SLOVAK GEOLOGICAL MAGAZINE

Utilizing Data on Specific Capacities of Wells and Water- Injection Rates in Regional Assessment of Permeability and Transmissivity

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Abstract.

Specific capacities from aquifer tests and water injection rates from water injection tests can be utilized for getting a true picture of spatial distribution of rock hydraulic properties even from less accurate and incomplete archival data inapplicable to an exact determination of hydraulic coefficients e. g. by straight-line method. After transformation to the approximative logarithmic parameters – permeability index Z and transmissivity index Y - the distribution characteristics of Z and Y are converted to the respective estimates of hydraulic conductivity and transmissivity. The conversion is simplified to the problem of an optimum estimate of the conversion difference than can be constructed analytically as a function of storativity, test time and well radius or found by analogy. The suggested procedure has proved its utility at regional hydrogeological research in many regions of various types.

Key words: hydrogeology, permeability, transmissivity, hydraulic conductivity, statistical evaluation, aquifer tests, water injection tests, methods of regional hydrogeology

(1 Fig., 2 Tabs)

At the present state of art, regional assessment of hydraulic properties of rocks is focused on the study of spatial and statistical distribution of permeability and transmissivity. In regional assessment, the number (density) of data is substantially more important than the absolute accuracy of the information on the hydraulic parameter value at an individual point of the studied space. To get a complete objective picture of the hydraulic properties of rocks on a regional scale, it is therefore indispensable to exploit to a maximum degree all, be it incomplete or less accurate, data on pumping from earlier wells and boreholes and to obtain from these data the necessary information in a way that would be adequate to their degree of accuracy. Such an effort resulted in defining approximative (comparative) logarithmic parameters - the permeability index Z and the transmissivity index Y, which can be determined in all cases where at least data on specific capacity of the well are available. The expression in the terms of such parameters signalizes a lesser degree of accuracy and rules out the risk of confusing the exactly determined hydraulic conductivity coefficient or the transmissivity coefficient with their inaccurate estimates. These parameters may be used advantageously also in a strongly inhomogeneous medium.

The number of wells in which specific capacity could be computed from archival documentation is, as a rule, considerably greater than the number of wells with complete data from aquifer tests making possible a more precise determination of aquifer parameters e. g. by interpretation of transient data. The specific capacity is therefore the starting point for getting sufficiently extensive and detailed information for regional assessment of permeability and transmissivity in hydrogeological practice.

The specific capacity or the ratio of the specific capacity to the length of tested interval was used at regional transmissivity and permeability evaluations by WALTON (1962), WALTON–CSALLANY (1962), ZEISEL et al. (1962), WALTON–NEILL (1963), CSALLANY (1963) and others. WALTON (1962, 1970) presented also charts for estimating transmissivity from specific capacity at selected values of storativity, time and well radius. Other authors studying the correlation between transmissivity T and specific capacity q suggested particular values of the ratio T/q for estimating transmissivity.

Determining specific capacity

The input for the derivation of the approximative logarithmic parameters is the specific capacity q defined as the ratio of the flow rate (withdrawn discharge) Q to the respective drawdown s in the well

$$q = Q/s \qquad (1)$$

Considering generally non-linear relation between Q and s, the discharge Q corresponding to the drawdown of one meter (s = 1 m) is substituted into the equation (1). Under water-table conditions, if only data for drawdown s_n greater than 1 m are available, the theoretical specific capacity q_1 for s = 1 m (standard specific capacity)

$$q_1 = q_n (2M - 1) / (2M - s_n) \qquad (2)$$

is derived (M = initial saturated thickness of an unconfined aquifer, s_n = observed drawdown of water table in the well, q_n = specific capacity calculated for $s = s_n$).

If the drawdown of unconfined water table in the well sw exceeds 1/10 of the initial saturated thickness M, the measured drawdown s_w should be adjusted by the formula (Jacob 1944)

$$s_c = s_w - s_w^2/2M \qquad (3)$$

substituting s_c into the calculations instead of the s_w. At the calculation of standard specific capacity q_1, e. g. according to the formula (2) is s = 1 m. Instead of the standard specific capacity value q_1 from the formula (2) we use in accordance with the correction (3) the adjusted value

$$q_{1c} = q_1 \;.\; 2M/(2M - 1) \qquad (4)$$

as the input specific capacity in calculating the approximative logarithmic parameters for a water-table aquifer.

