, économique, juridique, administratif, organisationnel et humain.

La protection des ressources en eau souterraine en milieu calcaire concernant particulièrement les bassins Adour-Garonne et Rhône-Méditerranée, les 2 Agences de l'eau ont lancé conjointement une étude en 3 phases :

- Première vision d'ensemble de la problématique : spécificités des procédures en milieu karstique, principales stratégies utilisées ces dix dernières années pour les 3 étapes de la procédure, éléments de blocage et de réussite, à partir des retours d'expériences disponibles dans la bibliographie et des témoignages des experts rencontrés ;
- Une quinzaine d'études de cas, essentiellement sur la base d'entretiens avec les acteurs impliqués dans la conduite des procédures, dans le sud de la France, les Alpes, le Jura, et la Normandie ;
- Des préconisations d'amélioration des démarches de protection, concernant toutes les étapes des procédures et tous leurs aspects : organisationnels, administratifs, méthodologiques, réglementaires et techniques. Elles s'adressent aux maîtres d'ouvrage, à l'administration et aux hydrogéologues agréés.

1. Introduction

Les terrains carbonatés occupent plus de 30% de la surface de la France métropolitaine. De nombreux points d'eau y sont exploités pour la distribution publique et l'alimentation humaine, que ce soit par captage de sources, certains depuis l'antiquité, ou par forages plus récemment. L'occupation croissante des zones karstiques et l'augmentation des besoins en eau imposent la mise en place de politiques de protection adaptées.

Bien que la réglementation française prévoie la mise en place obligatoire des périmètres de protection des captages d'eaux destinées à la consommation humaine, cette procédure accuse un retard important ; elle semble se heurter dans le cas des aquifères karstiques à de nombreuses difficultés.

Afin de proposer des solutions pratiques tant sur les procédures que sur les méthodologies, les agences de l'eau des bassins Rhône-Méditerranée & Corse et Adour-Garonne, ont lancé une étude commune visant à établir un bilan de la mise en œuvre des procédures de protection des captages d'eau potable spécifiquement en milieu karstique. Cette étude, réalisée en 2004-2005, fait le point sur la

situation existante (état des lieux, analyse des diverses causes du retard ou au contraire des stratégies « gagnantes ») et fait des propositions pour améliorer la situation.

2. Principales difficultés rencontrées dans les procédures de protection des captages en milieu karstique

Le bilan réalisé se base sur un retour d'expérience tiré d'entretiens auprès d'une douzaine d'experts, de la bibliographie et de l'analyse de 15 procédures sélectionnées dans 9 départements pour permettre de couvrir un panel de contextes hydrogéologiques, socio-économiques et humains suffisamment varié.

Les procédures de protection comprennent 3 étapes :

- l'étape amont concernant l'organisation et les stratégies départementales,
- l'étape de procédure aboutissant aux périmètres de protection réglementaires,
- l'étape d'application des mesures de protection, de gestion et de suivi des périmètres de protection.

Les principales difficultés sont présentées de manière synthétique pour ces différentes étapes. De manière générale, l'absence de méthodologies spécifiques à l'établissement des périmètres de protection en milieu karstique, tant au niveau départemental qu'au niveau national, se traduit notamment par une hétérogénéité des procédures d'un département à l'autre et même au sein d'un même département. L'existence de données hydrogéologiques sur de grands ensembles karstiques ou la présence d'un spécialiste du karst associé au groupe de pilotage départemental permettent une mise en place des procédures plus rapide et en meilleure adéquation avec la spécificité hydrogéologique d'un aquifère karstique.

2.1 Mise en place et déroulement des procédures

L'objectif des procédures : préserver le captage des risques de pollution provenant des activités exercées à proximité

Les objectifs attribués aux périmètres de protection visent fréquemment la protection de la ressource, et non seulement du captage. Les périmètres de protection n'ont cependant pas vocation à protéger la ressource contre toutes les pollutions potentielles en particulier lorsqu'il s'agit de pollutions diffuses.

En domaine karstique, les procédures qui visent de façon inappropriée la protection totale de la ressource peuvent aboutir à des propositions de périmètres de protection très étendus et, par voie de conséquence, à des contraintes socioéconomiques difficilement acceptables.

En outre, les différences de vision entre partenaires quant à la finalité de la protection constituent un des principaux écueils dans le déroulement des procédures.

Etudes préalables à l'établissement des périmètres de protection : à mieux cadrer et à rendre plus opérationnelles

L'absence fréquente de cahiers des charges type pour les études préalables induit une assez grande variabilité du contenu de ces études. Le plus souvent, elles s'attachent essentiellement à la caractérisation de l'aquifère karstique, les autres volets étant alors insuffisamment développés.

L'inventaire et la cartographie des sources potentielles de pollution sont systématiquement réalisés ; par contre, l'évaluation de la vulnérabilité est souvent incomplète, et il est très rare qu'une cartographie soit établie. L'insuffisance des approches en matière de vulnérabilité s'explique avant tout par l'absence d'une méthodologie, reconnue et standardisée, adaptée aux systèmes karstiques.

Les études peuvent ainsi apparaître déconnectées de la problématique de définition des périmètres. La réflexion sur les solutions complémentaires ou alternatives (traitement, dispositif d'alerte, interconnexion, ressource alternative) est souvent absente.

Intervention de l'hydrogéologue agréé : une adéquation à optimiser entre les moyens et les enjeux

Selon la réglementation, l'hydrogéologue agréé rend un avis technique, sur la base des études préalables, portant sur les disponibilités en eau, la définition des périmètres et les mesures de protection. L'inadéquation entre la mission de l'hydrogéologue agréé et les moyens qui lui sont accordés (2 à 3 jours de travail) est amplifiée pour les procédures en milieu karstique du fait de la complexité technique et de dossiers plus lourds à appréhender.

Un problème souvent soulevé concerne le manque d'argumentaire pour justifier la délimitation des périmètres et le choix des prescriptions ; l'absence de méthode de référence et de critères objectifs pour définir les périmètres en domaine karstique est sans doute à l'origine de la difficulté pour l'hydrogéologue agréé d'expliciter son avis.

L'avis de l'hydrogéologue agréé n'est pas traité de façon similaire dans tous les départements : il est soit considéré comme un élément de diagnostic parmi d'autres, négociable entre les services de l'Etat, la collectivité et l'hydrogéologue agréé, soit jugé indiscutable et retranscrit tel quel dans l'arrêté de Déclaration d'Utilité Publique.

Certains experts de niveau national considèrent qu'il serait bénéfique de consolider les connaissances des hydrogéologues agréés en matière de protection en général et de protection des aquifères karstiques en particulier.

Définition des périmètres : l'instauration d'une méthode de référence permettrait d'améliorer la lisibilité et la robustesse des périmètres

L'absence de méthode de référence pour définir les périmètres de protection en milieu karstique constitue un inconvénient notable. A défaut, les hydrogéologues agréés utilisent des critères de définition qui leur sont propres ; ces critères n'étant pas toujours exposés, il est difficile d'apprécier la pertinence des périmètres.

La prise en compte d'objectifs trop larges de protection de la ressource, ou l'utilisation de méthodes inadaptées au karst pour déterminer les périmètres, conduisent à des zonages surdimensionnés.

En revanche, le recours aux périmètres de protection satellites institués par la réglementation pour protéger les zones de forte vulnérabilité (avens, dolines, bétouilles), s'avère particulièrement bien adapté aux spécificités des ressources karstiques.

Prescriptions dans les périmètres de protection : la définition de mesures mieux ciblées et acceptables sur le plan socioéconomique est à favoriser

Les mesures de protection visent généralement tous les types de pollution : accidentelle, chronique, ponctuelle et diffuse. Les prescriptions sont ainsi trop générales et répètent les mesures de la réglementation en vigueur.

Quelquefois, les prescriptions sont au contraire jugées trop strictes, induisant des contraintes socioéconomiques trop fortes pour les collectivités concernées, surtout lorsqu'elles s'appliquent à de grands périmètres ; ce type de difficultés est à l'origine de la plupart des situations de blocage des procédures.

Les prescriptions spécifiques aux aquifères karstiques sont peu nombreuses ; citons principalement les aménagements des pertes et points d'infiltration rapide.

2.2 Mise en œuvre des prescriptions : à dynamiser par la création de dispositifs de contrôle et d'évaluation

L'aboutissement des procédures administratives ne constitue pas une garantie de mise en œuvre des prescriptions, du fait notamment de l'insuffisance de suivi et de contrôle.

Toutefois, des réflexions existent dans certains départements pour la création de structures ou d'instances chargées du suivi des mesures de protection et de leur efficacité.

2.3 En conclusion : des difficultés communes à tous les types de procédures, mais exacerbées sur les procédures en karst

Les problèmes mis en exergue ne sont pour la plupart pas spécifiques aux procédures de protection en milieu karstique ; toutefois, certains facteurs tendent à exacerber les difficultés en domaine karstique : ressources souvent mal connues, analyse de la vulnérabilité et des risques plus délicate du fait de la complexité de la structure et du fonctionnement des aquifères, de la taille des bassins d'alimentation, et donc du nombre de collectivités concernées.

3. Préconisations en faveur des démarches de protection des ressources karstiques utilisées pour l'alimentation en eau potable

3.1 Les objectifs de la protection

Réaffirmer la complémentarité entre les procédures réglementaires de protection des captages et les démarches générales de préservation des ressources souterraines

Les procédures réglementaires de protection des captages ne peuvent répondre à des ambitions plus larges que la protection contre les risques de pollution dus aux activités implantées à proximité (géographique et temporelle). Par ailleurs, vouloir se servir des périmètres de protection pour lutter contre la pollution diffuse n'est pas réaliste, car les mesures de protection ne seraient pas contrôlables par la police sanitaire.

Le rapport ministériel VILLEY-DESMESERETS *et al.* (2001) rappelle les limites de l'outil périmètres de protection, en particulier dans le cas des aquifères karstiques, où le principe d'éloignement, qui vise à assurer un temps de transfert suffisant, n'est pas valide.

Les périmètres doivent être considérés comme un des outils de protection d'une ressource captée, qui doit s'inscrire dans une démarche plus globale et être conjugué en tant que de besoin avec deux autres approches :

- l'application stricte de la réglementation générale ou de mesures de gestion adaptées à l'échelle du bassin d'alimentation du captage ;
- le recours éventuel à des équipements complémentaires : dispositif de traitement des eaux captées, système d'alerte à la pollution, équipement en ressource alternative ou interconnexion.

Il importe de mener de front la protection réglementaire rapprochée et la politique de préservation des ressources souterraines, tout en affirmant nettement la distinction : la première relevant de la législation sur la santé publique, et la seconde de la politique environnementale générale. Cette approche est cohérente avec les dispositions de la directive cadre européenne sur l'eau, qui impose le développement de stratégies d'actions à l'échelle des aquifères, dans le but d'atteindre l'objectif de bon état chimique des eaux souterraines.

3.2 Préconisations relatives à l'encadrement des procédures au niveau départemental

L'intérêt de l'existence d'un **groupe de pilotage départemental** a déjà été souligné lors d'un précédent bilan

sur l'avancement des procédures (Agence de l'eau Loire-Bretagne 1999).

Ses attributions comporteraient notamment l'élaboration d'un **cahier des charges type** pour les études préalables, adapté aux caractéristiques du karst dans le département, et la validation des cahiers des charges élaborés pour chaque nouvelle procédure.

Des **sessions de formation** sont à mettre en place pour les acteurs participant au groupe de pilotage départemental. Des outils pédagogiques sont à développer à cet effet, sous l'égide des ministères en charge de la santé et de l'environnement.

La présence d'un **hydrogéologue ayant des compétences en milieu karstique** dans le groupe de pilotage départemental est nécessaire pour assurer la cohérence et la pertinence technique des interventions du groupe.

Il est recommandé de réaliser des **études générales de connaissance des grands ensembles karstiques**, si possible en amont des procédures réglementaires, pour identifier les aires d'alimentation, caractériser le fonctionnement des systèmes karstiques, et cibler les zones particulièrement vulnérables.

3.3 Préconisations relatives à l'étape de lancement de la procédure

Il est recommandé que la collectivité pétitionnaire s'associe les compétences d'un assistant à maîtrise d'ouvrage pour toute la durée de la procédure.

Il est également préconisé qu'un **comité de suivi local** soit constitué en amont de chaque procédure, regroupant les élus locaux, des représentants des acteurs locaux, et au moins une personne des services de l'Etat ou du département, chargée du lien avec le groupe de pilotage départemental.

Un **état des connaissances hydrogéologiques** est à établir au début de la procédure, pour permettre de bien identifier les investigations complémentaires à mener dans le cadre des études préalables.

3.4 Préconisations relatives au contenu des études préalables

Les études préalables établissent un **diagnostic**, puis présentent et comparent des **scénarios** alternatifs pour la protection du captage et de la ressource et aboutissent à la définition d'une **stratégie opérationnelle de protection et d'un programme d'actions**.

Etablissement du diagnostic

Le diagnostic consiste à confronter :

- les éléments de constat : état de la qualité des eaux brutes et traitées, taille et caractéristiques de l'aire d'alimentation, filières de traitement en place, existence de ressources de secours, contraintes socio-économiques ;
- avec les résultats de l'évaluation des risques actuels et futurs de pollution accidentelle et chronique (issus du croisement entre vulnérabilité et aléas), qui permettent de cartographier différents types de zones, en fonction de l'intensité du risque ;
- et avec les possibilités d'équipements complémentaires : dispositif de traitement, station d'alerte, ressource de substitution, interconnexion.

Les trois points suivants sont essentiels pour l'établissement du diagnostic :

- pour chaque captage à protéger en zone karstique, bien identifier l'aire d'alimentation,
- établir une cartographie de la vulnérabilité intrinsèque selon une méthode standardisée,
- évaluer les risques de contamination à partir d'une analyse croisée entre la cartographie des sources potentielles de pollution et celle de la vulnérabilité.

Le prérequis est l'instauration d'une méthode de référence pour l'évaluation de la vulnérabilité dans le prolongement des réflexions conduites à l'échelle européenne ou française (par exemple en Franche Comté avec la méthode RISK).

Scénarios de protection

Le diagnostic établi dans les études préalables doit déboucher sur 2 ou 3 scénarios de protection, assortis d'**analyses de faisabilité technico-économique**.

Les scénarios de protection doivent composer tous les types d'actions : au niveau du captage, de la distribution, des périmètres de protection et de l'aire d'alimentation.

Il s'agit d'envisager les différents outils et moyens possibles pour sécuriser la qualité des eaux distribuées, en recherchant le meilleur équilibre et la complémentarité entre les mesures réglementaires à appliquer dans les périmètres de protection, les actions de maîtrise des risques à engager à l'échelle de l'aire d'alimentation et les solutions techniques au niveau du captage ou du système d'adduction, tout en tenant compte des enjeux et des contraintes socio-économiques.

Stratégie opérationnelle de protection

Trois principes sont à retenir pour élaborer une stratégie de protection équilibrée et réaliste :

- les mesures réglementaires dans les périmètres de protection ne sont qu'un des moyens utilisables pour atteindre l'objectif de conformité de la qualité des eaux distribuées ; les contraintes doivent être ciblées prioritairement sur les zones les plus vulnérables ;
- assurer une protection suffisante de l'aquifère de manière à réduire le degré de traitement nécessaire à la production d'eau potable ;
- ajuster les moyens à l'importance des enjeux de protection et aux contraintes socio-économiques.

La stratégie de protection définie à l'issue des études préalables devrait comporter entre autres un **programme d'actions hiérarchisé, chiffré et planifié** intégrant :

- les propositions de prescriptions réglementaires dans les périmètres de protection,
- les actions de réduction des risques de contamination à l'échelle de l'aire d'alimentation,
- le cas échéant, les équipements complémentaires à mettre en œuvre au niveau du captage ou du système d'adduction et les modalités de gestion des ouvrages de distribution en cas de pollution accidentelle ou de pics de pollution liés aux conditions hydrologiques.

Il est proposé qu'une **délibération de la collectivité maître d'ouvrage** entérine la stratégie de protection retenue.

3.5 Préconisations relatives à la définition des périmètres de protection

Les périmètres de protection sont à définir à partir de la cartographie de la vulnérabilité intrinsèque.

Il est important que la traduction de la carte de vulnérabilité soit guidée par un outil méthodologique, permettant de **relier de façon objective les classes de la carte de vulnérabilité aux périmètres de protection** rapprochée et éloignée.

Pour les procédures en milieu karstique, il est préconisé que le **périmètre de protection éloignée soit toujours calé sur l'aire d'alimentation du captage**. Cette mesure permet de formaliser l'aire d'alimentation dans le dossier réglementaire et dans les documents d'urbanisme. Il est envisageable que dans cette zone de vigilance, en fonction des risques identifiés, certains seuils des nomenclatures loi sur l'eau et Installations Classées pour la Protection de l'Environnement soient abaissés.

Par ailleurs, il est préconisé de mettre en place une **procédure allégée pour des modifications mineures des périmètres de protection**, en cas d'amélioration des connaissances ou d'évolution structurelle du système karstique.

3.6 Préconisations relatives aux prescriptions dans les périmètres de protection

Il convient de rappeler que la loi de santé publique du 9 août 2004 soutient l'application des mesures de protection dans le périmètre de protection rapprochée en facilitant la maîtrise foncière et en permettant à la collectivité propriétaire d'y prescrire des modes d'utilisation des sols adaptés.

Les prescriptions doivent être définies au cas par cas, en fonction des risques identifiés au préalable. Le niveau des prescriptions est à adapter au degré de vulnérabilité de chaque système aquifère ; les mesures de protection dans les périmètres de protection rapprochée ne doivent pas simplement calquer celles de la réglementation générale, ce qui annule l'intérêt du classement en Périmètre de Protection Rapprochée.

3.7 Préconisations relatives aux actions à l'échelle des aires d'alimentation

Les actions à l'échelle des aires d'alimentation sont particulièrement pertinentes pour la protection des ressources karstiques menacées par des pollutions diffuses, en particulier d'origine agricole. Elles peuvent néanmoins viser d'autres types de pollution. Les actions consistent en une application stricte voire un renforcement de la réglementation générale ; l'ampleur des mesures doit être modulée en fonction des enjeux et des contraintes socioéconomiques. Parmi les types d'actions à préconiser :

- mesures agri-environnementales : modification des pratiques ou des modes d'occupation des sols (dispositifs enherbés, conversion des cultures en prairies, cultures intermédiaires pièges à nitrates, réduction des intrants, etc.), à mettre en place notamment au travers de contrats d'agriculture durable,
- opérations coordonnées de mise aux normes des élevages situés en deçà des seuils réglementaires,
- choix de procédés d'épuration des effluents domestiques adaptés aux enjeux liés à la ressource (traitements complémentaires de déphosphatation ou décontamination),
- actions de prévention des pollutions accidentelles.

Au delà d'actions localisées, ces programmes, portés par des collectivités au périmètre cohérent avec celui des aires à protéger (groupements de communes, départements...), doivent promouvoir la **mise en cohérence des politiques locales d'aménagement du territoire avec les objectifs de préservation de la ressource**.

3.8 Préconisations relatives aux solutions techniques au niveau des installations d'adduction

Pour les captages en karst, compte tenu de la variabilité de la qualité des eaux, le développement des systèmes d'alerte, des systèmes de traitement et/ou du recours à une ressource alternative, revêt un intérêt particulier. En outre, l'instauration d'une norme de qualité plus sévère pour la turbidité oblige les collectivités à prévoir un dispositif permettant de gérer les pics de turbidité consécutifs aux épisodes de crue.

Les systèmes d'alerte peuvent s'appuyer sur la surveillance de la conductivité, des débits, ou directement de la turbidité.

En fonction des risques identifiés le système d'alerte peut être :

- un simple contrôle de paramètres basiques (débit, conductivité, turbidité),
- un dispositif plus sophistiqué fondé sur des indicateurs biologiques ou chimiques.

3.9 Préconisations relatives à l'intervention de l'hydrogéologue agréé

A la condition que des méthodes de référence soient utilisées pour l'élaboration de la cartographie de la vulnérabilité et sa traduction en périmètres de protection, il est recommandé que la proposition de périmètres et de mesures de protection soit faite à l'issue des études préalables par le bureau d'études. Cette proposition est validée par la collectivité maître d'ouvrage et le comité de suivi local, puis soumise à l'avis de l'hydrogéologue agréé.

Pour améliorer la continuité des procédures et fiabiliser l'intervention de l'hydrogéologue agréé, deux mesures sont préconisées :

Renforcer la formation des hydrogéologues agréés sur la protection des aquifères karstiques,

Recentrer la mission des hydrogéologues agréés : en tant que personnalité compétente et indépendante, il émet un avis technique sur les propositions de périmètres et de mesures de protection résultant des études préalables.

L'hydrogéologue agréé n'intervient qu'une seule fois dans la procédure, et son avis n'a pas à être modifié. L'avis est joint au dossier d'enquête publique, ainsi que, le cas échéant, le résultat de l'arbitrage du service départemental de l'Etat en charge de la santé (Direction Départementale de l'Action Sanitaire et Sociale - DDASS) faisant suite à l'avis.

3.10 Préconisations relatives à la mise en œuvre de la stratégie de protection

Les mesures d'incitation instaurées par les Agences de l'eau n'ont pas une efficacité suffisante et les moyens des services de l'Etat ne permettent pas d'envisager un renforcement du contrôle ; des relais territoriaux sont donc à mobiliser : les départements, au travers par exemple des services d'assistance technique aux exploitants d'eau potable (SATEP), pourraient développer des compétences dans ce domaine, en particulier auprès des petites collectivités.

4. Conclusions

Les deux Agences de l'eau Adour-Garonne et Rhône-Méditerranée & Corse se sont maintenant engagées dans la poursuite de cette démarche.

La première action concrète consiste en l'élaboration d'un guide méthodologique s'appuyant sur les résultats de l'étude, à destination des différents acteurs impliqués dans les procédures de protection (maîtres d'ouvrage et leurs exploitants, collectivités territoriales, Agences de l'eau, administration, hydrogéologues agréés, bureaux d'étude).

Ce guide pourrait paraître dès 2007, pour accompagner au plus vite l'accélération du rythme des procédures de protection souhaitée par le Plan National en Santé Environnement, qui vise l'achèvement de ces procédures en 2010 pour l'ensemble des captages du territoire.

Ce premier guide devra être accompagné d'un guide méthodologique concernant les thèmes suivants :

- identification et caractérisation d'un système karstique,
- traçages de reconnaissance et traçages quantitatifs en vue de simulations de pollutions dans un système karstique,
- délimitation des périmètres de protection en milieu karstique, à partir de méthode(s) d'évaluation et de cartographie de la vulnérabilité intrinsèque ;

Des ajustements réglementaires mineurs et éventuellement une incitation par voie réglementaire destinée à soutenir la mise en œuvre des préconisations seront également proposés.

Les deux Agences de l'eau ont d'ores et déjà engagé la diffusion d'une synthèse de l'étude, et inscrit le sujet à l'ordre du jour de différentes rencontres, notamment entre les 6 Agences et avec les deux ministères concernés (chargés de la Santé et de l'Environnement), afin de mutualiser les résultats et dégager rapidement les éléments clé des actions concrètes possibles à court terme sous la responsabilité des uns ou des autres.

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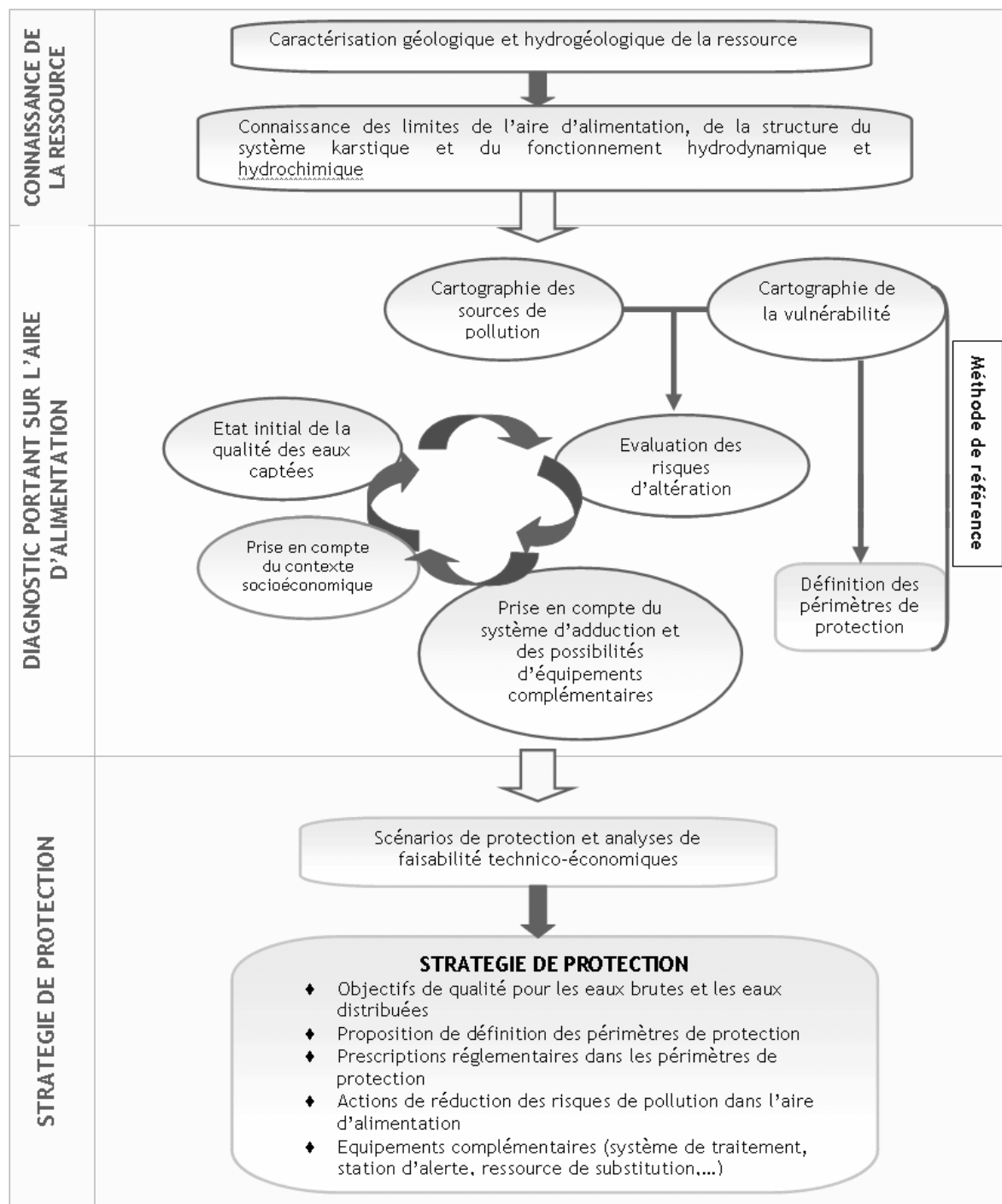


Fig. 1 : synthèse schématique des propositions relatives à l'élaboration de la stratégie

Spring monitoring and analysis of groundwater circulation in the Sibillini mountains aquifers (Adriatic side of central Apennine, Italy)

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Abstract

In this paper, hydrogeological studies on aquifers which recharge some Umbria-Marche ridge springs, emerging in the southern Adriatic side of Sibillini Mountains are analysed. The main springs are Capodacqua, Pescara and Foce; other shallow, local springs, emerging at 600-1500 meters above sea level, represent the infiltration altitude of rain waters. All the springs were monitored for a period of about three years, with a regular collection of data (temperature, electrical conductivity, etc.) and sampling for chemical and isotopic analysis. Capodacqua, Pescara and Foce feed the aqueduct network of the Ascoli Piceno province.

1. Introduction

The preliminary results of studies regarding the hydrogeology of the limestone Sibillini massif are presented in this paper. The Sibillini Mountains rise in the south-western part of the Marche region where the Umbria-Marche ridge and the Marche ridge form a single and complex morphological element (Fig. 1 and 2). The aim of the research is to contribute to the definition of the groundwater circulation in the limestone complex of Sibillini by monitoring the three most important springs present, and defining the volumes of water. To this end, the spring hydrography is studied and related to the trend of rainwaters and snowfall; the variations of the isotopic composition ($\delta^{18}\text{O}$ and $\delta^2\text{H}$) during the hydrogeologic year and the chemical-physical water parameters are analysed. The data processing allowed the marking of the limits of the recharge area of the springs and the definition of the hydrogeological basin characterised by the abundance of water. Recently a Rd detector has been installed near the Capodacqua spring, with the aim also to note the relationship between earthquake and variation of Rd content in the groundwater.

2. Lithostructural and hydrogeological setting

In the Sibillini area (Fig. 1 and 2) the Umbria-Marche sequence (CENTAMORE & MICARELLI, 1991) is present with the platform and dolomitic limestones of the Calcare Massiccio and the micritic limestones of the Corniola (9 in Fig.2) at the base. Between the Massiccio-Corniola and the Maiolica Formations, the marly Bosso Formation (8 in Fig.2) with low permeability may be present. The pelagic sequence follows, with the limestones of Maiolica (7 in Fig.2), the Marne a fucoidi (6 in Fig.2) and with the limestones, marly limestones, limestone marls of the Scaglia Formations (5 in Fig.2). Condensed series are sometimes present, where the contact between Calcare Massiccio and Maiolica is possible.

Schlier Formations and with various lithotypes according to different areas (4 in Fig.2) (Arenarie di Camerino in the Camerino basin; pre-evaporitic Laga Formation, Gessoso-solfifera Formation, post-evaporitic Laga Formation and Colombacci Formation in the northern part; Laga Formation with pre-evaporitic, evaporitic and post-evaporitic members in the southern part of the area). Finally the continental

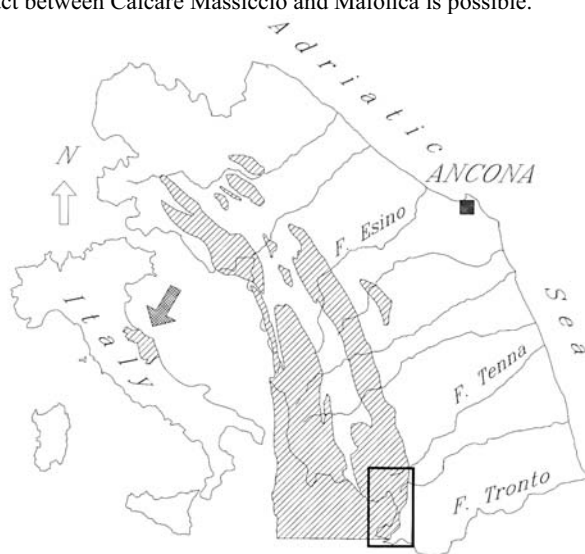


Fig. 1 – Location of the studied area. The Umbria-Marche limestone ridges are marked and the studied area is bordered.

The marly limestones and the marls of the Miocene sequence then outcrop (3 in Fig.2), with the Bisciario and

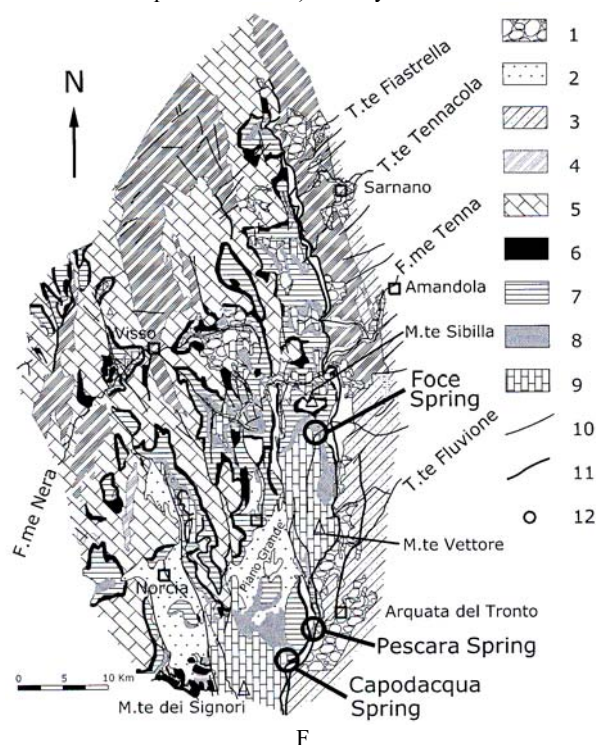


Fig.2- Hydrogeological schematic map of the Sibillini area. The monitored springs are also indicated. For the legend see paragraph 2.

Quaternary deposits, alluvial, glacial etc. (1 and 2 in Fig.2), are present in the Sibillini area. The structural setting of the complex is characterised by a plicate style with apennine and anti-apennine faults (10 in Fig.2) and overthrusts (11 in Fig.2). The main overthrust borders the Adriatic side of the Sibillini complex with a curved line, allowing the overlap of the Meso-Cenozoic limestone formations on the Scaglia cinerea, in the northern area and on the Laga Formation, in the southern area.

The aquifers of the Sibillini mountains (BONI *et al.* 1986; NANNI & VIVALDA 2004) essentially correspond to the three hydrogeological complexes of Calcare Massiccio, Maiolica and Scaglia. The Massiccio complex consists of the limestones of Calcare Massiccio and Corniola Formations and, in the presence of the condensed series of Maiolica Formation, it contains the basal groundwater flow of the limestone ridge. This aquifer is generally drained by linear springs along the rivers; single springs recharged by the Massiccio are rare but with high discharge (12 in Fig.2).

The Maiolica complex can be distinguished only in the areas with the Bosso Formation of the complete sequence (CENTAMORE & MICARELLI 1991).

The Scaglia bianca, rossa and variegata complex is located between the low permeable lithotypes of Marne a Fucoidi and Scaglia cinerea. This complex recharges a lot of springs in the area characterised by low discharge.

The permeability of the complexes is essentially due to fissures and karstic phenomena which are highly developed in the Massiccio Formation.

Aquifers are also present in the arenaceous units of the Miocene sequence in the Camerino basin and in the Laga basin with torbiditic arenaceous units. Finally, numerous springs with often permanent regime are present in the permeable colluvial deposits. These deposits can be found at the contact of the limestone complexes and the hilly area. Perennial springs may also be present at high altitude in the fluvial-torrential and glacial deposits.

3. Monitored springs

The monitored springs, Capodacqua (830 m), Pescara (885 m) and Foce (967 m), are connected with the basal aquifer of the Sibillini complex and feed the aqueduct network of Ascoli Piceno province (CIIP) by tunnels with a diameter of 2-2.5 meters. The Foce spring (Fig.2) emerges from the Calcare Massiccio and Corniola Formations in the overthrust zone in which there is the contact between the Sibillini complex and the terrigenous sequence of the Adriatic front. The Capodacqua spring (Fig.2) emerges in the contact area between the Corniola lithotypes and the Marnosa Arenacea Formation that acts as an aquiclude in the zone. Finally the waters of the Pescara springs (Fig.2), emerge in the area of the overthrust that brings the Maiolica and Scaglia Formations in contact. Other shallow, local springs, emerging at 600-1500 meters above sea level, indicate the infiltration altitude of rain waters.

4. Methods and equipments

In the tunnels of the monitored springs, some instruments for acquisition and memorization of chemical and physical parameters (temperature, electrical conductivity and hydrostatic level) have been installed. Datalogger connected to specific sensors with precision and sensitivity for this type of surveying, have been used. The survey system of the hydrostatic pressure has been realized in the Hydrogeology Laboratory of Ancona University by Honeywell differential low pressure sensors. The sampling frequency has been in accordance with the instrumental autonomy that allowed a backup with an interval of approximately 1 or 2 months. The

values, saved in the datalogger and unloaded on computer, have been elaborated with an electronic calculation sheet. The millivolt data of the hydrostatic pressure sensors have been correlated with the results of numerous tests of instantaneous discharge, obtained by a small currentometer and by fluorescent dyes. The meteorological data (rain and snow) are derived from the State Forestry Department and from the Civil Protection Service of the Marche Region. In the area of Montemonaco a rain and snow gauge with a datalogger has been installed for continuous recording.

4. Results

Spring discharge and meteoric events. The discharge of Pescara (average value 223 l/s, maximum 608 l/s and minimum 50 l/s), Capodacqua (average value 450 l/s, maximum 790 l/s and minimum 270 l/s) and Foce (average value 400 l/s) springs is strictly in accordance with the meteoric events. By comparing the discharge with precipitation trend (Fig.3), a rapid increase of the discharge after the rainfall can be noted. The melting snow together with the rain also affect the maximum value of the discharge which happens in summer-winter months. Consequently, the waters emerging in this period, are related to the seasonal spring recharge. This occurs by a flow that allows the rain and the rapid melting snow waters to reach the spring in a short time through the unsaturated zone.

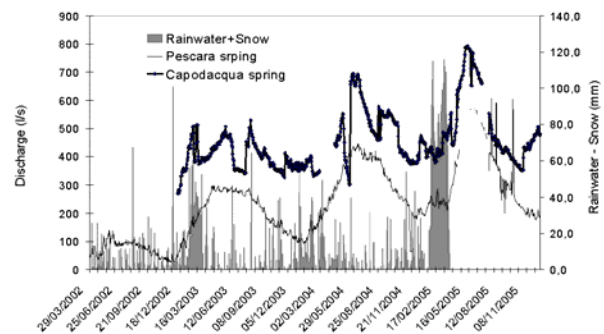


Fig. 3- Variations of discharge in Pescara and Capodacqua springs and precipitation trend during the survey period.

Hydrochemistry, temperature and electrical conductivity. The hydrochemical facies of Capodacqua, Pescara and Foce springs is typically calcium-bicarbonate with sulphate and magnesium enrichments in the Foce spring, probably in relation with leaching of the Triassic Anidriti di Burano Formation (NANNI & VIVALDA 2004).

