RESPONSE TO PUMPING OF WELLS IN SLOPING FAULT ZONE AQUIFERS

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ABSTRACT

A number of short term variable discharge pumping tests were conducted on sloping fault zones cutting serpentinites from a thrust complex in north Portugal. The fault zones were identified and geometrically characterized by geological mapping, interpretation of areal photographs and electromagnetic survey. The evolution of specific drawdowns obeyed two different laws: (1) Specific drawdown-time points fitted to straight lines in log-linear scatter plots. In these cases, test data were interpreted by the Cooper-Jacob method, as if fault zones were regular confined porous media aquifers; (2) Specific drawdown-time points fitted to straight lines in bilinear scatter plots. Because the available theoretical models were unable to match the observed data, a new approach was developed. The approach introduces the concept of impermeable barrier with variable effective position, a barrier that is represented by the upper limit of the fault zone aquifer. The mechanics of the method is still based on the concepts and formulae of confined porous media aquifers, including the framework of the image well theory. The calculated transmissivities ($T = 7 \times 10^{-7}$ to $2 \times 10^{-4} \text{ m}^2/\text{s}$) and storage coefficients ($S = 2 \times 10^{-5}$ to 6.1×10^{-4}) are within those commonly found for confined aquifers in fractured rocks.

Key words: electromagnetic survey, pumping tests, inclined fault zone aquifers, laws of drawdown evolution, hydrologic boundaries, aquifer's formation constants.

INTRODUCTION

Aquifers in crystalline rocks (igneous or metamorphic) are of three main types: weathering profile, fractured massif and major discontinuity aquifers (Verweij, 1995). The analysis of pumping test data from wells in weathering profile (horizontal porous media) or fractured massif aquifers is beyond the scope of this paper and can be found elsewhere (Theis, 1935; Cooper and Jacob, 1946; Newman, 1975; Barenblatt et al., 1960; among many others). The hydraulic behavior of an aquifer in a major discontinuity (fault zone or dyke) depends on the flow pattern that is generated once wells are submitted to pumping. If for instance a well is draining a major vertical fault zone showing a much higher transmissivity than that of the surrounding fractured massif, then the flow is initially planar (parallel to the fault's walls) and the drawdowns evolve as a function of $\sqrt{t}$, because the water is essentially being provided to the well by the major discontinuity. With continued pumping, the water starts coming also from the massif through small fractures. Initially, the flow in these fractures is preferably orthogonal to the discontinuity, the total flow is thus biplanar, and the drawdowns evolve as a function of $\sqrt[4]{t}$. But gradually the flow in the fractures becomes more and more important and of a radial like type. When this happens, the drawdowns in the discharge (observation) wells evolve as functions of $\ln(t)$. Mathematical models that explain the hydraulic behavior of vertical fault zone or dyke aquifers are known as single fracture models, and among them one may refer that of Boehmer and Boonstra (1986).

In this paper, I studied the response to pumping of wells fully penetrating a particular type of major discontinuity aquifers, those represented by sloping fault zones. It has been observed that drawdowns may evolve as functions of $\ln(t)$, suggesting radial flow to the wells, but sometimes they may also fit to linear functions of t , a response that has been linked to the

hydrologic boundary born with the aquifer's inclination. Such a dual behavior is striking and in an attempt to explain it a mathematical model has been developed that is described and tested in the next sections.

MODELLING PUMPING TEST DATA OF SLOPING FAULT ZONE AQUIFERS

Statement of the Problem

By combining the available geological information with the results of an electromagnetic survey, I studied the geometry (strike and dip) of four fault zones, in an area where thrustured serpentinites outcrop (Sobreda village, north Portugal); details of this work are presented later in this paper. I concluded that three of these structures are inclined, and in order to see how drawdowns would evolve in eleven wells draining them, I conducted pumping tests, which are also fully described in a section below. In six of these tests, drawdowns obeyed a logarithmic function of time, in the remaining five drawdown-time points fitted to straight lines in bilinear scatter plots. From comparison of diagnostic data plots with theoretical plots (Verweij, 1995), I selected the Cooper-Jacob method (Cooper and Jacob, 1946) for the analysis of the first group of tests, and concluded that no existing model would explain the second group of tests. In an attempt to fill in this gap, I developed the model "barrier with variable position".