In calculating the index Z, we could alternatively replace the correction (3) and (4) with a correction of the thickness M and substitute the corrected thickness

$$Mc = M - s_w/2 \qquad (5)$$

e. i. in the case for s = 1 m and M less than 10 m

$$Mc = M - 0.5 \text{ (m)} \qquad (6)$$

instead of the unaffected thickness M. As a matter of course, the uncorrected value of q_i is used in such a case.

At confined aquifers, the relationship between Q and s is linear up to certain drawdown. For greater drawdowns this relation is, however, non-linear even here. If a sufficient number of points for constructing the curve Q = f(s) is available, the standard specific capacity q1 is determined by a graphic extrapolation up to s = 1 m. If such an extrapolation is impossible, we estimate the standard specific capacity by a parabolic approximation, applying the relation (2) in the form

$$q1 = qn \;.\; (2H - 1)/ (2H - sn) \qquad (7)$$

where H = the height of the static level above the lower limit of the tested interval (Jetel 1993b) [q_n and s_n as in (2)]. This approximation is only a rough estimate of an unknown non- linear course of the curve Q = f(s). Nevertheless, it makes possible an objective reproducible correction of the decrease of specific capacity with drawdown.

Expressing the discharge Q in the formula (1) in $m^3 . s^{-1}$ and the drawdown in meters, the specific capacity is expressed in $m^2 . s^{-1}$. If we express the discharge in liters per second, the specific capacity will be expressed in liters per second per metre ($l . s^{-1} . m^{-1}$). In such a case the distinctive symbol qo is used.

Approximative logarithmic parameters

The approximative logarithmic parameter of permeability derived from specific capacity is the permeability index Z (Jetel 1964, 1968, 1974)

$$Z = \log(10^9 q/M) = 9 + \log(q/M) \qquad (8)$$

$$\text{or} \quad Z = \log(10^6 q^o/M) = 6 + \log(q^o/M) \qquad (8a)$$

where q = standard specific capacity in $m^2 . s^{-1}$, q^o = standard specific capacity in $l . s^{-1} . m^{-1}$, M = aquifer thickness in meters. With partially penetrating wells or in the case of an impossible or indefinite determination of the thickness M (e. g. in a fractured rock massif without distinct delimitation of aquifers) a substitutive parameter - the permeability index of the open interval

$$Z_L = \log (10^9 q/L) = \log (10^6 q^o/L) \qquad (9)$$

is used (L = the length of the open interval in the well below the static level).

At water-table aquifer with the thickness less than 10 m the corrections (4) or (6) are to be applied for the formulae (8) and (8a).

In a partially penetrating well, M is greater than L so that Z_L is greater than the value of Z from Eq. (8) that would be measured in a fully penetrating well. Let us mark the value of Z in fully penetrating well by Z_M, qL being the actual specific capacity substituted into Eq. (9). The symbol q_M will then indicate the theoretical specific capacity of fully penetrating well. Thus,

$$Z_M = \log(10^9 q_M/M) \quad (10)$$

If the inflow from the more distant unopened parts of an aquifer - in the extent greater than 3L - into the partially penetrating well is considered negligible, the theoretical value of Z_M is estimated as

$$Z_M = (4 Z_L - 0.48) / 4 \quad (11)$$

(JETEL , 1993b). The formula presented by TURCAN (1963)

$$q_L / q_M = b\left[1 + 7\sqrt{r_w / 2bM} \ . \ \cos(b\pi/2)\right] \quad (12)$$

where b = L/M, r_w = well radius, can be used for a more precise estimation. The value q_M is then substituted into Eq. (10).

The transmissivity index Y (JETEL – KRÁSNÝ 1968, JETEL 1974)

$$Y = \log(10^9 q) = 9 + \log q \quad (13)$$
$$\text{or } Y = \log(10^6 q^o) = 6 + \log q^o \quad (13a)$$

representing a logarithmic transformation of specific capacity is the approximative parameter of transmissivity. From eqs. (8) and (13) it follows that for fully penetrating wells

$$Z = Y - \log M \quad (14)$$

For partially penetrating wells with L smaller than M is the value Y from Eq. (13) an indication of certain effective transmissivity at the particular penetration degree and does not correspond to the total transmissivity of the aquifer at fully penetrating well. Provided that the permeability in the whole thickness M is roughly uniform, the representative value of the Y_M can be estimated from the specified Z_M value as

$$Y_M = Z_M + \log M \quad (15)$$

Logarithmic transformation of hydraulic conductivity k and transmissivity coefficient T (Jetel 1979)

$$Z_k = 9 + \log k \quad (16)$$
$$Y_T = 9 + \log T \quad (17)$$

(T in $m^2 . s^{-1}$, k in $m . s^{-1}$) and antilogarithmic transformations of the indices Z and Y (JETEL, 1985ab)

$$k_Z = \text{antilog}(Z - 9) = 10^{(Z-9)} \quad (18)$$
$$T_Y = \text{antilog}(Y - 9) = 10^{(Y-9)} \quad (19)$$

facilitate a direct comparison of the approximative logarithmic parameters with the exact hydraulic parameters on a common scale.

