ORKUSTOFNUN
Straumfræðistöð

EVALUATION OF GROUNDWATER LEVEL AND MAXIMUM YIELD OF WELLS IN A FRESH WATER LENS IN SVARTSENGI SOUTH-WEST ICELAND

Jón Ingimarsson
Jónas Elíasson
Snorri Páll Kjaran

OS SFS 7802

April 1978

ORKUSTOFNUN

Orðsending

Dags. 78.05.19

Til: Bókasafns

Frá: Jóni Ing. S.F.S.

Sendist til:

Varðar:

Hér fær bókasafn OS eitt eintak af grein sem lögð er fram á ráðstefnu Norræna vatnafræðifél sem haldin verður í ág 1978 í Helsinki. Efni greinarinnar er nánast sama og í skýrslu SFS frá í jan. '78 "Svartsengi framvinduskýrsla um ferskvatnsathuganir" SFS-7801 og ROD-7802.

Kveðja

Jón

Orkustofnun
Straumfræðistöð

EVALUATION OF GROUNDWATER LEVEL AND MAXIMUM YIELD OF WELLS IN A FRESH WATER LENS IN SVARTSENGI SOUTH-WEST ICELAND

Jón Ingimarsson
Jónas Elíasson
Snorri Páll Kjaran

OS-SFS-7802

Apríl 1978.

EVALUATION OF GROUNDWATER LEVEL AND MAXIMUM YIELD OF WELLS IN A FRESH WATER LENS IN SVARTSENGI SOUTH-WEST ICELAND

Jón Ingimarsson — National Energy Authority
Jónas Elíasson — University of Iceland
Snorri Páll Kjaran — National Energy Authority

ABSTRACT

The purpose of this study is to find the approximate maximum yield of wells in a fresh water lens in an unconfined aquifer. For that purpose the permeability coefficients of the formation are determined in different ways, using step drawdown tests, tidal effects, and groundwater models.

1. INTRODUCTION

A district heating service for the Reykjanes area is under construction. It is planned to supply all towns and villages in the area, including Keflavik International Airport and military base of the NATO Defense Forces, about 18000 people altogether. The scheme includes a central power plant in Svartsengi (see fig. 1) that is to deliver 200 kg/sec of 85 - 120 °C hot water to the distribution network. The maximum transport distance is about 25 km (fig. 2)

The heat is withdrawn from geothermal aquifers, 400 - 1500 meters below ground level. The geothermal fluid is brine, composed of about 2/3 sea water and 1/3 fresh water, its temperature is about 240 °C. It is flashed up the well casing and used in the power plant to heat up fresh water to the aforesaid temperatures. In the heat transfer process the brine is cooled to about 100 °C and discharged to the surface aquifers as waste.

The area considered is about 400 km^2 and everywhere within it all

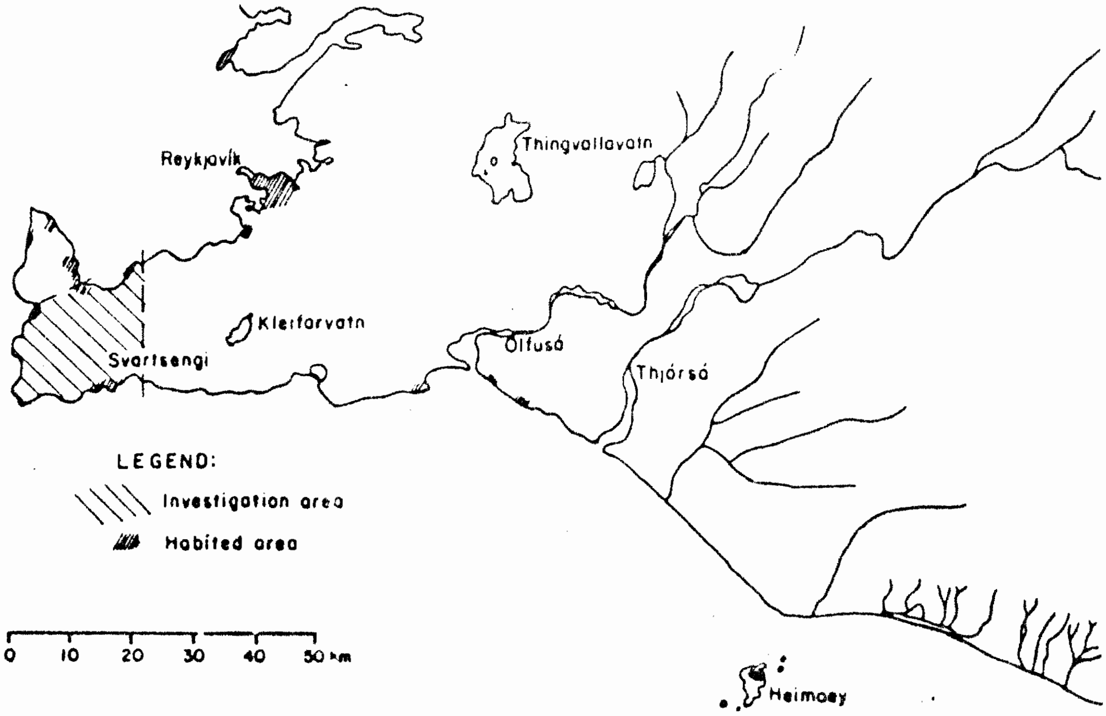
Reykjavík
Thingvallavatn
Kleifarvatn
Svartsengi
Ölfusá
Thjórsá
LEGEND:
Investigation area
Habited area
0 10 20 30 40 50 km
Heimaey

