DIRECT APPLICATIONS OF THE DAR ZARROUK PARAMETERS IN GROUND WATER SURVEYS *

BY

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Abstract

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The combination of layer resistivity and thickness in the so-called Dar Zarrouk parameters S and T may be of direct use in aquifer protection studies and for the evaluation of hydrologic properties of aquifers.

The protective capacity of a clayey aquifer overburden is proportional to its longitudinal unit conductance S which, in terms of aquifer protection, gets a dimension of time (e.g. infiltration time). Aquifer storage in fissured reservoirs may be determined from differential conductance measurements (ΔS). Combination of the expression for ΔS with an empirical expression for electric conduction in fissured media yields a simple formula for water content per unit surface area. Both principles and possible developments are illustrated for a set of carboniferous limestone basins.

Introduction

The concept of Dar Zarrouk parameters has been introduced by Maillet (1947). For a sequence of n horizontal, homogeneous, and isotropic layers of resistivity ρ_i and thickness h_i, the longitudinal unit conductance S and transverse unit resistance T are defined by

$$S = \sum_{i=1}^{n} h_i/\rho_i \quad \text{and} \quad T = \sum_{i=1}^{n} h_i \cdot \rho_i.$$

Both variables and the derived concept of Dar Zarrouk curves (Maillet 1947) are of prime significance in the development of interpretation theory for vertical electrical soundings (e.g. Orellana 1963, Zohdy 1965, 1974, 1975, Kunetz and Rocroi 1970).

For some specific applications in ground water surveys, S and T turn out to be powerful autonomous interpretation aids, aside of any exhaustive interpretation scheme. Such applications go beyond the usual hydrogeological

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applications of resistivity soundings, which commonly aim at the definition of an aquifer's geometry. Instead, full advantage is taken from the combination of thickness and resistivity into one single variable, which is used as a base for the evaluation of properties such as aquifer transmissivity, aquifer storage, and protection of ground water resources.

Aquifer Transmissivity

Aquifer transmissivity is defined as the product of aquifer thickness and hydraulic conductivity. It is usually denoted by the symbol T, which it has in common with the transverse unit resistance.

The physical correlation of these two Ts has already been tackled by a few authors and is merely cited here for the sake of completeness. The study of an alluvial aquifer in the Rhine valley, published by Duprat, Simler, and Ungemach (1970), is a representative example. Orellana (1972) cites a case study on Miocene deposits in Spain and, no doubt other examples of such applications could be found in literature.

Aquifer Protection

The application of S for ground water protection studies and aquifer storage evaluation is illustrated by two examples, drawn from a resistivity survey on Carboniferous limestone basins (Henriet 1974). This survey involved the measurement of a network of 273 Schlumberger soundings over an area of 216 km². A short geological outline, common to both examples, is given below.

Geological background

The investigated basins (fig. 1) are situated in the Condroz, south of the river Meuse (Belgium). Structurally, they form a twin set of complex synclines belonging to the Hercynian synclinorium of Dinant. The northern syncline and half of the southern one are drained by the Hoyoux river, the remaining half by the Néblon.

The main units of hydrogeological significance are the limestone and dolomite reservoir, the flanking and interlocking anticlines of Devonian sandstones, and a shaly transition layer forming a quasi impermeable reservoir lining. The limestone aquifer is covered by a silt/clay overburden, ranging from a few dm up to 10 or 20 m in thickness. Subsurface leakage of the upper Hoyoux basin into the lower one is largely impeded by structural and lithological barriers. As a consequence, three major ground water basins could be considered: the upper Hoyoux, the lower Hoyoux, and the Néblon basins. For the sake of convenience, they will be referred to as basins 1, 2, and 3, respectively.