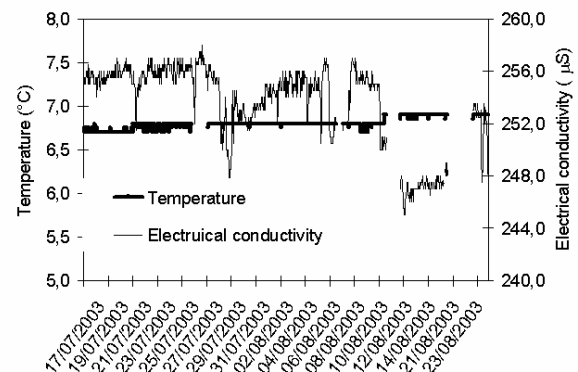


Fig 4- Relationships between temperature and electrical conductivity of waters in the Foce spring during the survey period.

In any case the water chemistry characterised by modest changes, with the addition of the spring waters at constant

temperature (Fig.4), confirms the hypothesis of considerable volumes of water in the aquifer.

The variation of electrical conductivity during the survey period, is negligible (Fig. 4) but, especially with regard to the Pescara spring, a small decrease of conductivity coincides with the maximum discharge in the spring-summer period (Fig.5). This may confirm the importance of the seasonal recharge.

The spring water temperature is approximately constant during the survey period (Fig.4). Even though the discharge follows the precipitation trend, the water temperature does not change. From the quantitative analysis, the temperature values in the Capodacqua, Pescara and Foce spring waters prove to be very low in comparison with those of other springs present in the limestone Apennine ridges and compared with the basal aquifer. This phenomenon means that, in accordance with the high altitude of the recharge basin, very cold waters infiltrate and immediately circulate and even the basal waters are characterised by low temperature. This may be in relation to the shallow position of the basal aquifer in this zone, caused by the particular structural setting of the Sibillini mountains.

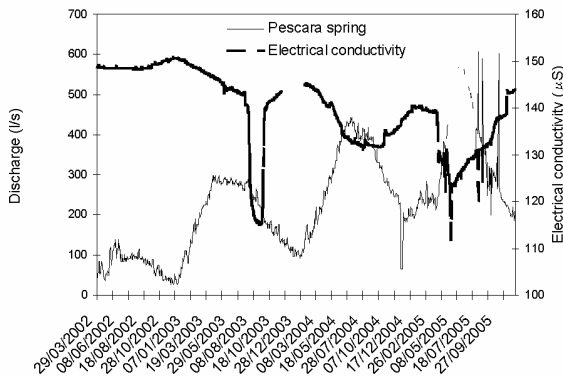


Fig. 5- Electrical conductivity of waters vs discharge during the survey period in the Pescara spring.

Groundwater isotopic composition. The study of the groundwater isotopic composition ($\delta^{18}\text{O}$ and $\delta^2\text{H}$), has been performed on the Pescara, Capodacqua and Foce springs, from September 2002 to February 2004. Springs behave in different ways, for example, the variability of the isotopic composition in Pescara and Capodacqua waters (Fig.6) shows a connection between discharge and isotopic values. The lowest isotopic contents are obtained at the end of 2002, characterised by the absence of snowfall during winter 2001-2002 and by scarce rainfall in spring 2002. Therefore during autumn-winter period 2002-2003, the two springs were essentially recharged by basal aquifer flow and essentially exhibits a low isotopic content. The highest contents instead are in accordance with the increase of the spring discharge. As we noted in figures 3 and 4, the increase of Pescara and Capodacqua discharge is due to the seasonal recharge, particularly the melting snow. Figure 6 shows that isotopic content rises together with the increase of discharge in springs which is strictly connected with the seasonal recharge (Fig.3). Therefore, the contents of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in the spring waters demonstrate the contributions of deep and shallow circuits to the springs. The isotopic values during the depletion phase and at the beginning of the recharge phase are related to deep circuit waters of the basal aquifer. The latter exhibits a multi-year recharge, which feeds the springs during the minimum flow period at the end of summer and autumn.

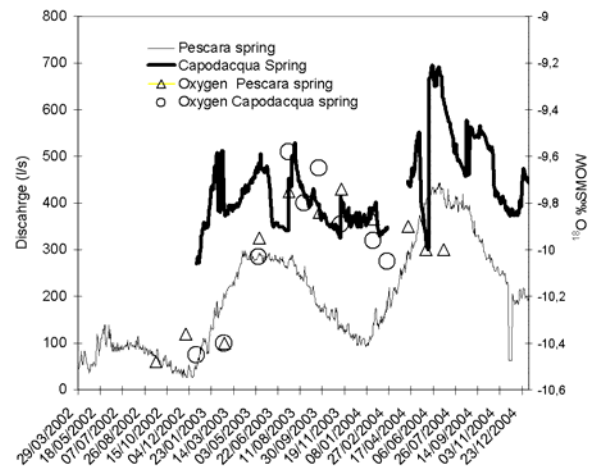


Fig. 6- Example of water isotopic composition variability in Pescara and Capodacqua springs during the survey period. In both springs the isotopic content is in relation to the discharge trend.

With regard to the negative values of $\delta^{18}\text{O}$ (Fig. 6) and $\delta^2\text{H}$ at the beginning of the recharge period, they are probably a consequence of the piezometric level rise in the spring area due to deep water release. This phenomenon may owing from the fast increase of the basal aquifer heads, induced by snow melting recharge. The isotopic values of Foce spring water (Fig. 7), characterised by a minimum discharge of 400 l/s, do not change during the period of minimum and maximum spring discharge. This reflects the different conditions of the hydrogeological basins of the three springs studied. In fact, the Foce spring is recharged by a larger hydrogeological basin as respect to the others, characterised by a more complex groundwater circulation. In this case a plausible hypothesis should be that seasonal recharge waters mix with the deep circuit waters overlapping the seasonal signal. The isotopic values connected with the increase in discharge in Pescara and Capodacqua springs, confirm the seasonal recharge contribution. This is characterised by shallow and fast flowpath and is due to the rapid infiltration of rain waters and melting snow.

The isotopic contents of the spring waters analysed during the survey period, have been correlated with other springs as seen in Fig.7. Here, if we observe the values of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ of the Pescara and Capodacqua springs in the autumn-winter period in the absence of seasonal recharge, we note they are the most negative and close to the northern Italian Meteoric Line (nIML). At the same time, these values are comparable to the Cacere spring values. This spring, located at 1380m a.s.l., exhibits hydrogeological features that allow the comparison between its winter isotopic contents with those of rain and snow of the Sibillini mountains during the same period. In fact, the Cacere water circuit is extremely rapid and very superficial. Therefore, by comparing the isotopic values of Cacere spring with the most negative values of Capodacqua and Pescara springs, we may confirm the hypothesis that the waters which recharge the basal aquifer derive from melting of the first snow fallen at high altitude (the maximum altitude of the studied area is 2200 m of M.Vettore). The isotopic values of Pescara, Capodacqua and Foce springs during the recharge period, are conform to the (LONGINELLI & SELMO, 2003) central Italian Meteoric Line (cIML). The alignment level of the isotopic values during recharge period with the cIML, confirms that the evaporation phenomenon in the Sibillini springs recharge area, does not affect the values of $\delta^{18}\text{O}$ and $\delta^2\text{H}$. This means that, like the other central Apennine hydrostructures (BARBIERI *et al.* 2003 and 2005), the infiltration is fast confirming the

results deduced in figures 3,4 and 5. In Fig. 6 there are also the isotopic values of the Rubiano A spring, located at an altitude of 850 m a.s.l. This spring has hydrogeological features comparable to Cacere spring; therefore the isotopic contents are probably similar to the values of rains which recharge small hydrogeological basins and springs, present at an altitude of about 800-900 m.

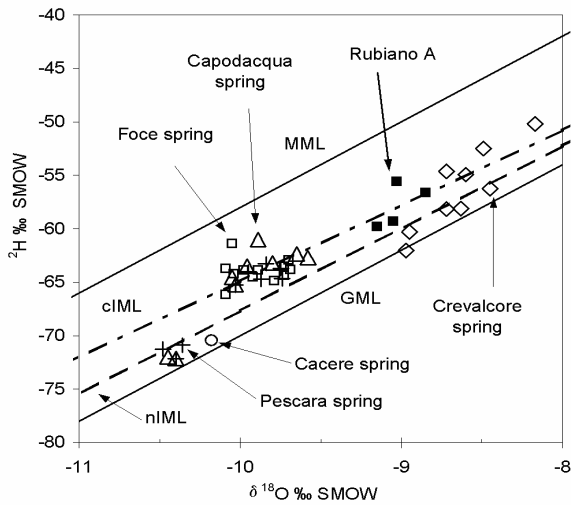


Fig.7- Isotopic values of the spring waters emerging from the Adriatic side of the Sibillini mountains compared with the Global Meteoric line (GML), central (cIML) and northern (nIML) Italian Meteoric Line (LONGINELLI & SELMO, 2003) and Mediterranean Meteoric Line (MML). The altitude of RubianoA. spring is 850 m and of Cacere spring is 1380 m. The correlation line of the isotopic values during the recharge period of the studied springs may close to the cIML.

Finally, comparing isotopic data of springs located at different altitudes in neighbouring areas considered in this study and in previous studies (CONVERSINI & TAZIOLI, 1993), has been observed a close relationship between $\delta^{18}\text{O}$ and $\delta^2\text{H}$ and altitude of the recharge basin. This relationship shows, within certain limits, that the mean isotopic aquifers recharge altitude agrees with the morphological and hydrogeological characteristics of the Sibillini area. This allows the location at 1700 and 1800 m a.s.l. of the springs infiltration zones analysed in this study (Fig.8).

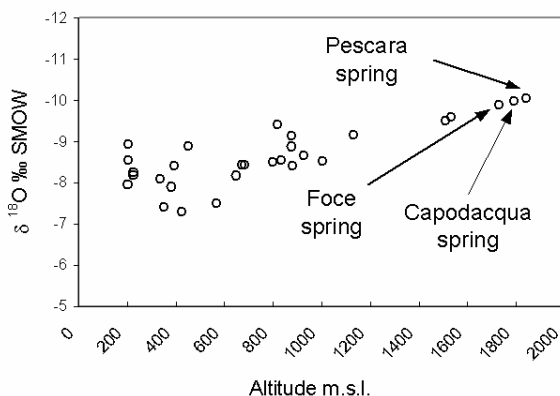


Fig. 8- Example of correlation between isotopic content in the water springs of Umbria-Marche Apennine and altitude of their alimentation basin.

5. Conclusion

This study was performed on the hydrodynamic characteristics and the chemical-physical parameters of the main springs located along the Adriatic front of the Sibillini

Mountains. The main findings are the alimentation from two different waters circuits, well recognizable in the Pescara and Capodacqua springs. The short and rapid circuit is essentially fed by the seasonal recharge. The deep circuit instead, recharges the springs during the minimum discharge period. Moreover, this study confirms the hypothesis concerning the priority of the seasonal recharge on the Sibillini spring alimentation. The rain and the fast snow melting reach the springs in a short time producing the highest discharges. The isotopic values in the Pescara and Capodacqua springs may confirm this hypothesis. The water supply to the springs from the basal aquifer is about 15-20% of the total discharge. Finally, the isotopic values allow the springs recharge zones location at an altitude of 1700-1800 m a.s.l., in accordance with the morphological and hydrogeological features of the Sibillini Mountains.

Acknowledgements

The authors thank the State Forestry Department, Meteomont service, the Protection Service of the Marche Region and the Water Consortium of Piceno (CIIP) for their availability and courtesy.

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Potential radon and radium sources of groundwaters of Gellért and József Hills (Budapest, Hungary)

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Abstract

Published archive data and current research results on radon and radium concentrations are used to outline the character and localisation of a high radioactivity spot in the larger area of Gellért and József Hills (Buda Thermal Karst System, Hungary). Geological models are elaborated for the evaluation of the importance and role of the possible impact factors.

The most prominent anomalous area lies at the northern part of the Gellért Hill. In the springs of Rudas Spa, ~600 Bq/l radon and ~1000 mBq/l radium concentrations were measured. Radon, and to a lesser extent radium concentrations, decrease towards the neighbouring Gellért and Rác Spas. The radioactivity of the József Hill springs is a magnitude lower than that of all the other springs mentioned above. Radium concentrations of the lukewarm (20-37°C) springs are about 50-100 mBq/l, while hot springs (37-60°C) show 200-300 mBq/l radium content. Both types have radon concentrations averaging 20-25 Bq/l.

While radium is assumed to be mostly of deep origin, radon needs a source in the discharge area. In the case of the József Hill, marls and soils may produce observed radon values, while on the Gellért Hill an extra, yet still unidentified radon source is suspected.

Continuous and batch radon and radium measurements, geological and geochemical studies focusing on the Gellért Hill zone are in progress in order to explain the radioactivity of the waters of the Buda Thermal Karst System.

1. Introduction

The radioactivity of the Buda Thermal Karst System has been known for almost a century (WESZELSZKY 1912). Spa waters with high radon content have been widely used as therapeutic media. However, detailed research was restricted to a few groups of springs, and the origin of radioactivity was not cleared. In this paper we

would like to present the scientific knowledge on the radioactivity of the thermal waters of the Gellért and József Hills, as well as our measurements and current hypotheses on the origin of the radon and radium content of these waters.

2. Materials and methods

The Buda Thermal Karst System is part of the Transdanubian Range, which is a geothermally influenced gravitational flow system (ÁLFÖLDI *et al.* 1968). One of the tectonically controlled, concentrated discharge zones is along the foothills of the Buda Mts (Fig. 1.). In this zone, upwelling thermal waters mix with descending cold water. Discharge occurs at a few groups of springs, the most important of these being at the Gellért and József Hills, which represent the probably most famous spa cultures in Hungary. While the topography of Gellért Hill is rough and infiltration through Triassic dolomite rarely limited, József Hill shows more gentle slopes, infiltration detained by marls and anthropogenic features. Published radon and radium values were measured by various methods (trace detection, liquid scintillation, bubbling methods) not dealt with in this paper. About 100 batch radon measurements, and 30 data for radium values have been available from the literature, covering several decades. For detailed measurement data see PALOTAI *et al.* 2005.

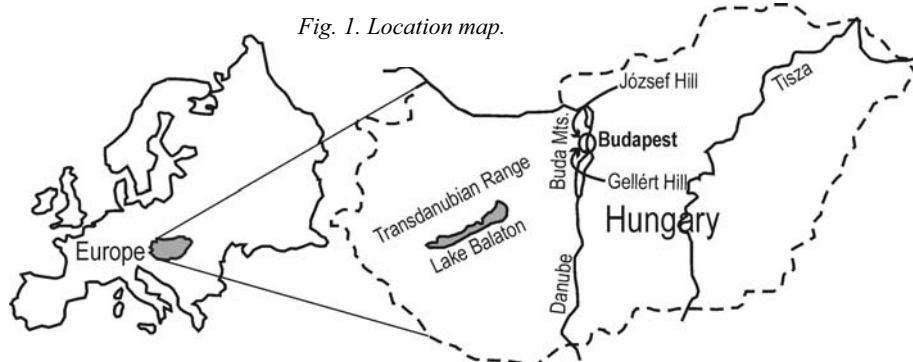


Fig. 1. Location map.

A majority of published radon measurements of the Gellért Hill spring group are summarized in Fig. 2, a non-linear, but approximately southeast to northwest trending section along the Danube. This profile shows a local maximum at the Rudas Bath area, decreasing towards the neighbouring Rác and Gellért Baths, i.e. to the north and south. In the centre of the anomaly, radon concentrations exceed 600 Bq/l, while at the peripheral zones these values remain below 100 Bq/l. Radioactivity values and temperature do not correlate.

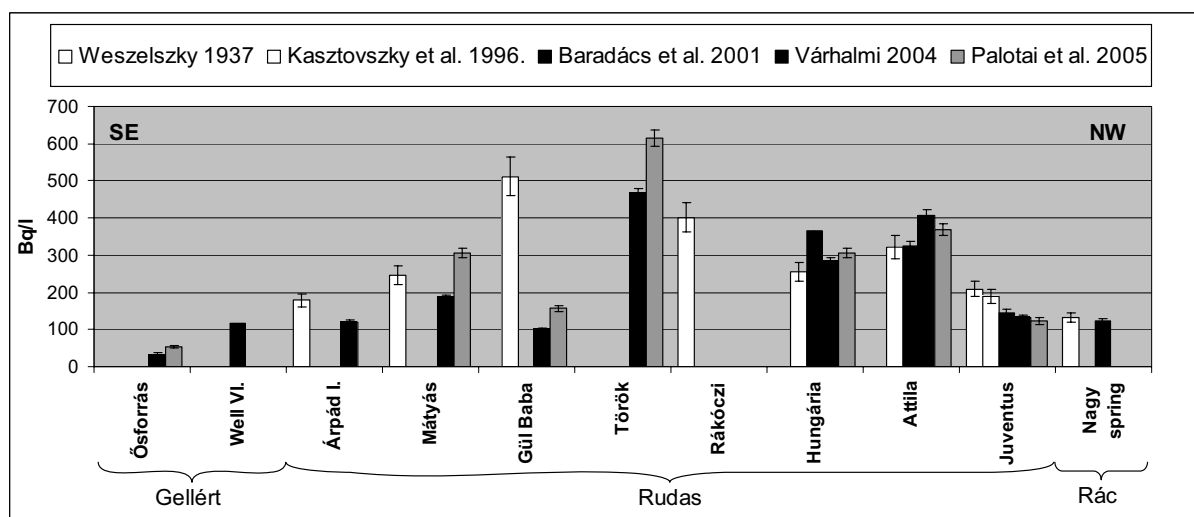


Fig. 2. Radon content of Gellért Hill springs, as published by various authors [Bq/l]. Horizontal scale is not linear. Data of BARADACS et al. (2001) are averages of etched track detection and bubbling methods results.

The radon concentrations measured in the József Hill spring group – also a famous spa area – show values even a magnitude lower. In this zone, hot (37–60°C) and lukewarm (20–37°C) spring discharge occurs in a very concentrated zone, the two types still being well separated. Results are summarized in Fig. 3. No

correlation is found between water temperature and radon concentration: both spring types show radon concentrations about 20–25 Bq/l. Only the Római spring shows a striking, very much localized anomaly, with concentrations 4–5 times higher than neighbouring springs.

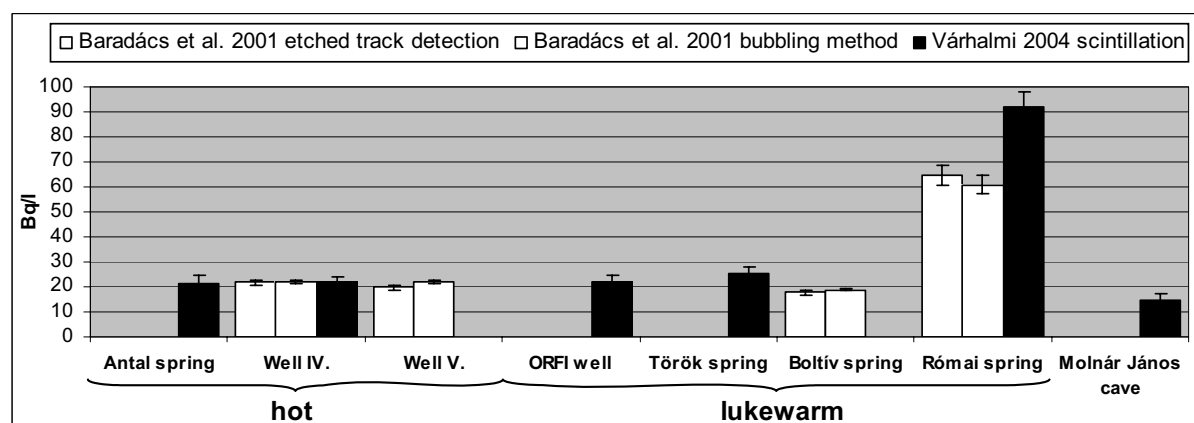


Fig. 3. Radon concentrations of the springs and wells in the József Hill area.

Examining the various published radium data (BARADACS et al. 1999, 2002; Fig. 4.), it can be observed that the radium content of the Gellért Hill wells and springs ranges between 500 and 1000 mBq/l, with a possible local maximum in the Rudas area. On

József Hill, hot waters bear a radium concentration of 200–300 mBq/l, while lukewarm springs show only 50–100 mBq/l radium content. No positive anomaly is found at the Római spring.

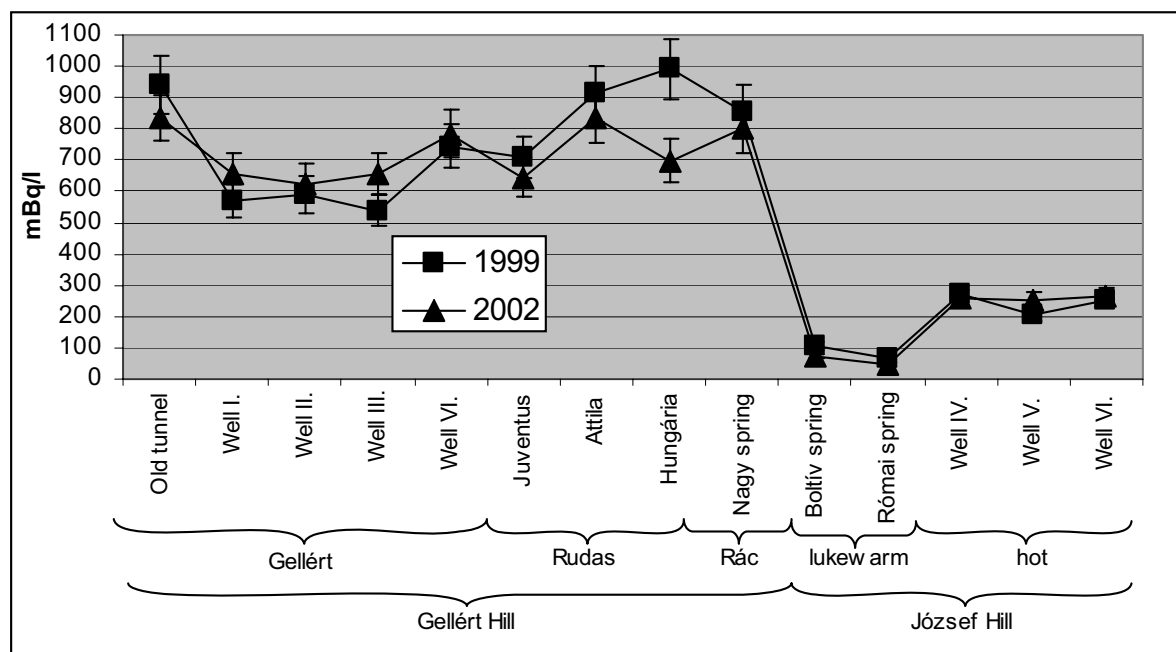


Figure 4. Radium content of Gellért and József Hill springs and wells (after BARADACS *et al.* 1999, 2002).

4. Results

Being a thermal karst with residence times in the magnitude of 100000 years (DEAK 1979), the studied system cannot be simply correlated with cold water karst systems showing rapid water transport, where radon is assumed to be of soil origin, as shown by SURBECK (2005).

Sporadic data for the deeper parts of the basin (Városliget II. well in: KASZTOVSZKY *et al.* 1996, PALOTAI *et al.* 2005) suggest an approximate radium concentration of 500 mBq/l, which, due to its half life of ca. 1600 years, is supposed to decrease during transport. The observed data (Fig. 4.) on József Hill support this theory, while on Gellért Hill a radium source close to the discharge area is needed to explain the measured values.

In the whole Buda Thermal Karst, dissolved radium rarely exceeds 1 Bq/l (Fig. 4.), so only 1-2 % of the observed radon concentrations (Figs 2. and 3.) could be originated from this source. The short half life of radon (3,82 days) indicates a source close to the springs and wells, i.e. in the very discharge area.

On József Hill, hot as well as lukewarm springs discharge from (and wells are filtered in) Upper Eocene – Lower Oligocene limestones and marls (ALFÖLDI *et al.* 1968). Discharge conditions being similar on the whole area, it is not surprising that observed radon values are basically the same in both types of springs: both gain radon originated in the rocks of their discharge areas. The anomalous behaviour of the Római spring needs further investigation and explanation: no special radon source has been yet found in the immediate surroundings of the spring.

The Gellért Hill springs discharge from Upper Triassic shallow marine dolomite (ALFÖLDI *et al.* 1968), supposed to be unable to support the observed radon values (Fig. 2.), which suggest a local, highly efficient source of radon, and also of radium. The nature of this

source can – at this stage of research – only be suspected. The distribution of radon (and radium?) concentrations, with the Rudas area at its centre, suggest a maximum of radon flux at the mentioned area. Radon values may increase in the vicinity of fault zones (e.g. CHOUBEY *et al.* 2001). In our case, the direction of the main fault zone of the hill (also determining geomorphology) (KORPAS *et al.* 2002) would result in similar radon magnitudes along the whole section shown in Fig. 2., making this solution unlikely.

Various solutions could explain this situation. Upper Cretaceous lamprophyres are common in the surroundings of the Buda Hills (HARANGI *et al.* 1996), but no such dyke has been found in the vicinity of Gellért Hill. Suspecting e.g. a buried lamprophyric dyke in the Rudas area might explain the observed phenomena, being a possible source of high amounts of radon (and radium).

Iron and manganese hydroxides produced in mixing zones of descending cold water and ascending thermal karst water make an excellent radon source (MACHE & BAMBERGER 1914, JOB & ZÖTEL 1969, SURBECK 2005). If the radioactivity of the Gellért Hill waters were produced by this mechanism, changes in the mixing rate of thermal and surface waters (induced e.g. by precipitation events) would induce a change in radon concentrations. To prove this hypothesis, spring deposits on Gellért Hill are currently being tested for iron, manganese, uranium and radium content as well as mineralogical structure, this research running parallel to continuous radon measurements in the Rudas area. Infiltration, and so mixing of surface and thermal waters on József Hill is much more limited than on Gellért Hill, making this theory seem more likely in the light of radon data. However, the origin of increased radium values on Gellért Hill cannot be explained this way.

5. Conclusion

Published data and current research were used on radon and radium concentrations in the larger area of Gellért and József Hills (Budapest) to outline the character and localization of their radioactivity. Our geological models (presently only as hypotheses) aim to explain the origin of observed radioactivity of these waters.

The most prominently anomalous area lies at the Gellért Hill, near to the springs of Rudas Spa. Here, ~600 Bq/l radon and ~1000 mBq/l radium concentrations were measured. Radon, and to a lesser extent radium concentrations, decrease towards the neighbouring Gellért and Rác Spas. The radioactivity of the József Hill springs is even a magnitude lower than that of all the other springs mentioned above. Radium concentrations of the lukewarm (20–37°C) springs are about 50–100 mBq/l, while hot springs (37–60°C) show 200–300 mBq/l radium content. Both types have radon concentrations averaging 20–25 Bq/l.

While radium seems to be of deep origin, radon is originated close to the springs, i.e. from discharge area rocks and possibly soils. On József Hill, this theory

perfectly fits measured data, while on Gellért Hill an extra, yet unidentified radioactivity source is needed.

Continuous and batch radon and radium measurements, geological and geochemical studies focusing on the Gellért Hill zone are in progress in order to explain the radioactivity of the waters of the Buda Thermal Karst System. These monitoring data over three months have shown that the time dependence of the radon concentrations are less than 10% in the springs of Rudas spa and that the radon clearly has a localized geological origin.

Current research focuses on two main topics: (1) to conduct continuous radon measurements in the centre of the Gellért Hill radon maximum in order to gain data for variability or stability of the radon concentration and its correlation with other parameters, such as CO₂ concentration, conductivity, total dissolved gas pressure, precipitation etc., and (2) to establish the factional basis for a geological model for the observed radon and radium concentrations, this being accomplished by batch water and rock radon and radium measurements.

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Contaminated dolostone aquifers in Southern Ontario: are they fractured or karstic?

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Abstract

Dolostone aquifers are an important drinking water resource for the highly populated southern Ontario. The region is also heavily industrialized causing these aquifers to be contaminated by point sources (e.g. DNAPL) and non-point sources (e.g. road salt, agriculture) pollutants. The aquifers are generally described as fissured-porous media and karstic properties are not included. In this paper, it is argued that the necessary conditions for the development of karstic features in the dolostone are met and opportunities to produce the karst occurred during the Pleistocene history. However at a later stage, Quaternary deposits on top of the dolostone modified significantly the aquifer boundary conditions. As a consequence, this would have choked the karstic flow regime of the aquifer (i.e. prevent high flow velocity, limit the transience in flow and solute transport).

More recently, municipal water supply wells pumping at high rate may have reactivated rapid flow through karst conduits, which would cause high flow velocities in some areas. This working hypothesis needs to be assessed further because it may have important consequences for groundwater quality and vulnerability assessments. A major difficulty lies in the limited direct observations of karstic features because the dolostone is generally covered by thick Quaternary deposits. Indirect methods are envisioned, including interpretation of large existing borehole datasets including geophysics, flow metering, televIEWing, packer testing, distributions of contaminants and natural tracers, and in the future tracing experiments are planned.

1. Introduction

Dolostone aquifers are common in southern Ontario where much of Canada's industry and a quarter of Canada's population exist and many of the aquifers are contaminated from various types of point and non-point sources. The hydrogeologic characteristics of these aquifers are being investigated within the framework of two alternative conceptual models: (i) fractured aquifer comprised of numerous interconnected fractures and (ii) karst aquifer where groundwater flow is dominated by a single or a few larger solution conduits continuous over large distance. It is expected that these two contrasting conceptual models would result in contaminant plumes with very different shapes and internal characteristics. The specificity of karst aquifers lies in their organised heterogeneity, which can be described as a network of high permeability conduits embedded in a Low Permeability limestone Volume (LPV) (KIRALY 1998). This structure is responsible for a duality in recharge conditions (concentrated in the conduits, diffuse in the LPV), in storage capacity (high in the LPV, low in the conduits), and in flow velocities (high in conduits, low in the LPV).

This study focuses on two of these dolostone aquifers used for municipal water supply and these aquifers are contaminated by dense non-aqueous liquids (DNAPL) point sources (e.g. TCE and metolachlor), and due to non-point sources such as road salt, agriculture and leaks from municipal sewer systems. At these two sites, glacial deposits up to 40 m thick occur with limited regional bedrock outcropping. Therefore the possible karstic nature of the bedrock is difficult to assess because direct observations of karstic features (e.g. swallow-holes, springs, caves, karrenfields) can be made only at a few locations in the general region.

The objectives of this paper are to assess the potential karstic nature of the dolostone aquifers, and to consider whether field data support the fractured versus the karstic conceptual model. Furthermore, additional field experiments are proposed for evaluating which conceptual model is correct.

2. Study sites

The sites are located in two important industrial cities of SW Ontario separated by a distance of 20 km, one in the city of Cambridge (CARTER *et al.* 1995, PARKER *et al. submitted*) and one in the city of Guelph. The area has surficial Quaternary deposits (3 to 50 m), which form a leaky aquitard, overlying the Silurian dolostone aquifers between 90 and 110 m thick (Lockport and Guelph formations). The Rochester shale below constitutes the lower aquitard. The dolostone aquifers are a major drinking water resource for the region. Hydrogeological investigations have been directed at groundwater abstraction and water resource assessment (e.g. LOTOWATER 1997) and at groundwater contamination issues (CARTER *et al.* 1995, TURNER 2001, BURNS 2005, PARKER *et al. submitted*). Aquifer recharge is mostly diffuse through the overburden. Natural discharge is most probably towards the major rivers (Grand and Speed rivers); however pumping wells both in the overburden and the dolostone have drastically changed the natural flow conditions.

At the two sites, point-source contaminant plumes are being investigated in detail using intensely-monitored core holes involving several types of logging and testing methods and the holes are completed with depth-discrete multi-level monitoring systems. Also, insights are gained from the distribution of environmental isotopes including atmospheric tritium, oxygen-18 and deuterium.

3. The karst aquifer hypothesis

Limited bedrock outcrops occur at or near the two sites and therefore direct observations of karstic features are not possible at these sites. However, it is of interest using other approaches to assess whether 1) the necessary conditions required for developing karst conduits are met, 2) if so, what reasonable inferences can be made about the conduit network geometry and what field observations could corroborate the conceptual model.

Theoretical background of karst conduits development

The specificity of karst aquifers develops through the positive feedback relationship between dissolution rate and discharge as described by several authors (e.g. GABROVSEK & DREYBRODT 2001, WORTHINGTON 2003): larger conduits grow at the expense of smaller ones because they accommodate more “carbonate-aggressive” flow. As a result, a hierarchical conduit network develops from recharge areas to a localized discharge area.

Speleogenetic models for limestone caves have been developed in detail over the past years. They show that the genesis of a karst conduit network has two stages: (1) from inception to breakthrough when rapid dissolution becomes active over the entire flow paths; this takes as little as 20'000-30'000 years (DREYBRODT 1996, PALMER 2003); (2) enlargement of conduits, which may vary between 150 cm/1000 yr for point source recharge and 20 cm/1000 yrs for diffuse recharge (PALMER 2003). These models also indicate that an initial continuous fracture aperture of 0.01 cm (100 μ m) is sufficient for developing large scale karstification. For dolostone dissolution, MORSE & ARVIDSON (2002) indicate that little is known: “in contrast to calcite, the attention devoted to laboratory determination of dolomite dissolution kinetics has been quite limited (...)”. Hence the timing of karst development is uncertain. However CHOU *et al.* (1989) showed that the dissolution rate for dolomite at conditions much below dolomite saturation (0.01 SI_{dolomite}) is approximately one order of magnitude lower than for calcite. Additionally, based on experiments conducted in a test vessel with rotating dolomite disks in aqueous carbonated solutions at constant P_{CO2} and temperature, HERMAN & WHITE (1985) showed a sharp drop in the dolomite dissolution rate for solution conditions still far from equilibrium and estimated that saturation will be reached after 1-2 years at least in these experiments. The above discussion is limited to the simplest case where only one solid phase, one gas phase (CO₂) and pure water are present. BISCHOFF *et al.* (1994) showed for natural systems that ionic strength effect or incongruent dissolution of dolomite may enhance solubility. This may have been the case at the Cambridge and Guelph sites since the dolostone there was previously overlain by gypsum and salt layers (the Salina formation). The lower solubility of dolomite as compared to calcite is expected to reduce breakthrough times because groundwater remains aggressive over longer flow distance. Additionally, the rate of conduit enlargement should be slower in dolomite, hence resulting in smaller conduit sizes.

The karst aquifer conceptual model

Development of karst conduit networks should be considered possible in these Ontario dolostone aquifers because: (i) the dolostone was exposed and infiltrated by acidic waters during part of Quaternary time, and (ii) eroded bedrock valleys (EYLES *et al.* 1997, COLE 2005) likely created substantial hydraulic gradients during the same time period.

For the studied sites, we are assessing the schematic karst conceptual model presented in Fig. 1a, which is based on several studies of karst aquifers found in the literature (e.g. WILLIAMS 1983, LEE AND KROTHER 2001, PERRIN *et al.* 2003). This model presents only the karstic features, but the fractures and matrix porosity are other important aquifer properties. Infiltration is diffuse through barren rock or thin glacial deposits and concentrated at some points through swallow-holes fed by glacial melt water. With this diffuse infiltration, an epikarst layer develops. This layer redirects water towards vertical enlarged fractures (typical density of 1/50 m). At the water table, sub-horizontal conduits along bedding planes direct water towards karstic springs at the bottom of the valley. Deep phreatic conduits are unlikely since the bedding dip is limited to 3° and the flow path lengths in this part of Ontario are limited to a few kilometers (WORTHINGTON 2004). The geometry of the sub-horizontal conduit network is most likely of branchwork type (PALMER 2003), controlled by bedding planes and to some extent by joint directions. However the actual location of conduits cannot be determined. The deposition of the Quaternary deposits during Pleistocene time caused cessation of the karstic evolution because it imposed new boundary conditions (i.e. “drowning”). Glaciation impacted to some extent the former land surface and the epikarst by glacial scouring of the shallow rock (FORD 1983, FORD 1987), but deep karstic features should have been preserved.

A proposed conceptual model for the present-state aquifer is presented in Fig. 1b. The karst is drowned by a large water table increase caused by the bedrock valleys filling with Quaternary deposits. The karst will hence behave in a “passive” way. This drowning has several consequences for the dolostone aquifer: (i) reduce the thickness of the vadose zone, (ii) lower natural hydraulic gradients and groundwater flow velocities, (iii) partial confinement, (iv) new imposed recharge type by

downgradient flow through the Quaternary aquitard, (v) new discharge zones, (vi) new flow directions, (vii) reduced dissolution processes (groundwater becoming saturated while circulating in the glacial deposits). If this model is correct, it implies several consequences for the present aquifer properties, as discussed in the next section.