Logarithmic conversion difference

The parameters Z and Y are simple functions of the specific capacity. Hence, the estimation of exact hydraulic coefficients may make use of the relation between transmissivity T and specific capacity q that may be expressed in terms of the logarithmic conversion difference (Jetel 1979, 1985 ab)

$$d = \log T - \log q \quad (20)$$
$$\text{e. i. } T/q = 10^d \quad (21)$$

with T and q in the same units. The first solution converting the indices Z and Y to hydraulic conductivity and transmissivity at transient flow presented by Carlsson and Carlstedt (1977) by introducing a coefficient related to the difference from Eq. (20) as follows

$$\alpha = 10^{-d} \quad (22)$$

Comparing Eqs. (20) and (13), we see that the relation between transmissivity coefficient T and transmissivity index Y is expressed as

$$T = \text{antilog}(Y+d-9) = 10^{(Y+d-9)} \quad (23)$$

(T in $m^2 . s^{-1}$). A parallel relation

$$k = \text{antilog}(Z+d-9) = 10^{(Z+d-9)} \quad (24)$$

applies to the hydraulic conductivity k expressed in meters per second.

A logical consequence of the combination of Eqs. (17), (20) and (23) is the relation

$$d = YT - Y \qquad (25)$$

The introduction of the conversion difference simplifies the estimation of the hydraulic coefficients from the specific capacity to the problem of an optimum estimate of an additive quantity - the conversion difference d.

Basic (primary) conversion difference for an ideal well

A fundamental component of the total conversion difference d in Eqs. (20) – (25) is the difference between log T and log q for particular calculation conditions on the assumption of an ideal well without any well loss (e. i. without any additional hydraulic resistance to the flow into the well and within it). We have termed this component the basic (primary) conversion difference d_o.

Under steady-state flow conditions, the Dupuit's formula implies that the basic conversion difference is given by (JETEL, 1982)

$$d = \log [\log (r_d/r_w)] - 0.436 \qquad (26)$$

where r_d is depression cone radius and r_w well radius. For non-steady state flow fulfilling the conditions of the Cooper–Jacob logarithmic approximation validity (COOPER–JACOB, 1946), the following relationship may be derived (JETEL, 1985 ab):

$$d_o = \log [0.183 - \log (2.25\, dt/r^2w)] \qquad (27)$$

where

$$D = T/S \qquad (28)$$

(D = hydraulic diffusivity in $m^2 . s^{-1}$, T = transmissivity in $m^2 . s^{-1}$, S = storativity, t = time after pumping started determining the radius of test influence at the particular moment). Hydraulic diffusivity D is estimated from preliminary assessment of T – e . g. by the TY from Eq. (19) - and from the expected range of storativity. The storativity S of a confined aquifer is computed from the estimate of specific elastic storativity S_s as

$$S = S_s . M \qquad (29)$$

(S_s in m_{-1}). The specific storativity S_s is estimated by Tab. 1 or by the formula

$$S_s = \rho g (mo_c w + c_r) \qquad (30)$$

where ρ = water density, g = gravitational acceleration, m_o = open porosity, c_w = water compressibility, c_r = bulk compressibility of rock skeleton. The time t in Eq. (27) is the time elapsed from the beginning of the test up to the instant for which specific capacity was measured. The r_w is the actual inner radius of well. In unconfined aquifer, the water-table storativity (specific yield) can be estimated at S = 0.24 for gravel and coarse sand, 0.22 for medium sand, 0.19 for fine sand, 0.15 for silty sand and 0.05 - 0.15 for loamy sand (MUCHA – ŠESTAKOV, 1987). In the near-surface zone of fractured rocks the values S = 0.02–0.05 may be recommended.

Eq. (27) is valid only before boundary effect or leakage appears. After the depression cone reached a recharge boundary,

$$r_d = 2\, x_b \qquad (31)$$

where x_b is the distance of the well from the recharge boundary, is substituted into Eq. (26). If leakage effect is assumed, we substitute

$$r_d = 1.12 \sqrt{TM / k} \qquad (32)$$

(M' = aquitard thickness, k' = vertical hydraulic conductivity of aquitard, T = estimated transmissivity of aquifer) into Eq. (26).