Concept of Barrier with Variable Position

Figure 1 shows an ideal sloping fault zone aquifer intersected by a vertical fully penetrating discharge well. The smashed material filling the fault's box is assumed permeable, homogeneous and isotropic and the surrounding rock mass impermeable. Once the well is

subject to pumping, the flow lines tend to be parallel to the aquifer's bottom and the depression cone to expand in all directions until its influence reaches the outcrop area of the fault zone (instant t_0). Further radial expansion of the cone is impossible because the aquifer is suddenly interrupted by the ground surface—in other words—the upper limit of the aquifer (segment A_0B_0 of the potentiometric surface) acts as an impermeable barrier. With continued pumping, the potentiometric surface lowers down. This displaces A_0B_0 towards the well (A_1B_1 for $t = t_1$, A_2B_2 for $t = t_2$, etc) and for that reason the barrier is said of variable position.

Backgrounds on Hydrologic Boundary Analysis

Pumping test data from bounded aquifers were first modeled by Ferris et al. (1962). Among the studied boundary scenarios, we retrieve that of a rock massif (impermeable barrier) cutting an horizontal water table aquifer. Under such hydrologic conditions, it has been found, for discharge wells fully penetrating the aquifer, that the dynamic water levels decreased faster once the depression cone touched the massif. In order to explain this drawdown evolution, Ferris et al. (1962) resorted to the image well theory—replaced the system *real well-aquifer-impermeable barrier* by one hydraulically equivalent *real well-aquifer-image well*—and predicted the drawdowns for the real well as if they were the result of simultaneous pumping in both wells.

The image well theory was applied in three consecutive steps:

Replacing the Real Barrier by a Virtual Plane of Zero Flow - The irregular and inclined contact surface between the rock massif and the aquifer was substituted by a vertical plane (zero flow plane) put at half distance between the outcrop area of the massif and the region where it intersects the bottom of the aquifer (average or effective

position). This initial procedure is required for a rigorous referencing of the barrier with regard to discharge or observation wells.

Replacing the Real Aquifer by One of Infinite Radial Extent - The water table aquifer is extended towards the area occupied by the rock massif so it acquires infinite radial extent.

Installing the Image Well in the Extended Water Table Aquifer - An imaginary discharge well (image well) is put at the same distance as the real well from the zero flow plane but on the opposite side, in a manner that both wells stand on a common line perpendicular to that plane.

Analysis of Hydrologic Boundaries in Sloping Fault Zone Aquifers

In case the aquifer is an inclined fault zone, radial expansion of the depression cone is not stopped by an impermeable rock massif but by the ground surface. Thus, the zero flow plane pretends no approaching to the contact surface between the massif and the water table aquifer, but instead to the upper limit of the fault zone aquifer. In Figure 1, the initial position of this limit coincides with segment A_0B_0 of the potentiometric surface and is approached by π_0 . As the potentiometric surface lowers down with pumping, segment A_0B_0 is displaced towards the well, being approached successively by $\pi_1, \pi_2, (\dots)$. As mentioned, adoption of a zero flow plane is necessary for exact referencing of the barrier's position. Initially (for $t = t_0$), the distance between the zero flow plane and the discharge well is L_0 ; in general this distance is given by: for $t = t_n > t_0$, $L = L_n < L_0$.

In the hydraulically equivalent system, the aquifer is by definition of infinite radial extent. To comply with this definition, Ferris et al. (1962) replaced the rock massif by an extension of

the water table aquifer. In the present situation, the assumption of infinite areal dimension implies that the fault zone aquifer is not inclined but horizontal, with the top surface buried to a depth that is marked by its intersection with the discharge well (h in Figure 1).

In agreement with the image well theory, the imaginary well is placed beyond the zero flow plane. Initially, the distance between the real and image discharge wells matches $2L_0$, but as time goes on and the zero flow plane gets closer and closer to the real well, that distance decreases, being generically given by $2L_n$.