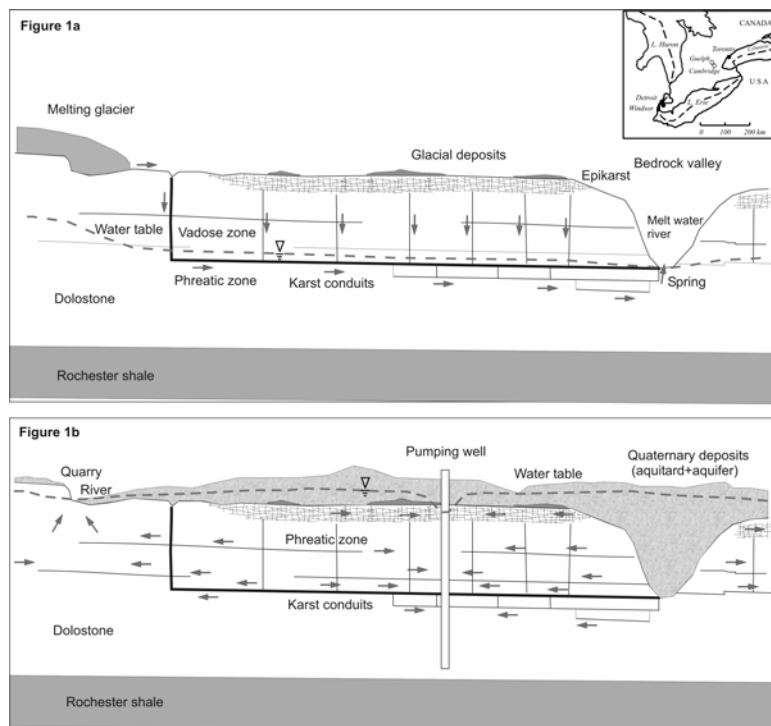


Fig. 1: A proposed model of karst development in Cambridge area. Note the drastic change in the groundwater flow field after the aquifer has been drowned (Fig. 1b).

4. Contaminated site field investigations

Each of the contaminated dolostone aquifers has been investigated previously by consultants (commissioned by the site owners) using conventional monitoring wells. The current studies being conducted by the University of Waterloo rely primarily on additional vertical boreholes continuously cored with contaminant analyses done on rock core samples and many types of complementary data are acquired from these coreholes including borehole geophysical and video logging, high-resolution temperature logging, heat-pulse flow metering, straddle packer testing, and pumping tests. After these data-acquisition activities, depth-discrete multilevel monitoring devices are installed in most of the holes to monitor hydraulic head and acquire groundwater samples for contaminant and other chemical analyses. The borehole studies are directed at identifying fractures intersecting the holes with emphasis on identifying the hydraulically active fractures and their transmissivity and effective hydraulic apertures. The contaminant distributions in each hole are determined from analyses of numerous rock core samples indicating contamination in the rock matrix and the multilevel systems provide contaminant distributions (point source and non-point source) in the fracture network where active flow occurs. Table 1 provides a summary of the various types of data being acquired from the two sites.

The probability of a borehole directly intersecting a karst conduit is small. For example, WORTHINGTON (2003b) estimates that the probability of intersecting a major conduit during drilling is comprised between 0.37 and 7.5 %. This range was obtained by dividing the plan area of a specific cave by the minimum rectangular area containing the cave for 10 well-mapped limestone caves from different karst regions. Therefore, if the boreholes provide no evidence of such conduits, the lack of karst is not proven. However, the three-dimensional distributions of contaminants indicated by the boreholes (from rock core and multilevel systems) are expected to provide evidence for or against the occurrence of karst conduits (Fig. 2). In essence, the 3-D contaminant distributions are solute tracer tests where the solutes (i.e. the contaminants) have been migrating in the dolostone aquifers for a long time (i.e. more than a decade). At the Cambridge site, the 3-D contaminant distributions are strongly influenced by the groundwater flow system imposed by the long-term pumping of municipal water supply wells. At the Guelph site, the nearest municipal wells have been shut down because of contaminations from other industrial sources and the natural groundwater flow system apparently governs the contaminant distribution at the sites.

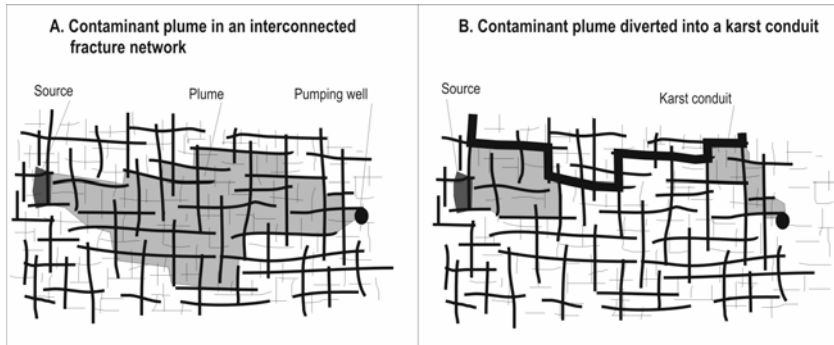


Fig 2. Conceptual model of contaminant plume in the case of A) an interconnected fractured network without karst conduits, B) the same network but with the presence of a karst conduit (horizontal view).

Table 1: Field observations that will help to determine the adequate conceptual model at the sites. Columns 3 and 4 list the expected observations for the two alternative models. The observations limited to boreholes are unlikely to intersect active karst conduits. (*recovery: corresponds to the ratio of recovered core to the total length drilled), (^FLUTETM is a specific technology for borehole multilevel monitoring; the installation rate can be related to the aquifer permeability at different depths)

	Data	Karstic conceptual model	Fractured conceptual model	Sites observations
Local scale at borehole site	Core visual inspection	Karstic features probably not recovered (low recovery* at some locations)	Higher recovery over the whole length of the core	In process
	Driller logs	Drill bit drops, large water loss/gain during drilling	No significant drill bit drops expected, limited water loss/gain	Bit drops indicated in some holes
	Video and acoustic televiewer logs	Presence of dissolution enlarged openings	No large voids expected	Presence of cm-dm openings at some depths
	Packer testing	High K at the top (epikarst), high at discrete intervals deeper	Random but heterogeneous	In agreement with the karstic model for some boreholes
	Hydrophysical logs	Indications of high flow velocity (temperature anomalies, flow metering)	High flow velocity less likely	High velocities observed near the top and at discrete depths (BURNS 2005)
	FLUTE TM K profiles	High K at some depths	Lower K	K reaching 2.0E-3 m/s at two discrete intervals (PLETT 2002)
	Point dilution	High flow velocities at specific depths	Lower flow velocities	No data
Regional scale hydraulics	Pumping tests	Wells with high specific capacity (SpC) and high discharge rates	Lower SpC and discharge rates	In process
	Transmissivity distribution	Highly heterogeneous with some preferential directions	Heterogeneous, a priori no preferential directions	Preferential directions shown by pumping and shutdown tests (LOTOWATER 1997)
	Interconnectivity	From small to large scale (> 1000 m)	Limited to small-medium scale	>1000m observed by LOTOWATER (1997) and BURNS (2005)
Regional scale transport	Non-point source contaminants	Spatially heterogeneous concentrations	More homogeneous concentrations distribution	Spatial heterogeneities observed for major ions and isotopes (PLETT pers.com.)
	Tracer tests	Short transit time (TT), sharp breakthrough curve	More dispersed response, longer TT	No data
	Point source contaminants	Narrowing plume with distance, transience in concentrations expected	Widening plume with distance	Wide plume at local scale
Regional scale geology	Visual observations at quarries/outcrops	Significant karstic features observed in quarries/outcrops	Features non existent	Karstic features in quarries being assessed for continuity
	"Hydro-karstic" features	Presence of springs, swallow-holes	Features non existent	Observations limited by the overburden

5. Discussion and conclusions

The two studied dolostone aquifers meet the necessary conditions for developing karst conduits. However direct observations of karstic features are limited by the presence of glacial deposits. Only indirect methods such as borehole logging and testing, well testing, and contaminant distributions can be used to assess the applicability of the karstic versus fractured conceptual model. Interpretations can be problematic because most tests are scale-limited and may only target the fractured-porous medium close to the borehole. This difficulty has resulted in the lack of agreement between authors concerning the karstic properties of these rocks. The following summarizes the debate in the literature based on studies at three sites in Silurian dolostone/limestone aquifers:

The Smithville site in SW Ontario (Silurian dolostone, Lockport formation) has been described by two different conceptual models: the first describes the aquifer as a porous-fractured aquifer (ZANINI *et al.* 2000) whereas the second includes also a well developed karst conduit network (WORTHINGTON 2002).

In 2000, serious contamination occurred at a municipal well tapping a Silurian limestone aquifer (Bass Island and Bois Blanc formations), Walkerton (SW Ontario): faecal bacteria caused the death of 7 persons and the illness of 2300 others. A first hydrogeological investigation concluded that an equivalent porous medium (EPM) model was adequate to describe the aquifer, leading to estimated flow velocities of 5-10 m/d. However, further studies including tracer experiments showed typical karstic flow velocities of 320-480 m/d (WORTHINGTON *et al.* 2002, WORTHINGTON 2003a).

Under the city of Milwaukee (Wisconsin state, US), deep tunnels for sewage water storage were constructed between 1987 and 1991 in the Silurian dolostone. Preliminary studies did not consider karstification as an issue but fractures enlarged by dissolution and karstic voids were encountered during the boring. This caused major problems in the tunnels such as persistent local rock falls causing serious safety hazards (killing one worker) and large groundwater inflows causing delays and extra-costs (DAY 2004).

At our two study sites, some borehole observations suggest that karstic features are present and well developed. Moreover within a 30 km radius, karstic features on dolostone outcrops have been described by several authors: dissolution caves on the flank of the Eramosa river valley at Rockwood (KUNERT *et al.* 1998), “foxhole-sized cavities and widened joints have created an area pocked with small sinkholes, 3.2 km southwest of Hayesland” (KARROW 1987) and about 20 km from Cambridge. Additionally, COLE (2005) describes dissolution features from borehole televising data at Guelph. However these aquifers cannot be treated as classical active karst aquifers since they have been choked by low to moderate permeability sediments laid down at the end of the last glacial period, resulting in a water table rise and a drowning of the dolostone. This drowning has reduced flow velocities and changed flow paths. Recently the implementation of pumping wells may have reactivated some of the rapid flow paths.

The presence or absence of karst conduits at the study sites is being assessed by several additional investigations:

Spatial and temporal variability of natural tracers, point source contaminants, and non-point source contaminants.

Large scale (>100 m) tracing tests for assessing flow velocities on inferred high permeability zones.

Specific pumping tests to better determine the directional heterogeneities in the hydraulic conductivity field.

Spatial correlation of different data sets such as acoustic televising and video observations, packer test hydraulic conductivity, flow metering, temperature logging.

Numerical modeling to assess the impact of the boundary conditions imposed by the Quaternary deposits and to develop further the concept of “drowned karst aquifer”.

More detailed understanding of the dolomite dissolution kinetics.

Assessment of geophysical information concerning the buried bedrock valleys topography and depths.

In conclusions, it is worth saying that even if the karstic properties of the rock mass (dissolution features, conduits) can be proven; this does not necessarily imply that the aquifer behaves as a karst aquifer because present boundary conditions may have changed the groundwater flow field. If these aquifers are considered karstic, the expected practical consequences are unexpectedly high flow velocities, very specific point-source plume shape and characteristics, and non-darcian flow in the conduits.

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Hydrogeological behavior of carbonate aquifers in Southern Italy: a preliminary conceptual model for the Acqua dei faggi test site

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Abstract

A research is carrying out in the Acqua dei faggi test site (Longano, Isernia, Southern Italy) in order to analyze the hydrogeological behaviour of the carbonate aquifer and propose a conceptual flow model. The research is based on hydrogeological and geophysical investigations.

The results suggest that the continuum approach can be applied to describe the flow on a scale ranging from metric to kilometric. Hence, on these scales the fracture spacing is sufficiently dense that the fractured medium acts in a hydraulically similar fashion to granular porous media. A discontinuous heterogeneity is caused by the presence of faults, which produced a large contrast in hydraulic conductivity. These faults often represent low-flow boundaries within a single aquifer system, which looks like a series of interconnected basins. The contrast in hydraulic conductivity can cause a partial discharge of groundwater (seasonal springs) and a concentrated head loss (higher hydraulic gradients). A layered heterogeneity results in hydraulic conductivity contrasts within the first tens of meters from the ground, depending on the increase in fracture density and/or in diffuse karstification towards the ground surface. Hence, the system as a whole acts like an anisotropic medium.

1. Introduction

The carbonate aquifers provide the main drinking water resources of Southern Italy, supplying an average volume of about $4 \cdot 10^9 \text{ m}^3 \text{ y}^{-1}$ (CELICO, 1983). The current conceptual model which represents the hydrogeological behavior of these aquifers can be summarized as follow (CELICO, 1986). The carbonate rocks have a very low primary permeability at core scale (10^{-9} to $10^{-12} \text{ m s}^{-1}$; MONJOIE, 1975) and the aquifer systems are permeable at medium-large scales, due to fracture and/or karst networks. The existence of higher karstified limestone rocks in the uppermost portion of carbonate aquifers was hypothesized, based on results of Lugeon tests in a site (CELICO *et al.*, 1986), but nothing is known concerning its distribution within the Apennines and its hydrogeologic behavior. The main springs generally occur at the contact with Miocene rocks or volcanic, marine and continental Pliocene-Quaternary deposits, which represent relative aquicludes. The hydraulic gradient usually ranges from $4 \cdot 10^{-3}$ to $4 \cdot 10^{-2}$ (CELICO, 1983; CELICO, 2002). Hydrogeologic investigations that were carried out mainly at regional or catchment scale suggested that fault zones sometimes act as barriers that impede groundwater flow.

The present paper describes the preliminary results of an interdisciplinary approach which is carrying out in the experimental site of Acqua dei faggi (Longano – Molise), in order to refine the existing conceptual model and support it with detailed investigations at site scale. The test site is mainly made of limestones. The rocks have very low primary permeability but are extensively fractured and subordinately karstified (CELICO *et al.*, 2006).

2. Data and methodology

The groundwater level was monitored hourly in 1 well and 3 piezometers, from October 2003 to December 2005. The springs discharge was monitored weekly, from October 2003 to December 2005, by means of a current meter. A pumping test was carried out in order to analyze the interconnection of fractures within the limestone rocks. Lugeon tests were carried out in order to calculate the hydraulic conductivity of the carbonate medium. Four Schlumberger vertical soundings (SEV) have been performed in the site test with the objective to reconstruct the shallow electrostratigraphy of the area.

3. Results

The piezometric surface slopes westwards, to different springs located along some faults, at different altitudes. The lowest spring is perennial, while the highest ones are seasonal. The hydraulic gradient ranges from $3.5 \cdot 10^{-2}$, within the protolith, to $6.6 \cdot 10^{-1}$, within the core of some faults (Fig. 1; CELICO *et al.*, 2006). A nervous hydraulic response can be observed in all piezometers and at springs (Fig. 2). Nevertheless, despite the nervous behavior, the fluctuations are simultaneous throughout the study site (CELICO *et al.*, 2006). The amplitude and the modalities of groundwater fluctuations significantly change with depth (Fig. 2). In recharge, when precipitation causes effective infiltration, the groundwater level rise has a different size in the deeper and in the upper portions of the aquifer that was investigated, even though no significant differences were detected in terms of amount and frequency of rainfall. During the recession phase, the potentiometric head falls with a velocity which changes with depth. For example (Fig. 3), in a piezometer it is always slightly lower than 0.09 m d^{-1} above 1004 meters a.s.l.. The rate ranges from 0.17 and 0.18 m d^{-1} between 998 and 1004 meters a.s.l., while it is always higher than 0.6 m d^{-1} below 998 meters a.s.l. Taking into consideration the progressive decrease in discharge of the spring during recession, this increase in velocity suggests a decrease in effective porosity of the carbonate medium with depth. The lower interconnected void space can also explain the higher groundwater level rises that were observed, in recharge, in the deeper portions of the bedrock. The high effective porosity in the uppermost portion of the carbonate substratum was confirmed by the detection of karst conduits within the first 4 meters of limestones below the topsoil.

The Lugeon tests suggest an hydraulic conductivity in a fault zone lower than $2.9 \cdot 10^{-6} \text{ m s}^{-1}$ (CELICO *et al.*, 2006). These values are slightly higher than those (average hydraulic conductivity of $7.5 \cdot 10^{-7} \text{ m s}^{-1}$) determined by carrying out Lugeon tests in the flysch deposits which represent the regional aquiclude of several carbonate aquifers of Southern Apennines (CELICO *et al.*, 2006). On the contrary, in the uppermost portion of the bedrock in the test site, the permeability of rock mass was too high to determine the hydraulic conductivity. The rock mass absorbed $40 \text{ l min}^{-1} \text{ m}^{-1}$ and the water acceptance exceeded capacity of the test system. No pressure was recorded on the manometer.

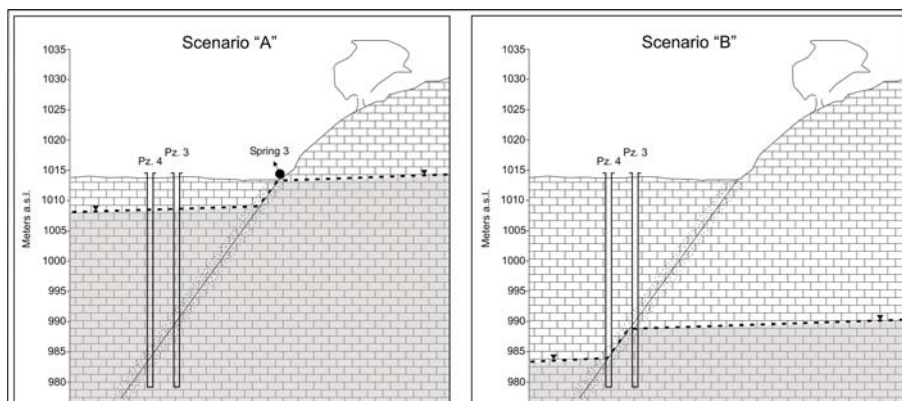


Fig. 1 - Evolution of the hydrogeological setting in a fault zone over time (Scenario “A” - hydraulic gradient monitored in the protolith; Scenario “B” – hydraulic gradient monitored in the fault core) (after CELICO *et al.*, 2006).

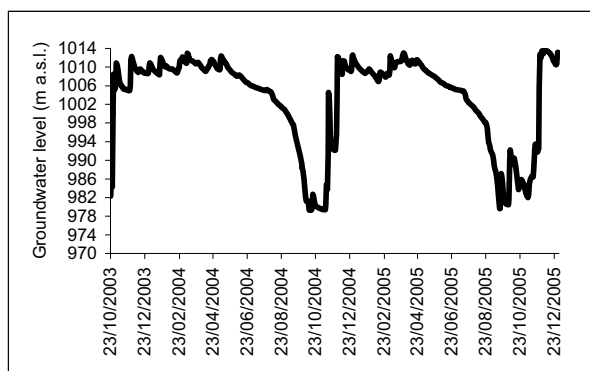


Fig. 2 – Example of groundwater level fluctuations in a piezometer.

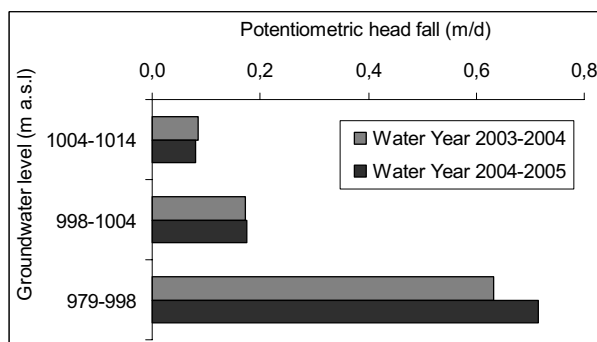


Fig. 3. Piezometric head fall in the test area.

The pumping caused 1.25 meters of drawdown in the well and 0.52 meters in the piezometer. Water level fluctuations in the piezometer in response to pumping show that the limestone aquifer is laterally and vertically well connected in the subsurface. It was calculated a transmissivity of $2.0 \cdot 10^{-3} \text{ m}^2 \text{ s}^{-1}$, and a storage coefficient of $2.3 \cdot 10^{-3}$.

The electrical soundings have been modelled using a simple 1D terrain structure. SEV-1 located in the centre of the valley is characterized by a 5 m thick coverage soil with $\rho = 33 \text{ } \Omega \cdot \text{m}$, below which resistivity rise to $565 \text{ } \Omega \cdot \text{m}$. The orthogonal SEV-2 showing the same pattern indicates the absence of significant shallow lateral heterogeneity. SEV-3 located north of the SEV-1 shows a more complex resistivity

distribution in the top-soil, with the presence of a 5 m deep electrical discontinuity characterized by the same resistivity ($565 \Omega \cdot m$) shown by SEV-1. Similar complexity in the top-soil is shown by SEV-4 located south of the SEV-1 but, underneath the coverage, a resistive layer is detected with a resistivity of $565 \Omega \cdot m$ in the first 2-3 m, below which, the resistivity increase to $1200 \Omega \cdot m$. The 5 m deep electrical discontinuity existing in whole area is consistent with borehole information and the inferred resistivity of the higher limestones is a strong indication for pervasive karstification phenomena (BOSÁK AND BENEŠ, 2003), according to the high effective porosity detected by aquifer recession characteristics.

4. Conclusion

The results suggest that the continuum approach can be applied to describe the flow on a scale ranging from metric to kilometeric. A discontinuous heterogeneity is caused by the presence of faults, that produce a significant contrast in hydraulic conductivity and a local significant increase of the hydraulic gradient. These faults often represent low-flow boundaries within a single aquifer system, which looks like a series of interconnected basins, with perennial and seasonal springs at different altitudes (CELICO *et al.*, 2006). A layered heterogeneity results in hydraulic conductivity contrasts within the first tens of meters from the ground, depending on the increase in fracture density and/or in diffuse karstification towards the ground surface. The electrical resistivity obtained by SEV soundings confirms this layered heterogeneities, showing resistivity in the uppermost limestones compatible with high karstification.

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Monitoring of organic carbon, natural particles and bacteria in a deep karst system, Yverdon-les-Bains, Switzerland

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Abstract

The Yverdon-les-Bains karst system was investigated in order to characterise the transport and interaction of organic carbon (OC), natural particles and bacteria. These parameters were monitored at a swallow hole draining an agricultural plain (input) and at two main groups of springs (outputs). Several tracer tests were carried out from the swallow hole in order to study its influence on the springs during different hydrologic conditions, but also to study particle and solute transport in a deep karst system. Transit times varied from 2 days during high-flow conditions to 13 days during low-flow conditions. Monitoring the dynamics of natural parameters made it possible to differentiate two types of turbidity at the springs: a primary turbidity signal (autochthonous) resulting from the remobilisation of particles inside the karst conduits due to a discharge increase, and a secondary turbidity signal (allochthonous) indicating the arrival of surface water sinking into the swallow hole, which is often microbiologically contaminated. These results suggest correlation between the arrival of faecal bacteria, OC and the secondary, allochthonous turbidity at karst springs. The two types of turbidity cannot always be clearly distinguished and often overlap during high-flow conditions. However, preliminary continuous measurements of particle size distribution (PSD) show clear differences between the primary and secondary turbidity signals.

Résumé

L'interaction et le transfert du carbone organique (CO), des particules naturelles ainsi que des bactéries ont été étudiés dans le système karstique d'Yverdon-les-Bains. Ces paramètres ont été mesurés en continu à une perte drainant une plaine agricole (input), ainsi qu'à deux principaux groupes sourciers (outputs). Des essais de traçages ont été réalisés depuis la perte afin d'étudier d'une part, son impact sur les sources lors de différents régimes hydrauliques et d'autre part, le transport de traceurs solubles et particuliers dans un système karstique profond. Le suivi en continu de la dynamique des paramètres naturels a permis de différencier clairement deux types de turbidité aux sources karstiques : une turbidité primaire (autochtone), typique pour la remobilisation des particules dans les conduits karstiques due à l'augmentation du débit, et une turbidité secondaire (allochtone) qui est caractéristique de l'arrivée de la composante fréquemment contaminée de la perte. Ces résultats présentent une bonne corrélation entre l'arrivée des bactéries fécales, du CO ainsi que de la turbidité secondaire allochtone aux sources. Les deux formes de turbidité ne sont pas toujours clairement distinguables puisque en conditions de hautes eaux, elles se chevauchent fréquemment. Cependant, les résultats préliminaires des mesures en continu de la distribution des tailles des particules (PSD) présentent des signaux spécifiques pour les deux types de turbidité.

1. Introduction

Karst aquifers are highly vulnerable to contamination (ZWAHLEN 2004), and pathogenic microorganisms represent the most frequent and problematic contaminants in such aquifers (AUCKENTHALER *et al.* 2002, FORD 1999). A secure use of karst spring water for drinking thus requires detailed knowledge of the transport processes of microorganisms, their interaction with the karst environment and the hydrodynamic setting of the karst aquifer itself.

Faecal bacteria (e.g. *E. coli*) are often used as indicators for the possible presence of microbial pathogens. However, the detection of such microorganisms is time consuming and laborious (MIKELL *et al.* 1996). Therefore, several studies aimed at identifying easy-to-measure hydrological and physicochemical parameters that correlate with microbial contamination. Some authors suggest that turbidity indicates the presence of pathogens (NEBBACHE *et al.* 1997, RYAN & MEIMAN 1996), while others use spring discharge (AUCKENTHALER *et al.* 2002). Pathogens in aquifers are generally allochthonous (GOLDSCHIEDER *et al.* in press), while turbidity may either be allochthonous or result from the remobilisation of sediments inside the karst conduits (autochthonous) (AMRAOUI *et al.* 2003, BOUCHAOUI *et al.* 2002). PRONK *et al.* (in press) observed increasing contents of faecal bacteria in the spring water while

discharge remained stable. Therefore, turbidity and discharge are not always reliable indicators of bacterial contamination.

The main objective of this study is to characterise the complex interaction and transport processes of organic carbon (OC), natural particles and bacteria in a deep karst system. Since summer 2003, these parameters as well as other physicochemical parameters are being monitored at the Yverdon-les-Bains karst system.

2. The Yverdon-les-Bains test site

Geological and hydrogeological framework

The test site is situated near the southwestern end of Lake Neuchâtel (Switzerland), between two major landscape and geological units (Fig. 1): the Jura Mountains, which rise to altitudes of 1500 m in this region, and the Molasse basin (Swiss plateau). The Jura Mountains consist of Jurassic and Cretaceous limestones and marls. The Malm limestones (Upper Jurassic) represent the main aquifer with a thickness up to 400 m; two other karstified formations are present in the Cretaceous series (Valanginian and Hauterivian).

On the Swiss plateau, impervious Molasse sediments (marl and sandstone) confine the karst aquifer. A complex arrangement of late-alpine faults and folds gave rise to two

hills (SOMMARUGA 1996) where the aquifer locally outcrops, forming two hydrogeological windows.

The Feurtille swallow hole is located at the western hill near the contact between the Molasse and the Malm limestone. The stream sinking into this swallow hole drains an agricultural plain and is often contaminated with high contents of organic matter, faecal bacteria and nitrate.

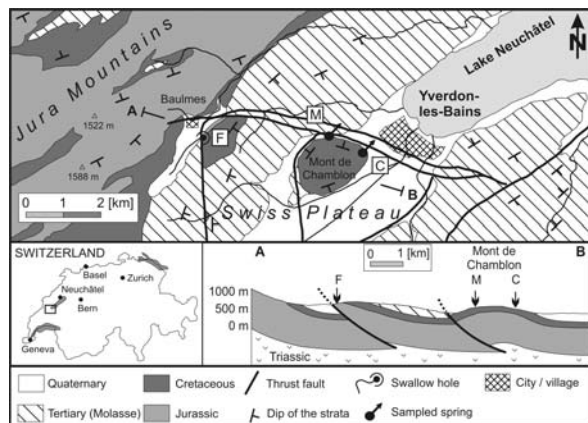


Fig. 1: Location and simplified geologic map and section of the Yverdon-les-Bains area (F = Feurtille swallow hole, M = Moulinet springs, C = Cossaux spring).

The Moulinet and Cossaux springs are situated at the foot of the eastern hill (Chamblon) at the contact between Molasse sediments and Cretaceous limestone. Previous investigations showed that the spring water consists of four components: thermal groundwater from several hundreds of metres depth; cold and slightly contaminated water from the Malm karst aquifer; cold and frequently contaminated water from the Feurtille swallow hole; and local groundwater from the Chamblon hill (MURALT 1999).

The Cossaux and Moulinet springs show significant differences in their physical, chemical and microbiological water compositions. However, the temporal variations of these parameters at the two groups of springs are very similar. Lower contamination levels and 2 °C higher temperatures occur at the Cossaux springs, which can be explained by a larger contribution of the thermal water component (MURALT 1999, PRONK *et al.* in press).

Tracer tests

Already at the end of the 19th century, a connection between the swallow hole and the springs was suspected. Since then, several tracer tests have been carried out (unpublished reports by SCHARDT 1910, 1920 and LOOSER 1990) and revealed transit times between 2 days during high-flow and 12 days during low-flow conditions.

In order to better characterise the hydrogeological relation between the swallow hole and the springs and to simulate particle and solute transport, three new tracer tests were carried out during low-, medium- and high-flow conditions. In order to easily compare the tracer tests, uranine was used as tracer for all the tests. For the medium-flow conditions, fluorescent 1 µm and 2 µm spheres were also injected.

The results are summarised in Table 1. During low- and medium-flow conditions, the tracer first arrived at the more distant Cossaux springs; during high-flow conditions this relation is inverted. The sum of the tracer recovery rates at the two springs remained constant at about 29 %.

Based on the tracer test results obtained during low-flow conditions, and using water and mass balance equations, PRONK *et al.* (in press) proposed a conceptual model of the

conduit network and calculated the groundwater flow rates in the system. Similar calculations have also been made for the results obtained during medium- and high-flow conditions. On this basis, the variations of the tracer recovery rates at the springs during different hydrological conditions can be explained by variable contributions of the four different flow components mentioned above.

Table 1: Results of the three tracer tests (t_1 = first detection, Q = discharge, R = recovery rate).

Hydrologic conditions	Cossaux springs			Moulinet springs		
	t_1 [h]	Q [L/s]	R [%]	t_1 [h]	Q [L/s]	R [%]
Low-flow	260.0	43	12.3	292.0	19	16.8
Med.-flow	81.9	52	3.2	86.9	120	25.3
High-flow	43.5	61	1.6	40.3	320	27.0

The results of the comparative particle-solute tracer test are shown in Fig. 2. The 1-µm and 2-µm spheres were first detected at the Moulinet springs after 85.4 and 94.4 hours respectively, while uranine first arrived after 86.9 hours (for details see GOLDSCHIEDER *et al.* in this volume).

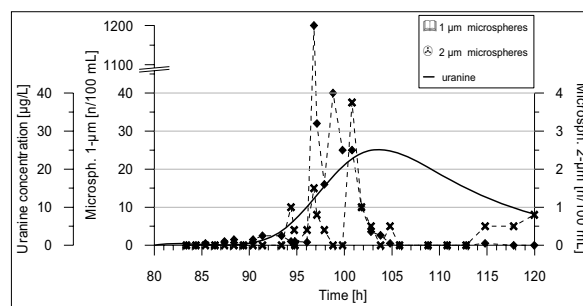


Fig. 2: Microspheres and uranine breakthrough curves at the Moulinet springs (analyses of the microspheres: N. GÖPPERT).

The maximum concentrations of the microspheres occurred before the uranine maximum. This behaviour of particle versus solute tracers has often been observed in porous media. It is generally explained by exclusion mechanisms, which can be further subdivided into charge, size and pore exclusion (GINN *et al.* 2002). Similar processes are expected to occur also in fissured and karst aquifers, although further studies are required to better understand particle transport in such environments.

3. Dynamics of natural parameters

Parameters and methods

Total organic carbon (TOC), turbidity and other physicochemical parameters were continuously monitored at the swallow hole and the two springs. Samples for chemical and microbiological analyses were taken manually every two weeks during more than a year (long-term monitoring). During selected hydrological events, a particle counter (KLOTZ Abacus mobil fluid) was installed at the Moulinet spring in order to measure the particle size distribution (PSD), and all sites were sampled hourly to daily for microbiology and chemistry (event-based monitoring).

Results

The monitoring of the Yverdon karst system started in summer 2003. The data obtained in 2003 already allowed some general conclusions to be drawn on the processes governing the behaviour of the springs: The dynamics of

discharge and physical parameters are highly related to the rainfall in the Jura Mountains, while the chemical and microbiological quality is impacted by the Feurtille swallow hole (PRONK *et al.* in press). As described above, the

temporal evolution of natural parameters at the two groups of springs are very similar. Therefore, only the data of the Moulinet springs are discussed here.

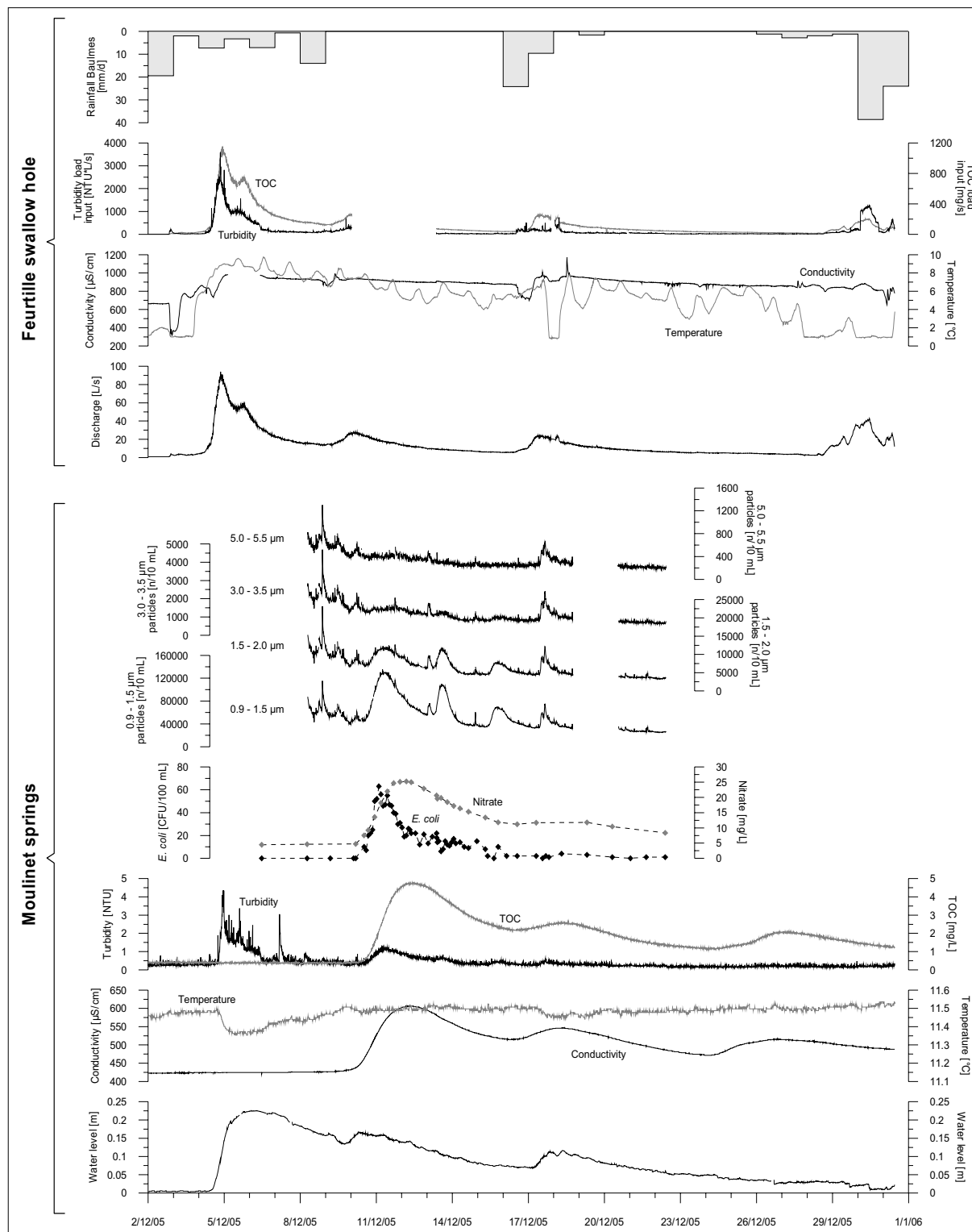


Fig. 3: Dynamics of natural parameters at the Feurtille swallow hole and the Moulinet springs. TOC and turbidity at the swallow hole are presented as load input, which is obtained by multiplying concentration and discharge.

A more detailed monitoring program started in 2005. Autumn 2005 was amongst the driest ever recorded in Switzerland. Low-flow conditions lasted from June to end

December, although a catastrophic flood event occurred in other parts of Switzerland. Nevertheless, some rainfall

events in the area generated minor discharge increases at the observed springs. One example is described here.

Beginning December 2005, an increase of the temperature causing snowmelt as well as a rainy period occurred in the area. A second rainfall occurred mid-December. Between the 4th and 6th of December, the discharge of the sinking stream rapidly increased from 1 L/s to more than 90 L/s, accompanied by high contents of turbidity and TOC. The conductivity increased from 600 to over 1000 $\mu\text{S}/\text{cm}$ (Fig. 3).

At the Moulinet springs, the discharge increased at the 4th of December, accompanied by a temperature decrease and a turbidity pulse, which can be explained by remobilisation of sediments in the conduit network. Conductivity, TOC, nitrate and *E. coli* remained stable until the 10th of December. After that day, a synchronous increase of these parameters was observed, as well as a slight increase of secondary turbidity. These variations can be explained by the arrival of storm-derived water, which entered the Feurtille swallow hole between the 4th and 6th of December. The time lag between the discharge increase and the arrival of contaminated swallow hole water at the spring is 5 to 6 days, which falls in the range of the tracer tests results. The maximums of the secondary turbidity and *E. coli* signals occurred about 1 day before the maximums of TOC, conductivity and nitrate. This finding is consistent with the earlier arrival of microspheres during the comparative tracer test.

The particle-size distribution (PSD) was also monitored during the arrival of the swallow hole component at the spring (Fig. 3). The small particle-size classes (0.9–1.5 μm and 1.5–2.0 μm) showed relatively smooth curves, synchronous and parallel to the *E. coli* and secondary turbidity signals, whereas the larger particle-size classes decreased continuously during the same period. The small increase of discharge observed between the 17th and 18th of December was also accompanied by a small primary turbidity peak. This increase of autochthonous turbidity is clearly visible on the particle graphs for all size classes.

