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EVALUATION OF NATURAL RECHARGE OF CHINGSHUI GEOTHERMAL RESERVOIR IN TAIWAN

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ABSTRACT

Thermal water production in the liquid-dominated Chingshui geothermal field is largely from a fractured zone in the Jentse Member of the Miocene Lushan Formation. The age of Chingshui geothermal water was determined with tritium data. An interference test in Chingshui geothermal reservoir was analyzed to determine aquifer porosity-thickness product which was then used to calculate the fluid-in-place by volumetric method. The paper combined the use of both tracer and interference tests to evaluate the natural recharge of Chingshui geothermal reservoir. The annual natural recharge was estimated by combing the use of the mean residence-time of Chingshui geothermal water and the fluid-in-place of Chingshui geothermal reservoir. Estimates for fluid-in-place, mean residence time and natural recharge for Chingshui geothermal reservoir are 2×10^8 m^3, 15 years, and 1.3×10^7 m^3 / year, respectively.

Keywords: Geothermal reservoir, tritium, interference test, recharge.

INTRODUCTION

The Chingshui geothermal field is located in the northeast portion of Taiwan, in the metamorphic terrain (Fig. 1). The geothermal exploration program in this area began in 1973 (Lee et al., 1980). Geological, geochemical and geophysical investigations have been carried out (e.g., Su, 1978; Tseng, 1978; Cherng, 1979; Hsiao and Chiang, 1979; Lee et al., 1980), and a number of gradient (less than 500 m deep) and deeper exploratory and production wells have been drilled (Cherng, 1979). Interference tests were conducted for the initial assessment of the field in 1979 (Chang and Ramey, 1979). The tritium concentration of geothermal water was measured in 1982, 1984 and 1985 to determine the age of Chingshui geothermal water (Chen et al., 1989).

The purpose of this paper was to evaluate the natural recharge of Chingshui geothermal reservoir using both tracer and interference tests. The age of Chingshui geothermal water was determined with tritium data. An interference test in Chingshui geothermal reservoir was analyzed to determine aquifer porosity-thickness product which was then used to calculate the fluid-in-place by volumetric method. The annual natural recharge was estimated by combing the use of the mean residence-time of Chingshui geothermal water and the fluid-in-place of Chingshui geothermal reservoir.

GEOLOGY

The Chingshui geothermal area is an area of hot springs situated along Chingshui River, approximately 13 km southwest of Ilan, Taiwan, as shown in Fig. 1. Fig. 1 also shows both surface and bottom-hole locations of these wells.Geologically, this area is composed of dark-gray and black slates, namely the Miocene Lushan Formation. Lushan Formation can be lithologically divided into three members: the Jentse Member, the Chingshuihu Member, and the Kulu Member. In general, the Jentse Member is composed mainly of metasandstones intercalated by slates, while the underlying Chingshuihu and Kulu Members consist mostly of slates (Tseng, 1978; Chiang et al., 1979).

The Chingshui geothermal area is located on a monocline structure, which is cut internally by numerous thrust faults that are lightly curved, and essentially trend NE-SW parallel to the bedding; the most important ones are the Tashi, Hsiaonanao and Hanhsi faults, shown in Fig. 2 (Su, 1978; Hsiao and Chiang, 1979). In the field itself, along the Chingshui River, there is the normal, N-S striking Chingshuihsi fault. Active tectonic movements most likely created the numerous faults and well-developed fractures around the Chingshui geothermal area. The best developed fractures in the slates are found near the most convex part of the Hsiaonanao fault along the Chingshui River.

There is clear evidence that the geothermal reservoir is fracture dominated. Due to the poor porosity and permeability of the slates, faults, joints, and other extensive fractures provide the conduits for the geothermal fluid flow. Predominant joints, which are almost aligned perpendicular to the strike of the strata, are found densely developed in the sandy Jentse Member. Fig. 3 shows the rose diagram for 67 joints measured at an outcrop of that member located near the Chingshui geothermal field (Tseng, 1978). The most prominent set of joints strikes northwest and dips between 65º and 80º to the southwest. A less conspicuous set strikes northeast and dips steeply northwest.

The trend of the Chingshui River is almost parallel to that of the joints. Its bed was cut through the slates, which present well-developed fractures. In the geothermal field, there are numerous hot springs and fumaroles along the river. It is reasonable to infer that the riverbed is the area where the major open fractures reach the surface.

Subsurface data indicate that geothermal production at Chingshui is largely from a fracture zone in the steeply dipping Jentse Member (Hsiao and Chiang, 1979). Structural analyses show that this member presents predominant, well-developed, steeply dipping joints that strike between N 25º W and N 40º W. Outcrops near the area of thermal manifestations also reveal that faults run parallel for almost 100 to 150 meters striking between N 30º W and N 35º W (Tseng, 1978).