Additional conversion difference

Under real conditions the total conversion difference d from Eq. (25) should be understood as the sum of the basic and additional differences

$$d = d_o + d_a \qquad (33)$$

The additional difference d_a expresses the deviation of actual conditions from the ideal well model. It represents the well loss, in other words the sum of differences caused by additional flow resistances arising close to the well screen, on the screen and inside the well. Theoretically it may be itemized into partial differences of various origins:

$$d_a = d_S + d_L + d_C + d_X$$

where d_S is skin difference, d_L = difference due to partial penetration, d_C is turbulent flow differencey and d_X comprises other unidentified differences. The skin difference d_S reflects the resistance corresponding to the skin effects s. l. (VAN EVERDINGEN, 1953, EARLOUGHAR, 1977) e. i. the changes in permeability in the near-well zone, the reduction of flow entrance area on the well screen etc. The difference d_L corresponds to the effect of partial penetration

$$d_L = \log(q_M/q_L) \qquad (35)$$

[qM determined e.g. by Eq. (12)]. It can be, however pre- eliminated at calculating the Z-value by Eqs. (10) – (12). The difference d_C due to the flow turbulence expresses the effect of the quadratically nonlinear resistance, mainly the turbulence within the well. It is of importance at discharges reaching tens and hundreds liters per second. It may be approximated as

$$d_C = \log \frac{\text{antilog } d_o + Q/r_w T^{0.25}}{\text{antilog } d_o} \qquad (36)$$

(antilog x = 10^x, Q = discharge, r = well radius, T = transmissivity). The formula has been derived (Jetel, 1985ab) from generalized empiric data given by CARLSSON–CARLSTEDT (1977) and GUSTAFSON (1974).

Analyzing the distribution of the Z and Y values

After computing individual values of permeability and transmissivity indices, we turn to statistical analysis aimed at determining the statistical characteristics of the distribution of Z and Y values in data populations corresponding to particular lithostratigraphic units, rock types or regions, first of all, minimum and maximum values, medians Md, sample arithmetic means M(Z) or M(Y) and sample standard deviations s_Z and s_Y with the estimates of general population standard deviations are to be determined. The statistical significance of the stated differences in sample means should be tested as well. It is necessary to delimitate the confidence intervals within which are located the true (population) means with the specified probability.

Simultaneously with the computation of the characteristics mentioned above it is useful to visualize the identified distribution by histograms and quantile (frequency) graphs (JETEL, 1985a). The quantile graph (cumulated relative frequencies graph) makes it possible to asses the conformity of the displayed data with the normal distribution model. If the graph indicates pronounced deviations from the normal model, it is necessary to verify the relevance of the extreme values to the studied population and to decide if the apparently homogeneous set of data is not an intersection of two or more subsets with theirs own distributions that must be studied separately.

An illustrative confrontation of individual data sets is possible by the "box-and-whisker plot" (e. g. GUSTAFSON–KRÁSNÝ 1994, JETEL, 1994, JETEL–VRANOVSKÁ, 1995) for which quartile values are to be computed.

After expressing the statistical characteristics of the distribution of approximative parameters Z and Y, these characteristics are to be transferred to the corresponding distribution characteristics of non-logarithmic hydraulic coefficients – hydraulic conductivity k and transmissivity T. After adequate transformations, the minimum and maximum values of the estimates of k and T as well as the medians Md(k) and Md(T) correspond directly to the minimum, maximum and median values of Z and Y. By contrast, it is the geometric mean of a non-logarithmic coefficient that is the statistical characteristic corresponding to the arithmetic mean of a logarithmic parameter:

$$G(k) = \text{antilog}\,[M(Z)+d-9] = 10[M(Z)+d-9] \quad (37)$$

$$G(T) = \text{antilog}\,[M(Y)+d-9] = 10[M(Y)+d-9 \quad (38)$$

where G(k), G(T) are geometric means of the respective coefficients, M(Z), M(Y) = arithmetic means of the indices Z and Y, d = total conversion difference. Normal (Gauss) distribution of the logarithmic parameters (Z, Y or other logarithmic transformations) indicates lognormal (Galton) distribution of the respective non-logarithmic coefficients (k, T). The variability is characterized also here by the standard deviation of logarithmic parameters (s_Z, s_Y) as it is practically identical with the standard deviation of non- logarithmic parameter logarithms (slog k, slog T).