4. Conclusions

The monitoring of natural parameters made it possible to better understand the dynamics and interaction of organic carbon, natural particles and bacteria in the Yverdon-les-Bains karst aquifer system.

During and following precipitation events, two types of turbidity signals can be observed at the springs. The primary or autochthonous turbidity coincides with increasing discharge and can be explained by the remobilisation of sediments in the karst conduits. This type of turbidity is characterised by high contents of all particle-size classes. The secondary or allochthonous turbidity signal indicates the arrival of storm-derived water that entered the swallow hole. Secondary turbidity is characterised by increases of the smaller particle-size classes only. During high-flow conditions, due to the short transit time between the swallow hole and the springs, the two turbidity signals often overlap. However, PSD monitoring allows autochthonous and allochthonous turbidity to be clearly differentiated.

The first arrival of faecal bacteria, which originate from the land surface (allochthonous), is synchronous to the increases of other typical allochthonous physicochemical parameters, such as TOC, nitrate and secondary turbidity. However, while the *E. coli* and the secondary turbidity curves are synchronous and parallel, TOC and nitrate curves show a positive time shift of the maximum concentrations,

as well as a more important tailing. Similar results were also obtained with the comparative tracer test (see GOLDSCHIEDER *et al.* in this volume).

Finally, this study suggests that discharge and turbidity are not reliable indicators for bacterial contamination in karst spring waters, whereas continuous PSD and TOC measurements can be recommended for water quality monitoring.

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Minimizing the effects of overexploitation in karst aquifers in dry years.

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Abstract

Since three decades, water from Iranian karst springs has been extensively used for irrigation in the spring, summer, and early fall. Winter is cold and rainy enough to minimize water demand. Winter crop is cultivated in fall based on predictions of an average water year. In a dry year, discharge from karst springs decrease significantly in the spring, leaving insufficient water for winter and spring crops. Overexploitation by pumping wells is unavoidable in a dry year, especially for providing drinking water and for irrigation of permanent trees. The pumping wells are usually constructed upstream of karst springs reducing the discharges of karst springs in subsequent months. The objective of this study was to propose a method to minimize the effect of overexploitation on the discharge of karst springs

The group of Atashkadeh karstic springs, with an average flow rate of 1700 l/s was selected as a case study. Wheat and barley were cultivated in the fall in anticipation of an average water year, but due to low precipitation in the following winter, the discharge of Atashkadeh springs was not enough to supply the water demand for the winter crops in the spring season and the next year's summer. The discharge of Atashkadeh Springs was predicted using the Double Continuum Porous Equivalent model (DCPE) under several alternative overexploitation conditions. An optimum combination of exploitation regime and a summer crop pattern were selected to limit the reduction of the karst spring flow rate up to the subsequent months of low water demand and earn the maximum net benefit.

1. Introduction

Karst formations outcrop in 11% of Iran's land, being more concentrated, up to 23%, in the south of the country. Karst water is used extensively as drinking water and for irrigation of agricultural lands in Iran. Precipitation occurs mostly during late fall, winter, and early spring. Water demand for agricultural purposes is mainly during the dry period. The water of karst springs decreases during a dry year to an extent that necessitates the reduction of agricultural farmland area in order to counterbalance any shortage of water during the high demand months, especially in May, June, July, and August. This shortage of water may be compensated by pumping karst water from the catchment area of the karst springs. The pumping of water reduces the discharges of karst springs. In this study, the Atashkadeh aquifer is simulated during a dry year under various alternatives of water pumping conditions. The alternative that causes the reduction of karst spring flow to occur during the wet season with the highest net benefit will be selected as the optimum scheme.

The study area is a part of the Podenow Anticline in central-south Iran, located in the Zagros Simply Folded Zone. KARIMI *et al.* (2005) studied the hydrogeology of the area (Figure. 1). The core of the Podenow Anticline is composed of the limestone Asmari Formation that is sandwiched between the two impermeable Pabdeh-Gurpi (marl, shale, and marly limestone) and Razak (silty marl to silty limestone) Formations. The thickness of the Asmari Formation in the study area is about 400 m and its contact with the Razak Formation is transitional. The study area is limited to the catchment area of the Atashkadeh springs (No. 1 on Figure 1). The Atashkadeh springs consists of 19 individual springs, emerging on the southern flank through the Transition Zone within an area 700 m by 700 m. The total discharge of the Atashkadeh springs is 1700 l/s. The catchment area of the study area is estimated to be 114 km², using hydrogeological and geological methods (KARIMI *et al.* 2005; ASADI 1998; TALAEI 2000). This catchment area is composed of mainly Asmari Formations and thin Transition Zone, bounding by the Razak Formation on the foot of the northern and southern flanks. The western and eastern boundaries are limited to the water divide of the adjacent springs. Firoozabad River flows through the U-shaped Tangab Valley. with a maximum depth of 100 m in a saddle-shaped region. The Tangab Dam is currently under construction at the beginning of the Tangab Valley. Karst water of the northern flank flows toward the southern flank in the saddle region (Figure 1). The most important karst features are dry valleys, caves, karrens, grikes, and springs.

Nine piezometers were constructed on both sides of Firoozabad River. Hydraulic conductivities of five-meter sections were measured by the Lugeon method in all the nine piezometers (ABNIROO CONSULTING ENGINEERS, 1992). The discharge, major ions, pH, specific electrical conductance, and temperature of all the springs were measured once every two to three weeks from April 1996 to September 1997. The discharge, specific electrical conductance, and temperature of these springs were measured daily during the wet season and every two or three weeks during the dry season (KARIMI *et al.* 2005). SAUTER (1992) and MOHRLOK & TEUTSCH (1997) have simulated karst aquifers using the Double Continuum Porous Equivalent (DCPE) model. NADRI (1999) calibrated the MODFLOW version of the DCPE model for the study area. The hydraulic parameters, storage coefficients, and exchange coefficients were calibrated to obtain the best fit between the measured and simulated values of the total discharge and base flow of springs, and the water level in the piezometers.

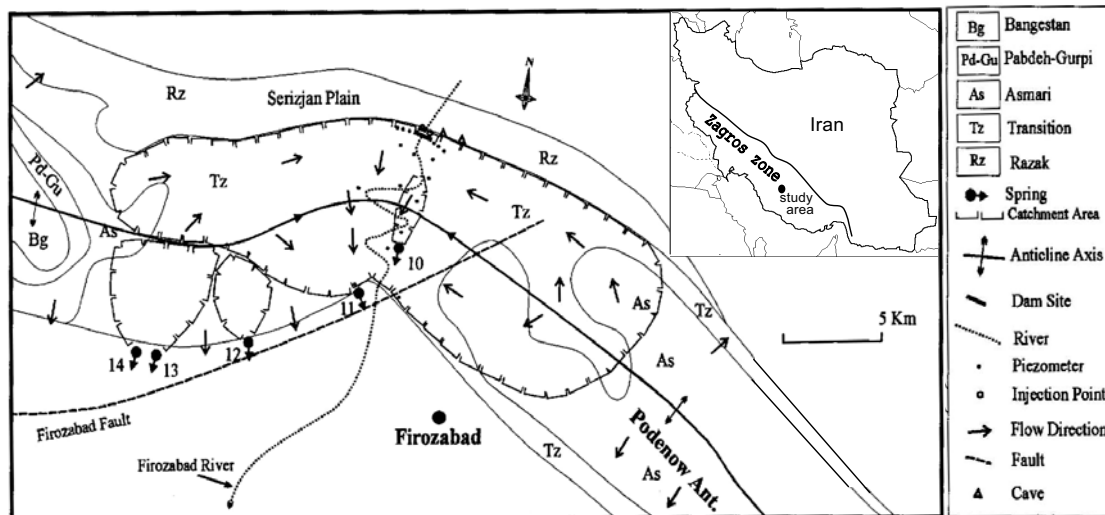


Figure 3. The probable catchment area of some of the karst springs in Podenow Anticline.

2. Material and Methods

Water from the Atashkadeh Springs is used to irrigate wheat, barley, corn, and rice, the main traditional crops in the study area. Wheat, barley, corn, and rice must be irrigated from November to June, November to May, June to October, and May to October respectively. ABNIROO CONSULTING ENGINEERINGS (1998) calculated the monthly water requirement, crop yield, and economic parameters of production such as cost of water, cost of production and net benefits per hectare for each crop. Three alternatives were selected to irrigate the land downstream the Atashkadeh Springs. In the first alternative (Alternative A), a year with average discharge of Atashkadeh Springs (1994-1995) was used to supply the monthly water demand of the four above-mentioned crops. Linear program was applied to determine crop patterns with the maximum net benefit. The monthly water demand being less than the monthly flow rate of Atashkadeh Springs was the main constraint. Alternative B was related to the dry year of 1996 – 1997. Wheat and barley were cultivated in early fall of 1996, in an area equal to Alternative A, based on predictions of an average water year, but the low amounts of winter precipitation reduced the springs' discharge to such an extent that water supply was insufficient in May, June, July, and August of 1997. The water shortage could be compensated by drilling several wells in the catchment area of the Atashkadeh Springs. Several pumping wells, located 7 km away from these springs were incorporated into the DCPE model and the Atashkadeh Springs' discharge was predicted considering the overexploitation of water by the pumping wells. A linear program was applied to determine the possible irrigation area for corn and rice under these new conditions, including the extra pumping cost. Alternative C was similar to Alternative B except that the traditional crop pattern of the study area was changed by adding melon to crops, because the rice water demand is very high (especially for a dry year).

3. Results

The Atashkadeh Springs' discharge and total water demand for Alternative A (average year 1994-1995) are presented in Figure 2. The linear program showed that the crop pattern of 932 ha wheat, 463 ha barley, 210 ha corn and 210 ha rice results in the maximum net benefit (Table 1). It was assumed that the 932 ha of wheat and 463 ha of barley were cultivated in December 1995, predicting an average year. Low precipitation reduced the Atashkadeh Springs' discharge significantly such that it could not supply the water demand of Alternative A (Figure 3), not even the 1119 l/s water demand for wheat and barely of May 1996. The exploitation of water by pumping wells in Alternative B reduced the Atashkadeh Springs' discharge, but the total water supply by the pumping wells and the Atashkadeh Springs' discharge were enough to compensate the water shortage for wheat and barley and the cultivation of 166 ha (Table 1) rice and 166 ha corn (Figure 4). The sources of water of pumping wells were partly provided by the aquifer storage and partly from the reduction of Atashkadeh Springs' discharge. The water harvesting of the aquifer storage resulted in lower hydraulic gradient and consequently the reduction of Atashkadeh Springs' discharge in the subsequent months. The decreasing of Atashkadeh Springs' discharge stopped in November, leaving no water shortage problems in fall (Figure 4). The changing of the traditional crop pattern and adding melon cultivation in Alternative C resulted in a higher income in a dry year (Table 1). The Pumping rates were 434, 91, 105, and 231 l/s in May, June, July, and August, respectively, but the declining of Atashkadeh Springs' discharge ended in November (Figure 5).

Table 1. Crop pattern and net benefit of Alternative A, B and C (The net benefit unit is 1000\$)

Crop	Area (ha) A	Area (ha) B	Area (ha) C
Wheat	932	932	932
Barly	463	463	463
Corn	210	166	180
Rice	210	166	90
Melon	-----	-----	410
Land Area	1395	1395	1395
Cultivated area	1815	1714	2075
Net Benefit	4440	3828	5331

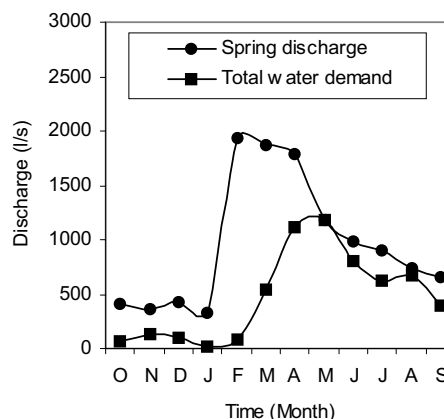


Figure 2. Springs' discharge and total water demand for Alternative A (1994-1995)

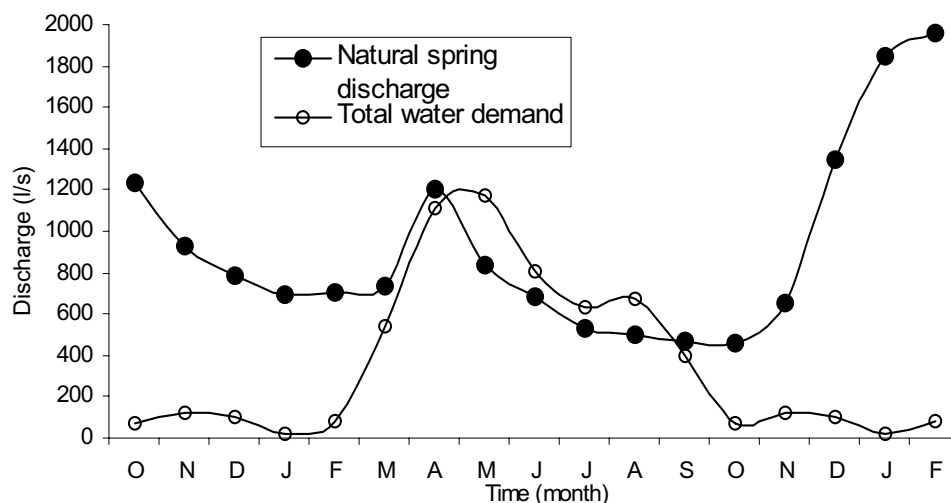


Figure 3. Natural springs' discharge of the dry year 1995-1996 and the total water demand for average water year of 1994-1995.

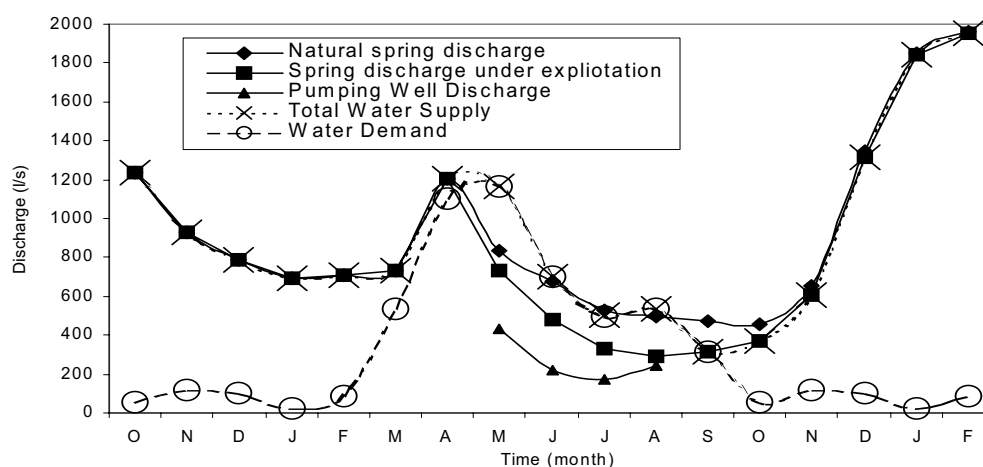


Figure 4. Natural springs' discharge, pumping well discharge, the springs' discharge under overexploitation, total water supply and water demand for Alternative B (dry year 1995-1996).

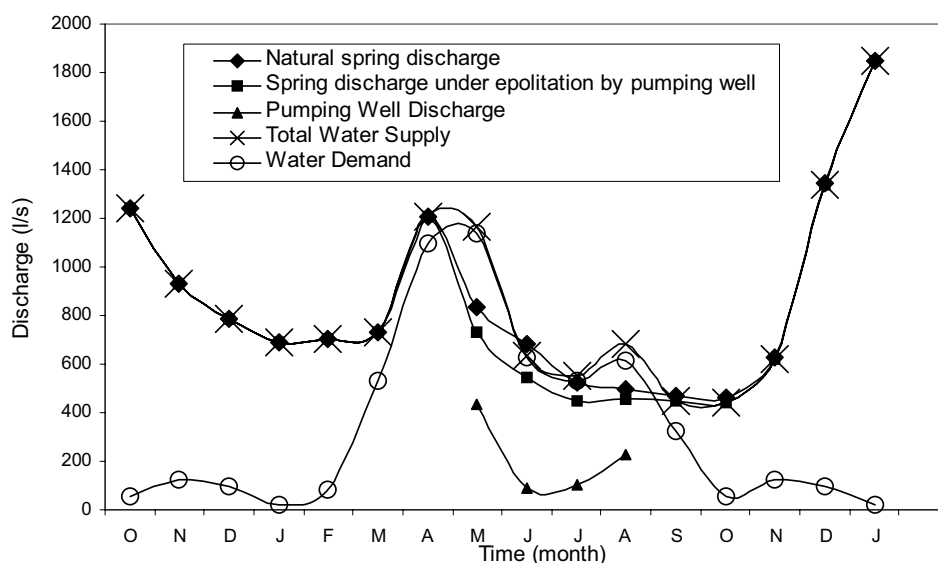


Figure 5. Natural springs' discharge, pumping well discharge, the springs' discharge under overexploitation, total water supply and water demand for Alternative C (dry year 1995-1996).

This method is used for a dry year with known values of spring discharges, but the flow rate of a karstic spring can be predicted for late spring and summer, already in early spring based on the previous spring's hydrograph. Therefore, the summer crop pattern, and the duration and discharge rate of pumping wells can be determined in early spring.

4. Conclusions

The discharges of karstic springs decrease significantly in late spring and summer, especially in dry years. The shortage of water can be compensated by exploitation of water by pumping wells. Although excessive harvesting from a karst aquifer reduces spring(s) discharge, the total output of pumping wells and spring(s) discharge is enough to overcome the water shortage. The reduction of spring(s) discharge continues in the subsequent fall, winter, and in case of severe overexploitation, even in the following years. The summer crop pattern, and the rate and duration of pumping can be determined to limit the effect of overexploitation on spring(s) discharge only in the subsequent months of low water demand, i.e. fall and winter. The discharge of karst spring(s) can be predicted in a dry year in early spring using the previous spring's hydrograph. A linear program can be applied to revise the spring crop pattern under various rates and durations of pumping, maximizing the net benefit. The traditional crop pattern can be revised under an overexploitation regime to achieve higher net benefit.

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A limitation of using EPIK method to define protection zones for karst aquifers

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Abstract

Groundwater in karst regions are very sensitive (vulnerable) to human activities. This vulnerability is related to the specific character of groundwater flow in karst aquifers, which is the presence of preferred pathways (pipe-like conduits) where contaminants can travel rapidly through the system. The physically based model SHETRAN was used to run particle-tracking simulations on a simplified hypothetical karst aquifer system. The objectives of the work were: Firstly, to simulate the aquifer system, with different model parameters (cases) to understand the sensitivity of each individual parameter on the particle movements within the aquifer as well as attempting to identify the most vulnerable parts of the aquifer to pollution. Secondly, to use the existing EPIK method to define protection zones for each case. The first case will be used as a reference case to compare the results with. In the second case the only parameter changed was the exchange coefficient between the surrounding matrix and pipe-like conduits. The hydraulic conductivity of the surrounding matrix was subjected to change in the third case. The obtained results show that any increase in the exchange coefficient or hydraulic conductivity of the matrix will enhance the chance that more particles will reach the outlet. The use of EPIK method on three different cases produced similar protection zones. This is because of the limitation of the EPIK method since it does not take in account the flow exchange between conduits and surrounding matrix as well as the variation of the hydraulic conductivity of the surrounding matrix.

Key words: Karst aquifer, Vulnerability, SHETRAN model, Pollution, Particle tracking, EPIK method, Protection zones.

1 Introduction

Flow in karst aquifers may occur through a connected network of conduits, matrix porosity or a combination of both. In the case of a well-developed conduit system, groundwater movement occurs almost exclusively through solution conduits, with velocities indicating turbulent flow. In the case of matrix porosity, diffuse flow occurs through the bulk of bedrock mass within small-interconnected pores and fissures, with laminar (linear) flow. Previous work on spring hydrograph analysis, in terms of variations in flow and chemistry, showed that spring water could be separated into two main types of flows: (1) Diffuse flow, where flow occurs within the matrix and (2) Conduit flow, where flow occurs within conduits (ATKINSON, 1977). Karst groundwater is particularly vulnerable to contamination because of the likelihood direct connection between surface activities and karst groundwater systems. The potential for contamination of karst groundwater from human impact sources is likely to be great if the recharge area is experiencing increased land use, particularly through urbanisation and agricultural activity. Karst regions are characterised by surface features such as sinkholes, swallow holes and losing streams. If surface water and contaminants rapidly enter the subsurface conduit networks through one of these features, they will rapidly be discharged via downstream springs (MARTIN & DEAN, 2001). Consequently, any contamination will affect the surface water quality of the spring runs, but will have little or no impact on water quality within the matrix because contaminants will be rapidly flushed from the spring. On the other hand, if matrix and conduits water mix, contaminants could infiltrate into the surrounding matrix, causing problems to the quality of water stored in the intergranular porosity.

All successful deterministic approaches to model (flow or transport) diffuse aquifers are based on the application of Darcy's law. However, within karst aquifers the use of Darcy's law is inappropriate. Modelling groundwater flow and transport in karst aquifers is very complex because of the difficulties encountered in establishing the fundamental spatial properties of the system. Such aquifers are characterised by heterogeneity, discontinuous porosity and large spatial variability of hydraulic parameters such as hydraulic conductivity and storage coefficient. As stated by JEANNIN 2001, most karst modelling techniques require an assessment of the geometry and hydraulic conductivity of the conduit network. However, such data are usually difficult or impossible to know.

The objectives of the paper could be grouped into three parts: Firstly, to simulate the aquifer system, with different model parameters (cases) to understand the sensitivity of each individual parameter on the particle movements within the aquifer as well as attempting to identify the most vulnerable parts of the aquifer to pollution. Secondly, to use the existing EPIK method to define protection zones for each case. Thirdly, to compare the results obtained from both methods.

2 Hypothetical aquifer

The case study used in this experiment is a simplified hypothetical karst aquifer ($800 \times 800 \text{ m}^2$), with an average thickness of 7 m. The aquifer is assumed to be unconfined with an impermeable base at the bottom and initial head of 111 m and fixed head of 114 m at both downstream (spring) and upstream respectively. The

catchment has been divided into 64 grids of 100 m in both x and y directions. Cross section of the aquifer and the imposed recharge to the system are presented in fig. 1 a and b.

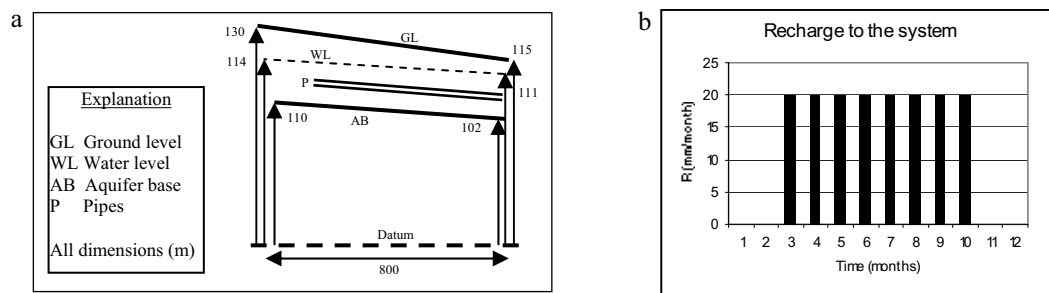


Fig. 1: (a) Cross-section of the hypothetical aquifer; (b) Imposed recharge to the system; (c) Grid distribution

2.1 Conceptual model

SHETRAN is a physically-based, integrated surface and subsurface modelling system, designed to simulate water flow, sediment transport and contaminant transport at catchment areas ranging from 1 to 2500 km² (Ewen et al, 1995). It has been developed by the Water Resource Systems Research Laboratory at the University of Newcastle upon Tyne. SHETRAN modelling system has three main components water flow, sediment transport and contaminate transport components. In this study, a new version of SHETRAN is developed to model karstic aquifers (ADAMS & PARKIN, 2002; RASHED & PARKIN, 2002). In addition to the mentioned components, the new karst model has four key elements, which have been proposed based on literature reviews of previous karst modelling (Parkin et al, 1999). These new elements, as illustrated in figures 2, are (1) Sub-horizontal cave network in the saturated zone; (2) Shafts; (3) Epikarst bypass flow generation and (4) Interactions with surface flows.

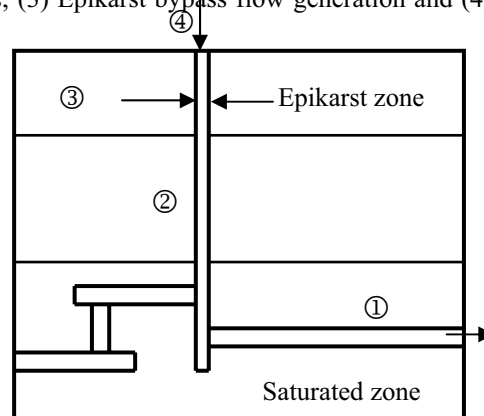


Fig. 2: Illustration of the karst model components

2.2 Application and methodology

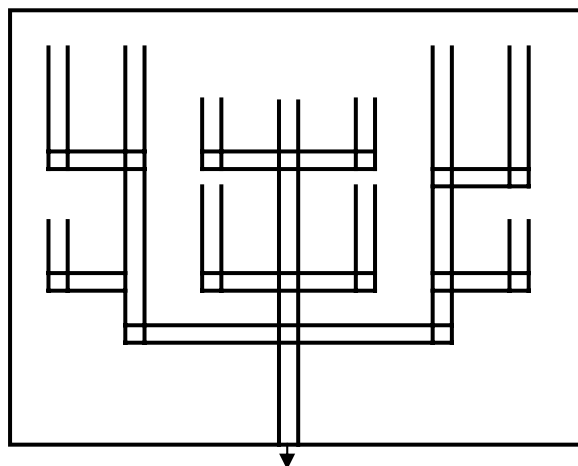
Three cases or scenarios have been studied each one contains 43 complex pipe network, ranging in diameter from 500 to 800 mm (figure 3). Because the aquifer is very simple so there was no epikarst layer or autogenic recharge (point source). The only source of water to the pipes is from the exchange between pipes and surrounding matrix. Since the first objective was to understand what affects can be made on the particle movements when some model parameters are subjected to change, only one parameter is allowed to be changed in each case. Table 1 presents the numerical values for the parameters used within the model for the three cases. As can be seen from the table, in case 2 the only parameter changed is the exchange coefficient between matrix and pipes, while hydraulic conductivity is the only parameter allowed to be changed in case 3. The methodology was (1) To release particles from the centre of the grid just one cell below the water table at the beginning of simulation; (2) To monitor the number of particles leaving the outlet.

Table 1: Numerical values used in the model

Parameters	K (m/d)	S _s (m ⁻¹)	D (mm)	L (m)	RC (mm)	EC (s ⁻¹)
Case-1	22	10 ⁻³	500-800	100	2	10 ⁻⁵
Case-2	22	10 ⁻³	500-800	100	2	10 ⁻⁴
Case-3	44	10 ⁻³	500-800	100	2	10 ⁻⁵

Where K is the hydraulic conductivity of matrix, S_s is the storage coefficient, D is pipe diameter, L is the pipe length, RC is the roughness coefficient and EC is the exchange coefficient between matrix and pipes.

Fig. 3: Complex pipe network (43 pipes)



2.3 Main findings

In order to make comparisons between the results from these cases, the results are presented in table 2. Results from table 2 show that in case 2 more particles reached the spring as a result of the increase of exchange coefficient that lead to more water to enter the pipes. Most of these particles travelled through pipes. For example, it took only 5 hours for particle No. 50 to reach the outlet in case 2, while around 6 days in cases 1 and 3. Interestingly, almost all particles reached the outlet in case 3, where high hydraulic conductivity is used. Suggesting that particle movements are more likely to be affected by the exchange coefficient and hydraulic conductivity values.

Table 2: Summary of the results obtained from different cases.

Comparison parameters	Case 1		Case 2		Case 3	
<u>Water table (m)</u>	<u>Max</u>	<u>Min</u>	<u>Max</u>	<u>Min</u>	<u>Max</u>	<u>Min</u>
Grid 16	109	106	106	104	111	107
Grid 50	112	110	109	107	113	111
<u>Spring flow</u>	<u>Peak</u>	<u>base</u>	<u>Peak</u>	<u>base</u>	<u>Peak</u>	<u>base</u>
Q (L/s)	76.2	24.4	152	35.1	108	41.8
<u>No. of Particles out</u>	No. of particles reached the outlet					
	After 9 year simulation		35		63	
	Within the first 2 days		21		3	
<u>Tracking paths (day)</u>	Particle reached the outlet after (days)					
	Particle No. 16		Not out yet		2146	
	Particle No. 50		0.2		6	
<u>Monitoring particles</u>	<u>Out *</u>	<u>In*</u>	<u>Out</u>	<u>In</u>	<u>Out</u>	<u>In</u>
Grids (particle & pipe)	15	15	27	3	30	0
Grids (particle only)	14	20	8	26	33	1

* Out stands for number of particles reaching the outlet.

* In stands for number of particles still moving in the domain.

3 The EPIK method

The EPIK method of vulnerability mapping takes into account the most knowledge of flow conditions in karst aquifers (DOERFLIGER et al, 1999). It is based upon vulnerability mapping of the catchment area of a source (spring or well), taking four criteria into consideration. These are Epikarst (E), Protective cover (P), Infiltration conditions (I) and Karst network development (K). DOERFLIGER et al, 1999 defined vulnerability as the intrinsic property of an aquifer (geological, geomorphologic and hydrogeological characteristics), which determines the sensitivity of groundwater to contamination by human activities. Both point and diffuse contamination are taken into account, as are the various types of contaminants at a global scale. The EPIK method is used for delineating protection zones in karst regions. The applicability of such method can be summarised into the steps: (1) Defining the catchment boundaries; (2) Assigning values to the four criteria; (3) The resulting vulnerability map by using GIS. The second step is the most difficult one because assigning values to the four criteria are very difficult task due to the lack of information about the nature of karst aquifers.

3.1 Definition of the criteria and their classes

The criteria are combined into a ranking by means of assigning numerical values to the classes. This assessment is very difficult because of the scarcity of data that might help to classify. The determination of the various classes of each criterion is performed with the help of several direct or indirect methods such as field investigation, geomorphologic studies, geophysics, tracer tests, interpretation of aerial photographs, shallow subsurface probe, water quality variation and hydrograph analysis. The various classes assigned to each criterion are presented in the following (DOERFLIGER et al, 1999).

Epikarst criterion (E)

E1=1 (highly developed epikarst, presence of Shafts, sinkholes or dolines and outcrops with high fracturing).

E2=3 (Moderately developed epikarst, presence of intermediate zones in the alignment of dolines, dry valleys and outcrops with medium fracturing).

E3=4 (Small or absent epikarst, non-existent epikarst (covered zones $P2$ or $P3$)).

Protective cover criterion (P)

P1=1 (Absent cover, less than 20 cm of soil).

P2=2 (Intermediate, soil cover between 20-100 cm).

P3=3 (Intermediate, soil cover between 100-200 cm).

P4=4 (High cover, soil cover greater than 200 cm).

Infiltration conditions criterion (I)

I1=1 (Concentrated, losing streams, streams feeding swallow holes or sinkholes and their catchment area).

I2=2 (Intermediate, catchment area for (I1), with a slope greater than 10%).

I3=3 (Catchment area for (I2), with a slope less than 10%).

I4=4 (Diffused, the rest of the catchment).

Karst network criterion (K)

K1=1 (Well-developed karst network, presence of well-developed karst networks, with decimetre or meter sized channels that rarely plugged and well connected).

K2=2 (Poorly developed karst network, presence of poorly developed karst network, with decimetre or smaller sized channels that usually plugged and poorly connected).

K3=3 (Mixed aquifer, presence of a spring emerging through porous terrain. Non-karst, only fissured aquifer).

3.2 Evaluation of the protection factor

The EPIK method is a multi-criterion weighting-rating method that assesses the groundwater sensitivity of a karst area in a strict manner. A multiplier, reflecting a relative importance weighting, is assigned to each criterion and then multiplied by the rate assigned to each criterion. The sum of these products will result to a final score (F_p). The lower the score, the high vulnerable the area is. The protection factor (F_p) can be calculated, for each grid, using the following equation:

$$F_p = \alpha E + \beta P + \gamma I + \delta K$$

Where F_p is the protection factor, E, P, I and K are the rate values assigned to each criterion (presented in the tables 4) and α , β , γ and δ represent the relative weighting coefficients for E, P, I and K respectively. The relative weighting coefficient emphasises the strong influence of one criterion to another on the vulnerability. For instance, epikarst and infiltration conditions strongly influence the sensitivity of the karstic groundwater; therefore, their relative weights have to be important. The choice of relative weighting coefficients is empirical. After carrying several empirical tests, DOERFLIGER et al, 1999 has suggested α , β , γ and δ equal to 3, 1, 3 and 2 respectively. Then by combining the values of the four criteria, the protection factor (F_p) can be calculated

using the above equation. The resultant F_p values, using the various possible combinations, were between 9 and 34. The higher the score, the less vulnerable the area is. According to these values, several classes of vulnerability can be identified. By grouping the F_p values, four major vulnerability areas (S1, S2, S3 and S4) can be identified (table 3).

Table 3: Allocation of protection zones (After Doerfliger et al, 1999).

Vulnerability areas	Protection factor (F_p)	Protection zone (S)
Very high	F_p lower than 19	S1
High	F_p between 20 and 25	S2
Moderate	F_p higher than 25	S3
Low	Presence of ($P4$)	S4 (The rest of the catchment)

3.3 Application to different cases

The EPIK method will be applied on cases 1, 2 and 3 of the hypothetical karst aquifer, attempting to identify protection zones for the spring area for each case. The rates of criteria, their relative weights and the calculated protection factors give the same results for the three cases (table 4). This can be attributed to the fact that the EPIK method does not, in any of the four criteria, take into consideration the hydraulic conductivity of karst network and their relation to the matrix (hydraulic head, hydraulic head distribution), as it is not easy to obtain such data. Furthermore, it does not take into account the variations of the vertical and horizontal hydraulic conductivity of the surrounding matrix (matrix around conduits). For example, in case 2, the exchange coefficient has been increased, so more water can enter to conduits and as a result more particles reached the outlet. Similarly, in case 3, the hydraulic conductivity of the surrounding matrix has been increased. Consequently, more water and particles entered to conduits and very quickly find their way to the outlet. This weakness in the EPIK method needs to be addressed in order to obtain more accurate protection zones for karstified aquifers.

Table 4: Calculation of protection factors and zones for three different cases.

Cases	$\alpha = 3$	$\beta = 1$	$\gamma = 3$	$\delta = 2$	F_p	Protection zone (S)
Case-1-2-3	E	P	I	K		
Group-A	3	3	3	3	27	S3
Group-B	3	3	3	1	23	S2

Group-A: Grids without pipes

Group-B: Grids with pipes

As stated by White (1999) several factors are likely to control the exchange of water between matrix and conduits. The most important two factors will be detailed hereafter. Firstly, the head gradient between the matrix and conduits plays an important role in deciding whether conduits are gaining or losing water. To illustrate this point, in low flow conditions, the conduits may act as a drain from the surrounding matrix, providing a baseflow condition for perennial springs. Whereas, in high flow conditions, the gradient may reverse, particularly if conduits are fed by allogenic recharge from sinking streams. In this case, the head within the conduit would be greater than the head of the surrounding matrix, causing the conduits to lose waters to the surrounding matrix. Secondly, the coupling between the conduits and the surrounding matrix is likely to be an important control of exchange between conduits and matrix. This link is more likely to be controlled by the hydraulic conductivity of the surrounding matrix and the size of the conduits. There will be extensive exchange if the matrix is extensively fractured or dissolved, resulting in increased permeability. For example, there was a dramatic increase in the spring flow as a result of increasing the hydraulic conductivity of the matrix (table 2). In terms of conduit sizes, small branching conduits will increase the surface area of conduits relative to their volume, increasing the likelihood of exchange of water with the matrix. Large conduits are, therefore, less likely to exchange water with the matrix than small branching conduits.

4. Conclusion

Particle tracking simulations were carried out using SHETRAN on a hypothetical karst aquifer system with different cases, attempting to identify the most vulnerable parts of the aquifer to pollution as well as to assess the sensitivity of changing model parameters on the particle movements within the aquifer. Then the EPIK method was used to define protection zones for each case. The results of the study can be summarised with two points: In terms of particle tracking simulations, particle movements are mostly affected by any changes in the exchange coefficient or hydraulic conductivity. As explained before, any increase in one of these parameters will enhance the chance that these particles will reach the outlet faster. In terms of defining protection zones, although three different cases have been studied to define protection zones using the EPIK method, the obtained results show similarity, indicating that an improvement to the EPIK method is needed to include the exchange coefficient and hydraulic conductivity in consideration within one of the four criteria.

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Integrating temporal hydrologic variations into karst groundwater vulnerability mapping – examples from Slovenia

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Abstract

Existing methods on karst groundwater vulnerability mapping take into account a variety of factors that control the infiltration of water and contaminants from the land surface towards the groundwater, such as overlying layers, concentration of flow, karst network development, and precipitation regime. Karst systems often show fast and strong hydrologic variations in response to precipitation events or snowmelt. Although such variations are crucial for contaminant transport, existing methods provide insufficient tools to cope with hydrologic variability. Examples from Slovene karst landscapes are presented, which are characterised by strong hydrologic variations, including groundwater level fluctuations, changing flow directions, intermittent lakes, some of which appear several times per year while others occur only very exceptionally, as well as temporary springs, swallow holes and estavelles. It is demonstrated why and how these variations are relevant with respect to contaminant transport and groundwater vulnerability. Finally, it is outlined how temporal hydrologic variability could be considered within the framework of karst groundwater vulnerability assessment and land use planning.