Fig. 1

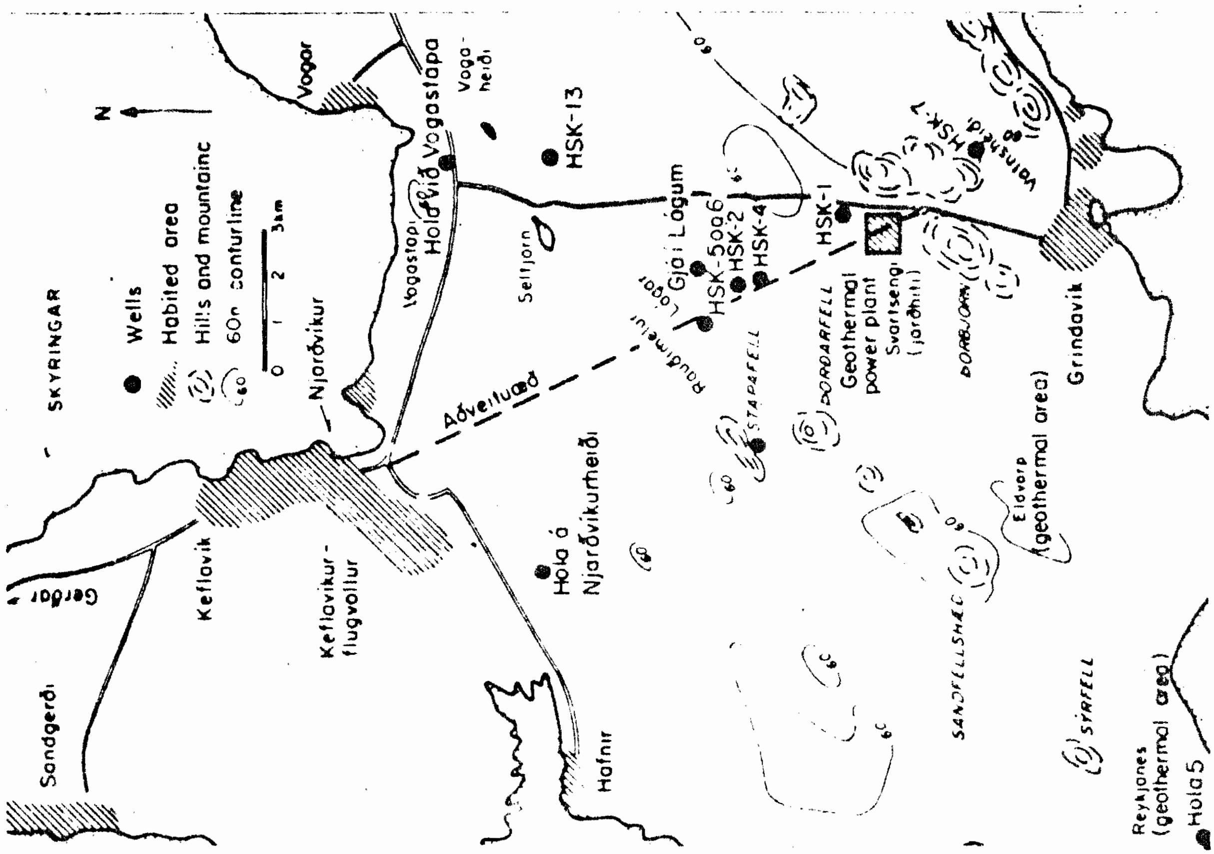
SKYRINGAR
Wells
Habited area
Hills and mountainc
60m conturline
0 1 2 3km
N
Vogar
Vogastapi
Hola við Vogastapa
Vogaheiði
HSK-13
Seltjörn
Gjá í Lágum
HSK-5og6
HSK-2
HSK-4
HSK-1
HSK-7
Njarðvíkur
Keflavík
Keflavíkurflugvöllur
Gerðar
Sandgerði
Hola á Njarðvíkurheiði
Hafnir
Aðveituæð
Rauðimelur
Lágar
STAPAFELL
ÞORBJARFELL
Geothermal power plant
Svartsengi (jarðhiti)
ÞORBJÖRN
Vatnsheiði
Grindavík
Eldvorp (geothermal area)
SANDFELLSHÆÐ
SYRFELL
Reykjanes (geothermal area)
Hola 5