The choice of the averaging method of individual k and T values depends on several factors. Large-scale permeability tests (WITHERSPOON et al., 1980, GUSTAFSON et al., 1989, GALE et al., 1989, BROCH–KJORHOLT, 1994) show that the geometric mean of permeability values determined from borehole tests agrees reasonably well with bulk rock mass permeabilities determined during macropermeability experiments. From this point of view, the geometric mean should considered as a true characteristic of mean permeability and transmissivity (cf. JETEL, 1985a). Yet where one-dimensional flow in a series connection of flow segments is assumed, it is the harmonic mean that conforms best to the effective permeability and transmissivity of such a series. In statistically homogeneous medium with lognormally distributed permeabilities, the geometric mean expresses according to GUTJAHR et al. (1978) and DAGAN (1979) the effective permeability for 2-dimensional flow while for 3- dimensional flow it is the effective mean

$$Ef(k) = G(k) \,.\, [1 + s^2_{\ln k}/6] \;-\; G(k) \,.\, [1 + 0.884\, s^2 \log k] \qquad (39)$$

If it is impossible to measure with confidence and to express quantitatively permeabilities smaller than a certain lower limit and an accepted minimum value is consequently substituted into the calculations, the geometric mean becomes overestimated and the median value will be its optimum estimate (BROCH–KJORHOLT, 1994).

Since the permeabilities (hydraulic conductivities) and transmissivities are, as a rule, distributed lognormally, there is still another characteristic of their mean value – the mathematical expectance of lognormally distributed values

$$EL(x) = G(x) \,.\, \psi n\,(t') \qquad (40)$$

where $$t' = 2.65\, s^2_{\log x} \qquad (41)$$

($s_{\log}$ x = standard deviation of the logs x). It is the mean value with the maximum likelihood for the given distribution, e. i. a hypothetical average generating with the maximum probability the observed empirical distribution. The values of the function are given by AITCHISON and BROWN (1957) (in more detail see JETEL, 1985a). At slog k less than 0.45 the EL (k) is close to the Ef(k) = Ef(x) from (39)

Besides the approximative logarithmic parameters Z and Y derived from specific capacities also the values of hydraulic conductivities k and transmissivities T determined by exact methods – e. g. by straight-line (semi-log) method from transient tests – can be used in the same manner for deriving the regional characteristics of permeability and transmissivity. In such a case, it is suitable to convert the values of k and T by the transformations (16) and (17) to the form compatible with the form of indices Z and Y.

Estimating hydraulic conductivity and transmissivity from the values of Z and Y

For converting Z and Y values to the corresponding estimates of hydraulic conductivity k and transmissivity T we use Eqs. (23) and (24) substituting the appropriate conversion difference d. To determine the difference d, the following procedures are available:

(a) In estimating k and T values from individually found values of Z or Y, the conversion difference is determined analytically – combining calculations by Eqs. (26) and (27) with estimates by analogy.

(b) In estimating the statistical characteristics of the distribution of k and T from calculated characteristics of the Z and Y values distribution, the conversion difference is estimated

(ba) analytically in the sense of (a) for particular minimum, median and maximum values of Z and Y,

(bb) from generalized regional estimates by means of regression equation d = f(Y) derived for particular found or constructed values of d, namely from the average estimates determined

(bba) analytically ad hoc for particular evaluated data,

(bbb) by previous works in the region,

(bc) by analogy with other regions.

The (bb) way is used where the converted values of Z or Y do not represent any actually measured values with data necessary to an analytical construction of d-value (e. g. et converting a computed mean).

At an analytical construction of d, the value of d_o is determined by Eq. (27) or (26). With high discharges from wells of small diameters the difference dC is computed by (36). If the correction on partial penetration has not been applied at calculating Z and Y by Eqs. (10) or (11) and the actual conditions substantiate the use of the partial penetration model, the difference d_L is estimated by (12) or from tables and charts (JETEL, 1985ab). The skin difference d_S cannot be analytically computed. In a first approximation it can be neglected. For a more accurate estimate we use an analogy with the wells in which the actual value of d could be determined from comparing Y with the transmissivity found directly from high quality data (e. g. by the straight-line method). From Eq. (25) the value of d is obtained so that

$$dS = d - d_o - d_C - d_L \qquad (42)$$

Another approximation is possible by substituting systematically a certain small constant value – e. g. $d_S = 0.1$ – for wells of standard construction.