1. Introduction

Groundwater vulnerability maps are increasingly used for protection zoning and land use planning. An overview of the most commonly used methods can be found in VRBA & ZAPOROZEC (1994), GOGU & DASSARGUES (2000), and MAGIERA (2000). Intrinsic vulnerability takes into account the properties of the hydrogeological system, while specific vulnerability additionally considers the behaviour of different contaminants. It is also possible to distinguish resource and source vulnerability maps. According to an approach proposed by COST action 620, intrinsic karst groundwater vulnerability mapping is founded on the assessment of factors that control the infiltration of water and contaminants from the land surface into the aquifer, such as overlying layers (O), concentration of flow (C), karst network development (K), and precipitation regime (P) (DALY *et al.* 2002, ZWAHLEN 2004).

Karst systems often show fast and strong hydrologic variations in response to precipitation events or snowmelt, including groundwater level variations, change from open-channel to pressurised flow, change from laminar to turbulent flow, changing flow directions and velocities, shifting groundwater divides, as well as temporarily active streams, swallow holes and springs (GOLDSCHIEDER 2002). Although it is generally acknowledged that such variations have an impact on contaminant transport, the existing methods of vulnerability mapping do not provide sufficient tools to cope with hydrologic variability.

This paper presents examples from Slovene karst landscapes that show important hydrologic variations. The most attention has been devoted to the examples from the Dinaric karst – the region of the Javorniki mountains, the Pivka valley and the Polje of Cerknica near Postojna. Based on these examples, the paper discusses in which cases and how such variability should be integrated into vulnerability assessment and land use planning.

2. Examples from Slovene karst

Karst regions in Slovenia extend over 44 % of the country (NOVAK 1993, GAMS 2003) and hold important groundwater resources that contribute 50 % to the drinking water supply (COST 65 1995). Most of these areas consist of thick carbonate rock sequences forming large karst

massifs and plateaus, intersected by poljes and valleys. Particularly at the intermediate karst poljes, shallow karst areas or in the contact karst areas, various forms of interaction between groundwater and surface water can be observed.

Groundwater level variations in karst aquifers are often in the order of tens of metres within a short time period. The amplitude of these variations depends on hydro-meteorological and hydrogeological factors. The first group of factors includes the type, amount, intensity and distribution of precipitation, and factors governing snowmelt, such as temperature and wind. The second group comprises catchment size, aquifer geometry, effective porosity, and the dimensions and connectivity of the karst conduits. Some Slovene karst systems show very high groundwater level fluctuations. In the famous cave system of Škocjanske Jame, that is included in the UNESCO world heritage list, water level can rise up to about 100 m. The highest variations, reaching up to 214 m, have been recorded in the Kras plateau (MARGON 2002).

Groundwater level fluctuations also result in variations of flow velocities and directions, catchment boundaries, and surface-groundwater interactions. Flow bifurcations can be observed in several Slovene karst areas, e.g. in the Polje of Cerknica and the Pivka valley, which is located on the Adriatic-Black Sea watershed (HABIČ 1989). The schematic section in Fig. 1 illustrates the groundwater level fluctuations in this area. During low-water conditions, groundwater from the Javorniki mountains and Pivka valley drains towards the Polje of Planina in the NE. In wet periods water level rises and a groundwater divide forms below the Javorniki mountains so that a part of the area drains towards the Pivka valley in the SW.

Due to groundwater fluctuations and weak connections between different karst conduits, several intermittent lakes of different size, temporal duration and frequency occur in this region. The largest one is the Cerčniško Jezero, which can extend over 25 km² and contain more than 28 million m³ of water (Figs. 1 and 4). The lake usually fills twice per year but remains empty in dry years; in wet years, it occurs several times per year and/or does not dry up entirely. Several springs, estavelles, surface streams and direct precipitation recharge the lake.

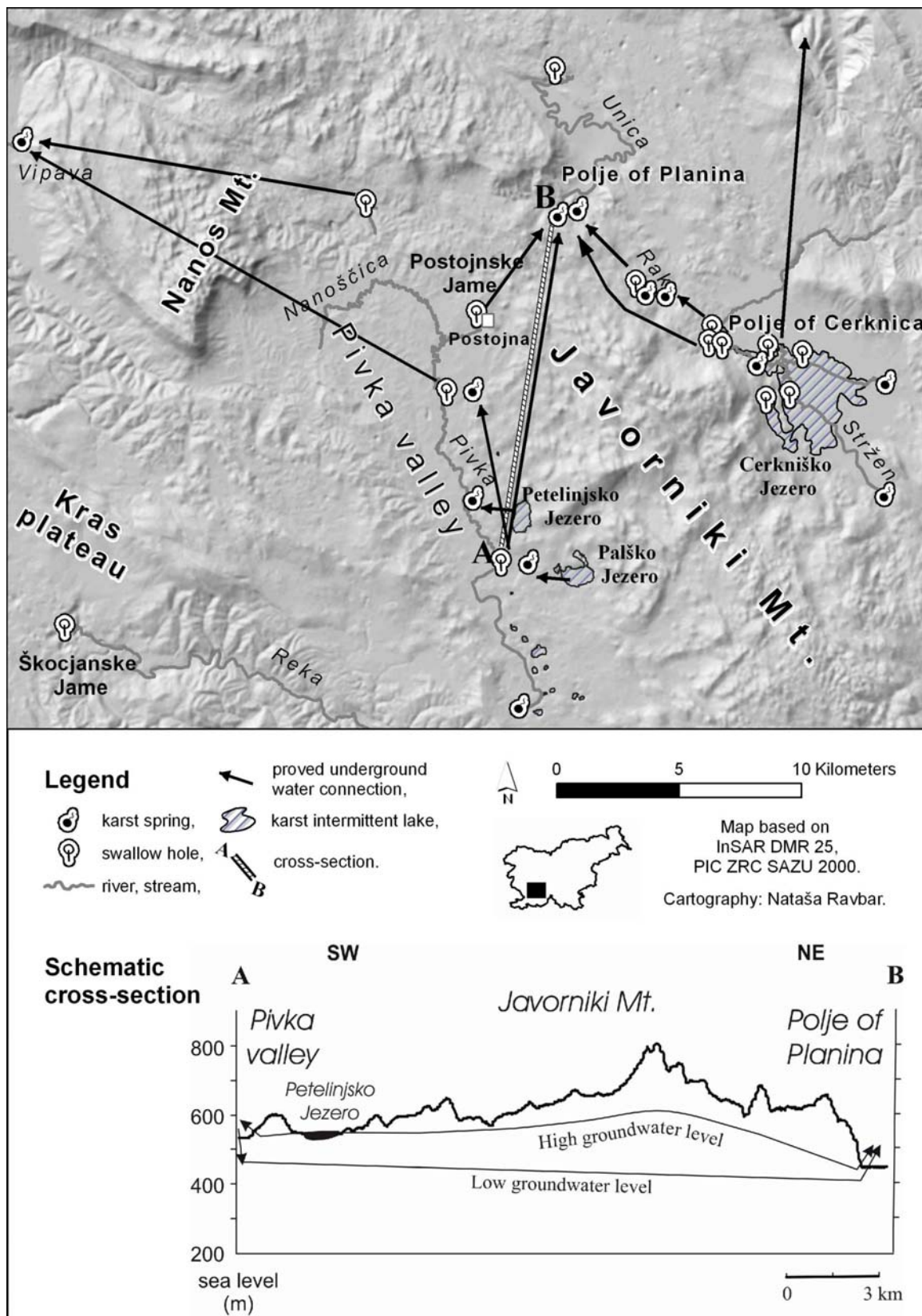


Fig. 1: Hydrological map of the Javorniki mountains, Polje of Cerknica and Pivka valley with the underground connections proved by tracer tests, and schematic section of the area during low- and high-water conditions.



Fig. 2 and 3: The intermittent lake Petelinjsko Jezero (location see in Fig. 1). At low groundwater level shallow karst hollow of the Petelinjsko Jezero is dry (left). When the level of groundwater rises the karst depression is flooded and forms a lake (right) (photo: N. Ravbar).

During low-water conditions, the estavelles act as swallow holes and contribute to the emptying of the lake, together with several other temporarily and permanently active swallow holes. The maximum outflow from the lake via these holes can reach 40-90 m³/s (KRANJC 2003).



Fig. 4: The Polje of Cerknica with Slovenia's largest intermittent karst lake (location see in Fig. 1, photo: N. Ravbar).

Several intermitted lakes can also be found in the Pivka valley, a closed and elongated depression, the southern part of which belongs to the zone of shallow karst. The lakes form in doline-like depressions with flat bottoms when the groundwater level is sufficiently high. The two largest examples are the Palško Jezero (1 km²) and the Petelinjsko Jezero (0,7 km²) (Fig. 1, 2 and 3). Numerous smaller lakes extend over a few hectares in surface (RAVBAR & ŠEBELA 2004). Some of the intermittent lakes appear once or several times per year, while others occur only in exceptionally wet periods, maybe a few times in a century. Similarly, some of the lakes contain water up to six months per year, e.g. the Petelinjsko Jezero (HABIČ 1975), while the majority of karst depressions are flooded only for some days during and after very high precipitation events.

As groundwater level is rising, water reaches the surface and activates several intermittent karst springs in the valley recharging the river and the lakes. Water also pours out of the aquifer through innumerable fissures and voids at the

bottoms or edges of depressions. These features are often small and not very distinctive, i.e. difficult to map. In periods of falling water level, the water sinks underground through the same fissures and voids, which consequently act as small estavelles. Additionally, numerous swallow holes are activated. The surface river also seeps and sinks underground at different locations in its riverbed. In dry periods, the swallow holes and estavelles remain dry (Fig. 5). The terminal swallow hole of the Pivka river is located at the NE edge of the valley.

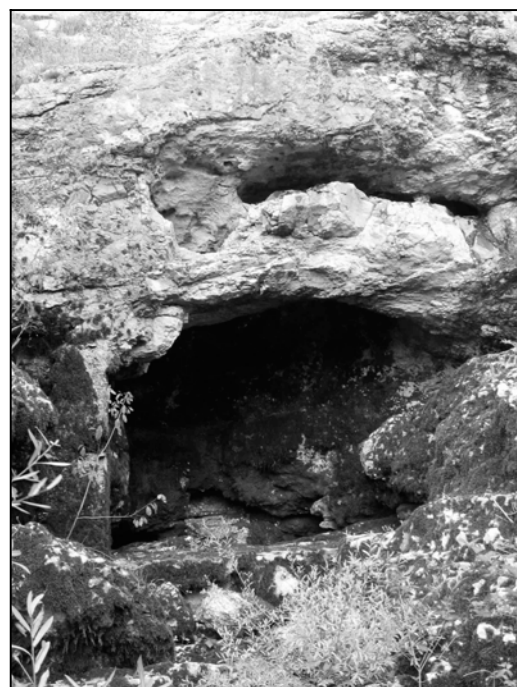


Fig. 5: The cave of Matijeva Jama is an estavelle that fills and empties the Palško Jezero (location see in Fig. 1). The water level fluctuations exceed 40 m. When it acts as a swallow hole, several m³/s sink into the cave (photo: N. Ravbar).

3. Implications for contaminant transport and groundwater vulnerability mapping

The described hydrological variability has many implications for contaminant transport and groundwater vulnerability mapping. Many common methods of karst vulnerability assessment, including the approach proposed by COST action 620 (see Introduction), take into account the thickness of the unsaturated zone. However, large water level fluctuations result in a variable thickness of this zone and, consequently, temporal variations of the groundwater vulnerability. Rising water levels means decreasing unsaturated zone thickness, decreasing protectiveness, and increasing vulnerability.

Most existing methods classify swallow holes, sinking streams and their catchment areas as zones of high or extreme vulnerability. The examples from Slovene karst illustrate that some swallow holes are frequently or permanently active, while others operate only during exceptional hydrological events, sometimes less than once per year. Only in the first case, a contaminant release would always and rapidly reach the groundwater without significant attenuation. The flow rate of sinking streams and their catchment areas may also vary drastically as a function of the hydrologic conditions.

Drainage divides and flow directions that change in response to hydrologic conditions also have implication for vulnerability mapping. The most important element of a source vulnerability map is the delineation of the spring catchment area. If the catchment boundaries vary for several 10 km, this raises a question, how to take in consideration these variations for source vulnerability mapping.

Furthermore, the flow and transport processes in the aquifer may drastically change in dependence on the prevailing hydrologic conditions. Some contaminants may be largely immobile during low-water conditions, as they might be stored in the unsaturated zone, in cave sediments or at the bottom of water-filled cavities. During high-flow events, these contaminants may be remobilised and cause problems (e.g. DNAPLs, LOOP & WHITE 2001).

There are different approaches to cope with hydrologic variability within the framework of karst groundwater vulnerability mapping. The PI method considers “average storm rainfall conditions” that occur several times per year (GOLDSCHIEDER 2002). COST action 620 proposed a precipitation factor (P) that classifies areas with high rainfall intensities as more vulnerable than such with gentle precipitation (DALY *et al.* 2002). The COP method presents a detailed assessment scheme for this P factor (VIAS *et al.* in press). However, none of these methods copes with the type of variations described here above.

Clearly, it is nearly impossible to create vulnerability maps for many different hydrologic situations (KRALIK & KEIMEL 2003). Furthermore, the characteristics of single hydrologic events are impossible to compile within one map. The concept of average hydrologic conditions also has drawbacks, because it would eliminate extreme events, which are particularly important for contaminant transport.

4. Outline of the proposed approach

In karst systems with relatively little hydrologic variability, the COST action 620 approach and the related or derived methods are sufficiently precise for vulnerability mapping. However, in the described Slovene landscapes and other karst systems with very high hydrologic variability,

this issue needs to be addressed, and existing methods should be supplemented.

Therefore, we propose to include the frequency and duration of hydrologic events into vulnerability assessment and adequate land use planning. Namely, it is difficult to explain a farmer why he should not use his land near a cavity that might operate as a swallow hole only a few times in a century.

Swallow holes that are permanently or frequently active (e.g. > 100 days/year) should be classified as more vulnerable than those that operate only exceptionally during extreme hydrologic events (< 1 d/y). This could be done, for example, by means of a new sub-factor ranging between 0 and 0.9, which could be integrated in the existing assessment scheme of the COP method.

Karst lakes, sinking streams and swallow holes that show systematic periodicity (e.g. active during snowmelt and dry during the rest of the year) require seasonally adapted land use practices.

The groundwater level fluctuations inside the aquifer are more difficult to deal with, and the required data are often not available. However, the protectiveness of the unsaturated zone in highly karstified rocks is generally considered to be fairly low. Variable thickness of this zone would consequently have limited impact on vulnerability. Therefore, the average groundwater level might be used for *resource* vulnerability mapping in most cases.

On the other hand, groundwater level fluctuations might alter catchment boundaries, which is crucial for *source* vulnerability mapping. Therefore, we propose to distinguish between an inner zone that is always or most often part of the catchment area, and an outer zone that contributes only during extreme events (a similar system is used in Ireland; GSI 1999). The inner zone should be classified as more vulnerable, which means that more stringent land use restrictions are required there. In some cases, these zones might be further subdivided. Similar as described for the swallow holes, a seasonal periodicity might be addressed with seasonally adapted land use responses.

5. Conclusions

Methods of groundwater vulnerability mapping should be kept as simple as possible in order to be actually applicable for protection zoning and land use planning. Integrating hydrologic variability into vulnerability mapping means an additional complication, which might not be recommended in areas where such variations are gentle.

However, many karst systems in Slovenia and elsewhere show drastic hydrologic variations, such as water-level fluctuations > 100 m, intermittent lakes, streams and swallow holes, as well as variable flow directions and drainage divides. These variations have a major impact on the vulnerability of groundwater to contamination. Existing methods do not sufficiently address this issue.

Therefore, an approach to integrate hydrologic variability into vulnerability mapping is outlined. A sub-factor for the frequency and duration of the activity of swallow holes could easily be integrated into existing assessment schemes. Variable drainage divides need to be considered for source vulnerability mapping. Seasonal variations can best be addressed through seasonally adapted land-use practices.

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La topographie 3D haute résolution : un outil pour l'étude des structures karstiques.

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Abstract

3D high resolution topography, a tool to study karst structure

The lidar (Optech Ilris 3D, precision ± 6 mm, angle min: 0.00115°) makes it possible to consider new scientific applications for studying karstic shapes. The laserscan has been used to obtain a precise topography of the Cathédrale room in Choranche cave (Vercors, France) and the “Salles Rouges” chamber in Orgnac cave (Ardèche, France). The grid step around 3 cm permits to measure different geostructural information like joints or bedding plane in the 3D numerical object of the chamber. It permits also to map the distribution of massive stalagmite of the floor and the ceiling. The comparison of geostructural data with ceiling and floor cartography permits to discuss the control of geostructure on stalagmite spatial distribution.

Résumé

Le lidar (Optech Ilris 3D, précision ± 6 mm, pas angulaire min : 0.00115°) permet d'envisager de nouvelles applications scientifiques pour l'étude des formes karstiques. Le laserscan a été utilisé pour obtenir une topographie précise de la salle de la Cathédrale dans les grottes de Choranche (Vercors, France) et des Salles Rouges à Orgnac (Ardèche, France). Le pas de mesure, aux alentours de 3 cm, permet d'étudier, dans le modèle numérique 3D de la salle souterraine, différentes informations structurales comme les joints de stratification ou des plans fracturés. Ce travail permet en outre de cartographier les dépôts stalagmitiques au sol ou au plafond. La comparaison des données structurales avec la cartographie des dépôts carbonatés offre une possibilité de mieux étudier le contrôle de la structure du massif sur la distribution spatiale des stalagmites.

1. Introduction : présentation de l'outil

Depuis peu, l'utilisation des Lidars en milieu karstique offre une nouvelle approche de la cartographie souterraine. Le Laserscan Optech ILRIS3D permet une topographie à haute résolution (distance précise à ± 6 mm, pas angulaire min de 0.00115°) d'objets naturels en surface ou en milieu souterrain avec un relevé de points par balayage avec des mailles pouvant aller de quelques millimètres à plusieurs décimètres. Ce maillage est paramétrable selon l'objectif de l'étude menée. À partir de ces points scannés, il est possible de construire un modèle numérique de terrain visualisable en trois dimensions. Cela a pour avantage de pouvoir observer un objet :

- sous un angle différent de celui auquel nous pouvions avoir l'habitude de le voir.
- qui n'était pas visualisable à l'œil nu ou alors de façon imprécise (exemple : plafonds de grands volumes souterrains comme ceux présents à Choranche ou à l'aven d'Orgnac).

Une analyse de la structure des formes et des formations endokarstiques, par une visualisation en trois dimensions, est alors possible.

2. Potentiel d'études sur l'identification de morphologies souterraines

Cet outil (lidar et logiciel dédié) de traitement numérique permet l'analyse de la structure d'objets karstiques généralement à macro-échelle. Il est possible de visualiser sur un même support, en trois dimensions, la topographie et le contexte lithologique. De fait, nous pouvons étudier quantitativement la morphologie des formes endokarstiques en incluant les relations de celles-ci avec leurs contextes lithologiques. Ainsi, nous pourrions trouver des indices pour mieux comprendre

leurs répartitions et leurs évolutions dans le temps et dans un espace donné.

Il peut être possible de repérer différentes phases de karstogenèse par l'identification de formes caractéristiques (exemple des banquettes limites, Fig. 1). Des mesures (degrés d'inclinaisons de parois de galeries, de coupes d'érosion) sont possibles sur logiciel dédié (Polyworks). Des logiques d'écoulements peuvent ensuite être identifiées et rattachées à différents contextes climatiques, géographiques ou environnementaux.

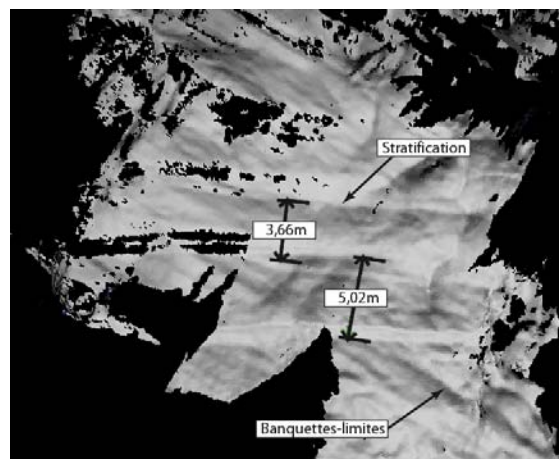


Figure 1 : Exemple de banquettes limites identifiées dans les Salles Rouges (réseau d'Orgnac, Ardèche, France) correspondant à une phase paragenétique. Des mesures sur logiciel dédié permettent de quantifier ici l'espacement des banquettes et d'évaluer la vitesse d'aggradation du remplissage sédimentaire.

De même, grâce à la précision du relevé topographique, il est possible de mettre en évidence des discontinuités

lithologiques dans la roche encaissante. Cela contribuera à mieux comprendre le choix préférentiel de l'implantation des formes endokarstiques ainsi que la localisation des stalactites et des stalagmites sous terre.

En effet, la structure de la roche joue un rôle fondamental dans l'implantation des stalactites sur les plafonds des grottes. Celles-ci poussent sur les discontinuités lithologiques (micro fractures, fissures, joints de strates). Dès lors, est ce que la taille de ces discontinuités joue un rôle important sur la morphologie, le fonctionnement des stalactites (diamètre, débit de l'écoulement dans la concrétion) et leurs nombres ? Ainsi, quelle pourrait être l'incidence sur la morphologie des stalagmites et sur les informations paléoenvironnementales et paléogéographiques qu'elles contiennent ?

3. La répartition et le fonctionnement des stalactites et des stalagmites

L'exemple de la salle de la cathédrale (grottes de Choranche, Vercors, France) permet d'appréhender la relation entre les édifices stalagmitiques au sol et leurs alimentations au plafond. Ici, la modélisation 3D permet d'étudier le contrôle de la structure géologique sur la distribution du concrétionnement (Fig. 2 et 3). L'outil se révèle alors précieux sur plusieurs points car il permet :

- d'envisager une quantification fine des volumes de dépôts carbonatés ;
- d'étudier finement les relations sols / plafonds en identifiant les zones d'alimentation ;
- d'envisager une analyse structurale de secteurs inaccessibles (plafonds de grotte).

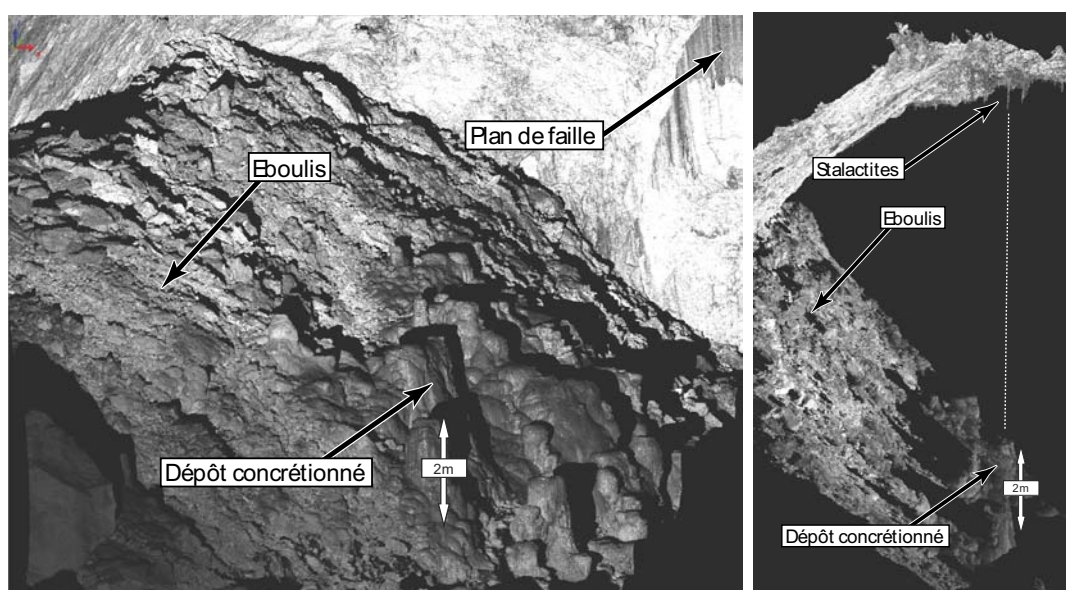


Figure 2 : Modèle numérique 3D d'une partie de la salle de la Cathédrale scannée avec le lidar Optech. Noter à gauche le talus ébouloux scellé par le volumineux dépôt stalagmitique. A droite, le modèle 3D tronçonné, permet d'identifier les stalactites à l'aplomb du dépôt stalagmitique.

Des études sont en cours de développement dans le cadre d'une thèse de géomorphologie visant l'étude 3D à plusieurs échelles de l'interaction entre structure géologique et formes karstiques. Par ailleurs, une thèse d'ingénieur soutenue par le CNRS va développer l'utilisation de la topographie 3D haute résolution dans la reconnaissance et la quantification des formes naturelles.

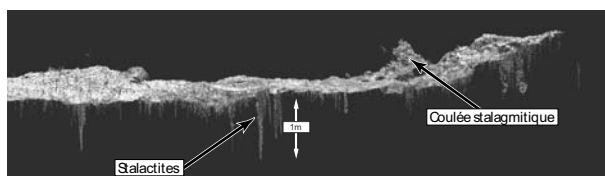


Figure 3 : Détail du modèle 3D du sommet de la salle de la Cathédrale (plus de 30m de haut). On différencie les coulées stalagmitiques des stalactites ($h < 1$ m) et quelques plans de discontinuités ayant guidé les écoulements de la zone non saturée.

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Maharlu karst basin behaves as double porosity media

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Abstract

Evaluation of alluvial aquifer yields with the conventional time-drawdown method is based on the assumption of Theisian or infinite radial flow (IRF) of ground water to a well. However, long-term yield of Karst aquifers is controlled by heterogeneities and boundary conditions, which lead to departures from the assumptions underlying the IRF. Aquifer-test responses in karst aquifers indicate internal heterogeneity at a scale below the resolution attainable with the available well control. Accurate prediction of long-term yield of karst aquifers therefore requires evaluation of aquifer heterogeneities. If they are to be treated mathematically, certain idealizations are imperative. Under circumstances where the rock matrix has very low permeability, one idealization has been to assume that flow can occur only in the fractures and not in the matrix, in other words the aquifer is treated as a "single porosity" continuum. Under conditions where matrix and fractures are both permeable the hard rock aquifer is treated as a "double porosity" continuum. Well test data is also analyzed based on the derivatives of drawdown with respect to the natural logarithm of time. This analysis has proved to be advantageous in well test interpretation, since it facilitates the identification of wellbore storage, inner/outer boundary effects and establishment of radial flow condition on the test. To illustrate application of these models, well test data from several pumped and observation wells in Maharlu karst basin in Fars Province of Iran are used. In all the cases double porosity behavior were observed. Reliable estimates of aquifer parameters are obtained by applying the derivative technique to the analysis of time-drawdown data.

1. Introduction

Fractures and fissures in rock formation strongly influence the fluid flow, conventional well-flow equations, developed primarily for homogeneous aquifers; therefore do not adequately describe the flow in fractured rocks. A complicating factor in analyzing pumping tests in fractured rock is the fracture pattern, which is seldom known precisely. The analysis is therefore a matter of identifying an unknown system. System identification relies on models, whose characteristics are assumed to present the characteristics of the actual system. We must therefore search for a well-defined theoretical model to simulate the behavior of the actual system and to produce, as closely as possible, its observed response.

2. Material and methods

Two approaches have been used for analyzing flow problems in fissured rocks. One is based on representing the fracture formation as a discontinuous medium, and considers individual fracture behavior and the exact geometry of the fissured system; the other assumes continuity and statistically homogeneous rock. GRINGARTEN (1982) give a detailed review of various approaches for analyzing the flow of fluid in fractured formations in connection with groundwater and petroleum engineering problems. Two models namely single-porosity and double-porosity are of more interest.

The concept of double-porosity medium is representative of the behavior of a fissured aquifer and was first introduced by BARENBLATT *et al.* (1989). Their model assumed the existence of two regions of different porosities and permeabilities (fissures and blocks) within the formation. The first porous medium consisted of relatively wide pores of the first order. The blocks in themselves were porous, consisting of grains separated by fine pores of the second order. The blocks fed liquid to the fissures and acted as a uniformly distributed source. The analytical solution for the above double porosity model results to type curves illustrated in Figure 1.

Type-curves are usually graphed as a dimensionless drawdown p_D versus a dimensionless time, t_D/C_D with each curve being characterized by a dimensionless number λ and ω that depends upon the specific reservoir model. Dimensionless parameters are defined as the real parameter times a coefficient that includes reservoir characteristics. When the appropriate model is being used, real and theoretical drawdown versus time curves are identical in shape but displaced with respect to each other, with the displacement factors for both drawdown and time axes being proportional to certain reservoir parameter.

Plotting real data as log-log drawdown versus time curve thus provides quantitative as well as qualitative information on the reservoir. Qualitative information, recognizing dominating flow regimes from their characteristic shapes help in selecting the most appropriate analytical reservoir model, and to break down the test data into segments that can be analyzed by specialized plots (semi-log), and usually provide more accurate results than quantitative log-log analysis.

There are difficulties inherent in the application of type curves for the determination of hydrodynamic coefficient, because they are nearly identical and superposition of field data plot on any of curves is possible. To overcome these difficulties a more diagnostic type curve known as derivative type curve may be used (BOURDET

et al. 1989) and (SPANE & WURSTER 1993). It is defined as the first derivative of dimensionless drawdown with respect to the natural logarithm of dimensionless time. It is very sensitive to variations in flow due to wellbore storage and aquifer non-idealities. As a result it has characteristic profiles that permit identification of flow regime types in an aquifer. Figure 1 shows the derivative type curve. In the early time, the derivative curves show a hump; this is a function of the dimensionless group $C_D e^{2s}$ and indicates wellbore storage and skin effects. Then, infinite acting radial flow occurs in the fissure system and the derivative reaches the 0.5 straight line. Later, transition begins and the derivative drops along a curve labeled $(\lambda C_D)/(1-\omega)\omega$ until it reaches a minimum, and then it follows a curve labeled $(\lambda C_D)/(1-\omega)\omega$ before returning to the 0.5 straight line corresponding to infinite radial flow in the total system $(C_D e^{2s})_{f+m}$.

3. Results

To identify an appropriate analytical model for Maharlu karstic aquifer in Fars province of Iran, a number of well test data are analyzed. All of test suggests a double-porosity model. One example is reported here:

A constant-rate pumping test of 1792.8 m³/day was performed in a well. The log-log time-drawdown (figure 2a) is easily matched with a number of type curves. This makes the evaluation of aquifer parameter rather subjective. But with the help of derivative curves various component of flow are more easily and correctly distinguished. The derivative curve (Figure 2c) consists of two straight horizontal line sections, and a depression. The straight lines indicate the Infinite radial flow (IRF), and the depression between the IRFs is indicative of matrix to fissure inter flow; i.e. Double porosity model. The infinite radial flow regimes appear as parallel line on semi-log plot Figure 2b. Table 1 list the aquifer parameters in three study sites.

Table 1. Aquifer parameters values.

well	T _f (m ² /day)	S _f	S _m	S _t
P1	691.30	9.56×10^{-4}	6.34×10^{-4}	1.59×10^{-4}
P2	775.52	3.82×10^{-5}	7.58×10^{-4}	1.14×10^{-4}
P3	977.5	2.94×10^{-3}	4.43×10^{-3}	7.37×10^{-3}

Conclusion

The interpretation of drawdown data recorded during a well has been used for many years to evaluate aquifer characteristics. Using type curve matching, conventional well test analyses have focused on homogeneous aquifers. Due to heterogeneities inherent in hard rock aquifer, these analyses are not adequate for hard rock aquifers. More recently, type curves, which indicate the drawdown response of pumping wells under variety of well and aquifer configuration, were introduced. However, due to unique shape of these curves, the result of interpretation is rather subjective. In this paper it was demonstrated that the drawdown derivative type curves, originally developed in Petroleum Engineering is a more diagnostic analytical tool for evaluation of hard rock aquifer. Since the study site is undergoing extensive development, it is recommended that hydraulic evaluation of karst ground water aquifers is conducted based on methods considered in this paper rather than conventional methods.

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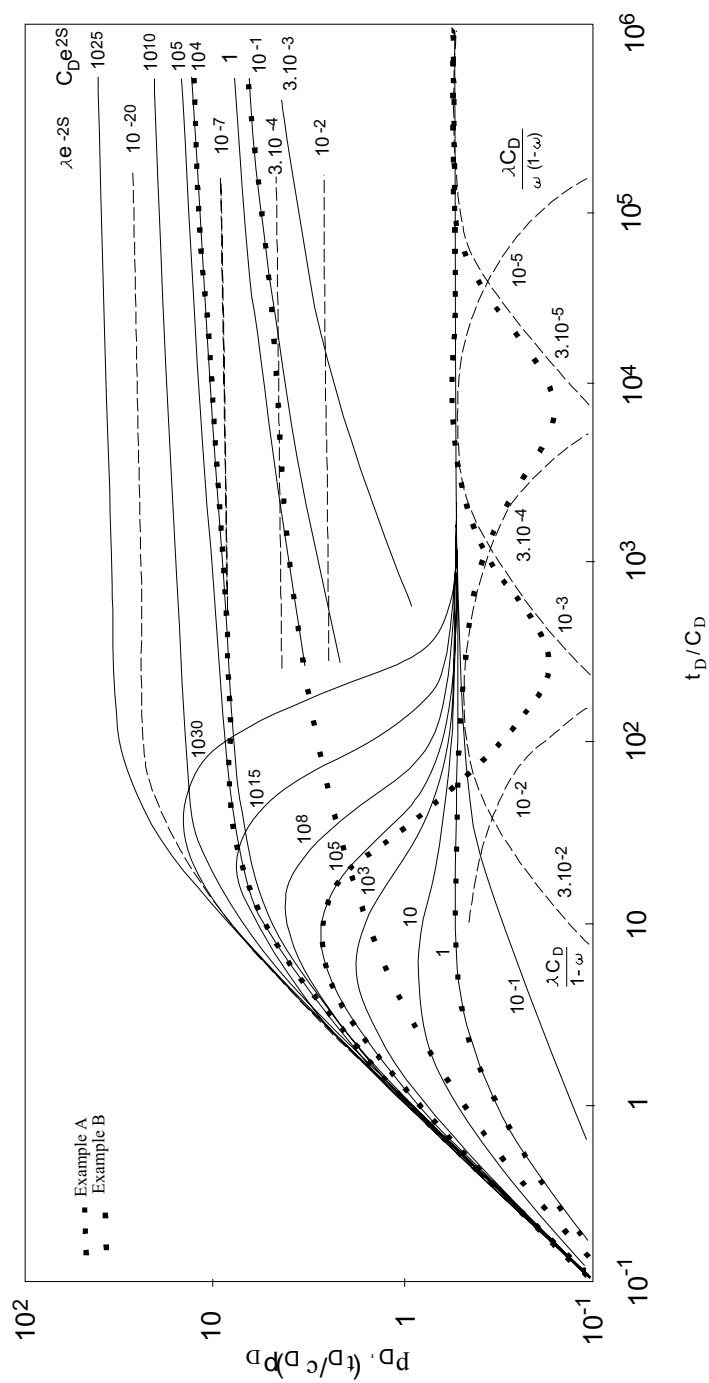


Figure 1. Derivative type curve of double-porosity reservoir(Bourdett et al, 1989).

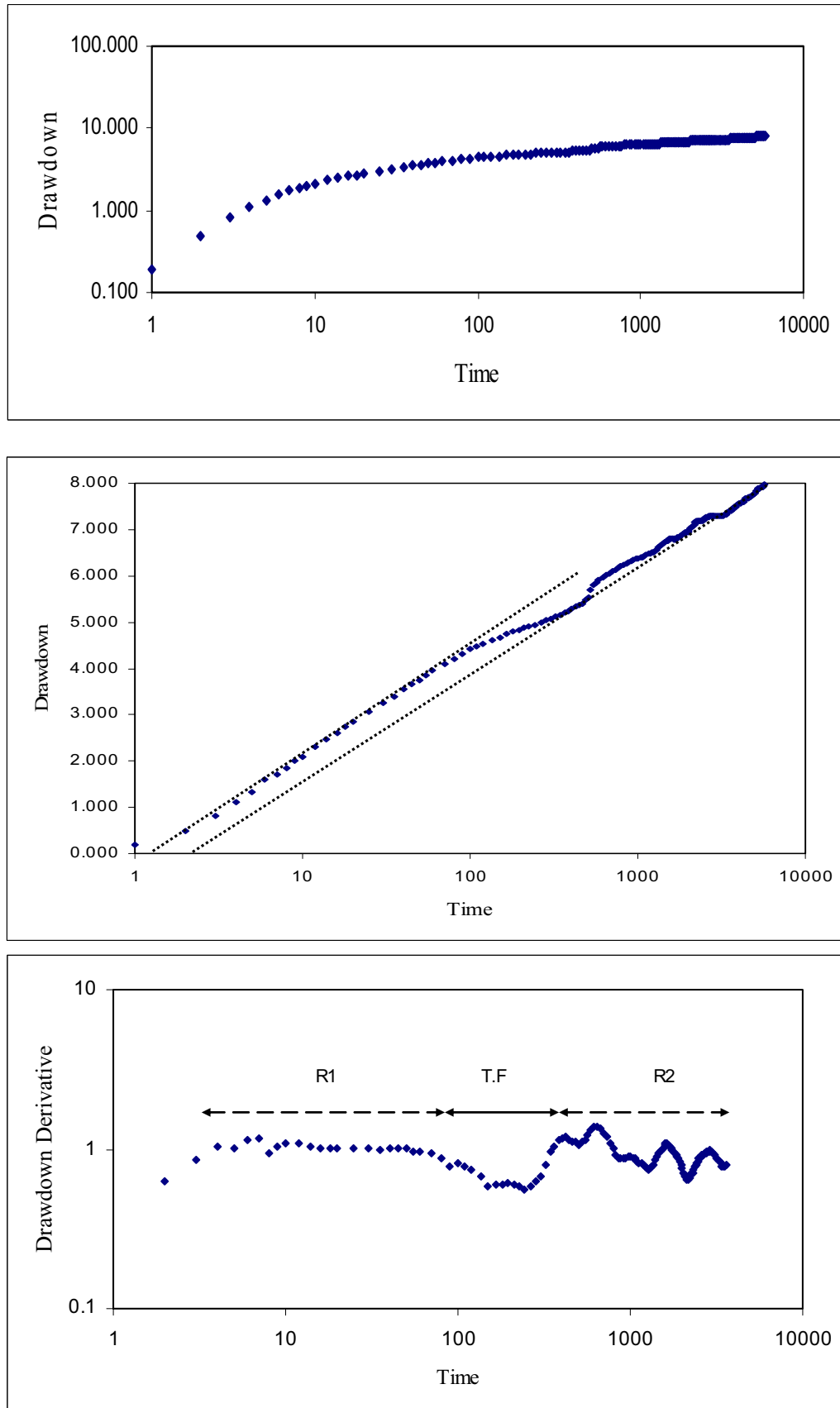


Figure 2 a,b,c.