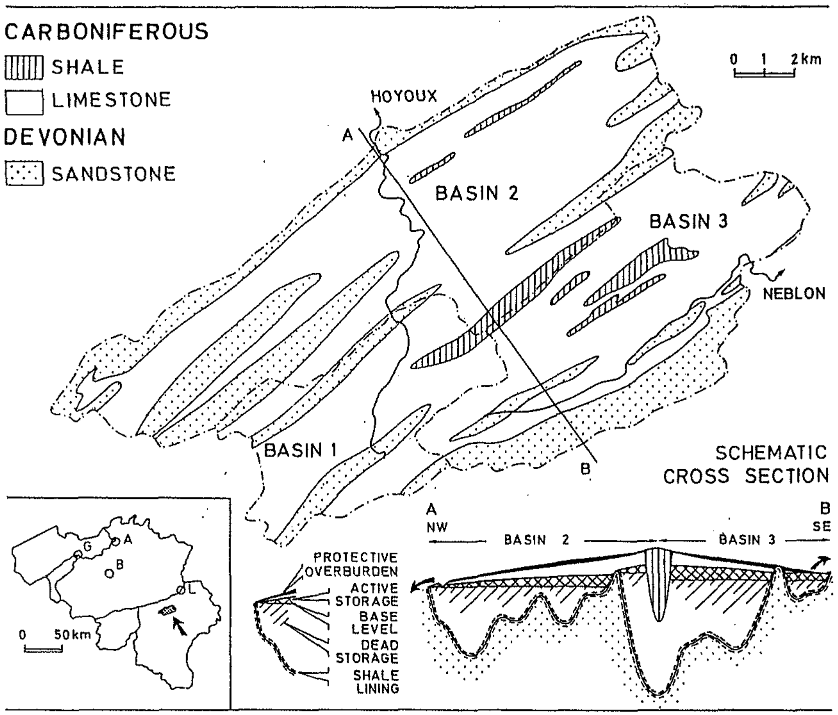

Fig. 1. Map of the investigated area, showing the outline and geology of the basins. A schematic cross section shows the basin structure, the reservoir protection and the main components of ground water storage.

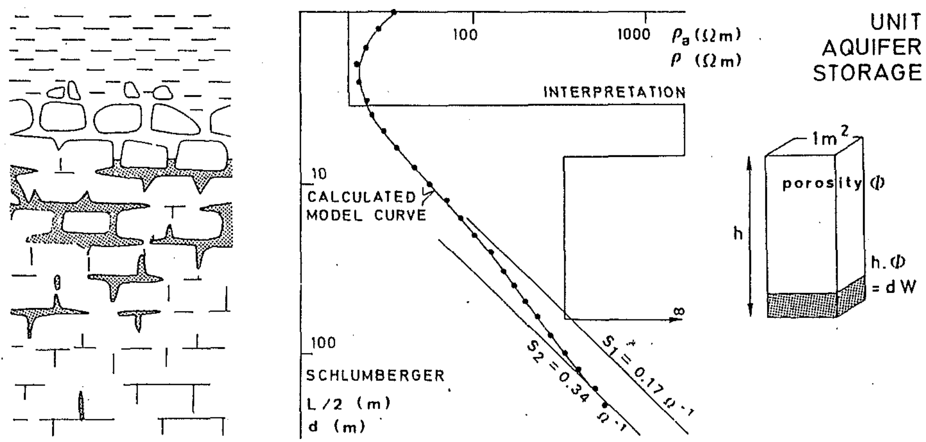

Fig. 2. A representative sounding curve on carboniferous limestone with the corresponding interpretation. The aquifer prism illustrates the concept of unit aquifer storage.

Annex 4-2

The vertical resistivity distribution on the limestone basins embraces four major units with alternating lower and higher resistivities: the clayey overburden, the unsaturated limestone, the saturated limestone, and the deeper compact limestone, characterized by a very high resistivity. A representative sounding curve, measured on such a sequence, is shown in fig. 2.

The time dimension of S

It is easy to conceive that the main natural protection of the limestone aquifer is provided by the clayey overburden. Its protective capacity, exerted by filtration and retardation of infiltrating solutions, is assumed to be proportional to its thickness and inversely proportional to its hydraulic conductivity.

On a purely empirical basis, the hydraulic conductivity of a clayey sediment could be linked to electrical resistivity through the concept of clay content. High clay contents generally correspond with low resistivities and low hydraulic conductivities, and vice versa. Hence, the protective capacity of the overburden could be considered as being proportional to the ratio of thickness to resistivity, or in other words to the longitudinal unit conductance S.