Drawdowns in Discharge or Observation Wells

For confined and of infinite radial extent aquifers, drawdowns in fully penetrating wells may be described by the Cooper-Jacob formula (Cooper and Jacob, 1946):

$$s_w = \frac{Q}{4\pi T} \ln \left(\frac{2.25Tt}{r_w^2 S} \right) = \frac{2.3Q}{4\pi T} \log \left(\frac{2.25Tt}{r_w^2 S} \right) \quad (1)$$

where s_w (m) is the drawdown in the discharge(observation) well, Q (m^3/s) the discharge rate, T (m^2/s) the aquifer's transmissivity, t (s) the time, r_w (m) the effective radius of the discharge well (or the distance between this and the observation well) and S (dimensionless) the aquifer's storage coefficient. Under the influence of an impermeable barrier, drawdowns increase faster with time, in agreement with the Ferris formula (Ferris et al., 1962):

$$s_w = \frac{Q}{2\pi T} \ln \left(\frac{2.25Tt}{r_w r_i S} \right) = \frac{2.3Q}{2\pi T} \log \left(\frac{2.25Tt}{r_w r_i S} \right) \quad (2)$$

where r_i (m) is the distance from the real discharge(observation) well to the image well. If the boundary were moved closer to the observation well, the drawdowns would evolve towards solutions set by the products $r_w r_{in}$, after periods of relaxation, i.e.:

$$s_w = \frac{Q}{2\pi T} \ln \left(\frac{2.25 T t_n}{r_w r_{in} S} \right) \quad (n = 0, \dots, \infty) \quad (3)$$

where r_{in} (m) is the distance to the image well at the instant t_n . Rearranging, we get:

$$s_w = \frac{Q}{2\pi T} \ln \left(\frac{2.25 T t_n}{r_w r_{i0} S} \right) + \frac{Q}{2\pi T} \ln \left(\frac{r_{i0}}{r_{in}} \right) = s_f + \delta s_v \quad (4)$$

where s_f are the drawdowns resulting from a constant r_i (Ferris drawdown) and δs_v the increments in s produced by the diminishments in that parameter (variation increments). If the barrier is moving softly because the cone of depression is expanding slowly, then the changes in r_i and s may be described by continuous functions of time.

The Shapes of Functions r_i and δs_v

When pumping proceeds beyond instant t_0 , the fault zone gets dewatered starting from its upper limit (A_0B_0 segment in Figure 1). Continuous dewatering is described indirectly by function r_i . Indeed, when for example $t = t_1$, $r_{i1} = 2L_1$ and the $A_0B_0A_1B_1$ polygon has just been dewatered. Initially, dewatering runs in slow motion because the hydrostatic pressure over the sloping aquifer is high, but as the pumping drops down the potentiometric surface dewatering gets faster and faster (e.g., for $t_2 = 2t_1$, $A_1B_1A_2B_2 > A_0B_0A_1B_1$). As a consequence of how

dewatering evolves in time, the fault zone aquifer is shrunk in a manner that reduction in length is dependent on its current value (i.e. as shorter the aquifer is as faster it shrinks), and for that reason r_i is described adequately by a first order differential equation of the type (Bronson, 1976):

$$-\frac{dr_i}{dt} = \psi r_i \quad (5)$$

where $\frac{dr_i}{dt}$ is the first derivative of r_i and ψ a constant of proportionality incorporating all physical parameters controlling the rate at which the aquifer shrinks. Aquifer dewatering is in a direct proportion of the pumping rate (Q) and inverse proportion of the aquifer thickness (b) and storage coefficient (S). Attending to the dimensions of Q and b , ψ must be given by:

$$\psi = \frac{Q}{b^3 S} \quad (6)$$

and by integrating Equation 5 we get:

$$r_i = r_{i0} e^{-\psi t} \quad (7)$$

meaning that r_i has the shape of a negative exponential. Replacing Equation 7 in Equation 3, we get:

$$s = s_f + \partial_{sv} = \frac{Q}{2\pi T} \ln \left(\frac{2.25 T t}{r_{i0}} \right) + \frac{Q}{2\pi T} \psi t \quad (8)$$

and conclude that the variation increments fit to linear functions of time.