Fig. 2

fresh groundwater to be found is on a surface lens that floats on sea water. The Lagar area has been selected as fresh water supply area, it is 20 km^2 and situated 3 km north-west of Svartsengi. The fresh water lens is about 50 meters thick here, and a supply of 150 - 300 l/sec is planned. The supply water is to meet the chemical standards for domestic consumption so practically no salt water intrusion into the supply water is tolerated.

2. HYDROLOGICAL CONDITIONS

In the investigation area there is a fresh water lens floating on saline water in an unconfined aquifer. The saline water´s density is 1.025 - 1.030 g/cm^3. The thickness of the lens is calculated by the Gyben - Herzberg relation as:

$$h = \frac{\rho_s}{\rho_s - \rho_v} \cdot H \qquad (1)$$

where H is the groundwaterlevel above mean sea level, ρ_s the density of the saline water, and ρ_v the density of the fresh water.

The groundwaterlevel in the Lagar area is on the average about 1.5 m above sea level which means 50 m thick lens.

The geology of the area is described by Sigurðsson et.al. 1978.

The aquifer is both inhomogeneous and anisotropic in all 3 - dimension.

Near the area there are 3 meteorological stations. From precipitation measurements the mean annual precipitation is estimated 1200 - 2000 mm. The evapotranspiration is estimated 400 mm per year. There is practically no surface runoff. The infiltration is therefore estimated 800 - 1600 mm per year. There is a great variance in the infiltration, on one of the stations "Grindavik" the annual measured precipitation is 1013 mm for the period

1931 - 1976 with a standard deviation of 331 mm. For the years 1950 - 1953 the measured mean annual precipitation is about 450 mm. We see therefore that it is of great importance to measure the movement of the interface between the fresh and the saline water, due to variations in infiltration. In the period March - November 1977 the interface was steady, although the groundwaterlevel did vary between 1.1 - 1.6 meters above mean sea level. Those investigations will continue.

3. ESTIMATION OF TRANSMISSIVITY AND PERMEABILITY

Three methods are used for the estimation of transmissivity and permeability, pumping tests, tidal effects and groundwater reservoir models.

PUMPING TESTS

Prior to the pumping tests we knew that the permeability is very high, f.ex. using a model for a circular island (Wilson 1969) we get the permeability $k = 5 \cdot 10^{-3}$ m/sec on the average for the area.

All the pumped wells are of the partially penetrating type, their lenght in water is 1/6 - 1/5 of the thickness of the lens. On the Lagar area there are two observation wells which penetrate into the saline water. They are only in 15 - 25 m distance from the supply points, one is a well and the other an open fracture. Therefore the usual approximation methods (Mucha 1972) are not valid. If the wells were in greater distance the drawdown might become too small compared to the measuring accuracy. The main purpose of the observation wells is to record the movement of the interface.