The regional mean estimate of the d-values for a region (a statistically homogeneous set of data) may be derived by computing the regression equation

$$d = a + bY \qquad (43)$$

Such average estimates derived from a particular equation of the type (43) are substituted into Eqs. (23) and (24). As examples, the empiric equations derived for the wells in alluvial aquifers in the Košice basin (Jetel, 1993b)

$$d = 0.07\,Y - 0.29 \qquad (44)$$

for the neovolcanics of the Slanské vrchy Mts. (JETEL, 1993a)

$$d = 0.13\,Y - 0.40 \qquad (45)$$

or for Paleogene in Hornád basin and Spišská Magura Mts. (JETEL – VRANOVSKÁ, 1995)

$$d = 0.23\,Y - 0.94 \qquad (46)$$

can be mentioned. Preliminary estimation of the d-values can be based on the generalized experience from various regions (Tab. 2). The total difference d in the wells in which it has been determined by Eq. (25) ranges from d = -0.30 to +1.15 with scarce exceptions. The highest values of d, da and dS are observed in deep boreholes of mineral deposit exploration in connection with high additional resistances due to mud fluid and imperfect perforation. The experience with the boreholes in the Hornád basin and Spišská Magura Mts. shows that in the intervals tested Tab. 2 The most frequent values of total, additional and skin difference at aquifer tests at first with open unlined wall the covering of the walls by perforated casing (screen) increased the additional difference da by 0.1–0.6 (0.25 on the average).

Interpretation of water injection test data

Water injection tests (pressure tests), originally a geotechnic method to studying rock basement tightness and grouting conditions, afford very useful data for exploring spatial distribution of permeability and are often the sole source of direct information on permeability in mountaineous regions without hydrogeologic boreholes. A direct computation of hydraulic conductivity from individual measured data is problematic with regard to some non-measurable input characteristics. However, for a regional assessment we can easily dispense with assigning a particular value of hydraulic conductivity to each tested interval. An approximative parameter – permeability index of tested interval Z' can be again used conversing the characteristics of its distribution to the corresponding characteristics of hydraulic conductivity.

Water injection rate (the volume of water injected during a time unit) is similarly to Eq. (1) transferred to the specific injection rate

$$g_h = Q_h/s_h \qquad (47)$$

where

$$s_h = p/\gamma + H_h \qquad (48)$$

is elevation of piezometric head in well, p is overpressure on the well orifice, γ is specific weight of water (9.8×10^{-3} $N.m^{-3}$) and H_h is the depth of static level below well orifice (in a non-saturated interval it is the depth of the tested interval base). Similarly to Eq. (9) the permeability index Z' at water injection test is defined as

$$Z' = \log(10^9 q_h/L) = \log(10^6 q^o{}_h/L) \qquad (49)$$

where q_h is specific injection rate from Eq. (47), qoh being expressed in $l \,.\, s^{-1} \,.\, m^{-1}$ after substituting Q_h in liters per second. In the documentation of injection tests the quotient of injection rate Q_h and length of tested interval L is often presented. Instead of q_h/L the quotient Q_h/Ls_h is then substituted into Eq. (49). The value of Z' from Eq. (9), yet with changed conversion difference.

Tab. 1 Specific elastic storativity S_s (from MIRONENKO–SHESTAKOV, 1978

H Depth (m)	Rock	S_s (m^{-1})
10–50	sand clay	0.007/H $4 \times 10^{-4} - 7 \times 10^{-4}$
50–200	sand clay sandstone, siltstone limestone, marl	$5 \times 10^{-5} - 2 \times 10^{-4}$ $1 \times 10^{-4} - 4 \times 10^{-4}$ $3 \times 10^{-5} - 1 \times 10^{-4}$ $1 \times 10^{-4} - 4 \times 10^{-4}$
deep-seated aquifers		$10^{-5} - 10^{-6}$ ± 1–2 orders of magnitude

Tab. 2 The most frequent values of total, additional and skin difference at aquifer tests

	d min. max..	Md(d) min. max..	d_a min. max	Md(d) min. max.	d_s min. max.	Md(d_S) min. max..
hydrogeologic boreholes less than 100 m deep	-0.15 +0.30 (+0.60)	+0.10 +0.20	-0.20 +0.50	+0.10 +0.25	-0.25 +0.60	0.10
deeper hydrogeologic boreholes	-0.10 +0.80 (+1.00)	-0.20 +0.50	-0.20 +0.60	+0.10 +0.30		
small-diameter boreholes of mineral deposits exploration	-0.10 +1.10 (+1.20)	+0.30 +0.60	-0.10 +1.00	+0.30 +0.40	-0.10 +1.00	+0.30 +0.40

d = total conversion difference, d_a = additional difference, d_S = skin difference M_d = the most frequent median values in large sets of data possible more seldom values are given in parentheses