Taking into consideration that hydraulic conductivity has a dimension of length divided by time, the expression for S, considered in terms of aquifer protection, yields a dimension of time. Such a time, born of a geophysical concept, may be used in different contexts:

— in a purely physical context, it can be regarded as an infiltration time
— in a sanitary context, it can be matched with bacterial decay times
— finally, from an operational point of view, it can be regarded as the time available for action, before a polluting spill reaches the ground water table.

Elaboration of a protection document

The longitudinal unit conductance of the clayey overburden is directly determined from the position of the 45° asymptote, fitting the first rising segment of the sounding curves. For the area involved, additional information is available from soil maps. By combining soil thickness data derived from these maps and clay resistivities measured during the survey, it is possible to define the detailed distribution pattern of the lower S values. The resulting information is compiled into a conductance map or S map (fig. 3), which forms the base document for the appraisal of natural protection of the considered ground water basins.

The final step is the translation of the S scale of the map into a protection scale. There is no straightforward solution to this problem, in view of the multitude of factors controlling the degree of protection, such as type and persistence of the polluting source. In a simple approach, the time equivalent of S values could be matched with bacteriological protection norms: if, for instance, the mean overburden resistivity amounts to 30 Ωm and hydraulic conductivity does not exceed 0.1 m/day, a conductance of 0.2 Ω^{-1} corresponds to an equivalent silt/clay thickness of 6 m and a filtration time of more than 60 days, which is commonly admitted as a safe level.

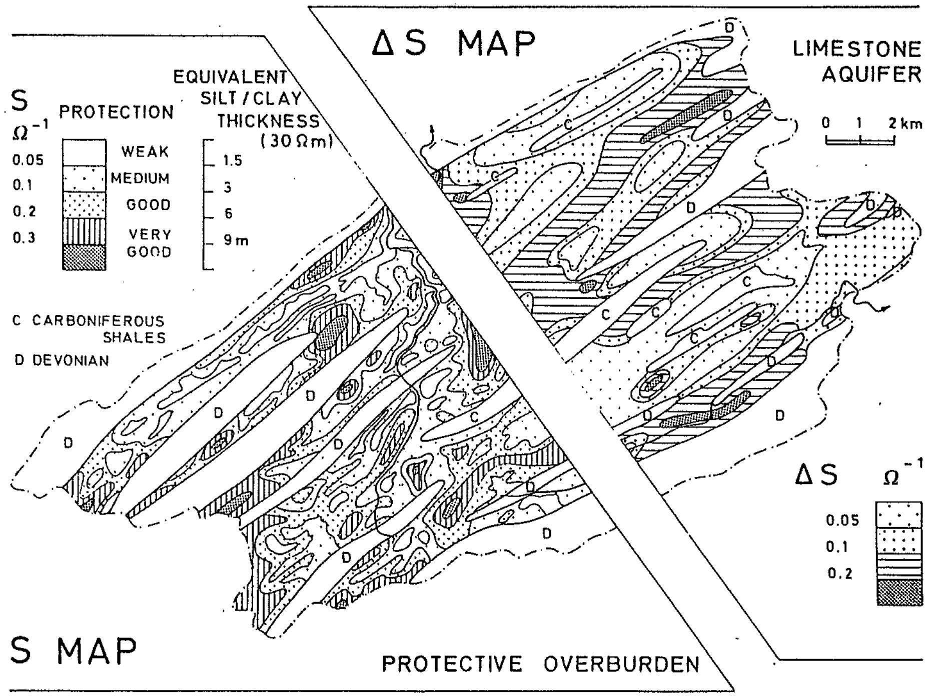

Fig. 3. Juxtaposition of conductance maps for the aquifer overburden (S) and the saturated ground water reservoir (ΔS). The first one is used as basic protection document, the second one for evaluation of ground water storage.

In spite of the highly pragmatic nature of this approach, it is assumed that the conductance of the overburden has a diagnostic value regarding protective capacity which is superior to that of any single parameter such as overburden thickness or hydraulic conductivity.

