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Geohydrological Properties of Tectonic Zones in Hard Rocks Obtained from Artificial Recharge Tests and Numerical Modeling

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Results from artificial recharge tests through wells in Angered, Gothenburg have made studies on geohydrological properties of tectonic zones possible. Measurements of piezometric head in bore holes were used to evaluate the transmissivity and storage coefficient of the zones by using a finite element model. The zones were considered to act as separate confined aquifers, and the values obtained were $T \equiv 3\text{-}4 \cdot 10^{-6}$ m²/s and $S \equiv 4\text{-}6 \cdot 10^{-5}$. The hydraulic conductivity of the rock mass was then found to be greater than $2\text{-}5 \cdot 10^{-8}$ m/s, which is in agreement with values of fractured, crystalline rock in Sweden obtained in other investigations.

Introduction

Lowering of the groundwater level and land subsidence caused by water leakage into deep-lying tunnels in the bedrock have been observed in the urban regions of Stockholm and Gothenburg in Sweden (Broms 1973; Lindskoug and Nilsson 1974; Jansson and Winqvist 1976; Broms, Fredriksson and Carlsson 1976).

Tectonic water-bearing zones and their geohydrological properties are of great importance when estimating the leakage into tunnels, the influenced areas, and also the effect on the piezometric levels in the soil. In Angered, a new suburb northeast of Gothenburg, decreasing piezometric head in bedrock and soil were observed during the excavation of tunnels in the area. Recharge tests were performed in order to restore the lowered piezometric head in the soil. Pumping tests could not be performed because of the lowered piezometric head and the low transmissivity. The recharge tests then made it possible to evaluate the geohydrological properties of the soil (Carlsson and Kozerski 1976; Carlsson 1978; Andersson, Bergman and Carlsson 1978). The results of these tests together with detailed geological documentation of the area (Wedel 1978) were used to estimate the geohydrological properties of two of the present tectonic zones. The purpose of this paper is to present the procedure used and the results obtained.

Hydraulic Conductivity of Fractured Rock

Crystalline Rock as a Water-Bearing Medium

Rock aquifers can be classified as

porous rock – fractured rock – karstified rock.

Aquifers of crystalline rock in Sweden are usually the second type, since they have a very low porosity. The fracture configuration determines the geohydrological properties of the bedrock and thus the tectonics of crystalline rock are of great importance to the geohydrologists.

The Relation Rock Mass Properties – Fracture Properties

The concept of hydraulic conductivity of fractured rock must be used with caution. It should be stated if the value is referring to the unfractured rock, to a certain volume of fractured rock or to the rock mass as a whole. On a regional scale, mean values of the hydraulic conductivity should be used, but for detailed investigations the properties of each fracture or set of fractures must be considered.

The hydraulic conductivity of fractured rock depends on

the width of the fracture
the roughness of the fracture
the kinematic viscosity of the water
whether fracture filling is present
the continuity of the fractures
the spacing of the fractures.

In normal conditions the first three of these are usually of such a magnitude that laminar flow occurs. The viscosity of groundwater varies little, since the temperature and chemical composition vary little at depths of less than 50 m.

For laminar flow of an incompressible fluid between smooth, parallel plates, the flow rate q may be expressed as (Snow 1969)

$$q = -\frac{b^3 g}{12 \nu} \frac{\partial h}{\partial x} \tag{1}$$

where b is the width of the fracture, ν is the kinematic viscosity of the water, and g is the gravitational acceleration. From this expression a "hydraulic conductivity" K_p for the space between the parallel plates can be derived:

$$K_p \equiv \frac{b^2 g}{12 \nu} \tag{2}$$

By analogy with this the hydraulic conductivity K_f of a fracture of width b can be expressed as

$$K_f \equiv C b^2 \tag{3}$$

where C is a constant depending on the viscosity of the fluid and the roughness of the fracture.

If fractures of a constant width b are uniformly distributed in the rock mass with a spacing d and are parallel to the hydraulic gradient, the mean hydraulic conductivity K_r is given by

$$K_r \equiv C \frac{b^3}{d} \tag{4}$$

Consequently, as mentioned by Snow (1969), the hydraulic conductivity of the rock mass is proportional to the cube of the width of and inversely to the spacing between fractures.

Hydraulic Conductivity of the Bedrock in Sweden

Values of hydraulic conductivity given for Swedish rock types are some kind of mean value for large rock masses, e.g. a value referring to the rock along 100 m of tunnel wall or to the rock mass surrounding a drilled well. But a few authors have given values for 2-3 m intervals in a bore hole.