Observation of groundwater resources in Switzerland – Example of the karst aquifer of the Areuse spring

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Abstract

The percentage of groundwater in the overall drinking water supply of the country is 84% (half of this is spring water). Responsibility for observation and monitoring of this important resource is shared between the Confederation, the cantons and water services with the cooperation of universities and private organisations. A new concept for the nationwide observation of groundwater resources aims to link the different national and cantonal groundwater networks more closely. The objective of these activities is to observe quantitative and qualitative variations in groundwater resources in the short term and to identify long-term changes sufficiently early. The discharge of the Areuse spring, one of the most significant karst springs in the Swiss part of the Jura Mountains, has been continuously observed since 1959. Since May 2005, the spring discharge data are accessible in almost real time on the internet. Temperature and electrical conductivity data will also be available shortly. High-water periods (e.g. 1987) and low-water periods (e.g. 2003) as well as seasonal variations (e.g. snow-melting) and long-term fluctuations (trends) are analysed on the basis of the 46-year long data series. Data of the Areuse spring are compared with those observed in boreholes within the spring recharge area.

1. Observation of groundwater resources in Switzerland

Responsibility for observation and monitoring of groundwater resources in Switzerland is shared between the Confederation, the cantons and water services with the cooperation of third party institutions and private organisations. The objective of these activities is to observe quantitative and qualitative variations in groundwater resources in the short term and to identify long-term changes sufficiently early, whether these are the result of natural or anthropogenic influences. The quantitative groundwater monitoring stations are concentrated in areas where the availability and demand for water is greatest, namely in the porous unconsolidated deposits in large alluvial river valleys that are often heavily urbanised (SCHÜRCH *et al.* 2004). Carbonate regions with significant subsurface drainage systems (about 16% of the Swiss land surface) dominate in the Jura region in the north-western part of Switzerland as well as in calcareous Alpine regions. These karst uplands are characterised as being water poor but obtain water from springs that are often highly productive. Nonetheless such systems are highly susceptible to all types of pollution. Discharge is continuously observed in carbonate regions nationwide at 16 springs and groundwater quality at 6 springs.

The Swiss Federal Office for the Environment has been observing groundwater levels and spring discharges nationwide since 1975 (FOWG 2004), a national groundwater quality monitoring network (NAQUA) has been developed in close cooperation with cantonal authorities since 1997. NAQUA measures a broad range of physicochemical parameter concentrations and performs specific microbiological tests (SAEFL/FOWG 2004). As a result, the evolution of the parameter and any deficits in groundwater protection can be identified. This allows existing and future protection measures to be improved. The national quantitative and qualitative groundwater observation program provides a representative overview as well as a characterisation of the groundwater situation at a

given date and of specific trends in individual major regions of Switzerland. NAQUA results indicate that many sources are contaminated with nitrate, pesticides and hydrocarbons. Although the concentrations detected do not represent a health hazard, they do indicate that precautions are required.

2. Reorganisation of groundwater observation nationwide

A new concept for the nationwide observation of groundwater resources aims to link the different national and cantonal groundwater networks more closely. Groundwater quality, groundwater level and spring discharge need to be monitored at the same observation station for integrated data treatment. The new national network for the observation of the groundwater resources consists of stations where groundwater quality and quantity are continuously monitored. As a result, the groundwater quality stations have been equipped for monitoring groundwater level and discharge. Selected stations will be equipped with a digital data acquisition system and with a transmitter for the online data transmission of groundwater level, discharge, temperature and electrical conductivity.

3. The karst aquifer of the Areuse spring

The basin of the Areuse spring is composed of two karst synclines: the La Brévine syncline and the Les Verrières syncline, separated by the Bois de l'Halle anticline. This anticline shows an axial depression south of La Brévine which allows an underground overflow from the La Brévine syncline towards the Les Verrières syncline (see Fig. 1) and towards the Areuse spring (see chap. 5). KIRÁLY (1973) established a hydrogeological map of the canton of Neuchâtel, including the isohypses of the Malm base which allowed the 3-D geometry of the aquifer to be understood. The limits of the basin are uncertain in the SW and NE, where it is in contact with the basins of the Doubs south of Pontarlier, of the Arvoux spring in Les Brenets and of the Noiraigue spring in Noiraigue.

The water-bearing formations are situated within the Malm limestone, which is some 350 m thick. They overlay the Argovian marls which form the lower confining bed of the aquifer. The surface area of the aquifer is about 130 km² or more, the mean elevation is 1,115 m; the elevation of the spring is 793 m.

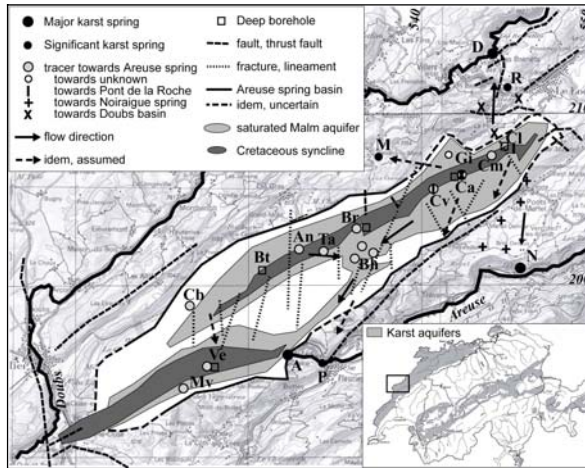


Fig. 1: The basin of the Areuse spring. Cb: Chez le Brandt, Ve: Les Verrières, Mv: Mont des Verrières, A: Areuse spring, P: Pont de la Roche spring, Bt: Le Brouillet, An: Aneta, Ta: Les Tallières, Br: La Brévine, Bh: Bois de l'Halle, M: Derrière Le Mont spring, Cv: Le Petit Cachot (de Vent/Moulinet), Ca: Le Cachot, Gi: Le Gigot, Cm: La Chaux-du-Milieu, Cl: La Clé d'Or, R: Rançonnière spring, D: Arvoux spring (Doubs), N: Noiraigue spring.

The first correct delimitation of the karst aquifer was performed by SCHARDT (1905), based on the geological structures, five tracer tests (An, Br, Cv, Ta and Ve, see Fig. 1) and a water balance ($P = 1,500$ mm, $E = 450$ to 600 mm, S around 140 km² for Q between 3.4 and 4.1 m³/s). Schardt's conclusions, including the uncertainties on the basin, were close to the actual knowledge.

A detailed study of the Areuse basin was performed by TRIPET (1973), based on an observation network in order to answer questions raised regarding water supply for drinking water and for hydraulic power generation. The network included five 77 to 450 m deep boreholes (Br, Bt, Ca, Cl and Ve) with more than 200 hydraulic conductivity measurements ranging from 10^{-7} to 10^{-6} m/s, and measurements of the groundwater level (see Fig. 2). Tripet had already simulated the Areuse basin with a 2-D analogue electronic model, including a single continuum hydraulic conductivity of 10^{-3} m/s, at least 3,000 times higher than the measured values. This shows that the description of a karst aquifer must include a discrete channel network.

KIRÁLY (1998) has developed a finite element model combining discrete channels and a continuum approach. The organised heterogeneity of the aquifer is schematised by a high permeability channel network immersed in a low permeability fractured limestone host rock. The use of a network with 1 km distant channels allows simulation of the observations (both piezometric levels in the boreholes and discharge at the Areuse spring), but it does not indicate where the permeable channels are located. It was possible to improve the simulation of both the peak and the recession flow at the spring by incorporating a permeable layer

(epikarst) into the 3-D model; this allows distribution of the infiltration over the basin surface.

A large number of tracer tests have been performed in and around the Areuse basin. The groundwater velocities range from 50 to 400 m/h or more. MÜLLER (1982) showed that a hydraulic connection between the western part of the La Brévine syncline and the Les Verrières syncline is likely. The tracer studies performed for the protection zone of the pumping well near La Clé d'Or (Cl) showed that the NE limit of the Areuse basin as defined by TRIPET (1973) should be adjusted. The area of Le Gigot (Gi) belongs to both the Doubs basin and the Areuse basin.

4. Observing the Areuse spring and its karst aquifer

The water level and discharge of the Areuse spring are measured in the nearby water course. They have been monitored since 1959 with a float connected to an analogue recording drum (water level recorder). Since May 2005, the float has been connected to a digital data logger and transmitter for online data transmission. Temperature, electrical conductivity and turbidity data will also soon be accessible in almost real time.

Since 1965, the observation boreholes have been monitored intermittently. Continuously records with an analogue recording drum were performed from 1967 to about 1975. The amplitude of the piezometric fluctuations attains about 30 m in Le Brouillet (Bt) and La Clé d'Or (Cl) and 60 m in Les Verrières (Ve). They show that the water level in the boreholes as well as the Areuse spring react similarly and quickly to significant precipitation events (see Fig. 2). The reaction in the boreholes is slightly delayed (0.3 to 2 days) as compared to the spring, and shows some singularities related to the heterogeneity of the aquifer. The recession after a precipitation event is the quickest in Les Verrières (Ve) and the slowest in La Clé d'Or (Cl). Due to their low hydraulic conductivities, the observation boreholes do not cross the highly permeable channels, but their rapid piezometric fluctuations suggest that they must be situated close to or even crossing the epikarst.

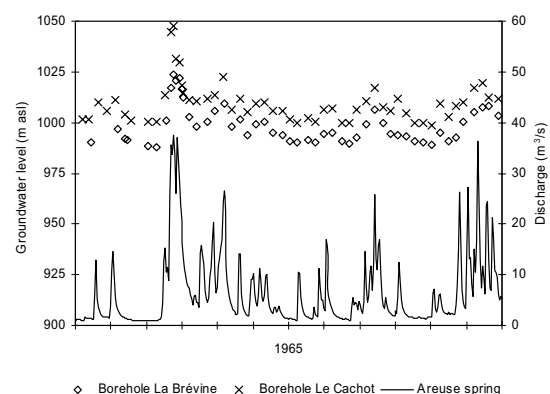


Fig. 2: Groundwater level in the boreholes at La Brévine (Br) and Le Cachot (Ca) and discharge of the Areuse spring for the year 1965.

5. Seasonal variations, high-water and low-water periods of the Areuse spring

The discharge of the Areuse spring fluctuates between 0.3 and 55.5 m³/s with a mean value of 4.9 m³/s. Figure 3

shows the classified monthly mean values as well as the long-term minimum, mean and maximum of the monitoring period 1959-2004. The long-term mean discharge of the Areuse spring generally has two maxima, in April and December, and two minima, in January and August. The Areuse spring is mainly recharged by intensive rainfalls and by snow melt.

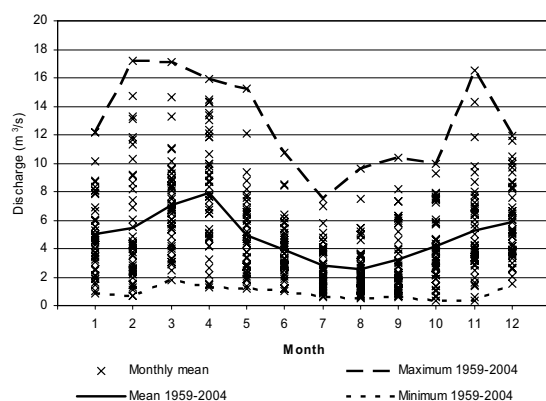


Fig. 3: Monthly mean discharge of the Areuse spring for the period 1959-2004.

The annual maximum discharge has been classified for each month for the period 1959-2004 (see Fig. 4).

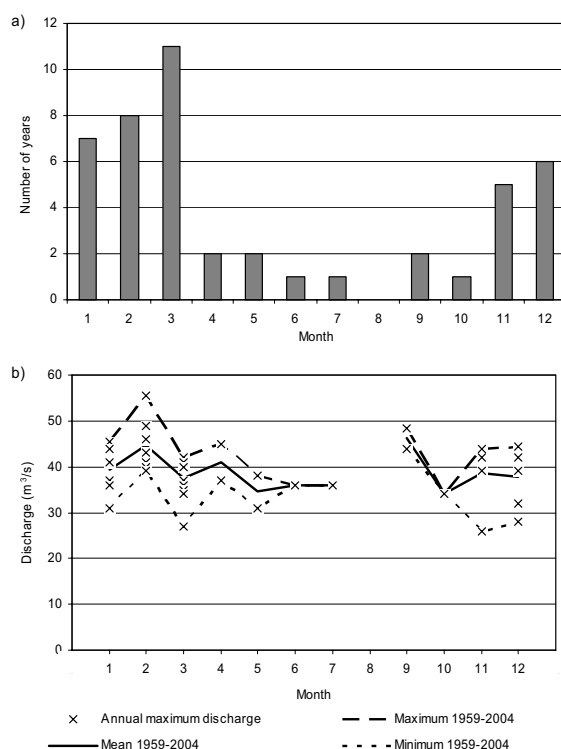


Fig. 4: Annual maximum discharge of the Areuse spring. a) Occurrence for each month and b) Classified for each month.

Figure 4a shows the number of years that the annual maximum discharge occurred for each month. The annual maximum discharge mainly occurs between November and

March with a peak in March. Over the measuring period, the annual maximum discharge has never been observed in August. Figure 4b shows the annual maximum discharge for each month. The record high-water periods for the Areuse spring both occurred in February ($55.5 \text{ m}^3/\text{s}$ on 15 February 1990 and $49.0 \text{ m}^3/\text{s}$ on 9 February 1966). High-water periods at the Areuse spring are probably induced by warm front events from the Atlantic Ocean with intensive rainfall and snow-melt. Table 1 shows the annual maximum discharge values with the corresponding recurrence interval and probability.

Table 1: Annual maximum discharge of the Areuse spring, recurrence interval and probability calculated with Gamma distribution and standardised method.

Recurrence interval (years)	Probability (%)	Annual maximum discharge (m^3/s)
2	50.0	39.0
5	20.0	44.0
10	10.0	47.0
20	5.0	49.0
50	2.0	52.0
100	1.0	54.0

The annual minimum discharge has also been classified for each month for the period 1959-2004 (see Fig. 5).

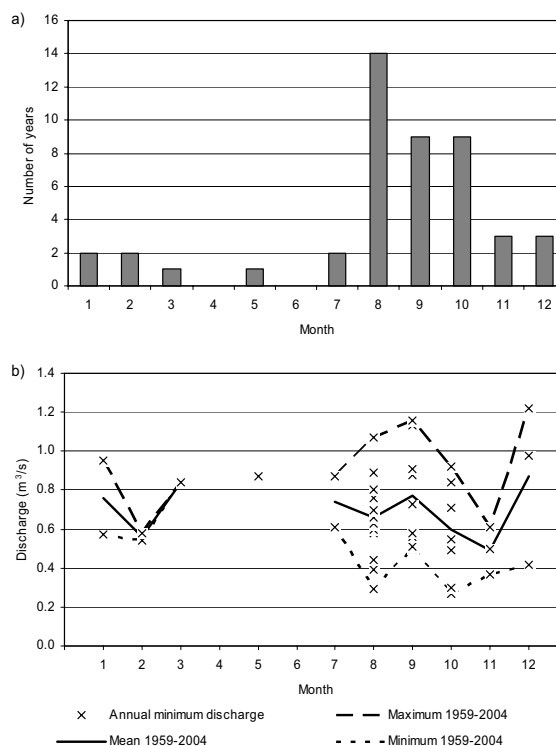


Fig. 5: Annual minimum discharge of the Areuse spring. a) Occurrence for each month and b) Classified for each month.

The annual minimum discharge occurs mainly between August and October with a peak in August (see Fig. 5a). The two record annual minimum discharge events of the Areuse spring were on 21 October 1962 ($0.27 \text{ m}^3/\text{s}$) and 28

August 2003 (0.29 m³/s). The annual minimum discharge also occurred during very dry winter periods between November and March. But it has never been observed in April or June.

Table 2 shows the annual minimum discharge values together with the recurrence interval and probability.

Table 2: Recurrence interval and probability of the annual minimum discharge of the Areuse spring calculated with Pearson distribution and standardised method.

Recurrence interval (years)	Probability (%)	Annual minimum discharge (m ³ /s)
2	50.0	0.67
5	20.0	0.49
10	10.0	0.40
20	5.0	0.34
50	2.0	0.28
100	1.0	0.24

6. Long-term fluctuations, trends

For the basin of the Areuse spring, BOUZELBOUDJEN et al. (1998) describe a statistical random precipitation character at the regional scale and, nevertheless a tendency to rise at the local scale for the period 1901-1992. The discharge of the Areuse spring shows no significant long-term trend for the observation period 1959-2003. However the annual maximum discharge seems to have increased over this period (see Fig. 6). This increase of the maximum discharge corresponds to one of the climate change scenarios described by BOUZELBOUDJEN et al. (1998).

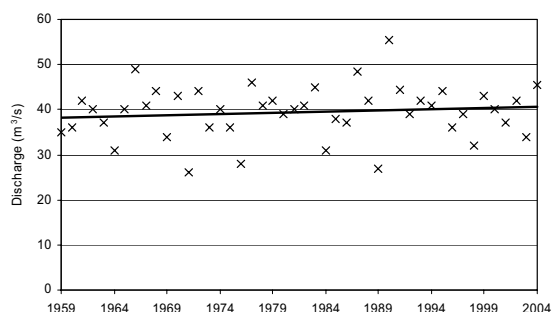


Fig. 6: Annual maximum discharge and long-term trend of the Areuse spring discharge.

The long-term discharge data series for the Areuse spring enable identification of above-average and below-average recharge periods. High groundwater recharge was observed in 1968, 1970, 1977, 1987, 1995, 1999 and 2001. Exceptionally low groundwater recharge was observed in 1964, 1971, 1976, 1989 and 2003. The groundwater recharge deficits, which have been caused by the dryness of the years 1964, 1976, 1989 and 2003, have remained until successive recharge surpluses of the following years have refilled the aquifer. This shows that successive precipitations recharge only gradually the karst aquifer of the Areuse spring.

GRASSO & JEANNIN (1998) have developed a mathematical model to evaluate effects of factors other than discharge (climatic changes, human activities) on water chemistry. The chemistry variations have been simulated and subtracted from the observed chemistry variations. The

residual difference between both chronicles was then assumed to relate to climatic changes. For the Areuse spring, the variation of temporary hardness with respect to discharge has been simulated. The residual difference showed that climate-dependent processes, such as variation in CO₂-production in soils, influence temporary hardness variation, but they are subordinate with respect to discharge.

7. Conclusions

Field mapping, tracer tests, monitoring of groundwater levels and spring discharge, 2-D and 3-D modelling have provided the necessary data to characterise the karst aquifer and the flow regime of the Areuse spring. The knowledge of the hydrogeology of the karst aquifer in combination with the long-term data series are the basis for the management of the groundwater resources of the Areuse karst aquifer. Furthermore, the vulnerability of the groundwater resources due to changes of the land-use and of the meteorological conditions can be observed, and, if necessary, suitable measures for groundwater protection can be adopted.

Long-term monitoring of the Areuse spring discharge allows observation of any long-term changes (trends) as well as of periods with low and high groundwater recharge. After very dry periods, the Areuse spring still has a discharge of about 0.3 m³/s, which is crucial for energy production in the nearby hydropower plant or for irrigation.

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Assessment of the Herbicide Atrazine within a Karst-Influenced Drinking Water Source: Rough River Lake, Kentucky, USA

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Abstract

Atrazine, a widely used herbicide in US corn production, impacts quality of drinking water for many rural water supply systems. Four systems at Rough River Lake, Kentucky, USA have had high levels of atrazine in raw water and in some cases finished water has exceeded the legal limit of 3 parts per billion (ppb). We are conducting an assessment of atrazine levels in lake's 1,700 km² catchment, much of which is in limestone karst. One system obtains its entire water supply from two major karst springs, with a combined catchment area of about 100 km².

The project's objectives are to create and support a citizen/agency stakeholder network, conduct field water quality sampling, and provide educational workshops. Grab and stratified samples were collected from eighteen locations (including all major tributaries) within the study area on a fourteen-day cycle during the spring 2005 growing season and twenty-eight day cycle during the following fall and winter months. Atrazine concentrations rose at nearly all sites within weeks of the onset of spring application in May, then generally fell by late July. Five locations had over 3 ppb for at least two sampling rounds. Two sites exceeded 10 ppb.

1. Introduction:

Water is critical to sustaining life on this planet. Water quality and accessibility challenges are key environmental problems that must be addressed in the 21st century. Twenty five percent of the world's population receives its drinking water from twenty percent of the earth's surface considered karst terrain (FORD & WILLIAMS, 1989; KARST WATER INSTITUTE, 2001).

Atrazine is an herbicide used to control broadleaf weeds in the production of corn, grain, and other plants (ASHTON & MONACO, 1991). Atrazine was first sold 45 years ago and since has become widely used because of its cheap price and efficient results in producing high-yield crops (COLBORN & SHORT, 1999; KENTUCKY DEPARTMENT OF AGRICULTURE, 2004). Potential health problems from long-term exposure to atrazine are found in the literature (APPLE & HUDAK, 2001). Studies of atrazine have occurred to see what effects the chemical has on both human health and the environment (RICHARDS & BAKER, 1995; KETTLES *et al.*, 1997; STEVENS *et al.*, 1999; COADY *et al.*, 2004). This study focused on atrazine and the process by which it moves through a watershed located in a karst landscape.

The study area is a 1700 km² watershed located in a rural area with land use dominated by row crop and pasture agriculture. The area is mostly karst, with various springs, sinkholes, and blue-holes in the area. Sixty percent of the corn produced in the area is located in Hardin County. The remainder of the corn production being split between Breckinridge and Grayson counties. With this heavy no-till corn production comes the risk of atrazine contamination in the local watersheds.

Four rural water suppliers acquire water from the watershed. The water treatment plants include Hardin County Water District Number One, Grayson County Water District, the City of Hardinsburg, and the City of Leitchfield. One of the four water treatment plants receives its entire water supply from two karst springs. The other three receive their raw water from the Rough River Lake reservoir.

The UNITED STATES GEOLOGICAL SURVEY (2002) and the KENTUCKY ENVIRONMENTAL QUALITY COMMISSION (2001) research found that the Green River basin contains higher atrazine levels than allowed by current EPA maximum contamination levels (MCL), currently 3ppb. Traces of atrazine were found in all springs sampled within the Green River basin, with maximum levels as high as 5ppb during the spring application season. (UNITED STATES GEOLOGICAL SURVEY, 2002).

This study focused on addressing important questions on how atrazine moves in karst landscapes. The project used the following tools to address the problem, field sampling, stakeholder formation and support, and educational workshops.

2. Methodology

Local University of Kentucky Agriculture Extension Agents chose the sample locations for both the lake and major tributaries. Eighteen grab sample locations and four stratified locations were selected for the project, and sampling began February 2005. Samples were taken on a fourteen-day cycle, starting February 2005 through June 2005, then to a twenty-eight day cycle for the remainder of the non-growing season. Having started in February 2005, it allowed the project to gain background readings before the spring application began in late April and early May. An average of 12 samples were taken

for each site out of 36 total samples, which included finished and stratified samples. Environmental Protection Agency water sampling protocol were used and samples were analyzed using immunoassay test kits. Field duplicates were chosen randomly for each sampling round to satisfy data quality control procedures.

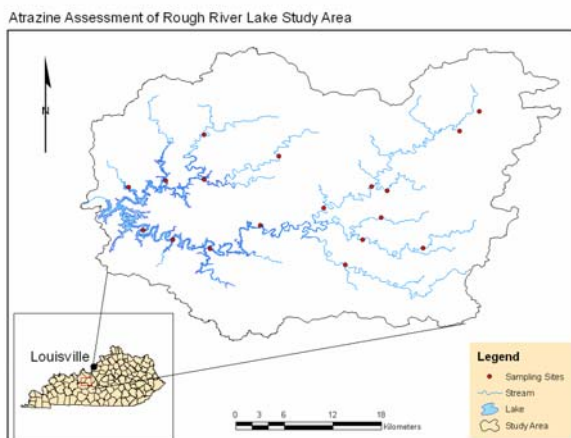


Fig. 1: Study Area

3. Results

Application of atrazine began in the watershed at the end of April and the first recordings of atrazine over the MCL of 3ppb occurred during the sampling round in May 2005. Four hundred nineteen samples were collected during the study period. Of those 419 samples 37 or 8.8% were over the 3ppb MCL. One location showed a concentration of 10.33ppb and another at 11.18ppb, the latter with 30% of the samples taken were over 3ppb. One location, a rural water treatment plant receiving its entire water supply from two karst springs, had concentrations as high as 4.69 ppb. These high concentrations were recorded only during one event, May 23, 2005, after the application.

Stratified samples taken showed that stratification was occurring as the atrazine moved from the tributaries into the lake. Over time atrazine concentrations increased as the chemical moved down in depth.

Summary: February 2005-December 2005

LOC	mcl (ppb)	# of samples	Max concentration recorded	# of samples exceeding mcl	% of samples exceeding mcl	mean
CC001	3.00	13	1.75	0	0.00%	0.21
MC002	3.00	14	3.13	1	7.14%	0.42
LM003	3.00	15	1.33	0	0.00%	0.26
HS004	3.00	15	4.95	1	6.67%	0.58
LC005	3.00	13	0.32	0	0.00%	0.08
RR006	3.00	13	4.43	1	7.69%	0.63
GC007	3.00	14	2.18	0	0.00%	0.93
GC007.A	3.00	6	3.48	1	16.67%	1.07
GC007.1	3.00	11	2.01	0	0.00%	1.07
GC007.3	3.00	11	3.69	1	9.09%	1.41
GC007.5	3.00	11	4.75	3	27.27%	1.87
LL008	3.00	11	2.11	0	0.00%	1.16
LL008.1	3.00	12	2.09	0	0.00%	1.19
LL008.2	3.00	1	1.24	0	0.00%	1.24
LL008.3	3.00	10	2.64	0	0.00%	1.51
LL008.5	3.00	10	6.00	3	30.00%	2.23
NF009	3.00	15	4.38	1	6.67%	0.59
CH010	3.00	14	3.27	1	7.14%	1.45
CH010.A	3.00	13	1.76	0	0.00%	0.60
MP011	3.00	11	6.43	2	18.18%	1.14
LE012	3.00	14	3.41	1	7.14%	1.27
LE012.A	3.00	13	3.52	2	15.38%	1.18
LE012.1	3.00	11	3.81	2	18.18%	1.72
LE012.2	3.00	9	3.30	2	22.22%	1.96
LE012.3	3.00	11	3.76	3	27.27%	2.06
WC013	3.00	12	2.66	0	0.00%	1.44
WC013.1	3.00	11	2.68	0	0.00%	1.49
WC013.2	3.00	1	1.15	0	0.00%	1.15
WC013.3	3.00	11	6.51	2	18.18%	2.16
WC013.5	3.00	10	10.33	1	10.00%	2.31
RR014	3.00	13	11.18	4	30.77%	2.42
MC015	3.00	15	3.54	1	6.67%	0.41
NF016	3.00	14	4.99	1	7.14%	1.45
RS017	3.00	13	4.09	1	7.69%	0.65
PS018	3.00	14	4.69	1	7.14%	0.62
PS018.A	3.00	14	3.95	1	7.14%	0.55

*Sample sites with .A are finished water samples.

** Sites with .1, .2, .3, .5 are stratified samples taken at 1, 2, 3, or 5 meter depths.

Table 1: 2005 atrazine concentration summaries.

4. Conclusions

There is a definite connection with atrazine's movement within a karst landscape. The lower concentration of atrazine in the spring feed water treatment plans may be due to the use of best management practices and federal set aside programs for producers. Best management programs for the use of atrazine in karst areas require applicators to stay at least 150 meters from sinkholes and streams during atrazine application.

The building and support of the watershed stakeholder group has lead to a proactive approach to dealing with atrazine. Future studies may focus on areas where high levels of atrazine occurred and source areas may be defined. Further study also may yield that only one or two producers are the cause of the pollutant and financial incentives may then be put in place to help producers switch to non-atrazine products, thus leading to a solution to the problem.

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Microsphères de verre : un traceur pour la simulation du transport sédimentaire

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Résumé

L'utilisation des traceurs classiques n'est pas une méthode satisfaisante pour la simulation du comportement des particules naturelles dans les eaux souterraines karstiques. Pour l'obtention de résultats directs sur le transport sédimentaire, des microsphères de verre (30 μm) ont été employées. Un test d'aptitude a été réalisé en laboratoire ainsi que sur le terrain. Les résultats démontrent que les microsphères ont des propriétés similaires aux particules mobiles naturelles. De plus, ce nouveau traceur est facilement applicable et analysable. Un traçage dans la grotte de Milandre (Jura suisse) permet de faire des interprétations sur le transport de cette fraction granulaire, qui joue un rôle important dans le fonctionnement des systèmes karstiques.

Abstract

The issue of natural particle transport in karst systems cannot be adequately tackled using classical tracers. Employing glass beads within the silt fraction (30 μm) may be an appropriate tool for investigating sediment propagation. Their nature and characteristics were screened regarding sediment tracing by means of lab and field experiments. In combination with the easy applicability and a low-tech analysis solution, glass beads are thus qualified for simulating the behaviour of the mobile grain fraction in karst groundwater. The new tracer was tested at the Milandre fluviokarst system (Swiss Jura Mountains) and allows for providing preliminary interpretation with respect to sediment transport in conduit flow systems.

1. Introduction

Le fonctionnement hydraulique des systèmes très karstifiés est caractérisé par une grande variabilité de l'énergie inhérente, créant des fluctuations de débit et des paramètres physico-chimiques des eaux karstiques. L'écoulement dans les conduits est souvent turbulent. Ces conditions permettent à la matière solide d'être acheminée dans la zone saturée pour y être transportée, et finalement évacuée à l'exutoire du système.

À la source, ces particules en suspension se manifestent par la turbidité de l'eau, qui est soumise à une grande variation temporelle et généralement corrélée au débit. En étiage, la turbidité reste modérée : une partie des particules est déposée dans des zones stagnantes. Lors des crues, la quantité de particules par unité d'eau peut atteindre des valeurs considérables. Cela étant dû, soit à l'arrivée de matière provenant de la surface et portée par les eaux infiltrantes (apport allochtone), soit par la mobilisation du sédiment dans les conduits (apport autochtone).

Ce changement entre des phases d'accumulation et d'érosion, dans le milieu souterrain, tient un rôle important dans la possibilité d'utilisation des eaux karstiques comme eau potable : lors des crues, l'augmentation de la turbidité dégrade la qualité de l'eau. De ce fait, la compréhension du transport sédimentaire est un objectif essentiel de l'hydrogéologie karstique.

Des mesures de la variation de la turbidité, de la constitution et distribution des grains naturels ou des paramètres chimiques et biologiques, ne suffisent pas pour répondre à ces questions : en général, les conditions initiales sont inconnues. Les microsphères fluorescentes classiques, dont la densité est inférieure à celle de l'eau, ne sont pas des traceurs appropriés pour l'objectif donné.

Une méthode de traçage sédimentaire est ainsi présentée. Afin d'illustrer les bénéfices et les restrictions

d'usage, un test d'aptitude est nécessaire pour identifier les caractéristiques de ce nouveau traceur.

2. Caractéristiques du traceur

Les microsphères employées comme traceur sédimentaire sont disponibles sous forme d'une poudre industrielle, distribuée sous le nom de Spheriglass® (POTTERS INDUSTRIES INC. 1997). Il s'agit des sphères de verre disponibles en différentes tailles : un diamètre de 30 μm a été sélectionné pour cette étude (figure 1).

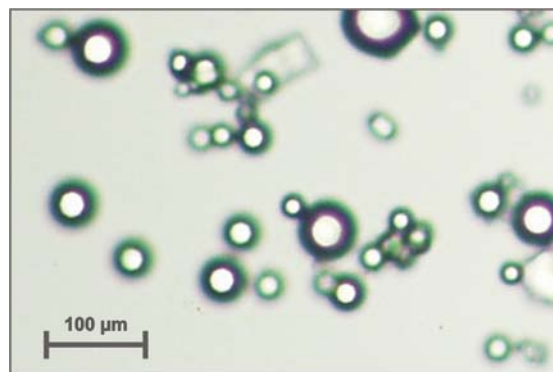


Figure 1: Microsphères de verre sous microscope.

Les propriétés physico-chimiques fournies par le fabricant sont présentées dans le tableau 1. Il s'agit de particules de silice ayant l'avantage d'avoir une structure amorphe et d'être stable. La constitution, la densité et la taille correspondent aux caractéristiques des grains naturels de la fraction silteuse. Ces particules sont à la fois, suffisamment mobiles pour rester en suspension dans le courant, et capables de sédimenter dans les zones stagnantes.

Tableau 1: Propriétés physico-chimiques selon le fabricant (POTTERS INDUSTRIES INC. 1997).

composition	SiO ₂ (> 70 %)
structure	amorphe
taille (Ø)	30 µm
variance (x)	27-36 µm
densité	2.4-2.6 g/cm ³
charge	négative (?)
fluorescence	aucune

Pour permettre une interprétation quantitative des futurs traçages avec des microsphères de verre, les caractéristiques de la mobilité du traceur doivent être déterminées. A part le transport advectif-diffusif, seuls les processus de sédimentation et d'adhésion influencent le comportement des microsphères, ceci étant dû à leur nature persistante.

Un essai de laboratoire a été effectué pour illustrer la sédimentation des microsphères sans et avec agitation mécanique. La figure 2 montre le dispositif et les résultats de cette expérience pour laquelle le traceur a été mis en suspension et renversé dans un tube vertical. La concentration en microsphères a été déterminée à trois niveaux différents selon de temps.

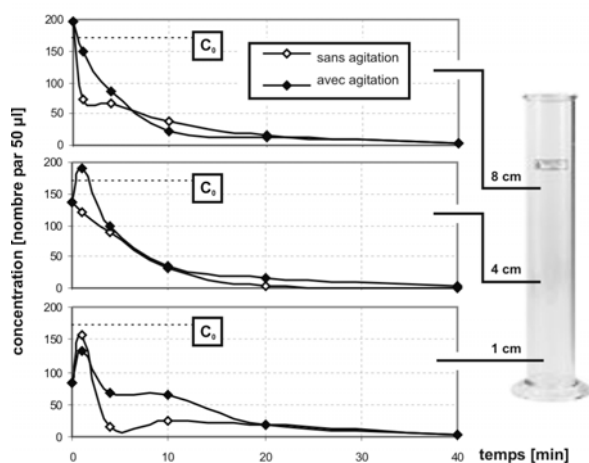


Figure 2: Essai de sédimentation.

En première approximation, la concentration en microsphères dans l'eau du tube diminue de façon exponentielle. Pour le niveau intermédiaire (4 cm) et au fond du tube (1 cm), cette décroissance est superposée par l'arrivée des particules provenant du haut. Enfin, la plupart des sphères se déposent après plus de 20 minutes. Ce délai est nettement plus grand que celui estimé d'après la loi de Stokes.

Il est remarquable d'observer que la concentration est nettement différente à toutes les hauteurs du tube, directement après avoir versé l'eau ($t=0$). La raison invoquée est l'effet de flottabilité liée aux bulles d'air qui se créent lors du remplissage du récipient : les particules restent ainsi plus longtemps en suspension. De plus, cet effet est renforcé par l'agitation mécanique du système. En milieu naturel, le même effet est produit par un flux turbulent. Cela permet aux particules d'être transportées sur des grandes distances.

L'effet potentiel de fixation adhésive aux surfaces d'aquifère a été aussi évalué en laboratoire. L'essai démontre l'interaction entre les microsphères et les

différents matériaux disponibles dans un système karstique (figure 3). Des flacons ont été remplis de différents matériaux trouvés dans le lit d'un cours d'eau souterrain (cf. traçage) et de l'eau provenant du même karst, tracée par des microsphères. Ces échantillons ont été agités et des prélèvements effectués durant 24 heures.