Aquifer Storage

The main aquifer in the highly fissured limestone shows up as a thick low-resistivity horizon, sandwiched between the overlaying unsaturated bedrock and the compact limestone basement, both of high resistivity. On sounding curves, it causes a sigmoid bend, framed by two steep segments (fig. 2). Such

Annex 4-2

a situation has also been depicted by other authors (e.g. Astier 1971). The 45° asymptotes, fitting the steep segments, yield conductances S_1 and S_2. Their difference ΔS can roughly be equalized to the longitudinal unit conductance of the aquifer.

Unit aquifer storage from ΔS

Calculation of aquifer storage from ΔS is done by introduction of an empirical formula for the resistivity of saturated jointed rocks. This formula, ascribed to Nechai (1964), relates the bulk resistivity ρ of the rock to the joint porosity Φ, the rock matrix resistivity ρ_p and the ground water resistivity ρ_w.

$$1/\rho = 2\Phi/3\rho_w + (3 - 2\Phi)/(3 - \Phi)\rho_p.$$

As the resistivity of massive limestone is very high, the second term of the expression may be neglected:

$$1/\rho = 2\Phi/3\rho_w.$$

Developing the expression for ΔS with this simplified formula leads to an expression in which the thickness h of the aquifer is multiplied by the porosity Φ.

$$\Delta S = h/\rho = 2h\Phi/3\rho_w = 2dW/3\rho_w.$$

The product $h \cdot \Phi$ is nothing but the volume of water dW, contained in an aquifer column of unit cross section. This is represented in a schematic way in fig. 2, where all fissure water contained in the considered prism has been moved down to the bottom. The height of the water volume equals $h \cdot \Phi$.

As follows, the aquifer storage per unit surface could be defined in a unique way by ΔS and ground water resistivity, without prior knowledge of porosity:

$$dW = 1.5\,\Delta S\,\rho_w.$$

Total storage within sounding depth

Calculation of total aquifer storage is a matter of straightforward integration over the whole reservoir area. A convenient way is a graphic procedure, starting with the compilation of a ΔS map (figs. 3 and 4). Surfaces enclosed by respective contour lines are measured by planimetration and plotted versus corresponding ΔS values. The area, enclosed between this cumulative distribution curve and the coordinate axes, is the integral of ΔS for the whole reservoir area. Multiplication by ground water resistivity and the appropriate empirical constant yields the total storage W.

The quantity W strictly represents the total volume of low-resistivity phase occupying rock voids in the saturated zone within sounding depth. It includes various components, such as detrital clay and ground water in different potential-energy states (capillary water, gravitational water). It should, incidentally, be noted that in the considered carbonate reservoirs mean ground water resistivity and clay resistivity do not substantially differ.

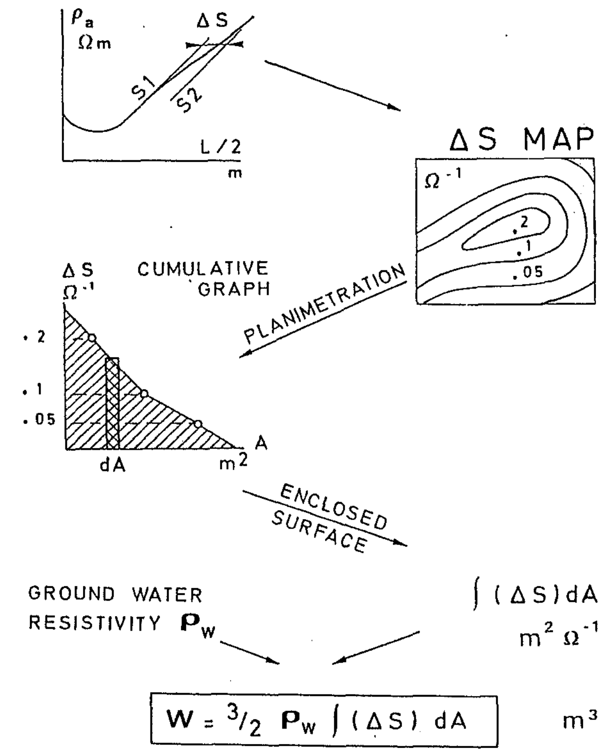

Fig. 4. Flow diagram for the integration of ΔS over the whole reservoir area A and subsequent derivation of total storage W.