Carlsson and Olsson (1977) have performed water-loss measurements in bore holes in five areas of Sweden. The tests have comprised water injection between two packers some 2 m apart. They report the hydraulic conductivity to be 10^{-7} – 10^{-5} m/s in fractured crystalline rock at depths of less than 50 m (see Fig. 1).

Carlsson and Carlstedt (1976) have made a statistical analysis of pumping-test data from wells to obtain average values of transmissivity and hydraulic conductivity for different Swedish rock types. The results of their work are shown in Fig. 1. They also stress the importance of the degree of tectonization.

Wedel (1978) reported values of hydraulic conductivity for the bedrock in the

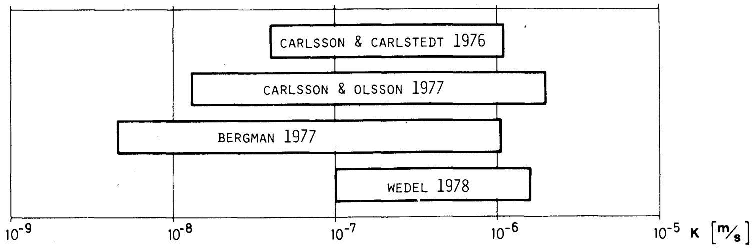


Fig. 1. Reported values of the hydraulic conductivity of crystalline rocks in Sweden.

Angered area of $10^{-7} - 10^{-6}$ m/s. In this case injection tests and tracers have been used.

Bergman (1977) has calculated average values of hydraulic conductivity from measurements of water leakage into 73 tunnels and rock caverns. Most of them were post-grouted. This accounts for the relatively low conductivity value of this report.

Mathematical Models of Fracture Flow

Background

Groundwater flow in fractured rock can be treated in two different ways (Parsons 1972). One is *the discontinuum approach,* where the geometry and hydrology of each fracture or set of fractures are described specifically. The other is *the continuum approach,* where the rock mass as a whole is considered practically homogeneous, and therefore the ordinary geohydrological parameters of porous media can be used. Which approach to use depends on the scale of the work to be done. For regional groundwater analysis the continuum approach is most suitable, but e.g. when groundwater flow to a well is studied the discontinuum approach would be better. Still, in most cases the continuum approach has to be applied because of lack of necessary data.

The mathematical treatment of fracture flow can, of course, be more or less complex. The advantage of the continuum approach is that ordinary mathematical models developed for flow in porous media can be used. If one set of fractures dominates the flow pattern, this is simulated by an anisotropic hydraulic conductivity usually defined by the direction and magnitude of the greatest and smallest conductivity. This approach has been described by Snow (1969) and Parsons (1972).

The other way to treat flow in fractured media is the discontinuum approach, which means that the actual conditions within these discontinuities are treated mathematically. The first step is then to study flow in one plane, open fracture. This case has been described by Wittke (1969). He has treated different degrees of roughness of the fracture.

A large plane tectonic zone having greater hydraulic conductivity than the surrounding rock mass can sometimes be treated as an aquifer itself. Especially for fault zones and overthrusts, as in the Angered case, this approach can be justified. We have used it in order to quantify the geohydrological properties of such a zone.

The next step towards a more accurate treatment of groundwater flow in fractured rock is to consider more than one set of fractures. Castillo, Karadi and Krizek (1972) have solved a problem involving a two-dimensional, unconfined flow in a medium with two crossing sets of joints. An even more sophisticated description of reality can be obtained when a porous medium with open fractures is considered. This approach can also be applied to an impermeable rock with both large and small fractures. The effect of the system of small fractures is similar to that of the porous matrix. This problem has been studied in connection with oil production, and both two-dimensional and three-dimensional flows have been considered, by among others Gringarten and Witherspoon (1972) and Closmann (1975). Three-dimensional flow in non-porous rock with three crossing fracture planes has been studied numerically by Wittke, Rissler and Semprich (1972).

Computer Program Used

In recent years numerical methods have been used increasingly to solve groundwater flow problems. The methods used are the finite difference method (FDM) and the finite element method (FEM), because they both are suited for computer calculations.