The pumping tests which have been performed are step drawdown tests. According to Jacob (1947) the drawdown (Sw) in a well in a homogeneous and isotropic aquifer can be described

$$S_w = B(t) \cdot Q + C \cdot Q^2 \qquad (2)$$

where $B(t)Q$ is the drawdown in the aquifer just outside the well and CQ^2 is the well-loss.

In equation 2 it is assumed that the groundwaterlevel when no pumping takes place is exactly known and S_w measured from there. The measuring equipment used in the pumped well have the accuracy ± 0.01 m but the total drawdown with 0.05 m^3/sec. discharge is 0.30 - 3.00 m depending on the wells, therefore we use:

$$S_w = A + B(t) \cdot Q + C \cdot Q^2 \qquad (3)$$

The best fit parabolas to equation 3, are found by using stepwise multiple regression. Typical results are:

$$S_w = 0.002(\pm 0.027) + 1.6(\pm 2.2) \cdot Q + 364.2(\pm 41.5) \cdot Q^2$$
$$\text{and } S_w = -0.036(\pm 0.096) + 12.5(\pm 8.6) \cdot Q + 439.8(\pm 145.6) \cdot Q^2$$

where S_w is in meters and Q in m^3/sec. The figures in the parenthesis are 95% confidence limits and the multiple correlations coefficients for the parabolas are ≥ 0.998. The variance in the measured drawdowns is mainly due to CQ^2. CQ^2 often describes about 90% of the total variance. The drawdowns at low discharges must therefore be measured with accuracy.

One sees that the confidence limits of the calculated B and C coefficients are rather great and it is also difficult to repeat drawdown tests with accurate results. The reason is not only the great inhomogenity of the aquifers but also the high accuracy requirements that can only be met by using special equipment. Therefore 6 automatic waterlevel gauges have been installed in observation wells within the area. One of the results brought forward so far is that the groundwater level can change as much as one centimeter in an hour resulting from rainfall only. This variation clearly does influence the drawdown measurements in the pumping tests. By measuring the drawdown with automatic pressure recorders and also using the water level gauges it is possible to eliminate the drawdown due to pumping alone.

With a constant pumpage the factor B · Q for artesian aquifer can be described (see Mucha 1972)

$$B(t) \cdot Q = \frac{Q}{4\pi k(l-d)} \left[M(u,\beta_1) - M(u,\beta_2) \right] \qquad (4)$$

where $$u = \frac{r_w^2 \cdot S}{4k \cdot b \cdot t} \qquad (5)$$

r_w is the radius of the well
S is the storage coefficient
k is the permeability coefficient
b is the thickness of the aquifer
t is the time since pumping started
$\beta_1 = \frac{l}{r_w}$, see fig. 3
$\beta_2 = \frac{d}{r_w}$

$$M(u,\beta) = \int_u^{\infty} \frac{e^{-y}}{y} \cdot \left[\mathrm{erf}\left(\frac{l-Z}{r_w}\sqrt{y}\right) \right] dy$$

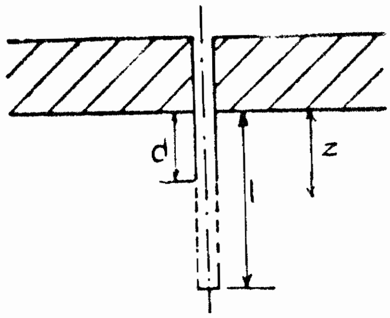

Fig. 3

In an unconfined aquifer for fully penetrating wells, the drawdown S_w is adjusted into an equivalent drawdown S_{weq} according to the relation (see Jacob 1944)

$$S_{weq} = S_w - \left(\frac{S_w^2}{2b}\right)$$

Mucha recommends: $S_{weq} = S_w - \left(\frac{S_w^2}{2l}\right)$

for relatively short pumping periods for partially penetrating wells. Those adjustments are of no interest in our investigation, ($S_w^2/2l < 0.005$ m). By inserting typical observation data into equation 4 we get an estimation of the permeability coefficient, k. Taking B = 2 sec./m^2, l = 8 m, d = 2 m, r_w = 0.1 m, b = 55 m,

$S = 0.1$ and $t = 900$ sec, we get: $k = 0.02$ m/sec.