The hydraulic conductivity estimated from injection rate is often expressed as

$$k = C_f \cdot g_h/L \qquad (50)$$

where C_f is a dimensionless constant (shape factor). Particular values of C_f were derived for various assumed patterns of flow about the test zone (HVORSLEV, 1951, LOUIS–MAINI, 1970, ZIEGLER, 1976, GALE et al., 1982, CHAPUIS, 1989 and others), but they yield very discrepant results. As follows from the analysis of the relations between injection rate and hydraulic conductivity, the shape factor C_f cannot be constant being actually a function of transmissivity and storativity (JETEL, 1993b). The conversion of injection rate to the parameter Z' implies the use of the conversion difference d for estimating hydraulic conductivity. The injection test conditions differ distinctly from both the steady-state flow described by Dupuit formula and the unsteady constant-discharge flow expressed by COOPER-JACOB approximation. As shown by DOE and REMER (1980), the constant-head radial flow model should be used to the analysis of unsteady flow at injection tests.

It is the choice o proper radius of influence r_d that is the key to the optimization of the conversion difference value for injection tests. With this in mind

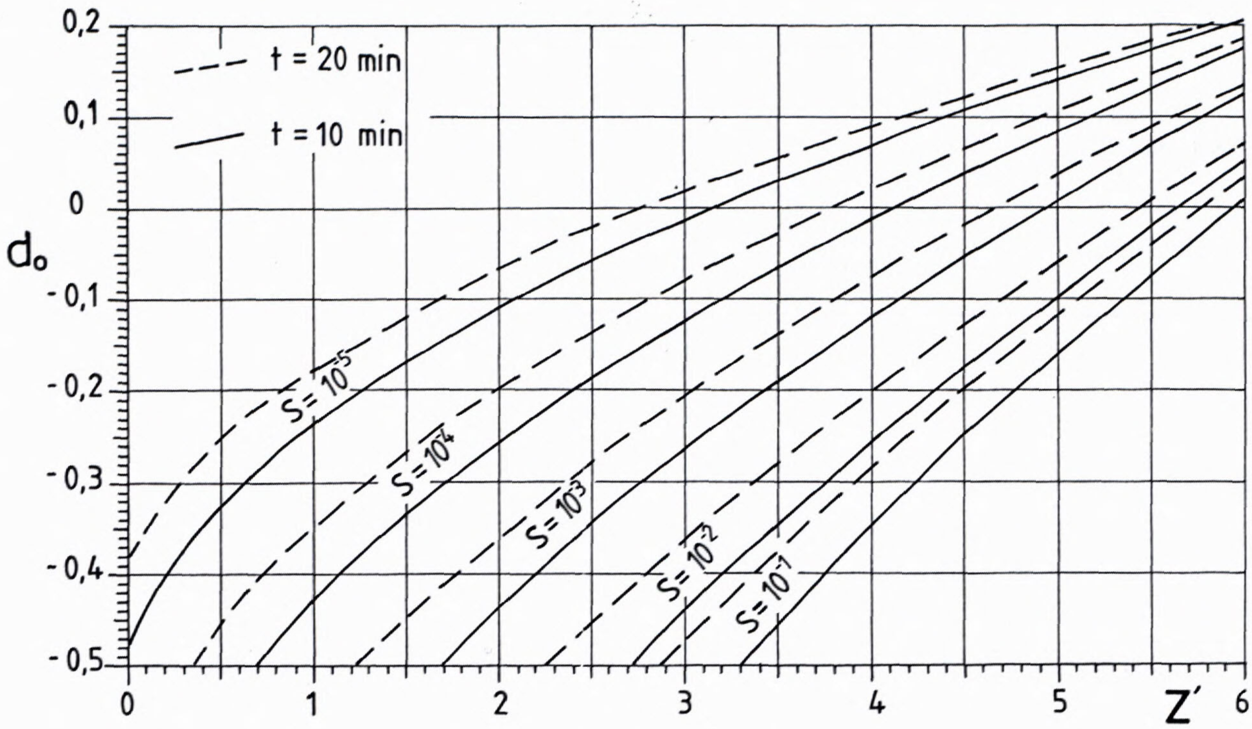

Fig. 1. The basic conversion difference d_o at water injection test as a function of permeability index Z', test time t and estimated storativity of tested interval S

we adopted the approach suggested by DOE and REMER (1980). The error in using steady- flow formula (Dupuit equation)