Logarithmic vs Linear s Curves

The shape of Equation 8 is constrained by Ψ and t —for increasing t values, if $\ln(t) \gg \Psi t$ the s curve fits to a log-shaped line whereas if $\ln(t) \ll \Psi t$ the curve fits to a straight line. The time for which the (s,t) points start fitting to a straight line in a bilinear scatter plot is set by the following inequality:

$$\psi > \frac{\ln(t)}{t} \quad (9)$$

or, taking into account the equation for Ψ and rearranging, by:

$$S < \left(\frac{Q}{b^3} \right) \times \left(\frac{t}{\ln(t)} \right) \quad (10)$$

The graph of Equation 9 was drawn for a number of t values, assuming that the pumping proceeds at a constant rate of $Q = 1$ L/s (Figure 2). It is readily seen that low S and(or) b values induce linear drawdown evolutions within shorter periods of time than do high scores of those parameters.

Residual Drawdowns in Discharge or Observation Wells

In the end of a pumping test, the water levels in the discharge(observation) wells rise or recover. Theis (1935) has demonstrated that the residual drawdowns are equivalent to those observed if the pumping would still continue but another (imaginary) pump would start injecting water at the same rate into the aquifer through an imaginary well placed just in the

same point of the real well. Under this conditions, if τ is the duration of the test and t is the time passed since the test has finished, then the residual drawdowns (s') during the recovery period may be given by:

$$s'_w = \frac{Q}{4\pi T} \ln\left(\frac{\tau+t}{t}\right) = \frac{2.3Q}{4\pi T} \log\left(\frac{\tau+t}{t}\right) \quad (11)$$

For the cases where the pumping test is conditioned by the presence of an impermeable barrier with variable position, the recovery period implies an inversion in the sense of movement of the zero flow plane, meaning that r_i increases during that period. The Theis formula is given by:

$$s'_w = \frac{Q}{2\pi T} \ln\left(\frac{\tau+t_n}{t_n} \times \frac{r_{i(\tau+n)}}{r_{in}}\right) \quad (12)$$

where $r_{i(\tau+t)}$ is the distance from the real discharge(observation) well to its image at the instant $\tau+t_n$ and r_{in} is the distance from the imaginary recharge(observation) well to its image at the instant t_n . Because the sites where the real discharge and imaginary recharge wells are installed, as well as those of their images, are by definition coincident (Theis, 1935), $r_{i(\tau+t)} = r_{in}$ for any given t , and thus Equation 11 simplifies to:

$$s'_w = \frac{2.3Q}{2\pi T} \log\left(\frac{\tau+t}{t}\right) \quad (13)$$

Determination of T and S in case Drawdowns Evolve as Linear Functions of t

Calculation of the aquifer's formation constants, in case s evolves as a function of t , needs anticipated knowledge of b , the aquifer's thickness. First, calculate T from the s' vs. $\log[(\tau+t)/t]$ plot using the common straight line slope method (Cooper and Jacob, 1946). Second, calculate Ψ from s vs. t plot applying the same method. Third, calculate S from Equation 6.

IDENTIFICATION AND GEOMETRICAL CHARACTERIZATION OF SLOPING FAULT ZONE AQUIFERS

The Sobreda village is located in the eastern part of the Trás-os-Montes province (north Portugal). The local geology is characterized by ultramafic rocks (mostly serpentinites) of the Morais massif, a pile of thrust sheets emplaced on top of autochthonous sequences (Cambrian slates and greywakes) during the Hercynian orogeny (Figure 3).

Structural Geology

A structural map (scale 1/50000) of the Sobreda region has been drawn from field observations by Pereira et al. (1998); the main structures (faults and thrusts) are shown in Figure 3. Additional structural information is provided by Pacheco (2000), which interpreted as probable faults the lineaments observed in areal photographs at the scale 1/20000 (dashed lines).