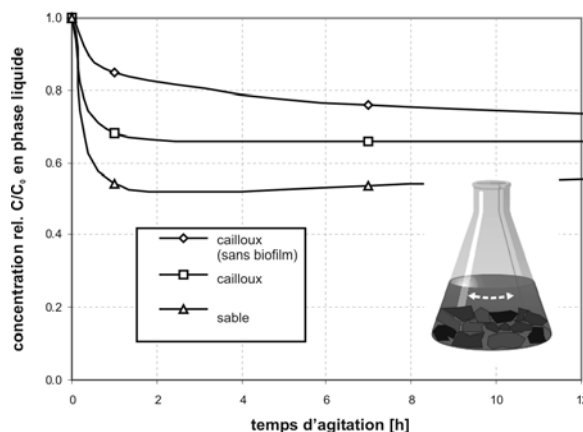


Figure 3: Essai d'interaction avec trois types de matériaux.

La plus grande baisse de concentration en microsphères dans le liquide, c'est-à-dire l'interaction la plus prononcée, est observée avec le matériel sableux (composé de calcaires, de marnes et d'oxydes). L'effet est plus faible pour des cailloux calcaires, et diminue nettement pour des cailloux sans biofilm naturel (enlevé par H₂O₂). Ces effets de fixation seraient sûrement beaucoup plus faibles en milieu naturel avec des temps de contact plus limités. Néanmoins, ce processus doit être pris en compte pour les microsphères et pour les particules naturelles.

En conséquence, les microsphères de verre sont considérées comme comparables aux particules sédimentaires naturelles au regard de leur nature et de leurs caractéristiques de transport. La courbe de restitution obtenue pendant un traçage peut en principe être utilisée pour étudier le transport sédimentaire.

3. Méthode d'analyse

Un autre critère pour l'applicabilité d'un traceur concerne sa détection analytique. Les microsphères de verre sont analysées par comptage dans un volume d'eau prélevé. Le comptage s'effectue à l'aide d'un simple microscope à lumière blanche pour lequel la procédure suivante s'est avérée la plus pratique : l'échantillon, prélevé et stocké en flacon de verre ou plastique, est homogénéisé et un volume de 50 µl est extrait par pipette. Ce volume est étalé sur un porte-objet et couvert par une lamelle. Ensuite, les sphères sont comptées individuellement en scrutant la surface du carré (agrandissement de la lentille préférablement entre 10 et 30 fois).

Les microsphères de 30 µm sont facilement identifiables à leur forme sphérique et régulière et à leur opacité. Les particules artificielles et naturelles (minérales ou matière organique) sont discernables les une des autres sans aucune ambiguïté grâce aux propriétés optiques des microsphères.

Le nombre de sphères adéquat pour le comptage est entre 10 et 100 par échantillon. Une concentration plus élevée nécessite la dilution de l'échantillon avant la préparation. Avec une concentration plus faible que

10 sphères par 50 µl, le risque d'erreur relative liée au comptage devient critique. Dans ce cas, il faut augmenter la concentration par filtration. Ainsi, le seuil de détection peut être baissé à volonté, à condition que le volume d'eau tracée disponible soit suffisant (minimum 1 par volume).

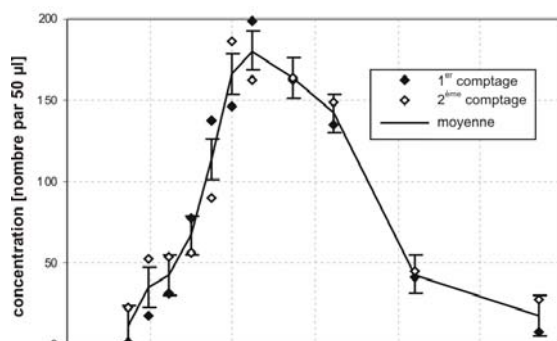


Figure 4: Analyse double avec estimation de l'erreur.

Les résultats de deux comptages pour les mêmes échantillons sont présentés dans la figure 4, ainsi que les moyennes arithmétiques associées (sans dilution ou filtration). Les barres d'erreur de ces dernières correspondent à un écart-type de 1. Cet exemple montre que l'ensemble d'une série de valeurs est bien reflété par deux séries de comptages, malgré une erreur individuelle significative.

A côté de l'analyse, d'autres sources d'erreur existent au niveau du prélèvement et de la préparation.

4. Application par traçage

Suivant les tests réalisés en laboratoire, un essai de traçage doit valider l'applicabilité du nouveau traceur sur le terrain. Ainsi, les microsphères ont été injectées dans un système karstique dans le Jura tabulaire suisse. La grotte de Milandre est un lieu propice à l'expérimentation grâce à l'accès facile à un cours d'eau souterrain d'un fluviokarst d'une longueur de plus de 4 km.

L'apport de masse solide dans ce système est considérable (JEANNIN *et al.* 2005). Cela constitue la condition préalable pour une grande concentration de particules déposées et en suspension. PERRIN & KOPP (2005) ont mesuré des débits entre 20 l/s en étiage et plus de 300 l/s en crue pour le cours d'eau souterrain. Ces variations sont accompagnées par des changements significatifs de turbidité.

La figure 5 montre la distribution de taille des particules naturelles dans l'eau du système lors des moyennes eaux (conditions pendant le traçage). Cette microgranulométrie est responsable d'une turbidité de 5 NTU environ. Bien que la plupart des particules naturelles tombent dans la fraction colloïdale (< 1 µm), le maximum de la charge sédimentaire en volume se trouve autour de 30 µm. Ainsi, les microsphères utilisées représentent une gamme importante du spectre microgranulométrique du site d'étude, lequel est également représentatif pour d'autres sites jurassiens (ATTEIA & KOZEL 1997).

Afin d'identifier le comportement spécifique des microsphères par rapport au flux d'eau, celles-ci ont été employées en combinaison avec le bromure (Br⁻), traceur soluble et conservateur. Les deux traceurs ont été mélangés dans un récipient et simultanément déversés dans le cours

d'eau. Par la suite, des prélèvements à la main ont été effectués environ 160 m en aval et les échantillons stockés dans des flacons en plastique pour l'analyse en laboratoire.

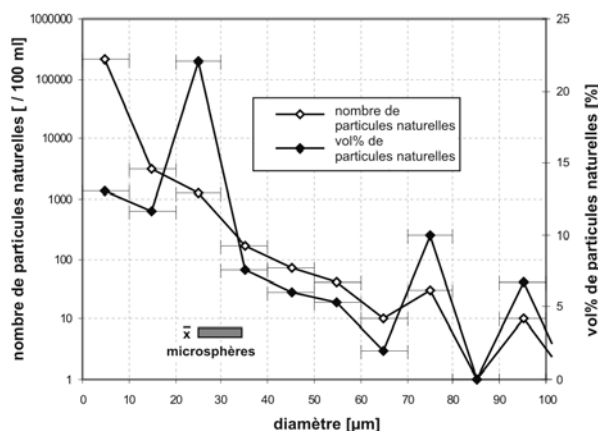


Figure 5: Distribution de taille granulaire pour la charge sédimentaire en suspension.

Les paramètres caractéristiques pour la réalisation du traçage sont représentés dans le tableau 2, ainsi que les valeurs-clés des résultats. Ceux-ci sont basés sur les courbes de restitution obtenues par l'analyse des échantillons (figure 6).

Les courbes des traceurs soluble et particulaire diffèrent significativement. Tandis que la courbe de restitution du bromure est serrée et peu dispersée, ce qui correspond à un régime de flux en conduit sur courte distance, l'arrivée des microsphères s'étale sur une période de plus de deux heures. Le maximum de la concentration déterminée (par rapport à la concentration d'injection) atteint seulement le tiers de celle des microsphères. Cependant, les deux traceurs arrivent pratiquement en même temps, indiquant une vitesse maximale de 20 m/min, respectivement 1200 m/h. La vitesse dominante pour les pics des courbes est 15 m/min (870 m/h) pour les microsphères, et 13 m/min (800 m/h) pour le bromure.

Tableau 2: Conditions initiales et résultats-clés du traçage comparatif.

	30 µm	Br ⁻
débit	65 l/s	
distance	160 m	
masse d'injection	2 kg	10 kg NaBr
temps de transit (min)	8 min	
vitesse (max)	1200 m/h	
temps de transit (pic)	11 min	12 min
vitesse (pic)	870 m/h	800 m/h
concentration relative (max) C_{max}/C_0	5.3×10^{-4}	1.8×10^{-3}
taux de restitution (après 2 heures)	95 %	96 %
temps 90 % restitution	95 min	15 min

La restitution de masse est quasi-complète pour les deux traceurs, mais les 90 % sont atteints beaucoup plus rapidement pour le bromure. En observant les pics principaux, le déficit de masse des microsphères s'explique par le retard d'une certaine quantité de particules par rapport au bromure. Cela se manifeste par un deuxième "pic",

beaucoup plus irrégulier, qui s'étale sur une durée de deux heures environ, soit 5 fois plus longue que l'effet de la dispersion hydraulique.

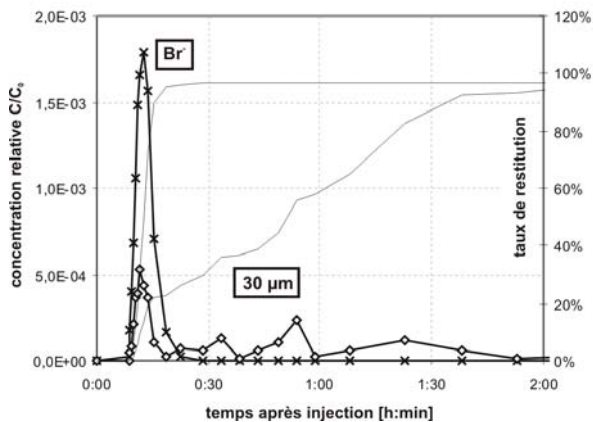


Figure 6: Courbes de restitution pour le traceur conservateur (Br⁻) et les microsphères (30 µm).

5. Conclusions

Des microsphères de verre (30 µm) ont été testées comme traceur sédimentaire. En conclusion, elles accomplissent toutes les exigences imposées à ce type de traceur:

- Composition, taille et densité correspondent à la charge sédimentaire et arrivent à reproduire le comportement d'une certaine gamme de grains naturels.
- L'identification analytique est fiable sans interférence avec d'autres matières ou traceurs.
- Les résultats obtenus d'un traçage peuvent être exploités d'une façon quantitative, grâce à une erreur modérée et un seuil de détection adaptable.
- Finalement, du point de vue pratique, le traceur est bon marché, applicable en grosses quantités et inoffensif à l'environnement.

L'utilisation de ce traceur peut contribuer à mieux comprendre le comportement des grains naturels (figure 7) et aussi mieux caractériser le système karstique.

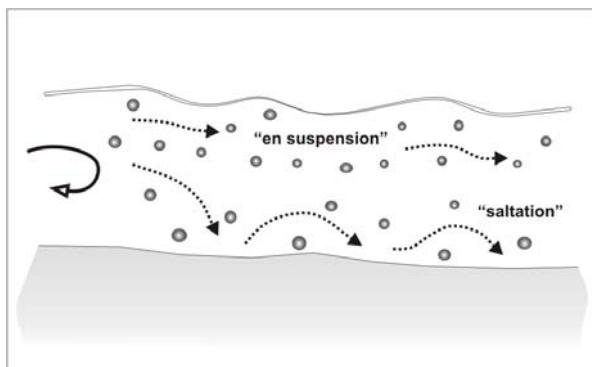


Figure 7: Modèle conceptuel pour le transport sédimentaire issu des résultats du traçage.

Le taux de restitution des microsphères étant très haut pendant l'essai de traçage, montre qu'elles ne restent pas déposées en permanence dans les conditions données. Pour expliquer ces résultats, on peut utiliser un modèle

conceptuel comme celui de la figure 7. Il postule que la saltation d'une partie des sphères due au flux turbulent, cause un retardement. Des petites inhomogénéités en taille et en densité des microsphères peuvent favoriser cette séparation entre suspension (pic principal) et chute gravitaire (pic secondaire irrégulier, figure 6). Le retard d'une partie du traceur serait sûrement plus prononcé pour des conditions de basses eaux ou pour un tronçon du cours d'eau avec zones stagnantes (WHITE & WHITE 1997).

Plusieurs autres domaines d'emploi sont envisageables pour ce type de traceur. Il serait intéressant de comparer son comportement sur le même site en différentes conditions hydrologiques ou pour différentes sections de la rivière souterraine. D'autres questions spécifiques seraient l'évaluation de la filtration en zone vadose des milieux hétérogènes (fracturés ou karstifiés), ou en milieux poreux, à une échelle d'étude en laboratoire. La possibilité d'adapter le seuil de détection en fonction de la quantité d'eau filtrée permet d'utiliser cette méthode pour des distances très grandes dans des systèmes karstiques, ou pour des milieux à forte atténuation en particules.

Cependant, les interprétations possibles sont restreintes : les microsphères ne couvrent qu'une gamme limitée du spectre des particules en suspension ou sédimentées. Elles ne simulent effectivement pas le comportement des microorganismes; par contre, elles informent sur le transport de ceux-ci par fixation aux surfaces mobiles, processus très important dans le cadre de la protection des eaux (DUSSART-BAPTISTA *et al.* 2003). Une meilleure utilisation de ce traceur est recommandée en combinaison avec des mesures des paramètres naturels des systèmes karstiques, notamment avec la turbidité.

Remerciements

Les microsphères ont été généreusement offertes par l'entreprise PQ Potters Europe GmbH, Allemagne. Mes collègues Cécile Allanic, Ray Flynn, Ivann Milenkovic, Alain Pochon, Michiel Pronk, Ludovic Savoy et Marie-Eve Wyniger sont chaleureusement remerciés pour leur aide à la réalisation de cette étude.

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Comparative tracing experiments to investigate epikarst structural and compositional heterogeneity

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Abstract

Comparative tracer testing may be used to evaluate the vulnerability of water to specific contaminants by comparing reactive tracer response to that of a simultaneously injected non-reactive "conservative" substance. Conversely, knowledge of tracer reaction with specific materials permits information about subsurface heterogeneity to be inferred. A series of tests completed in the vadose zone overlying a limestone aquifer employed a cocktail of particles along with reactive and non-reactive solute tracers to investigate transport rates between the ground surface and monitoring points approximately 10 m below ground. Short pulse tests revealed both solutes and particulate contaminants could travel at rates of over 10 m/h. Comparison of particle (micro organisms) and non-reactive solute tracer breakthrough revealed that particle tracers experience pore exclusion resulting in higher peak relative concentrations which arrive earlier than those of the solute. Prolonged tracer injection during subsequent experiments confirmed the response observed and illustrated that over 40 % of flow paths between injection and monitoring points were inaccessible to particles, but could allow solutes to pass through them. Similarly, the difference in response between different reactive tracers demonstrated tracers reached monitoring points via multiple flow paths and suggests geochemical heterogeneity plays an important role in influencing tracer behaviour. The results of this investigation highlights the complexity of water flow through the vadose zone of fractured limestone and the vulnerability of groundwater in karst aquifers to contamination when soil cover is thin to absent.

Résumé

Des traçages comparatifs peuvent être utilisés pour évaluer la vulnérabilité de l'eau souterraine à des contaminants spécifiques, en injectant simultanément des traceurs réactifs et un traceur conservateur. La connaissance du comportement des traceurs spécifiques renseigne sur les caractéristiques du sous-sol. Des traçages ont été réalisés dans la zone vadose d'un aquifère calcaire, consistant en injectant à la surface un mélange de traceurs et en prélevant dans une galerie située à 10 m de profondeur. Les résultats montrent que des contaminants solubles et particulaires sont transportés à une vitesse supérieure de 10 m/h. Les courbes de restitution des particules (micro-organismes) montrent que leurs pics arrivent plus en avance et avec des concentrations relatives plus élevées, par rapport à celui du conservateur soluble, démontrant un effet d'exclusion de pores. Une injection en continue confirme ces résultats et illustre que plus de 40 % des passages d'écoulement, entre la surface et les points de prélèvement, sont inaccessibles aux particules. De plus, les différences dans la réponse des traceurs solubles et à la fois réactifs confirment qu'il y a une multitude de passages pour atteindre les points d'échantillonnage. Cela suggère également que l'hétérogénéité géochimique joue un rôle important sur le comportement des traceurs lors de leur transport dans l'epikarst. Cette étude met en évidence la complexité de l'écoulement de l'eau à travers la zone vadose d'un aquifère calcaire, et souligne la vulnérabilité de l'eau souterraine dans les aquifères fracturées en l'absence de sol.

1. Introduction

Tracing techniques are widely used for proving hydraulic connections, assessing groundwater flow parameters and quantifying mixing in the subsurface. These tests require conservative tracers be employed in order to obtain breakthrough curves that fully mimic the behaviour of the traced water. In contrast, comparative tracer tests compare a conservative tracer's response, which reflects water flow, to that of other tracer types with specific characteristics. This has been done in order to define the properties of the tracers and to evaluate their suitability for use in the field (BEHRENS *et al.* 1992, BÄUMLE *et al.* 2001).

This information, in turn, can be used to simulate the behaviour of specific reactive contaminants. Real contamination events in karst settings can rarely be investigated using appropriate sampling and monitoring strategies, particularly in fast flowing systems. In contrast, injection and sampling of tracers that act as contaminant surrogates under well-controlled conditions can provide

important information on the transport of relevant contaminants.

An alternative application of comparative tracer testing involves using tracers, whose interaction with particular materials is well characterised, to obtain an insight into conditions in inaccessible regions of the subsurface. This includes conditions in limestone bedrock. Consequently, comparative tracing may be used to characterise the structure and composition of the uppermost part of a limestone formation, i.e. the epikarst.

This paper presents selected results and conclusions from a 3-year project investigating solute and particle transport in heterogeneous unsaturated limestone bedrock. The collective responses of various tracers injected into the epikarst below a test site under variable injection conditions provide an insight into mass transport processes in the vadose zone of limestone aquifers as well as an indication of the role of heterogeneity along the flow path connecting injection and sampling intervals.

2. Field site and experimental setup

Experiments were completed at the Gänsbrunnen test site in the Swiss Jura Mountains, Canton of Solothurn. An artificial gallery located about 10 m below ground surface provided access to the vadose zone of a fissured moderately karstified Cretaceous limestone formation (Fig. 1). During heavy rainfall prior to the test programme, percolating water was observed entering through several discrete outlets in the ceiling of the gallery at rates of up to several l/min. These percolation points were related to fractures and bedding planes arising from structural heterogeneities in the limestone, some of which were subsequently widened by karstification

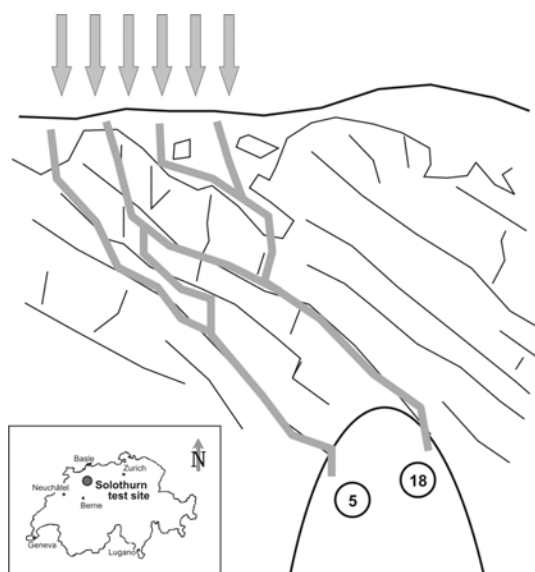


Fig. 1: Schematic illustration of the Solothurn test site. Irrigation and tracer application occurred at ground surface above an artificial gallery site. Samples were collected at discrete percolation point (outlets 5 and 18).

A series of tracer experiments completed at the site under controlled flow conditions at the ground surface above the gallery involved injecting a cocktails of reactive and non-reactive solute tracers and particle tracers in a diffuse manner over a 12 m² plot. Conservative solute tracers included Br⁻, I⁻ and Cl⁻, while reactive solutes consisted of a range of dye tracers, anions and cations. Particle tracers consisted of inorganic fluorescent microspheres and micro organisms (bacteria: *R. eutropha*, bacteriophages: H40/1 and T7).

Tab. 1: Irrigation and injection key characteristics.

	Injection type	Irrigation pattern	Flow conditions
Experiment type I	pulse	after injection	transient
Experiment type II	pulse	before + after injection	steady-state
Experiment type III	step	before + during injection	steady-state

Irrigation rate and tracer injection mode varied between experiments (Table 1). Irrigation rate varied from one experiment to the other but remained constant during each

experiment. Experiments lasted for at least 6 hours. Tracers were applied either as a pulse injection or as a continuous injection step.

Water samples were regularly collected from different percolation points in the gallery's ceiling during irrigation and stored in cool dark sample boxes prior to analyses. All analyses were completed at the Neuchâtel University, except microsphere analyses, which were determined by W. Käss, Umkirch/Germany. Tracer relative concentration (C/C₀) variation with time permitted breakthrough curves to be generated for each sampling point; the use of relative concentrations facilitated comparison of responses for different tracers as well as quantitative interpretation.

3. Results and interpretation

Solute transport and attenuation

Fig. 2 represents conservative solute breakthrough (iodine) at two percolation points (outlets 5 and 18) in the gallery under steady-state flow conditions, as reflected by stable discharge rates at sampling points in the gallery. These conditions, coupled with an instantaneous pulse injection provide a promising means of determining flow and transport parameters.

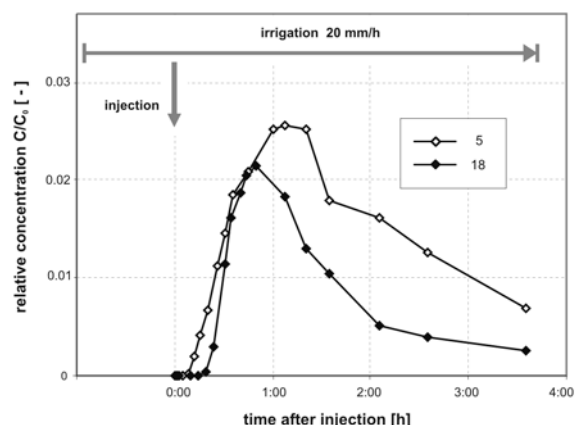


Fig. 2: Iodide (I⁻) solute tracer breakthrough curve for monitoring points 5 and 18 (experiment type II).

First detection of iodide tracer occurred within less than 10 minutes after injection. Peak arrival times indicate dominant flow velocities in the range of 10 m/h. Flow rate weighted mass recovery within the sampling period was 59 % for point 5 and 34 % for point 18; this was partially due to incomplete breakthrough at the end of the experimental period. Differences in tracer breakthrough at percolation points, which are placed at a distance of a few meters only, highlight the large-scale spatial heterogeneity of the epikarst zone. No water flow occurred in the gallery between these discrete percolation points. In the following, all figures refer to sampling point 5.

Conservative solute response illustrates the rapid arrival of diffusely-infiltrated water despite the moderate levels of karstification visible in the tunnel and the lack of visible preferential infiltration points at the ground surface. These results suggest that contaminants can be flushed into the system in the course of heavy rainfall events and thus rapidly penetrate into the vadose zone. On the other hand, long tailing indicates that a significant proportion of active flow paths have considerably longer residence times. This duality between flow components becomes very important

for assessing potential contaminant attenuation and groundwater vulnerability.

Low flow rates, and mobilisation of stored water, become particularly important when dealing with some reactive contaminants, for which a prolonged residence time favours additional attenuation. Comparative tracer testing employing reactive tracers shed considerable light on these processes. These tracers were regarded as surrogates for real organic and inorganic contaminants that may potentially impact karst groundwater quality thus allowing attenuation processes in the epikarst zone to be better evaluated. Tab. 2 summarises the responses of selected solute tracers relative to the conservative solutes (SINREICH *et al.*). Significant additional peak attenuation and mass loss are considered to be the result of sorption and/or degradation processes, according to the tracer-specific properties.

Table 2: Summary table of key parameters for selected reactive tracer breakthrough curves (all values relative to conservative solute) at sampling point 5.

	Nature of tracer	Rel. max. conc. C/C_0	Relative recovery
Sulpho G	organic	0.23	0.19
Pyranine	organic	0.26	0.32
PO_4^{2-}	inorganic	0.17	0.23
NH_4^+	inorganic	0.012	0.014
K^+	inorganic	0.008	0.010
Sr^{2+}	inorganic	< 0.004	< 0.0001

Particle transport and attenuation

Knowledge of mass transport rates through the vadose zone overlying aquifers also provides important information that is critical in determining whether prolonged residence times will afford a degree of protection from microbiological contamination to underlying water resources. Understanding particle response is helpful for interpreting the structure and relative importance of different flow paths through the system. Conservative and reactive solutes may use the same flowpaths depending on reaction rates; these rates ultimately control responses observed at observation points. Particle flow is fundamentally different because of the potential limitation of pore volume access. The comparative application of both solute and particle tracer types may thus discriminate between those fractions of the effective porosity that exclude particulate contaminants and those which allow them to pass through.

Fig. 3 displays breakthrough curves of some of the particle tracers employed for experiment type II (SINREICH *et al.*). Solute and particle breakthrough curves have distinctly different shapes. H40/1 and *R. eutrophia* breakthrough curves are less disperse and have peak concentrations that occur significantly earlier than those of the conservative solute, indicating exclusion processes. Similar responses are observed of T7 and the various sizes of fluorescent microspheres although they are considerably more attenuated. The lower concentrations of the latter can be explained by first-order kinetic sorption of these more reactive particles along the same flow paths.

Despite the earlier arrival of particle tracers compared to conservative solutes, this alone does not necessarily prove that exclusion processes are operating. ZHANG *et al.* (2000) effectively demonstrated that peak concentrations shift towards earlier times when first-order kinetic adsorption

operates. Under these circumstances, comparative plotting of solute and particle tracers with absolute concentrations is of little value in terms of visualizing exclusion processes. Comparison of relative concentrations eliminates this problem.

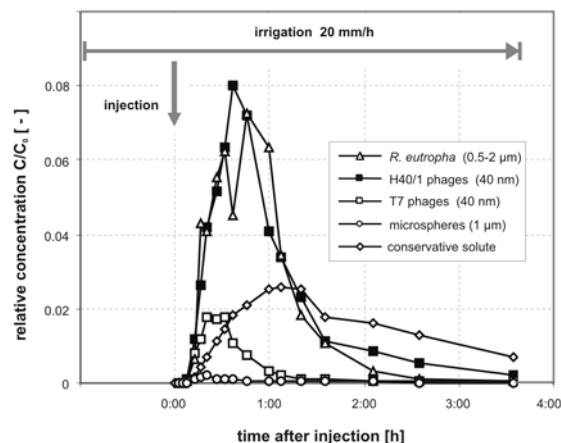


Fig. 3: Breakthrough plot of various particle tracers in comparison to the conservative solute.

The bacterial tracer *R. eutrophia* and the bacteriophage H40/1 reach maximum relative concentrations that are twice as large as those of conservative solutes, but display comparable dispersion one to each other. These data provide convincing evidence of the operation of exclusion processes during the tracer test. Moreover, the breakthrough curve for the 40nm diameter H40/1 phage tracer being similar to that of *R. eutrophia*, which is approximately 80 times larger, provide convincing evidence of pore exclusion is responsible for this behaviour (SINREICH & FLYNN). Such a result could not be obtained if size exclusion were the driving process for the particle response observed (Fig. 4).

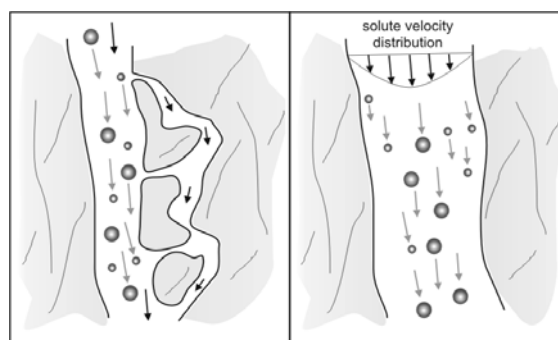


Fig. 4: Schematic description of pore (left) and size exclusion (right) processes in fractured rock.

Based on the above results it is apparent that, despite attenuation, micro organisms may be transported through the vadose zone more rapidly than non-reactive solute tracers.

Multi-component-flow and storage

In order to better characterise the different flow components below the test site, an alternative injection signal was applied, again under steady-state flow conditions. This involved applying the tracer cocktail to the irrigation plot at a constant rate for nearly 6 hours

(experiment type III). This approach allowed for a better quantitative estimate of the portion of the total pore volume that was accessible to certain tracer types (SINREICH & FLYNN).

Conservative solute and H40/1 phage breakthrough curves observed using this setup are plotted in Fig. 5. A very fast flow component is indicated by the steep rise of phage concentrations. 90 minutes after tracer injection started, H40/1 concentration plateaus out with about 0.6 times the input concentration and stops increasing despite sustained tracer injection. This suggests that all pore volumes accessible for particles are filled with tracer and that 40% of the phage tracer is permanently retained by the system under these conditions. Indeed, the portion of faster flowing water is certainly lower than 60 % of the total flux, since these channels also contain particles excluded from other flow paths accessible to water.

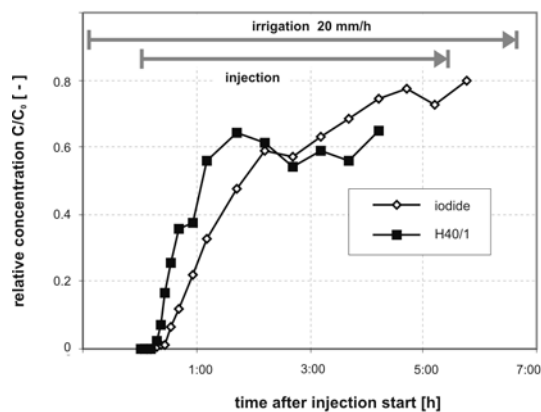


Fig. 5: Iodide and H40/1 concentration evolution during continuous tracer injection experiment.

Conservative solute concentrations rise more smoothly than particles and asymptotically approach a plateau of constant relative concentration of 0.8 approximately 5 hours after the start of injection. This portion corresponds to the sum flow components travelling along paths that can reach sampling points over the duration of the injection. A proportion of these pathways are accessible to solute and particulate tracers, whereas others are accessible to solute tracers alone.

At the end of the continuous tracer injection period, there was still a near-constant contribution of about 20 % of the total flux that was tracer free, which represented the proportion of total flow travelling along pathways that are insufficiently fast to permit tracer to reach sampling points within the experimental timeframe. The contrast in solute and bacteriophage responses suggest that particulate tracers are unable to reach the sampling points via these routes.

It is hypothesised that the conservative solute concentration would continue to rise asymptotically towards unity, were tracer application to continue. Part of the reason for this response is suspected to be a consequence of older water, present in the system prior to irrigation, being forced out of storage by the hydraulic action of the water being added at the ground surface. The detection of non-reactive tracers in the sample discharge water prior to tracer addition during this experiment indicates that some of the tracer applied during the previous experiment remained stored in the system during the intervening six month period.

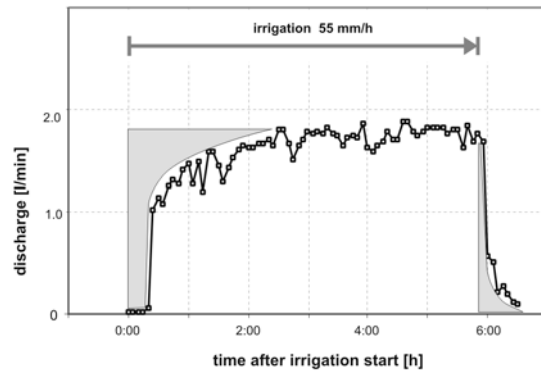


Fig. 6: Percolation discharge and related estimations for storage volume.

This is consistent with data presented in Fig. 6. Irrigation at a rate of 55 mm/h resulted in a steady-state inflow of 1.8 l/min at the percolation point 2.5 hours after irrigation started. Following this interval, flow into the gallery remained constant until irrigation ceased. After this point inflow rates into the gallery declined rapidly. Indeed the volume of water draining following the end of irrigation was considerably less than the volume flowing into the gallery prior to onset of steady-state flow. These results suggest that additional water may have been taken into storage in the interval between the start and end of the experimental period.

Rock-tracer interaction

Fig. 3 presents evidence of the strong attenuation of some of the particle tracers employed during pulse injection experiment, such as T7 and the microspheres. Even for the less reactive H40/1 phage type, the step injection experiment shows a permanent attenuation rate of 0.4 (Fig. 5).

These results are suspected to be a consequence of particle filtration along flow paths between injection and sampling points. Filtration depends on both the sticking efficiency of the particles and their collision frequency with reactive surfaces (ELIMELECH et al. 1995). Sticking efficiency is mainly a function of the physical-chemical surface properties of both the particles and the rock, such as charge and hydrophobicity. This in turn is a function of the chemistry of the water in the system. Collision frequency, i.e. the frequency of contact between particles and rock surfaces, is mainly controlled by physical aspects of the system such as particle size and the aperture of pores and fractures, as well as the flow characteristics.

The upper part of Fig. 7 represents the same experiment as Fig. 5, but over a longer period and with an additional breakthrough curve for the T7 phage and electric conductivity data; the latter acts as a surrogate for the ionic strength of the water being sampled. The high ionic strength in the irrigation water during the experiment was due to the high load of ionic tracers. The gradual increase in ionic strength provided much important additional information about particle filtration in the rocks underlying the site (FLYNN & SINREICH).

A strong decrease of phage concentration occurred at 40 minutes after injection start for T7 and 4 hours later for H40/1 (Fig. 7). T7 concentrations declined more gradually with increasing electric conductivity until no phage were detected in samples despite constant tracer application. This

behaviour contrasted with that of H40/1, which maintained a constant concentration until a critical deposition point was reached when concentrations declined dramatically. Associated laboratory-based batch experiments demonstrated that phage inactivation/ adsorption at the air-water interface were negligible processes suggesting that both viruses were predominantly adsorbed to rock surfaces under the prevailing chemical conditions (FLYNN & SINREICH). This is corroborated by a dramatic release of both phage types following the application of low ionic strength water to the system at the end of tracer injection. The lower part of Fig. 7 schematically displays the effect of changing sticking efficiencies during the different phases of the experiment.

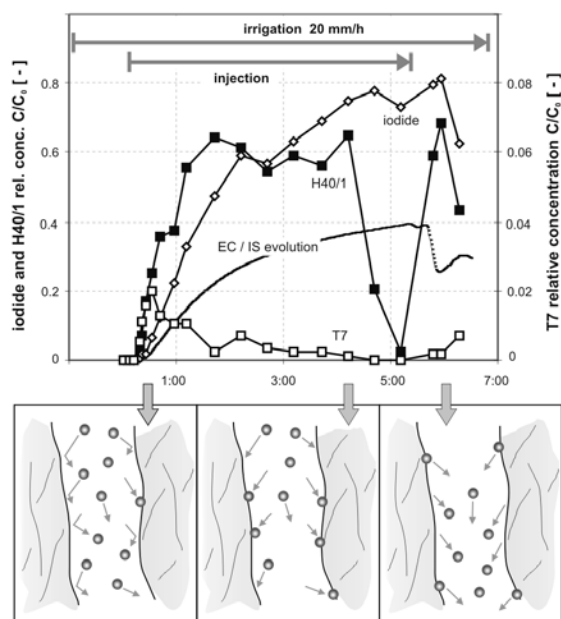


Fig. 7: Breakthrough curves for selected tracers in experiment type III. Full sorption of T7 and H40/1 phages occurs when reaching critical ionic strength (IS) resp. electric conductivity (EC) values (above). According conceptual scheme for three experiment phases (below).

These results illustrate, on the one hand, how changing hydrochemical conditions influence the various particle tracers employed, with T7 being much more sensitive to ionic strength increase than H40/1. On the other hand, the experiment strongly indicates that all of the particle tracers employed have very high collision frequencies, i.e. virtually all particles can collide with fixed surfaces.

Solute tracer responses are also believed to reflect geochemical variability in the epikarst along flow paths connecting the the irrigation zone with sampling points (Fig. 8). The organic dye tracer Sulforhodamine G has a well characterised sorption affinity to other hydrophobic materials such as organic matter (KÄSS 1998). Breakthrough curves for this tracer demonstrated a reduced tendency to react along flow paths with short travel times compared to tracers arriving later. This in turn indicates that the surfaces along these later paths contained more hydrophobic materials, possibly due to fissures filled with soil matter comparable to those observed in outcrop close to the test site. Following the same line of reasoning, phosphate tracer appears more attenuated along flow paths with low residence times.

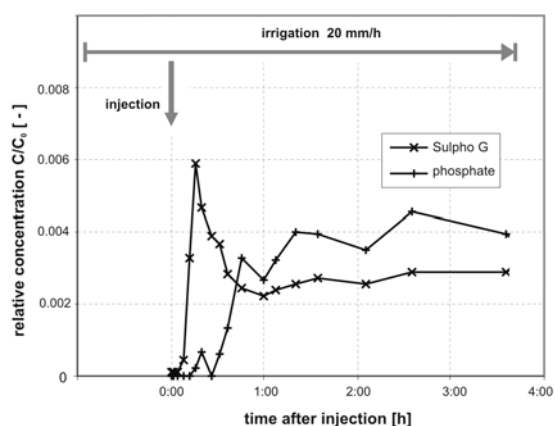


Fig. 8: Sulforhodamine G and phosphate breakthrough curves for experiment type II with prevailing attenuation for differing flow components.

Preliminary hydrochemical modelling for experiment type III results indicate ionic exchange as the predominate process responsible for the gradual change in phosphate and other ions in tracer samples (Laurent André, pers comm.).

4. Conclusions and conceptual model

The facilities available at the experimental site, together with specific experimental setup, permit specific flow and storage characteristics of the vadose karst zone to be investigated in detail. Since the soil cover above the limestone is very thin, the main effects on water drainage and tracer transport can be attributed to the epikarst layer. Its dual function for both focusing recharge into preferential vertical flow paths and for storing infiltrating water makes it an important sub-system of the vadose karst zone.