By using Theis (1935):

$$B(t) \cdot Q = \frac{W(u)}{4\pi k \cdot b} \cdot Q \tag{6}$$

$$\text{where } W(u) = \int_u^{\infty} \frac{e^{-y}}{y} \cdot dy$$

we get: $k = 0.01$ m/sec.

Typical results for the permeability coefficient from pumping tests in the Lagar area is: $4 \cdot 10^{-3}$ m/s $\leq k \leq 4 \cdot 10^{-2}$ m/s.

TIDAL EFFECTS

Tidal effects have been registered in the area upto 3 km distance from the shore. Approximating the tidal wave as a simple sinuousodical wave the groundwater level at a distance x is (Todd 1967)

$$h_x = h_o \cdot e^{-x \cdot \sqrt{\frac{\pi S}{t_o kb}}} \cdot \sin\left[\frac{2\pi t}{t_o} - x \cdot \sqrt{\frac{\pi S}{t_o kb}}\right] \tag{7}$$

where h_o is the tidal amplitude, t_o is the period of the tide.

The radio $\frac{kb}{S}$ can be determined as:

$$\frac{k \cdot b}{S} = \left(\frac{x}{t_1}\right)^2 \cdot \frac{t_o}{4\pi} \tag{8}$$

where t_1 is the time lag, or as:

$$\frac{k \cdot b}{S} = \left(-\frac{x}{\ln\left(\frac{h_x}{h_o}\right)}\right)^2 \cdot \frac{\pi}{t_o} \tag{9}$$

The result obtained using eq. (8) sometimes differs from that of eq. (9). The difference is due to nonlinear terms which are neglected in the mathematical treatment. The full nonlinear mathematical procedures have not been developed as the leakage path in question is outside the actual Lagar area where the geology is slightly different. By estimating the storage coefficient to be equal to the effective porosity of the magnitude 10^{-2} - 10^{-1} we get the typical results for the permeability coefficient: 10^{-3} m/s $\lesssim k \lesssim 10^{-2}$ m/s.

LINEAR GROUNDWATER RESERVOIR MODELS

The differentialequation governing the flow of groundwater in the fresh water lens in an inhomogeneous and anisotropic aquifer, assuming no flow in the saline water, is given by:

$$\frac{\partial}{\partial x}\left(\frac{\gamma}{2}k_{xx}\frac{\partial H^2}{\partial x} + \frac{\gamma}{2}k_{xy}\frac{\partial H^2}{\partial y}\right) + \frac{\partial}{\partial y}\left(\frac{\gamma}{2}k_{yy}\frac{\partial H^2}{\partial y} + \frac{\gamma}{2}k_{yx}\frac{\partial H^2}{\partial x}\right) = - I \qquad (10)$$

where:

H is the groundwaterlevel above mean sea level

$\gamma = \frac{\rho_s}{\rho_s - \rho_v}$, where ρ_s and ρ_v have been defined before

k_{xx}, k_{xy}, k_{yy}, k_{yx} are the anisotropic permeability coefficients

I is the infiltration rate

By using geological information on relative permeability and anisotropy directions and appropriate boundary conditions equation (10) can be solved. The linear groundwater reservoir model described in Eliasson and Kjaran (1976 a, b) is used. By matching the calculated and measured groundwaterlevel, the absolute permeability coefficients can be determined. The permeability coefficients thus determined in the Lagar area are given by: $k \simeq 8 \cdot 10^{-2}$ m/s.

4. THE MAXIMUM YIELD OF WELLS

Many theoretical formulas have been presented to give the upcoming of the interface of a fresh water lens as a function of the well discharges. Upcoming is linear in Q upto a critical rise. The critical rise, Z_{cr}, is between 0.5 - 0.75 times m - 1, see fig. 4 (Bear - Dagan 1964 and Muskat 1946). Bear and

Dagan recommend $Z_{max} \leq 0.25 \cdot (m - 1)$ to be used.