The presence of inclined fault zones in the Sobreda region is marked undoubtedly by the thrust outcropping 350 m to the north of the village, an oscillating thrust striking to an average direction WNW-ESW and plunging roughly to SSW. The other faults strike preferably to NNE-SSW and NW-SE (Figure 3). To support the identity of the inclined fault zones as sloping aquifers, an electromagnetic survey was conducted in the central part of Sobreda (rectangle in Figure 3) using a GEONICS EM34-3 equipment.

Distribution of Ground Electrical Conductivities and Identification of Anomalies

The ground electrical conductivities (GEc) were measured at 315 sites, covering an area of approximately 0.2 km² (500 × 400 m) and assuring an average density of 1 site per square of 25 m side. They were obtained with the 40 m length cable by setting the electrodes to horizontal and vertical dipoles (investigations at 30 and 60 m depths, respectively). Results are shown in Appendix 1.

During the survey, I found GEc in the intervals 0.2–180 $\mu\text{S/m}$ (30 m depth investigation) and 0.2–36 $\mu\text{S/m}$ (60 m). In both cases, conductivities were clustered into at least two groups, one with the highest GEc (anomalies), the other with the lowest GEc (noise). The cleavage between anomalies and noise was set by quantile-quantile (Q-Q) diagrams. Basically, these diagrams plot the quantiles of a population (the observed values sorted in ascending order) against their theoretical normal counterparts. If the population is normal or gathers a finite number (k) of normal populations, the points in the scatter plot fit to 1 or k straight lines. Our presumption is that, in case the anomalies and the noise do belong to different normal

populations, the Q-Q plots drawn for the 30 and 60 m depth surveys will both incorporate 2 straight line segments.

Figures 4a,b are the Q-Q plots drawn for the GE_c. In the 60 m depth plot, the expected two straight line segments (dashed lines) intersect at GE_c = 8 μ S/m. The 30 m depth plot incorporates 3 (instead of 2) segments. I assumed that the first segment represents the noise whereas the other two represent moderate and strong anomalies. Consequently, the separation between noise and anomalies is set by GE_c = 10 μ S/m.

Spatial Distribution of Anomalies

The spatial distribution of anomalous conductivities is depicted in Figure 5. The anomalies are separated from the noise by contours of equal GE_c, respectively GE_c = 10 μ S/m (emplacement at 30 m depth) and GE_c = 8 μ S/m (60 m) as suggested by the Q-Q plots. The unfilled polygons limited by a thick dash line (shallower investigation) show a picture of three zones of anomalous conductivity, termed FZ1, FZ2 and FZ3 and following the E–W, N–S and NW–SE trends. These zones are confirmed by the deeper investigation (filled polygons limited by a thin dash line), which has scanned an additional one, termed FZ4 and following an oscillating trend roughly approaching the NNW–SSE direction.

Assignment of Zones of Anomalous Conductivity to Fault Zone Aquifers

I bear in mind that the conductivity soundings simply indicate tabular subsurface zones of elevated conductivity, eventually resulting from the presence of clay minerals in the smashed material filling the fault's box. But I believe that in this case, the FZ1-4 zones are a direct consequence of the elevated permeability, i.e. they are fault zone aquifers. The argument is

that previous surveys with the EM34-3 equipment proved efficient in the selection of sites for construction of public water boreholes (Pacheco et al., 1997; Sousa Oliveira and Pacheco, 1998). In the work of Pacheco et al. (1997), the electromagnetic survey succeeded a very detailed geological mapping (scale 1/10000). By setting together the geological and geophysical data, the surface and underground geometries of a thrust cutting Silurian phyllites and quartzites of the Balugas region (Padrela mountain, north Portugal) could be delineated with great precision. A well has been drilled nearby the outcrop area of the thrust. From comparison between the thrust geometry and the geographical location of the well, it has been demonstrated that this became productive only where it crossed the sloping structure. The work of Sousa Oliveira and Pacheco (1998) covered the granitic area of Souto Velho (Chaves, north Portugal). Now, the electromagnetic survey succeeded the interpretation of areal photographs at the scale 1/15000, and it has been noted that the "fracture" networks interpreted from both the photographic and geophysical data were very similar. Several nodes of interception between fractures were marked as sites for construction of productive boreholes and in one case the well has been effectively drilled, with success.