In this work we have utilized a FEM-program called GEOFEM-G developed at Chalmers University of Technology, Gothenburg, which performs 2-D confined aquifer analysis (Runesson and Wiberg 1977). GEOFEM-G is easily available at the Gothenburg Universities' Computing Centre and has at least five special characteristics worth mentioning, i.e.

input data are given in a free format
superparametric elements are used
the design of the element net has hardly any limits
the time-stepping procedure works automatically
the output data are written in matrix format, which makes them easy to read.

Necessary input data are the geometry of the element net, material properties (T and S), and boundary conditions. The output is the head at the nodes, the groundwater flux within the elments, and flow caused by imposed boundary conditions.

Geohydrological Properties of Tectonic Zones in the Angered Area

Geology

The geological and hydrogeological conditions of the Angered area have been described in detail by Wedel (1975, 1978). The area is situated about 60-70 m above sea level and forms a clay plain with small hills of bedrock within and around the plain, as shown in Fig. 2.

The bedrock consits of gneisses of different composition and configuration. The bedrock topography is dominated by the imbrication in small nappes separated by thrusts. These are dipping westerly, and the bedrock is further divided into blocks by nearly vertical joints in the WNW and NE directions.

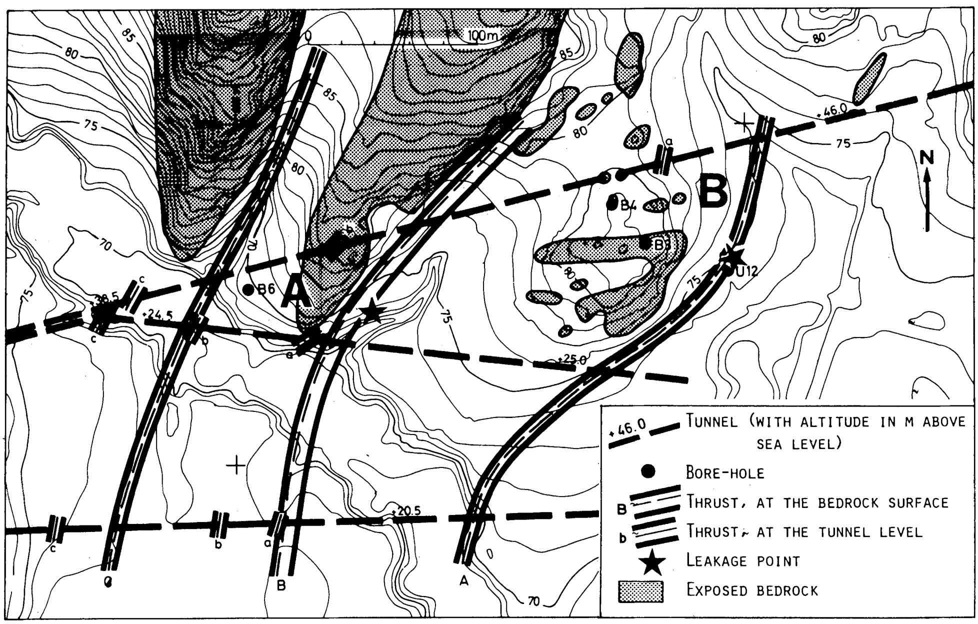


Fig. 2. Map of the Angered Area (partly from Wedel, 1975 and 1978).

The soil is mostly an overconsolidated clay with many thin silty layers. Usually the clay lies directly on the bedrock, but in some places there is a layer of frictional material on the bedrock surface. This layer has a maximum thickness of 1-2 meters, but generally the thickness is less than 0.5 meters. The thrusts usually form the bottoms of the small valleys and depressions within the Angered area, but many exceptions exist. Lindskoug and Nilsson (1974) and Wedel (1978) have pointed out the general importance of the different tectonic zones for the occurrence of groundwater. The groundwater exchange between bedrock and soil takes place where the thrusts are covered by frictional material. These layers form small confined aquifers with a piezometric head close to the ground surface under undisturbed conditions.

Three tunnels at the levels 38-46, 24.5-25, and 20-21 m above sea level were constructed in the Angered area, see Fig. 2. The water leakage into the tunnels on the whole is rather small (Lindskoug and Nilsson 1974). No pregrouting was made, but the tunnels were grouted where necessary during construction. The largest leakage was observed where the tunnels crossed the thrusts.

Artificial Recharge Tests

In two sub-areas A and B in Fig. 2, in Angered, tests with artificial recharge through wells were carried out in order to restore the lowered piezometric head in the soil. The wells were 2'' perforated steel-tubes driven down to the bedrock surface. The tests are described in detail by Carlsson (1978).