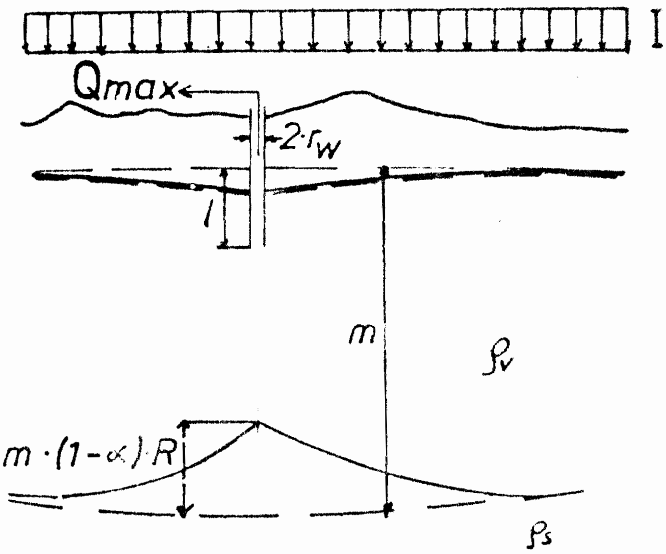

Fig. 4

In the Lagar area the drawdown should be less than 0.30 m using this criterium.

By using Dupuit´s approximation, linearizing the differential-equation and assuming no flow to be in the saline water we get for the maximum well discharge for a homogeneous and isotropic aquifer.

$$\frac{Q_{max}}{k \cdot m^2} = \frac{2\pi}{\ln\left[\left(\frac{1}{\pi}\cdot\frac{Q_{max}}{k\cdot m^2}\cdot\frac{k}{I}\right)^{1/2} \cdot \frac{1}{(r_w/m)}\right] - 1/2} \cdot \Delta(1+\Delta)\alpha(1-\alpha)$$

$$\cdot \left[1 + 7 \cdot \left(\frac{r_w}{m} \cdot \frac{1}{2}\alpha\right)^{1/2} \cdot \cos\frac{\pi}{2}\alpha\right] \cdot R \qquad (11)$$

where

Q_{max}: The maximum well discharge, m^3/s

r_w: Radius of the well, m

I: The infiltration rate, m/s

m: Thickness of the fresh water lens, m

k: The permeability coefficient, m/s

Δ: $= (\rho_s - \rho_v)/\rho_v$

α: Penetration degree = l/m

l: Depth of well, m

R: Risk coefficient

See fig. 4 for further explanation of the symbols. The last term in paranthesis in equation (11) is a correctionfactor due to partial penetration of the well according to Muskat (1946). R is a kind of risk coefficient depending upon how much upcoming one allows, see fig. 4. Using the criterium by Bear and Dagan mentioned before R = 0.25. Equation (11) is plotted in fig. 5 for one set of the dimensionless parameters k/I, R and Δ. It can be seen from the figure that we get optimum penetration around $\alpha \simeq 0.4$. Let us take a small example for one well in the Lagar area. Typical results are: $k = 10^{-2}$ m/s, I = 800 mm/year, $\Delta = 0.028$, R = 0.25, m = 55 m, $r_w = 0.1$ m, l = 11 m, we then get: $k/I = 4 \cdot 10^5$, $\frac{r_w}{m} \simeq 1.8 \cdot 10^{-3}$, = 11/55 = 0.2. From fig. 5 we then get: $Q_{max}/k \cdot m^2 = 1.24 \cdot 10^{-3}$ giving $Q_{max} = 1.24 \cdot 10^{-3} \cdot 55^2 \cdot 10^{-2} \cdot 10^3 = 37.5$ l/s.