Geometry of the Fault Zone Aquifers and Relation to the Observed Structures

FZ2 is a vertical 30 m thick fault zone striking almost parallel to the NNE-SSW faults observed in the field by Pereira et al. (1998). The shiftings between unfilled and filled polygons indicate that FZ1 and FZ3 are inclined. Given the strike and dip, FZ1 probably represents a replica of the WNW-ESE thrust shown in Figure 3. FZ3 has uniform dip and thickness, respectively given by 45°SW and 35.4 m, and strikes parallel to the NW-SE probable faults identified by Pacheco (2000). FZ4 is visible solely at the deeper investigation.

Because it follows a trend close to FZ3, I assumed that it also inclines 45° towards SW, and therefore that its average thickness is 33.2 m.

PUMPING TESTS IN WELLS DRAINING SLOPING FAULT ZONE AQUIFERS

The Sobreda village has 25 drilled wells. In 11 (E1–E11, Figure 5), I executed short term pumping tests, which lasted for 60–2120 min (on average, 400 min).

Verweij (1995) suggested a sequence of steps to be involved in proper interpretation of pumping test results, that include preparation of data, comparison of diagnostic data plots with theoretical plots, selection of the analytical method for parameter calculation and determination of the aquifer formation constants. In this study, I followed his recommendations.

Preparation of Data

Drawdowns (s) and residual drawdowns (s') were computed from the measured dynamic water levels (DL, Appendix 2). All tests except E7 have been conducted under variable discharge rates (Q); the Q values decreased according to logarithmic (E1 and E11) or exponential (the other tests) functions of time (t) as depicted in Table 1. In order to account for the variations in Q , drawdowns were converted into specific drawdowns. This was done by dividing the s values by their associated Q values (s/Q for each t).

Selection of Analyzing Methods

The drawdown response of a well to pumping depends on the well conditions, aquifer conditions and time, and in general may be described by one or more theoretical models. To identify the correct aquifer condition(s) during a test, the observed drawdown evolution(s) must be compared with known theoretical responses. This is done by construction of diagnostic plots, where drawdowns are plotted against time in bilinear, log-linear and log-log diagrams, and subsequent comparison of these data plots with the available characteristic drawdown-time relations (Kruseman et. al, 1990; Horne, 1990).

I built diagnostic plots for all tests and compared them with the list of characteristic drawdown-time relations provided by Verweij (1995). In some cases (tests E1, E3, E4, E7, E10 and E11), after a short period of time the (s/Q,t) points fitted to straight lines in the log-linear diagrams; Figure 6a includes the cases where specific drawdowns were very high (left axis) or very low (right axis) and Figure 6b incorporates the other situations. According to Verweij (1995)'s list, the fitting of (s,t) points to one straight line in a log-linear diagram is characteristic of wells draining homogeneous confined aquifers, and may be analyzed by the Cooper-Jacob method (Equation 1).

For tests E2, E5, E6, E8 and E9, specific drawdowns also fitted to straight lines but in the bilinear and log-log diagrams (Figures 7a,b), and according to the list of Verweij (1995) such response occurs if wellbore storage is an important source of drawdown or when a confined aquifer is limited by a closed barrier. In case the hypothesis "wellbore storage" is assumed true and specific drawdowns are used in the diagnostic plots, the slopes (bilinear plot) or intercepts-y (log-log plot) of the fitted straight lines must be given by $1/\pi r^2$ (Custodio and

Llamas, 1983), where r is the borehole radius. These theoretical constraints are represented in Figures 7a,b by the dotted lines (in all cases, $r = 70$ mm) and apparently contradict the starting assumption. Under the initial belief that the hypothesis "closed barrier" is faithful, the same slopes or intercepts- y must in absence of recharge match $1/A\eta$, where A is the area of the aquifer and η its porosity (Custodio and Llamas, 1983). With the available information, it is impossible to set together the data plots with the appropriate theoretical plots, but the hypothesis is rejected by the observed residual drawdowns (Figure 7c). When a closed barrier is the main source of drawdown to a pumping test, total recovery is never attained, which in a plot like Figure 7c would imply that $s' = 0$ for $(t + \tau)/t < 1$. This is never the case.