Each of the tests was carried out with a constant recharge rate in the early stage and with a constant recharge head in the later stage.

Drinking water from the Gothenburg water supply system was used as recharge water. The change of piezometric head in bedrock and soil was measured during the transient and steady state.

Field investigations followed by evaluation of the changes of piezometric head with time and distance made is possible to construct mathematical models of the aquifer, i.e. the frictional layer between clay and bedrock as shown in the section through sub-area A in Fig. 3. The transmissivity and storage coefficient and the position of the leakage between soil and bedrock were evaluated analytically. In average the transmissivity and storage coefficent obtained were $5 \cdot 10^{-6}$ m^2/s and $3 \cdot 10^{-4}$ respectively (Carlsson and Kozerski 1976).

The change of piezometric head was recorded during and after the injection tests which made it possible not only to locate the points of the leakage but also to estimate the rates during and after the injection. This is shown in Fig. 4 for sub-area A. The leakage rates thus determined were $5 \cdot 10^{-5}$ m^3/s and $6.7 \cdot 10^{-5}$ m^3/s in sub-areas A and B respectively. The leakage between soil and bedrock seems to occur in small, well-defined areas where the thrusts in the bedrock are covered by the frictional layer below the clay.

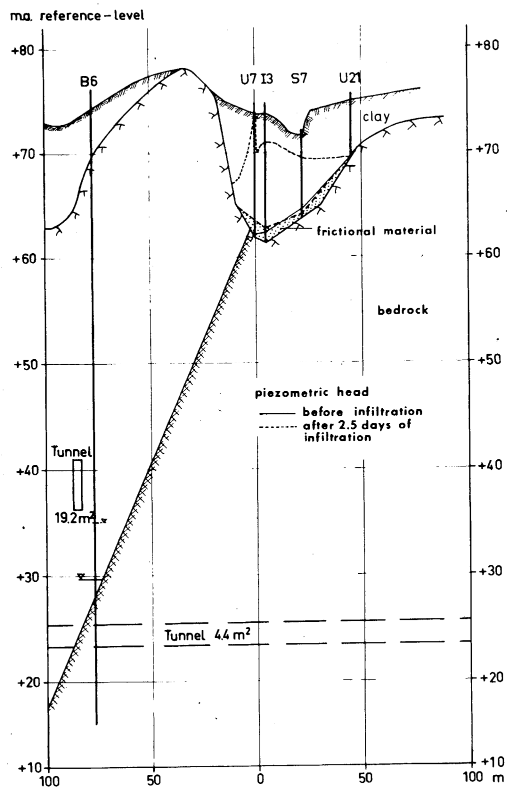
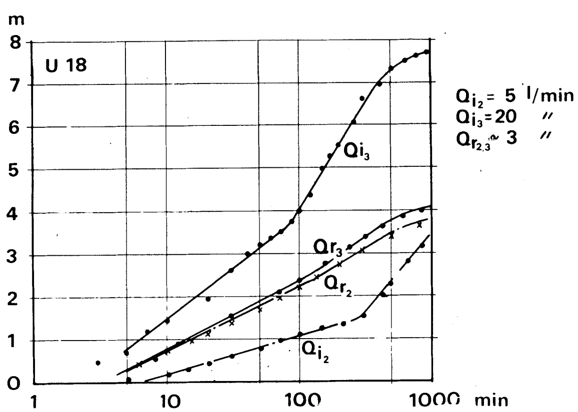


Fig. 3. Section in W-E-direction through sub-area *A* (from Carlsson and Kozerski 1976).



Q_i = injection rate
Q_r = calculated leakage rate between soil and bedrock.

Fig. 4. Change of piezometric head observed in one observation well during and after two different injection-tests in sub-area A.

The aquifers in the two sub-areas are small and limited, 6,000 m^2 and 4,000 m^2 respectively. Thus the influence of the injection tests was limited. It was estimated that about 40% of the injected amount of water was transferred into the bedrock – increasing both the piezometric head of the bedrock and the leakage to the tunnels.

Calculation of Transmissivity and Storage Coefficient

In Fig. 5 a chart of the procedure for estimation of transmissivity and storage coefficient of the tectonic zones is presented. The geometry of a tectonic zone was determined from drillings and observations made in the tunnels.