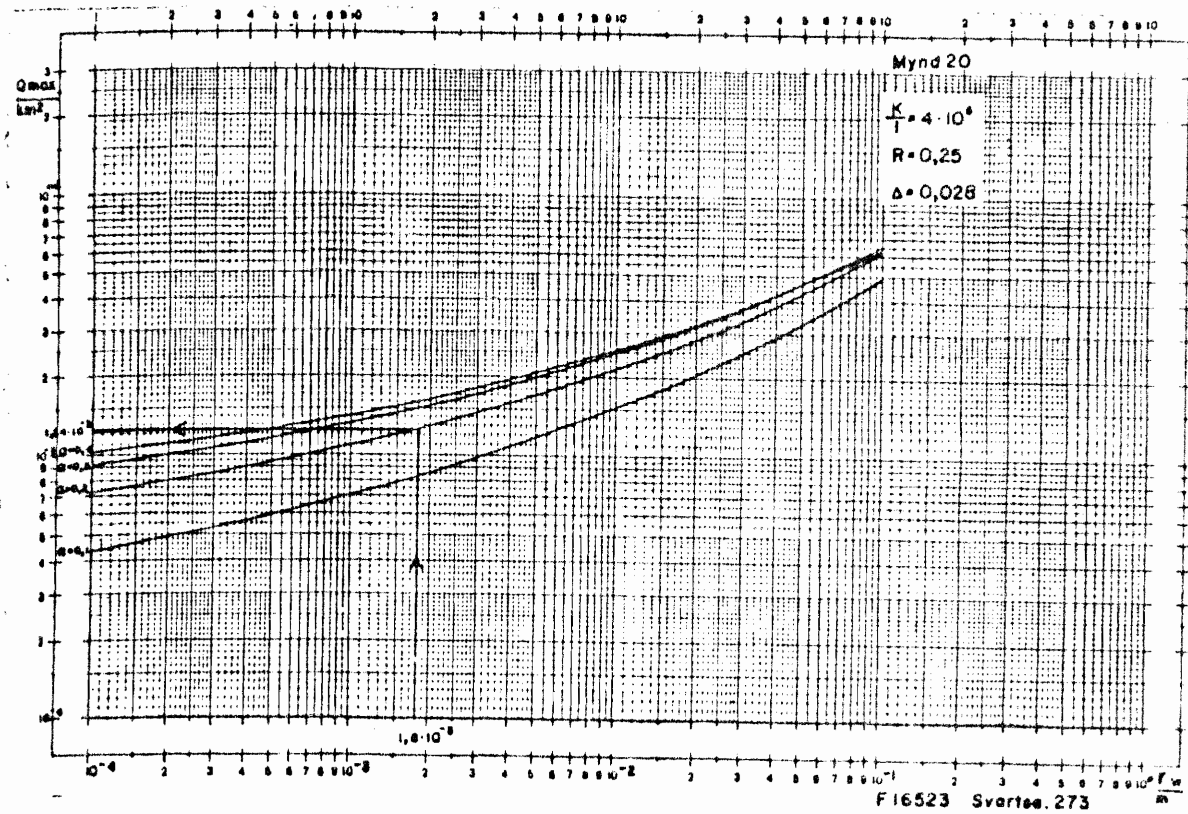

Fig. 5

Bear and Dagan (1964) solved the differential equations for the flow in the fresh and saline water by the method of small perturbation. Their solution agrees well with equation (11) for small upcoming, that is to say for $R < 0.25$.

ACKNOWLEDGEMENTS

This paper is based on a project sponsored by the Sudurnes Regional Heating to which the authors are grateful for permission to publish these results.

REFERENCES

Bear, J. and Dagan, G. 1964. The Transition Zone between Fresh and Salt Waters in Costal Aquifers. Technion- Israel Institude of Technology. Technical Report No.5 1964.

Elíasson, J. and Kjaran, S.P. 1976. Practical aspects of the Linear Groundwater Reservoir Models. Nordic Hydrological Conference 1976.

Elíasson, J. and Kjaran, S.P. 1976. Reservoir Mechanism Analysed by the Method of Galerkin and Orthogonal Eigenfunctions. Nordic Hydrology 7, 1976.

Jacob, C.E. 1944. Notes on determining permeability by pumping tests under watertable conditions. U.S. Geol.Survey 1944.

Jacob, C.E. 1947. Drawdown test to determine effective radius of artesian well. Trans. ASCE, Vol.112, 1947.

Mucha, I. 1972. Evaluation of Pumping Tests in Partially Penetrating Wells. Nordic Hydrology No.4, 1972.

Muskat, M. 1946. The Flow of Homogeneous Fluids through Porous Media. McGraw - Hill 1946.

Sigurðsson, F. et.al. 1978. Integrated Hydrogeological Survey of a Fresh - Water Len . Presented at this conference.

Theis, C.V. 1935. The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using ground water storage. Trans.Amer. Geophys. Union, Vol.?, 1935.

Todd, D.K. 1959. Groundwater Hydrology. John Wiley and Sons, Inc. New York, 1959.

Wilson, E.M. 1969. Engineering Hydrology. Macmillan and Co.Ltd. 1969.