$$Q_s = 2\pi k s_h L / \ln(rd/rw \qquad (51)$$

(r_d = radius of influence, r_w = radius of ideal well free of well loss, L = length of tested interval) rather than the transient technique can be quantified by preparing flowrate data versus time plots using the equation of constant-head unsteady radial flow (JACOB–LOHMAN, 1952)

$$Q_t = 2\pi T s G(\chi) \qquad (52)$$

where G (χ) is the constant-head well function. When substituting s_h from Eq. (48) for the drawdown s, Q_t will be the injection rate in the time t. The average flowrate Q_{ta} over the period of time flowrate data were taken is determined by integrating the flow over the period of time in question. By comparing step-by-step the Q_{ta} values representing the average Q_t from Eq. (52) in the final pseudostabilized phase of the test with the Qs values from Eq. (51) at varying r_d/L and k, the optimum r_d/L value is found as the value yielding zero difference between Q_{ta} and Q_s.

The analysis made by DOE–REMER (1980) indicated the essence of relations between variables in question. It is however insufficient to a generally applicable determination of optimum values of r_d for estimating hydraulic conductivity. In particular variants the values L = 3 m, 2 r_w = 79 mm and s_h = 30 m were taken as constants corresponding to the most cases interpreted hitherto. More detailed information on the computations and their results is given by Jetel (1993b). Substituting the values of r_d for which $Q_s = Q_{ta}$ into Eq. (26), the optimum values of d_o were found as a function of the Z' and storativity of the tested interval. The results are shown in Fig. 1. The computed relations are described approximately also by the following regression equations:

– for the test time 10 min

$$d_o = 0.395 Z^{1/2} - 0.130 \log S - 1.330 \qquad (53)$$

– for the test time 20 min

$$d_o 6= 0.348 Z^{1/2} - 0.118 \log S - 1.155 \qquad (54)$$

The equations (53) and (54) predict fairly well the actual values of do at $S = 10^{-5} - 10^{-4}$ but they cannot be recommended for S exceeding 10^{-3} where they begin to fall off conspicuously from the directly determined values.

As to the choice of storativity values, in the uppermost parts of the near-surface zone close to the water level it is recommended to assume the values of the order of $S = 10^{-2}$, exceptionally 10^{-1}. In deeper parts the relation

$$S = S_s . L \qquad (55)$$

can be used with the data from Tab. 1.

The comparison of the determined optimum values of r_d/L with the values of C_f suggested by the authors mentioned above shows that the constant C_f could be valid in very narrow ranges of S and k only.

The conversion of the values of Z' is applied mainly when converting the statistical characteristics of Z' values to the respective characteristics of hydraulic conductivity distribution e. g.- in particular depth zones. To adjust the determined value of do for additional resistances we recommend to add to the value of do from Fig. 1 as a conventionally accepted estimate $d_a = 0.05$ or, at high injected rates exceeding 10–20 liters per minute, $d_a = 0.10$.

Statistical evaluation of water injection data

After the conversion to the values of Z', the results of water injection tests are statistically evaluated in the same way as with aquifer tests. With regard to regular decrease of mean permeability with depth, the data sets are defined not only by regions and lithostratigraphic units but also according to chosen depth zones (JETEL, 1985 ac, 1994) and by the position in the relief (valley bottoms or slopes). It is extremely important to determine the parameters of exponential regression equation

$$k(H) = k_o \exp(-AH) \qquad (56)$$

where k(H) = mean hydraulic conductivity expected at the depth H, k_o = theoretical hydraulic conductivity for H = 0, A = exponential decrease coefficient. This type of equation presents an optimum approximation of the course of permeability changes with depth. Thus, the parameters k_o and A are the cardinal characteristics of hydraulic behaviour of rock massif at permeability decreasing with depth. However, in different depth zones these parameters can change (A usually decreases with depth).

Conclusions

When evaluating the archival data of aquifer tests or of water injection tests, the data on specified capacities and water injection rates can serve as invaluable information base. By means of conversion to the approximative logarithmic parameters Z and Y it is possible to arrive to consistent estimates of regional characteristics of permeability a transmissivity. The key to an optimum estimate of hydraulic coefficients corresponding to the approximative indices Z and Y is the determination of proper conversion difference d. The delineated procedure proved to be useful at regional hydrogeological research in many regions of various types. Its application is conditioned by sufficient number of data allowing to compute at least specific capacities (the ratio discharge (drawdown) or water injection rate. It is possible to include in the evaluated data – after appropriate logarithmic transformation - also the values of hydraulic conductivity or transmissivity determined directly by exact methods, namely by transient techniques (e. g. straight-line method).

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