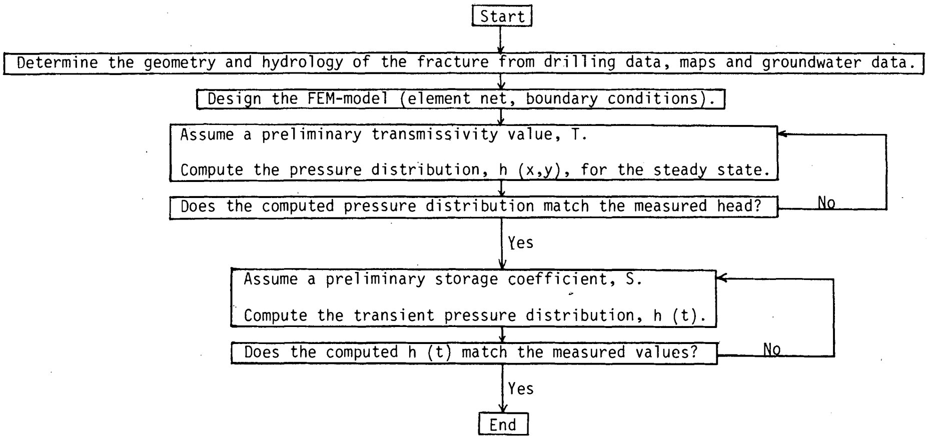


Fig. 5. Flow chart of the procedure used.

The zone is considered to act as a confined aquifer. The size of the element net and the boundary conditions are determined from the intersection of the tunnels with the zone and from the leakage point between soil and bedrock. Fig. 6 shows one element net used in sub-area A. The boundaries of the net in connection to the tunnels are chosen in such a way that they represent a flowline in steady state. In regions with a steep hydraulic gradient smaller elements are used.

Steady state was reached during the later stage of the recharge tests. The piezometric head is in this stage independent of the storage coefficient. By assuming different transmissivities we can compare the piezometric head obtained in the nodes at the bore holes with the measured head.

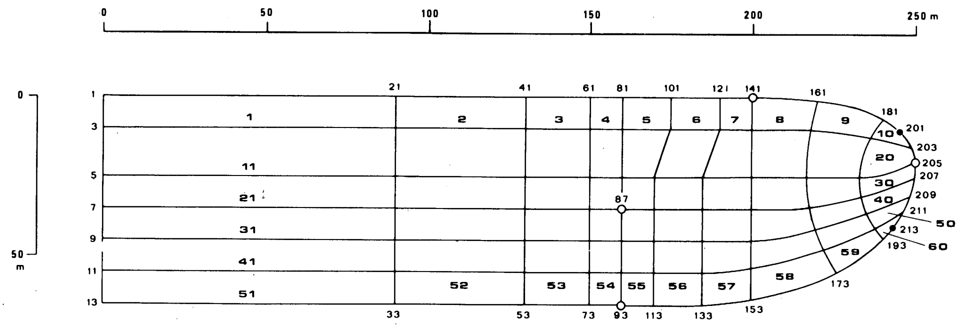


Fig. 6. Element net used for the tectonic zone in sub-area A. Nodes 93 and 141 are the intersections between tunnels and the zone, node 87 is the bore hole B6, and node 205 is the leakage point between soil and bedrock. All boundaries are impermeable.