Calibration

The lack of resolving power between effective ground water storage and dead storage of similar electrical behaviour forms the most obvious limitation of the ΔS method. In the Condroz survey, this problem could be tackled by comparison with an effective storage evaluation from ground water recession graphs recorded at the outlet of basin 2.

A prerequisite to a valid comparison is the correction of the geophysical results by withdrawal of storage volumes, which do not contribute to spring flow at the basin outlet. The estimation of such dead volumes requires a few basic assumptions:

— the average sounding depth of Schlumberger electrode lay-outs of 400 m length in the considered area is rated at 150 m,

Annex 4-2

— the distribution of the low-resistivity phase within sounding depth is assumed to be random,

— the active reservoir boundaries have to be known. As piezometric data are scarce, it is the envelope of valley bottoms which is chosen as closest approximation for the ground water table or top of the active storage volume. Its base is, by definition, the basin outlet level.

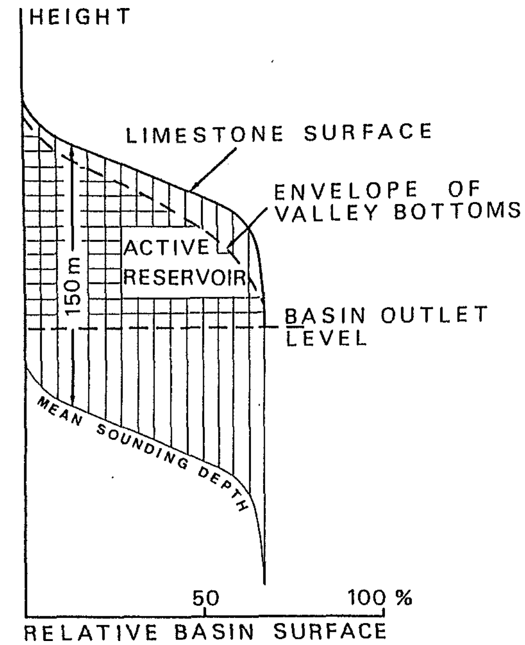

Fig. 5. Hypsometric diagram showing the volume fraction of the active ground water reservoir (horizontal shading) within the rock slice, involved in the geoelectrical measurements (vertical shading).

The reduction process is illustrated by fig. 5, where the top full line and top dashed line, respectively, represent the hypsometric curves of the land surface and of the envelope of valley bottoms. The horizontal dashed line is the base level and the lower full line is the average sounding depth. This graph clearly shows that the bulk volume of the active reservoir—which is proportional to the area with horizontal shading—amounts to about half the total volume of the rock slice involved in the geoelectrical measurements, represented by the area with vertical shading.

After application of this reduction process to the storage results for basin 2, the total volume of low-resistivity phase occupying rock voids within the considered boundaries turns out to be twice the effective storage derived from spring flow analysis. The order of magnitude of this result seems plausible. This should corroborate the validity of a storage evaluation by the ΔS method. Furthermore, it justifies the use of this result for calibration, allowing predictive storage evaluations for the neighbouring basins 1 and 3, where spring flow analyses are lacking.

Concluding remarks

It should be noted that a document such as the ΔS map proves useful not only as an intermediate step in the calculation of total storage W, but also, in a more qualitative way, as a picture of the spatial distribution of storage potential in a ground water reservoir.

Comparing the respective merits of the geophysical and the hydrological storage evaluation methods, one should bear in mind that ground water recession graphs consist of time series of discharge data, collected at a single point, the basin outlet. In other words, the observations are spread out in time, but localised in space. On the other hand, geoelectrical observations are localized in time, but spread out in space, over the whole basin. In a way, a geophysical storage evaluation from ΔS brings in an additional dimension to hydrological methods.

The range of applications of the method under its present form is bound to certain field conditions, which, however, are not uncommon in nature. Modifications are not excluded, for instance in the empirical expression for the bulk resistivity of saturated rocks or in carrying out the integration or calibration processes. Full benefit of this interpretation method can only be drawn by moulding its principle on local field conditions.

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Annex 4-2

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