Experiments at the Solothurn site provide an insight into phenomena that can be encountered in the vadose zone of karst aquifers including the following:

- Flow heterogeneity within epikarst,
- rapid transport of potential contaminants despite diffuse infiltration and moderate karstification,
- attenuation potential for reactive solute and some particle contaminants,
- exclusion processes leading to earlier arrival of particles (including micro organisms) relative to solutes,
- multi-component-flow with fast preferential and slower pathways,
- long-term storage providing prolonged time periods for solute and micro-organism attenuation,
- very high collision frequency of particles in conduit along fracture flow paths,
- the role of geochemical heterogeneities in generating irregular breakthrough curves.

These phenomena could be highlighted by means of evaluating different injection scenarios with different tracer types. Regarding the epikarst as a multi-component system and using well defined tracer experiment input and output functions allows for quantification of the components reaching sampling point via different flow paths. For a multi-component flow, solute breakthrough can be split into several partial breakthrough curves representing separate populations of flowpaths (BEHRENS *et al.* 1992).

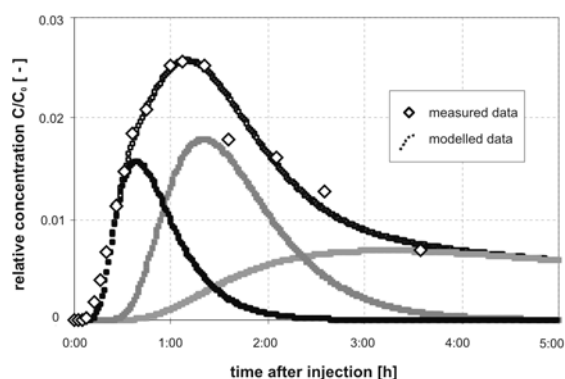


Fig. 9: Preliminary multi-flowpath model for solute tracer breakthrough.

Breakthrough curves presented in Fig. 9 reflect analytical solutions for non-reactive solute transport, simulating overall breakthrough using three flow components based on concepts developed in the above experiments. Reactive solute tracer response can be deduced by adding a kinetic and/or equilibrium sorption terms to the various components of flow.

A comparable hydrogeological conceptional model for transport through epikarst has also been established in terms of a multi-reservoir model using discharge data (PERRIN & KOPP 2005). Fig. 10 schematically summarises the epikarst at the experimental site as a three-reservoir system transporting irrigation/rainfall water at the surface and discharge it to the gallery at 10 m below ground surface.

The model tentatively proposes that the fast flow component results from preferential fracture/conduit flow. Slower flow components result from flow through finer fissures and/or fissures filled with soil material. A very slow component arises due to water flowing through very fine fissure network or limestone porous matrix.

The conceptual model is considered preliminary without integrating all data obtained from the experimental series. It neglects potential lateral boundary effects, which were suspected to be of minor influence.

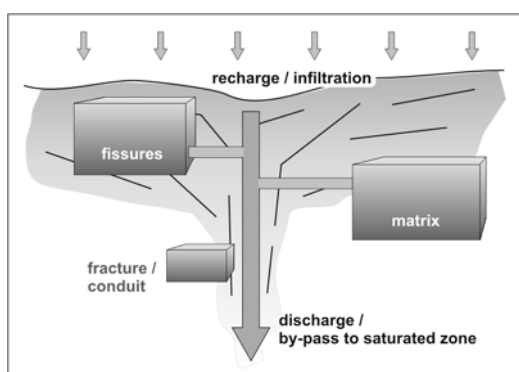


Fig. 10: Conceptual three-reservoir model at the epikarst test site.

Some specific conclusions became apparent in the scope of this research with respect to groundwater protection, which may also be valid for the saturated zone of a karst aquifer:

- Specific transport and attenuation for individual contaminant types, as well as the contamination

scenario, should be taken into account when dealing with groundwater vulnerability assessment.

- Exclusion processes should be accounted for to explain early particle arrival. Solute response may not always be an appropriate indicator for microbiological transport and attenuation.
- Size alone has proved inadequate for understanding particle fate and transport at the site, as illustrated by the comparative behaviour of fluorescent microsphere tracers and similarly sized *R. eutrophia* bacteria, with attenuation rate of the spheres 30 times that of the bacteria.
- The chemical composition of water plays an important role in particle attenuation. Rapid changes in hydrochemistry are frequent in the course of recharge events in karst aquifers. The increase of microbiological contamination during recharge events may thus partly be explained by desorption of previously adsorbed particles due to the presence of lower ionic strength water rather than the transport of fresh infiltration containing micro-organisms. This phenomenon will depend strongly on the rate of die-off of adsorbed micro organisms.

Understanding attenuation processes in the vadose zone overlying karstified aquifers is essential if water quality is to be adequately protected. The results presented from this site are believed to provide information that may also valid for the epikarst zone of other moderately karstified and diffusely recharged limestone terrains.

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Integrated vulnerability and risk mapping – a multi-purpose groundwater management tool

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Abstract

Numerous empiric and mathematical methods had arisen to ascertain the vulnerability of groundwater resources towards contamination. An integrated approach on vulnerability and risk assessment emanates from European COST Action 620 dealing with the protection of karst aquifers. It implies a vulnerability-hazard-risk approach that allows for generating maps for different purposes. Intrinsic vulnerability assessment conjoins the consideration of four parameters considering precipitation, runoff, unsaturated and saturated zone. The integration of point recharge and preferential flow phenomena make the assessment applicable to karst environments. In order to provide contaminant-specific vulnerability maps as well, physico-chemical processes, by which particular contaminant types are dominantly affected, may be incorporated into the assessment. Finally, a land use module distinguishes different impact scenarios, which allows risk maps to be compiled. Taking into account both the vulnerability of a hydrogeological system and specific contamination scenarios makes the method applicable to real questions arising from groundwater protection and land use strategies. The proposed approach is thus believed to be an appropriate and flexible tool for managing diverse activities in karst terrains.

Résumé

De nombreuses méthodes mathématiques et empiriques ont été développées pour l'évaluation de la vulnérabilité des ressources en eaux souterraines faces aux contaminations. Une nouvelle approche, émanant du projet européen Action COST 620 visant à la protection des aquifères karstiques, intègre l'évaluation de la vulnérabilité et du risque. L'établissement de ces cartes implique de tenir compte à la fois de la vulnérabilité, du danger et du risque. L'évaluation de la vulnérabilité intrinsèque dépend des quatre paramètres suivants : précipitation, ruissellement, zones saturée et non-saturée. L'intégration du type de recharge et le phénomène de flux préférentiel rendent l'évaluation applicable aux environnements karstiques. En tenant compte des processus physico-chimiques affectant les contaminants, il est possible de fournir des cartes de vulnérabilité spécifiques. En dernier lieu, les informations sur l'utilisation du sol et les différents scénarios d'impact complètent la réalisation de cartes de risques. Cela rend la méthode applicable aux cas concrets, dans le cadre décisionnel de la protection des eaux souterraines et de l'occupation du sol. L'approche proposée est considérée comme outil universel et flexible pour la gestion d'activités diverses sur des terrains karstiques.

1. European COST Action 620

Groundwater vulnerability and risk mapping has been proven to be a valuable practical tool for land use planning and groundwater protection, by helping to avoid the contamination of water present beneath pollution-sensitive land.

New standards for integrated vulnerability and risk assessment for carbonate (karst) aquifers were set by the European COST Action 620 (ZWAHLEN 2004). It had become apparent that many existing methods did not adequately consider heterogeneous geo-environments, i.e. karst-specific phenomena, for assessing vulnerability of groundwater resources towards contamination. Furthermore, vulnerability maps alone are not sufficient to answer real questions arising from groundwater protection and land use strategies. Integrated risk assessment result in a set of maps that may provide basic information to groundwater resources managers and decision makers.

COST 620 proposed an intrinsic/specific vulnerability and risk assessment, following an origin-pathway-target approach. It has been designed for karst, but what is relevant for the most heterogeneous medium is also valid

for less contrasted geo-environments, such as fissured hardrocks and granular sediments. The main goal is to answer three questions (BROUYÈRE *et al.* 2001): (i) when will a contaminant appear at the outlet ? (ii) to which level of concentration ? (iii) during which time ?

The principle which guided vulnerability assessment was the cumulative effect of superimposed layers: topsoil, subsoil, unsaturated hardrock and saturated zone which can act one after the other one. Such a *multi-layer approach* allows for considering the stratigraphic heterogeneity of a geological setting. This is particularly important, when thin layers become very efficient in terms of reactive contaminant retention.

Key parameters were defined that determine different processes occurring in the layers and being able to protect groundwater towards contamination. Such processes are related to transport and storage (advection, dispersion, dilution) as well as attenuation (retardation, degradation) in case of reactive and/or non-persistent contaminants. A *multi-process approach* enables each type of contaminant being adequately tackled.

Different kinds of output maps arise from the vulnerability-risk assessment. Each of them can be used, or is appropriate, for a different purpose. The *multi-purpose*

approach may be completed by employing the assessment for specific contaminants that are of concern in the investigation area.

The COST 620 vulnerability and risk approach has been implemented, among others, to a carbonate catchment in Southern Spain (ANDREO *et al.* 2006) and to a karst basin in the Swiss Jura Mountains with focus on contaminant-specific assessment (SINREICH *et al.* 2005). For the latter site, we illustrate the qualitative implementation of the integrated risk approach. Vulnerability and risk maps are presented for a micro-biological contamination, i.e. for faecal coliforms, indicating pathogenic bacteria.

2. Vulnerability assessment

Intrinsic vulnerability mapping provides a generalised view of vulnerability. However, due to the numerous kinds of contaminants to which groundwater catchments are exposed, specific vulnerability mapping is also required. The impact of heavy metals, hydrocarbons, pesticides or micro-organisms, resulting from agriculture practices (irrigation, tillage, stock rearing etc.), accidents and industrial pollution often leads to groundwater contamination. While intrinsic groundwater vulnerability refers to hypothetical conservative and persistent substances, each particular contaminant is still characterised by its own nature and behaviour. Thus intrinsic vulnerability inadequately characterises contaminant fate and transport as it accounts only for the inherent geological and hydrogeological setting of an area, but makes no allowance for the nature of the contaminants concerned. In contrast, specific vulnerability of groundwater refers to a particular contaminant or group of contaminants, accounting for the properties of the contaminants in addition to the intrinsic vulnerability of an area (Figure 1).

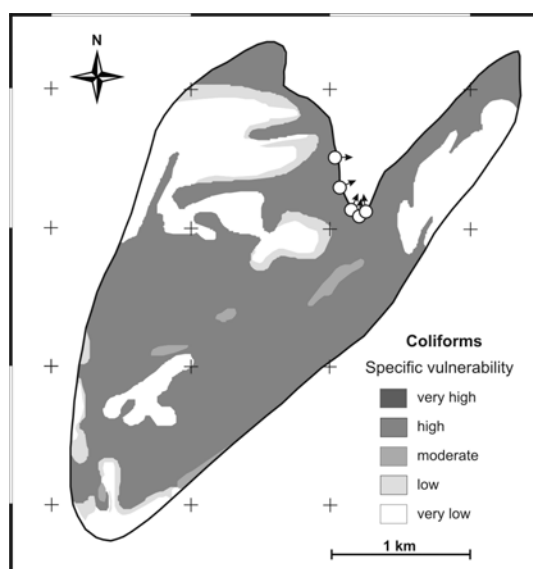


Figure 1: Specific vulnerability map.

The COST 620 intrinsic vulnerability concept considers four relevant factors (DALY *et al.* 2002, ZWAHLEN 2004):

- the layers overlying a karst groundwater resource (O-factor),
- the concentration of flow due to runoff and point infiltration (C-factor),
- the influence of precipitation regime (P-factor),

- and the saturated karst layer in case of source protection (K-factor).

In addition to these four factors, specific vulnerability assessment comprises another factor that takes into account

- the specific attenuation capacity of a particular contaminant type (S-factor).

3. Risk assessment

Figure 2 presents the results of a hazard mapping for bacterial contamination in the catchment (CLAVIEN 2003).

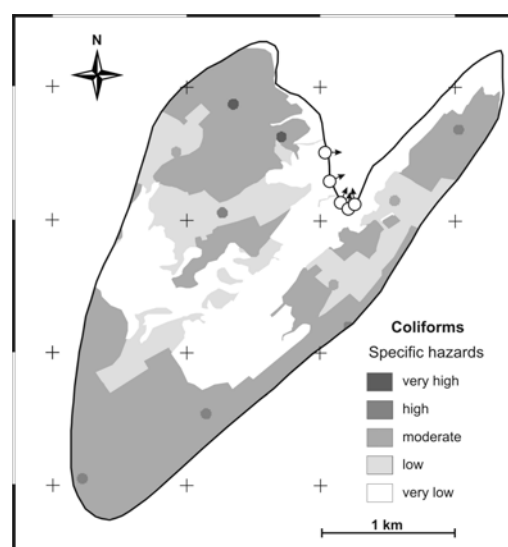


Figure 2: Hazard map.

A hazard is defined as the potential source of a contamination caused by human activity. In rural areas, pathogen contamination of drinking water supplies represents one of the major problems in terms of land management. Pathogens and their coliform indicators derive from cattle rearing and manuring representing diffuse hazards, as well as from septic tanks and stabling representing point contamination sources.

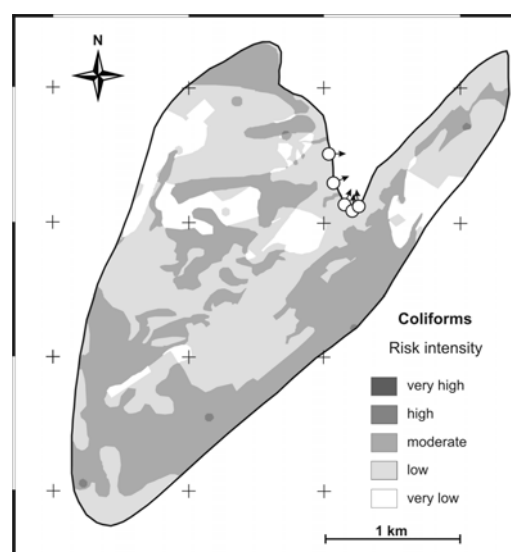


Figure 3: Risk intensity map.

COST 620 defined two types of risk, risk intensity and total risk. Risk intensity (Figure 3) can be evaluated by the intersection of vulnerability and hazard maps. Risk intensity maps provide an overview, on which surfaces a contamination of groundwater is likely to occur or not.

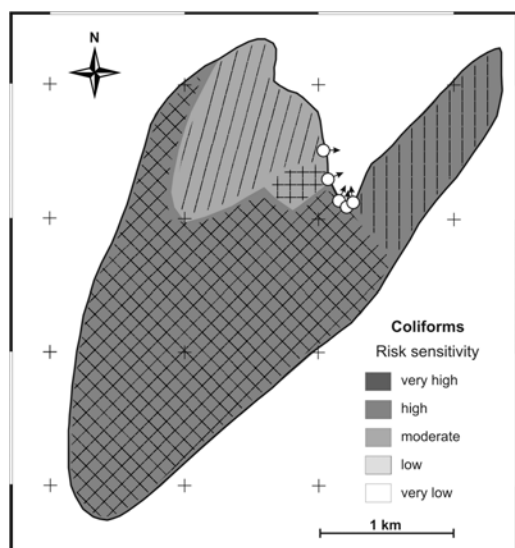


Figure 4: Risk sensitivity map.

COST 620 also highlighted the importance of risk sensitivity being incorporated into the risk assessment procedure. In doing so, total risk can be ascertained, which is a linkage of the degree of a potential contamination event with the costs of the consequences if the event actually occurred (REES 2002). Hence, risk sensitivity assesses the social and economic value of a groundwater resource, or vice versa, the damage that may result from a given risk intensity. Groundwater valuing can be addressed to an in-situ resource value or an abstraction value for water supply, whereas various approaches exist for both of them (NRC 1997).

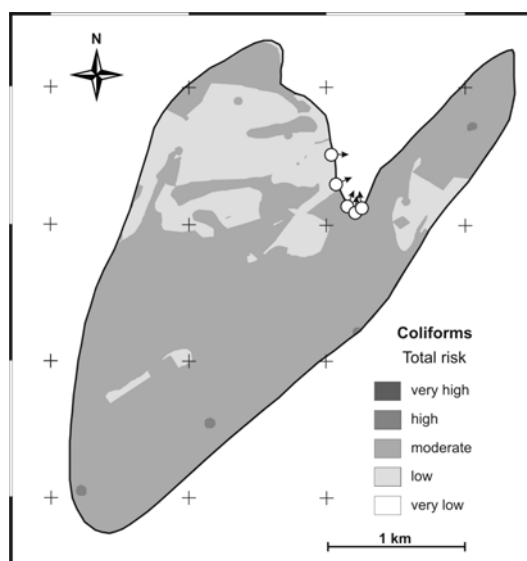


Figure 5: Total risk map.

This study considered two main factors for the assessment of risk sensitivity: contaminant toxicity and

resource importance for drinking water supply (Figure 4). The classification was done for each sub-basin. Risk sensitivity is high for a catchment of captured springs and harmful contaminants, while non-exploitable resources and harmless substances result in low sensitivity classes.

Finally, a total risk map can be compiled (Figure 5) by combining information from Figure 3 and 4.

4. Integrated multi-purpose approach

The combination of individual maps in the frame of the risk assessment is displayed in Figure 6. This set of maps provides a comprehensive and flexible tool for land-use management and water protection.

Intrinsic vulnerability maps generally display a worst case scenario and fail to take into account the positive effects deriving from specific contaminant properties such as retardation and degradation. This ignores the additional attenuation potential of contaminants under certain conditions and thus may overestimate the vulnerability of an area. Contaminant-specific vulnerability and risk maps are more sophisticated and contain more information than intrinsic maps reflecting a more detailed potential migration picture of a particular contaminant. They may be a valuable component of an integrated risk assessment, playing an important role for a comprehensive environmental management. However, contaminant-specific assessment should only be applied when a particular need can justify the supplemental costs of obtaining necessary additional data.

The logic behind the vulnerability assessment is consistent with experimental results obtained from tracer experiments in the unsaturated and saturated karst zone. Comparative tracing tests employing reactive tracers as surrogates for real contaminants provide a useful means for point validation of vulnerability and risk maps. Global catchment validation can be done by means of natural tracers. Their response at the outlet of a karst system needs to be consistent with the spatial statistics of vulnerability classes. Similarly, real contamination events can be used to validate risk maps.

Specific vulnerability and risk assessment may become more effective by using quantitative methods. Field measurements (tracer testing) and flow/transport modeling are currently being adopted for calibration and validation of this approach (POCHON *et al.* 2004, SINREICH *et al.* 2005). This is anticipated to allow specific contamination scenarios to be simulated.

Vulnerability and risk maps are not self-explanatory but need to be interpreted. This in turn may lead to misunderstandings, in particular on the part of non-specialists. In order to act as an efficient tool for decision makers, and to avoid misinterpretation the meaning of the different map types needs to be clarified. With currently available GIS tools, several maps can be produced relatively easily. They may be used for a variety of purposes including

- optimisation and reduction of groundwater protection zones for dominant contaminant groups (vulnerability maps),
- identification of land mismanagement, reorganisation, best practices and future planning (risk maps).

Both vulnerability and risk maps are screening tools. While vulnerability maps are a static tools and normally don't change with time, hazard maps may be adapted to change in land use in order to obtain dynamic risk map. The

implementation of best management practices may only be successful with considering the specific hydrogeological aspects of a karst aquifer, as has been shown by CURRENS (2002).

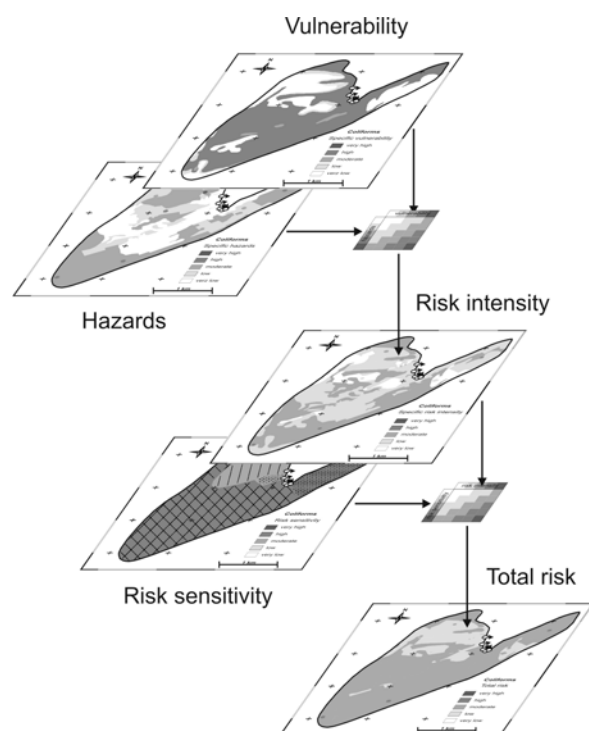


Figure 6: Procedure of the integrated vulnerability and risk assessment.

A one-dimensional perspective is not adequate for an integrated approach. Socio-economic aspects will be more and more merged with hydrogeological knowledge towards multi-purpose and multi-disciplinary approaches. According to REES (2002), one has to consider "risk mitigation as an economic good".

Acknowledgements: Experts from 15 European countries were involved in the COST Action 620. The vulnerability and risk approach that is implemented in this study is a collective work of all COST 620 members. We are particularly grateful to our colleagues from Working Group 2 on "Specific vulnerability".

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Statistical comparison and control of different vulnerability mapping methods in Bauschlotter Platte aquifer (Germany)

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Abstract

Seven methods for vulnerability mapping have been applied on a karst aquifer in southwest Germany. The different methods were compared and screened by statistical techniques. The results of the statistical analysis shows that DRASTIC, PI and COP have a high correlation between the main considered factors of the unsaturated zone and the mapped vulnerability classes, whereas GOD and EPIK give minor importance on the protection capacity of the overlying strata. The DRASTIC method can be applied on areas without point infiltration conditions. Among the tested vulnerability mapping methods COP and PI are better representing the hydrogeological conditions of the test site. DRASTIC gives also good results for the main part of the area, but this method fails if there is allogenic and point recharge.

Résumé

Dans le site de Bauschlotter Platte, un aquifère fissuré et peu karstifié, sept méthodes de cartographier la vulnérabilité à la contamination ont été appliquées. Les résultats des différentes méthodes ont été comparés par statistiques. On constate une forte corrélation entre la carte de vulnérabilité et les facteurs principaux de la zone non saturée avec les méthodes DRASTIC, GOD, PI et COP, tandis que GOD et EPIK donnent moins d'importance à la capacité de protection fournie par la zone non saturée. La méthode DRASTIC peut être appliquée dans des zones sans infiltration ponctuelle. Parmi les méthodes de vulnérabilité appliquées, COP et PI s'avèrent les plus appropriées aux caractéristiques hydrogéologiques de l'aire d'étude. DRASTIC donne aussi de bons résultats pour la plupart de la zone étudiée, sauf dans les zones d'infiltration ponctuelle.

1. Introduction and objectives

The main goal of this work is to determine, which of the methods applied in the test area is more suitable to assess the groundwater resource vulnerability according to the hydrogeological characteristics. According to this topic several works by different techniques were made to compare the vulnerability maps in others test sites, i.e. CORNIELLO *et al.* (1997), GOGU *et al.* (2003), NEUKUM & HÖTZL (2005) or VÍAS *et al.* (2005).

By means of the aquifer characteristics, it is possible to determine the most influencing factor in each vulnerability mapping method. Generally, the principal factors controlling vulnerability have been established by VRBA & CIVITA (1994): recharge, soil, unsaturated zone and saturated zone. In the COST Action 620 (ZWAHLEN, 2004) the main factors to assess the resource vulnerability in carbonate aquifers were determined: overlying layers (capacity of attenuation in the unsaturated zone, O factor); flow concentration (conditions of the surface to bypass unsaturated zone, C factor) and precipitation (transport). For source vulnerability also the karst network is considered (K factor). Flow concentration (C factor) is specific of carbonate aquifers and is directly related to the development of the epikarst. Therefore in karst aquifers with well developed epikarst, C is one of the most important factors. In low karstified aquifers the epikarst is low developed and the C factor is less important but the O factor controls the degree of vulnerability.

VÍAS (2005) correlated the aquifer characteristics and the main vulnerability factors by means of statistical analysis in two aquifers in southern Spain (one karstic and one fissured aquifer). It was demonstrated that the C factor is more correlated with the vulnerability in the karstic aquifer whereas the O factor controls the vulnerability in low karstified aquifers. Taking into account that every factor includes a group of variables, it is possible to assume the influence of the main factors and variables that control the vulnerability in each kind of aquifer (Figure 1).

The geological sequence of the test area consists of Triassic sedimentary rocks (Muschelkalk and Keuper). The main part of the area is occupied by limestones, marls and dolomites of the Upper Muschelkalk with a karstification, which is considered to be rather moderate to low developed. Accordingly only a few karst features are developed at the surface. However, tracer tests at dolines and sinking streams have proven very fast subsurface connections over large distances with first arrival mean velocities up to 180 m/h (TENHÄFF & KÄSS 1987). The Triassic rocks are covered sporadically by Quaternary loess and alluvial sediments with highly variable thickness. In addition the whole area is covered by soils with varying thickness between 0.3 and 1m. The recharge is mainly diffuse and preferential infiltration points are found only in a few dolines where sinking streams feeding the aquifer temporarily.

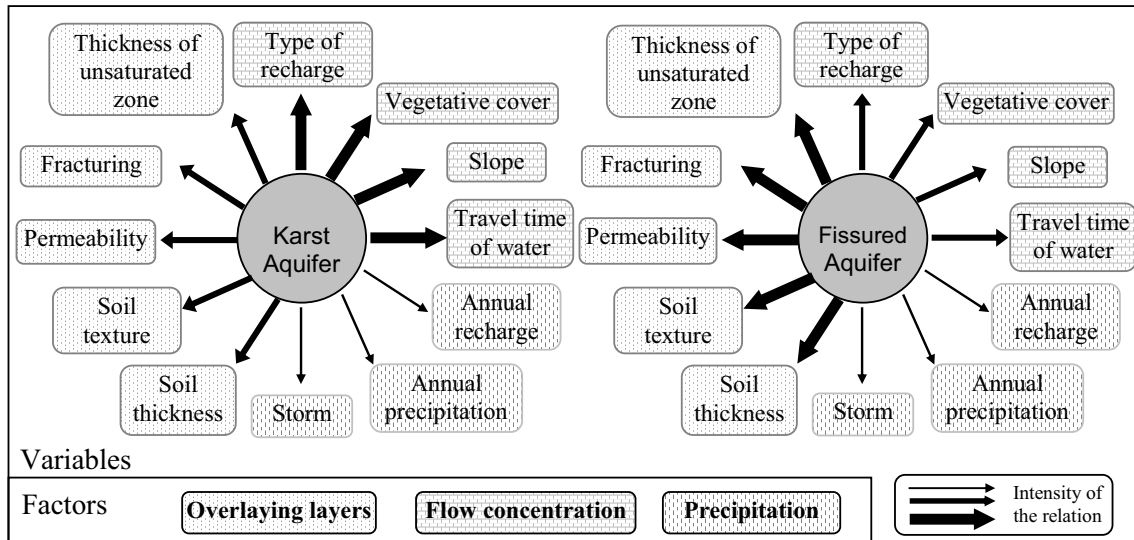


Figure 1. Influence of main variables and factors on the resource vulnerability of carbonated aquifers both karstic and fissured.

2. Methodology

To analyse the obtained vulnerability maps of the Bauschlotter Platte, two statistical techniques have been applied: correlation and regression coefficients.

Statistical tools can be used to address four objectives: detecting differences, predicting measurements, explaining trends and exploring data interactions (KUFS, 1992). To explain trends in the spatial pattern of variability in a dependent variable, the relative importance of each independent variable is used.

The applied statistical parameters are Pearson and Regression coefficient. The Pearson Product Moment r , (equation 1) estimates the correlation between two variables. Their values are in the range from +1.0 to -1.0. These values mean a perfect correlation of points on a line but 0.0 means no relationship between the variables.

$$r = \frac{n(\sum xy) - (\sum x)(\sum y)}{\sqrt{[n \sum x^2 - (\sum x)^2][n \sum y^2 - (\sum y)^2]}} \quad [1]$$

x = independent variable, y = dependent variable and n = number of cases

To analyse the strength of the relationship between variables the determination coefficient (R^2) has been also used. This statistic is the square of the Pearson coefficient and it establishes the proportion of variance in the dependent variable that is accounted for the independent variable. The determination coefficient has been classified according to the variance into three regular intervals (table 1).

Values of Pearson coefficient (r)	Values of Determination coefficient (R^2)	Class of correlation
0 - 0.59	[0 - 0.36)	Low
0.60 - 0.79	[0.36 - 0.64)	Moderate
0.80 - 1.0	[0.64 - 1]	High

Table 1. Values and classes of both coefficients: Pearson and Determination.

Regression coefficient b (equation 2) allows assessing the relative weight of every independent variable (vulnerability factor) on the dependent variable (vulnerability index) by a linear regression model (minimum square). The values of b are influenced by the variable values; therefore the results are normalized to reduce errors in the analysis of the coefficient.

$$b = \frac{n(\sum xy) - (\sum x)(\sum y)}{n \sum x^2 - (\sum x)^2} \quad [2]$$

x = independent variable, y = dependent variable and n = number of cases

Independent of the applied statistic (correlation or regression), the data have to be pre-processed to get a map of unique conditions subareas. This consists of one or more zones where a unique combination of every factor values is used to calculate the vulnerability index. In practice, this stage is performed using the GIS "combine" function by ARCGIS. Otherwise the results can be influenced by the dataset. Redundant data have been removed from the datasets to avoid any effect on the correlation coefficient values.

3. Results

Seven methods of intrinsic vulnerability were carried out in Bauschlotter Platte aquifer: AVI (STEMPVOORT *et al.*, 1993), COP (VÍAS *et al.*, 2006), DRASTIC (ALLER *et al.*, 1987), EPIK (DOERFLIGER *et al.*, 1999), GLA (HÖLTING *et al.*, 1995), GOD (FOSTER, 1987) and PI (GOLDSCHIEDER *et al.*, 2002). The spatial proportion of each class in the resulting vulnerability maps are very variable (figure 2). COP assesses most of the area to be low vulnerable. In contrast GLA does estimate most of the area to be very high vulnerable (figure 2). All other methods are somewhere in between these two extremes.

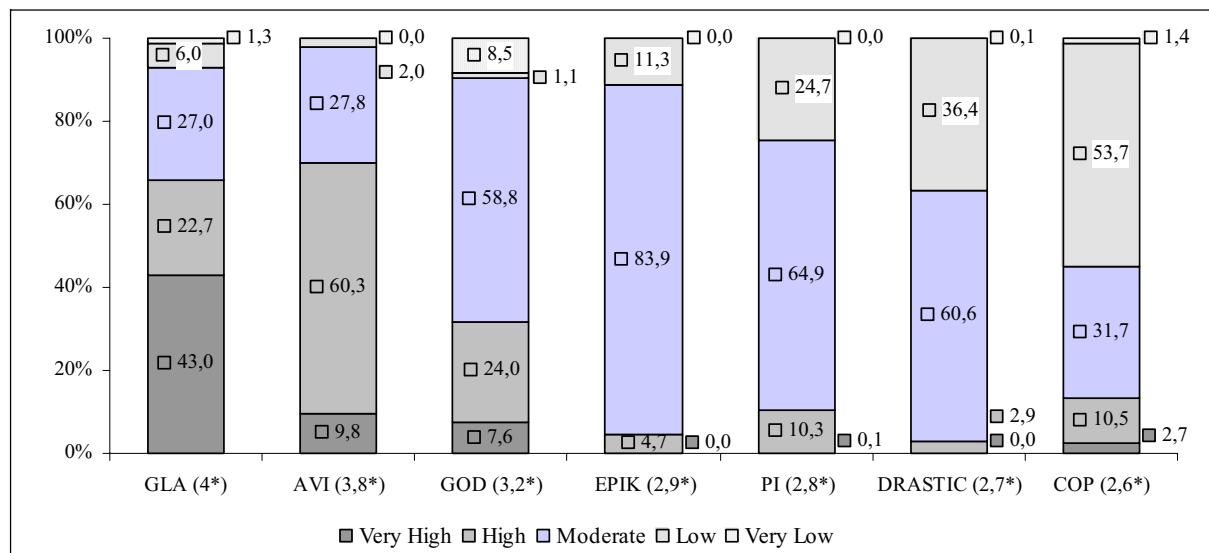


Figure 2. Percentage of surface for every vulnerability classes in Bauschlotter Platte aquifer by several methods. * = mean value of vulnerability from 5 (max) to 1 (min).

The statistical indicators, i.e. r and b coefficients, show the highest values in the test area for those factors, which are connected to the unsaturated zone (D in DRASTIC, G and D in GOD, O in COP and P in the PI-Method) (table 2).

Factor	Correlation		Regression
	r	R^2	b
COP - C	0.51	0.26	0.44
COP - O	0.86	0.75	0.82
COP - P	0.00	0.00	0.12
DRASTIC - D	0.68	0.47	0.72
DRASTIC - R	0.24	0.06	0.270
DRASTIC - A	0.20	0.04	0.24
DRASTIC - S	0.26	0.07	0.271
DRASTIC - T	0.13	0.02	0.15
DRASTIC - I	0.48	0.23	0.51
DRASTIC - C	0.43	0.19	0.12
EPIK - E	0.67	0.44	0.65
EPIK - P	0.28	0.08	0.18
EPIK - I	0.69	0.48	0.69
EPIK - K	0.17	0.03	0.22
GOD - G	0.68	0.46	0.639
GOD - O	0.58	0.33	0.10
GOD - D	0.60	0.36	0.64
PI-P	0.88	0.77	0.80
PI-I	0.57	0.33	0.41

Table 2. Statistical coefficients (r , R^2 , b) for every factor. Analysis between factors - index for each method applied.

EPIK is the exception by having the highest correlation with parameters, which are associated to the flow concentration (E and I). This is mainly a result from the differing weighting factors applied in EPIK which stress on infiltration condition and epikarst development (DOERFLIGER *et al.*, 1999).

The results observed are logical for the majority of the applied methods because the Bauschlotter Platte aquifer is a moderate to poorly karstified aquifer and diffuse infiltration is the main recharge process.

For GLA and AVI methods it has not been possible to obtain results due to their algorithm to calculate the vulnerability index. Both methods are using solely the overlaying layers (O factor) to assess vulnerability but the flow concentration (C factor) is not incorporated into the assessment schemes or others factors. The importance of every parameter might be varying from layer to layer of the unsaturated zone. Therefore, the correlation between the factors and the vulnerability index will show also strong spatial variations between the surface and groundwater table.

4. Conclusions

The statistical coefficients r and b show good and coherent results between the factors and the vulnerability maps with respect to the characteristics of the aquifer. The Bauschlatter Platte is a low to moderate karstified aquifer and the highest correlation is found for the factors which are connected to the Overlying layers and linked to the protection capacity of the unsaturated zone.

Due to the mainly diffuse recharge of the karst aquifer with minor flow concentration in the epikarst zone those methods are most preferable which give the main influence to the protection capacity of the unsaturated zone. These are COP, DRASTIC and PI. It is important to mention that DRASTIC is not applicable for dolines and sinking streams and their near catchments as it does not involve allogenic and point recharge.

The EPIK results reflect the weighting factors of each vulnerability parameter, giving the strongest influence on epikarst development and punctual infiltration conditions. It is therefore not really adequate for this kind of karst aquifer investigated in this study. GOD gives the main influence on groundwater confinement and to the thickness of the unsaturated zone. The overlying strata have the weakest importance for the vulnerability map according GOD.

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Description of the cover pictures:

Front cover:

Kapuzbaşı spring, Kayseri region, Taurids, central Turkey. This impressive karst spring outflows from the autochthonous Jurassic and Cretaceous carbonate formations of the Aladaglar Massif.

Photo: Jacques Mudry.

Back cover:

Up: The Boka Spring near Bovec in the Soča Valley in the Slovenian Alps, one of the most beautiful karst springs in the world. The spring discharge is up to 100 m³/s, and the two-stage waterfall is 144 m high. Photo: Nico Goldscheider.

Medium: Karst drainage gallery in Triassic carbonate rocks for the drinking water supply of the city of Hall in Tirol, Austrian Alps. Photo: Nico Goldscheider.

Bottom: Injection of 200 g of uranine into the entrance of the Schwarzwasser cave, probably the largest estavelle in the Alps, located in the Hochifen-Gottesacker karst area, Austro-German Alps.

Photo: Nico Goldscheider.

Description des photos de couverture :

Recto :

Source de Kapuzbaşı, région de Kayseri, Taurides, Turquie centrale. Cette impressionnante source karstique émerge des formations carbonatées jurassiques et crétacées autochtones du Massif d'Aladaglar.

Photo: Jacques Mudry.

Verso :

Haut : La source de Boka, près de Bovec dans la vallée de Soča (Alpes slovènes), l'une des plus belles sources karstiques du monde. Le débit de la source atteint 100 m³/s, et la cascade à deux étages a une hauteur totale de 144 m. Photo: Nico Goldscheider.

Milieu : Galerie de drainage karstique dans les roches carbonatées triasiques, pour l'alimentation en eau potable de la ville de Hall au Tyrol, Alpes autrichiennes. Photo: Nico Goldscheider.

Bas : Injection de 200 g d'uranine à l'entrée de la grotte de Schwarzwasser, probablement la plus grande estavelle des Alpes, situé dans la région karstique du Hochifen-Gottesacker, Alpes germano-autrichiennes. Photo: Nico Goldscheider.

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