The lack of a feasible method to interpret tests E2, E5, E6, E8 and E9 induced the creation of a new model, the model barrier with variable position. With the available information, it is impossible to check the data plots against theoretical plots and then diagnose if the approach is applicable to the tests. But now, neither the predicted drawdown evolutions nor the residual drawdown evolutions are in contradiction with the observed data.

Association Between Wells and Fault Zone Aquifers

Most wells are 80–100 m deep. Although no reliable information exists on depths of interception between wells and fault zones, the following associations are assumed:

E1, E4 and E10 drain FZ1;

E3, E5, E7 and E11 drain FZ3;

E2, E6, E8 and E9 drain FZ4.

The arguments are: (1) Proximity between well locations and filled polygons (Figure 5);

(2) Drawdown evolution. During the pumping periods, drawdowns in E1, E4 and E10 (the FZ1 wells) evolved as functions of $\ln(t)$ whereas in E2, E6, E8 and E9 (FZ4 wells) they evolved as linear functions of t (Figures 6a,b and 7a); (3) Response of observation wells. E3 reacted to pumping in E5 and vice versa, E3 reacted to pumping in E11 (Appendix 2).

Calculation and Evaluation of Aquifers' Formation Constants

Formation constants (T,S) were calculated for FZ1, FZ3 and FZ4 (Table 2) using the methods selected above. Transmissivities (7×10^{-7} to $2 \times 10^{-4} \text{ m}^2/\text{s}$) are coherent with hydraulic conductivities of fractured igneous and metamorphic rocks (Domenico and Schwartz, 1990). They tend to be systematically lower when computed from residual drawdown data, decreasing at the most to 20% (E3 test) the values obtained from drawdown data. Further, they tend to drop down from FZ3 (where T are in the interval 1.4×10^{-6} – $2.0 \times 10^{-4} \text{ m}^2/\text{s}$) to FZ1 (where $T = 7 \times 10^{-7}$ - $5.5 \times 10^{-6} \text{ m}^2/\text{s}$) and again to FZ4 (where $T = 8 \times 10^{-7}$ – $2.6 \times 10^{-6} \text{ m}^2/\text{s}$). The calculated storage coefficients ($S = 2 \times 10^{-5}$ – 6.1×10^{-4}) are among those of porous media confined aquifers. FZ4 show rather conservative S values ($(8 \pm 2) \times 10^{-5}$) whereas FZ3 is associated to their lowest (2×10^{-5}) and highest (6.1×10^{-4}) scores.

Additional Support to the Model Barrier with Variable Position

A striking observation supports the identity of the model barrier with variable position. As shown in Figures 8a,b, drawdowns evolve as functions of $\ln(t)$ or t depending on whether the discharge well is E3 or E5. This local control of the drawdown evolution may be explained by the model's equations: (E3 test) In agreement with Equation 10, drawdowns would start evolving as linear functions of t if the test has lasted for $t > 690 \text{ min}$, instead of $t = 280 \text{ min}$;

(E5 test) Linear drawdown evolutions are predicted for $t > 240$ min, a time that unexpectedly goes beyond the end of the test. However, recall that, for the calculation of Ψ (Equation 8), I used transmissivities pertaining to the recovery period (the ones available), assuming that they would not differ much from those related to the pumping period. In case this assumption is not warranted, the computed storage coefficients (Equation 6) are in error by a factor T_r/T_p , where T_r and T_p are the transmissivities calculated from the residual and specific drawdowns, respectively. In Table 2, only E11 complies with the assumption $T_r \approx T_p$. For the other tests, the T_r/T_b ratios range from 0.2 to 0.6, meaning that the real storage coefficients are 20 to 60% the ones shown in Table 2. Reusing Equation 10 with the corrected S values, I predicted linear drawdown evolutions for $t > 39$ – 135 min (on average, $t > 86$ min), which is acceptable.

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Table 1

Caption: Laws of discharge rate variation.