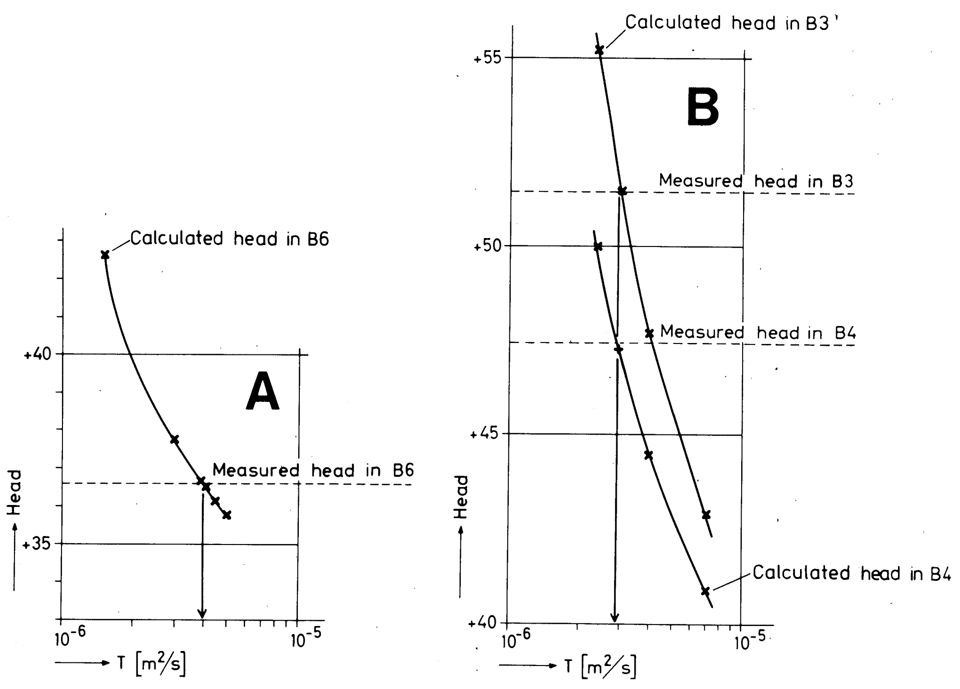


Fig. 7. Calculated piezometric head in steady state in the nodes representing bore holes for different values of the transmissivity of the tectonic zones.

In Fig. 7 the relation between transmissivity and piezometric head is shown for nodes at bore holes. The calculations are made under the assumption that no leakage to the zones occurs except in the leakage-points between soil and bedrock determined by the recharge tests. The true conditions comprise leakage from groundwater in the soil or directly from precipitation *via* other leakage-points.

The piezometric head in the tectonic zones before artificial recharge indicates a certain "natural" recharge. In sub-area A this recharge has been calculated under the assumption that it is equivalent to a recharge in the earlier defined leakage-point. Thus the recharge from soil to bedrock in sub-area A under conditions affected by the leakage to the tunnels is calculated to be about $1.5 \cdot 10^{-5}$ m^3/s or 460 m^3/year.

The calculation of the storage coefficient S was made with the estimated transmissivity and the same element net as earlier mentioned. The piezometric head under transient conditions for different S-values is compared with the measured head as illustrated in Fig. 8.

The transmissivity and storage coefficient of the tectonic zones (thrusts) calculated are 3-4 $\cdot$ 10^{-6} m^2/s and 4-6 $\cdot$ 10^{-5}, respectively. It should be pointed out that the obtained values are calculated, assuming saturated homogeneous and isotropic conditions in the zones.

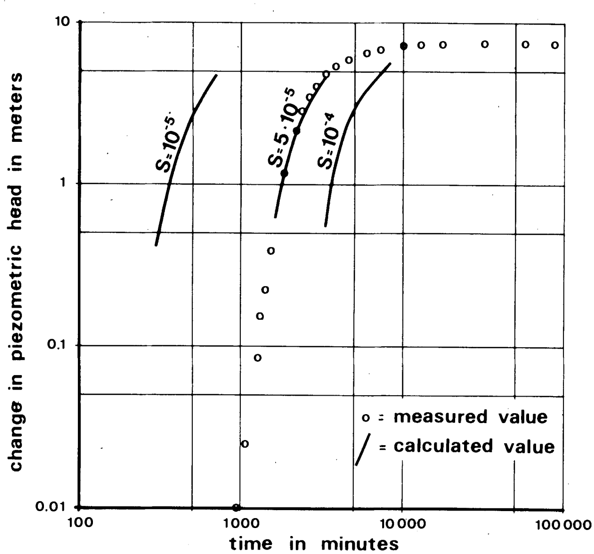


Fig. 8. Registered change in piezometric head in transient state and calculated head with different storage coefficients in the tectonic zones in sub-area A.

Discussion

According to Wedel (1978) the horizontal spacing of the thrusts is about 80-150 m in the investigated area. If we assume the rock mass between the thrusts to be completely impervious, an average hydraulic conductivity of $2\text{-}5 \cdot 10^{-8}$ m/s is obtained for the rock mass regarded as a continuum. This value is about half the value given by Wedel (1978). The difference is explained by the occurrence of the earlier mentioned vertical joints crossing the thrusts.

It should be pointed out that the obtained values of transmissivity and storage coefficient are representative only of the thrusts in Angered. The configuration of fractures varies considerably, and thus a great variation of the hydraulic properties exists. This is strikingly illustrated by the values $T = 2 \cdot 10^{-4}$ m²/s and $S = 6 \cdot 10^{-4}$ determined from pumping tests of a fracture zone in granite by Wesslén, Gustafson and Maripuu (1977).

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