Pumping Test	$Q \times 10^{-3}$ (m ³ /s)
E1	$0.945-0.245\log(t)$
E2	$0.826\exp(-0.01t)$
E3	$0.667\exp(-0.001t)$
E4	$0.921\exp(-0.0007t)$
E5	$1.018\exp(-0.01t)$
E6	$0.811\exp(-0.004t)$
E7	1.000
E8	$0.731\exp(-0.006t)$
E9	$0.653\exp(-0.008t)$
E10	$0.620\exp(-0.002t)$
E11	$1.069-0.133\log(t)$

Q - Discharge rate.

t - Time in minutes.

Table 2

Caption: Results of the pumping tests: discharge(observation) wells and associated fault zone aquifers; pumping times, average discharge rates and formation constants (T and S).

General				Pumping Period					Recovery Period
Tested Well	Observation Point	Drained Aquifer	b (m)	Duration (min)	Average Q x 10 ⁻³ (m ³ /s)	T _p x 10 ⁻⁶ (m ² /s)	r _w (m)	S x 10 ⁻⁴	T _r x 10 ⁻⁶ (m ² /s)
E1	E1	FZ1		400	0.40	1.2			0.7
E4	E4			485	0.76	5.5			2.8
E10	E10			200	0.50	5.2			
E3	E3	FZ3	35.5	280	0.57	5.9	24.8	0.5	1.4
	E5					6.1			
E7	E7			150	1.00	203.0	53.3		
E11	E11			2120	0.67	8.0			7.4
	E3					28.1			
E5	E5			110	0.59			0.2	3.1
E2	E2	FZ4	33.2	60	0.59			1.0	0.8
E6	E6			180	0.55			0.8	2.6
E8	E8			120	0.50			0.7	1.4
E9	E9			105	0.42			0.6	1.1

b - Aquifer's thickness;

r_w - Distance between the discharge and observation wells;

T_p - Transmissivities determined from the specific drawdowns;

T_r - Transmissivities determined from the residual drawdowns;

S - Storage coefficient.

Figure 1

Caption: Schematic representation of an ideal sloping fault zone. Symbols described in the text. It is assumed that $t_2 - t_1 = t_1 - t_0$, meaning that when the origin of time is reset to t_0 , $t_2 = 2t_1$.

Figure 2

Caption: Instants for which drawdowns in discharge or observation wells fully penetrating sloping fault zones are expected to start evolving as linear functions of t (Equation 10).

Symbols: Q - Assumed discharge rate; b and S - Aquifer's thickness and storage coefficient.

Figure 3

Caption: Geographical location (Hayford-Gauss coordinates) and structural map of the study area (Sobreda village, north Portugal). Faults and thrusts were drawn from Pereira et al. (1998), probable faults from Pacheco (2000).

Figure 4a

Caption: Q-Q plot of ground electrical conductivities (investigation at 30 m depth).

Figure 4b

Caption: Q-Q plot of ground electrical conductivities (investigation at 60 m depth).

Figure 5

Caption: Spatial distribution of zones of anomalous conductivity in the Sobreda village; the school and church are points of easy reference.

Figure 6a

Caption: Log-linear plots of tests E1 and E7 (very high and very low specific drawdowns, respectively).

Figure 6b

Caption: Log-linear plots of tests E3, E4, E10 and E11 (moderate specific drawdowns).

Figure 7a

Caption: Bilinear plots of tests E2, E5, E6, E8 and E9. The specific drawdowns would fit to the dotted line if wellbore storage was the main source of drawdown.

Figure 7b

Caption: Log-log plots of tests E2, E5, E6, E8 and E9. The specific drawdowns would fit to the dotted line if wellbore storage was the main source of drawdown.

Figure 7c

Caption: Recovery plots of tests E2, E5, E6, E8 and E9. The null residual drawdowns would imply $(t+\tau)/t < 1$ if a closed barrier was the main source of drawdown.

Figure 8a

Caption: Local control of drawdown evolution (see Figure 8b): when E3 is the discharge well, specific drawdowns evolve as functions of $\ln(t)$ in wells E3 and E5.

Figure 8b

Caption: Local control of drawdown evolution (see Figure 8a): when E5 is the discharge well, specific drawdowns evolve as functions of t in wells E3 and E5.