AGE DETERMINATION OF GEOTHERMAL WATER

In order to determine the age of Chingshui geothermal water, the tritium concentration of geothermal water was measured in 1982, 1984 and 1985 in Chingshui geothermal field (Chen et. al.,1989). Water samples were first distilled before mixing with liquid scintillator. A 40-ml distilled water sample was mixed with 60-ml liquid scintillator in a 100-ml Teflon vial. The vials were shaked vigorously for homogeneity, cooled at 12 □ and stored overnight to eliminate phosphorescence. The minimum detectable activity (MDA) can be calculated as follows,

$$MDA\ (pCi\ /\ L) = \frac{4.66\sqrt{B\ /\ t_c}}{2.22\,EV} \quad (1)$$

where B is background counting rate, cpm; t_c is counting time, min; E is counting efficiency, □; and V is sample volume, L. For a count time of 1,000 min, background counting rate of 2.71 ± 0.25 cpm, and a counting efficiency of 14 □, a minimum detectable activity of 20 pCi/L was achieved using the sample volume of 0.040 L. Table 1 shows the measured data. A close examination of Table 1 indicates that the tritium concentration does not correlate with the location of wells. Instead, the tritium concentration appears to decrease with the sampling date. Table 1 also shows the average concentrations of tritium measured for the Chingshui geothermal water at various sampling dates.

From IAEA worldwide survey data (IAEA, 1973, 1975, 1979, 1983, 1987, 1991), Fig. 4 shows the tritium concentration in precipitation at Hong Kong (114.2^0 E, 22.3^0 N) which is very close to Taiwan (121^0 E, 23.5^0 N). Fig. 4 also shows the least-squares regressed line for the tritium concentration in precipitation of Hong Kong. The date April 1, 1963 was set for time zero when the peak concentration of tritium was observed. The regressed equations for the tritium concentration in precipitation at Hong Kong are as follows,

$$\textit{from}\ 4-1-1963\ \textit{to}\ 1-1-1978$$

$$\log C_{HK} = -5\times10^{-12}\,T^3 + 9\times10^{-8}\,T^2 - 0.0006\,T + 2.405 \quad (2)$$

$$\textit{from}\ 1-1-1978\ \textit{on}$$

$$\log C_{HK} = -7\times10^{-5}\,T + 1.3795 \quad (3)$$

when C_{HK} is the regressed concentration of tritium in precipitation at Hong Kong, TU; and T is the time since April 1, 1963, day. A tritium unit (TU) is the equivalent of 1 tritium atom in 10^{18} atoms of hydrogen. One TU is equivalent to 3.238 $pCi\ /\ L$ of water.

Fig. 5 shows a big difference between the average concentrations of tritium measured for Chingshui geothermal water at various sampling dates and those reported in the precipitation at Hong Kong. The big difference is caused by the decay of tritium since the infiltration of precipitations, or, the mean residence time of Chingshui geothermal water. The mean residence time of Chingshui geothermal water can be determined by minimizing the difference between the corrected concentrations of tritium and those reported in the precipitations of Hong Kong. Decay corrections for the measured concentrations of tritium due to the mean residence time of Chingshui geothermal water can be made as follows,

$$C_0 = C \times EXP\,((-0.693\,\Delta t)/(12.43 \times 365)) \quad (4)$$

where C is the measured concentration of tritium, TU; C_0 is the concentration of tritium corrected for the decay with time Δt, TU; and Δt is the mean residence time of Chingshui geothermal water, day. Notice that the time used in the decay correction for tritium is also the mean residence time of Chingshui geothermal water.

Various values of the mean residence time of Chingshui geothermal water, ranging from 2,000 to 7,000 days were tested for the decay correction for tritium. The best estimate for the mean residence time was determined by minimizing the sum of squares of residuals defined as follows,

$$\sigma = \sum_{i}^{8} \left[\left(C_{HK} \right) - \left(C_0 \right) \right]_i^2 \quad (5)$$

when C_{HK} is the concentration of tritium in precipitation at Hong Kong, TU; and C_0 is the corrected concentration of tritium corrected for decay with the mean residence time of Chingshui geothermal water, TU. Fig. 6 shows the variation of the sum of squares of residuals, σ, as a function of the mean residence time of Chingshui geothermal water. According to Fig. 6, σ attains the minimum at the residence time of 5,550 days. Fig. 5 shows that with decay corrections due to the mean residence time, the average concentration of tritium measured for Chingshui geothermal water at various sampling date agree well with those reported in the precipitation at Hong Kong.

INTERFERENCE TEST

During the 1979 interference test, well 16T was produced, and pressure responses were observed in wells 4T, 5T, 9T, 12T, 13T, and 14T (Chang and Ramey, 1979). Because in all wells the drill bit had drifted following geologic structures, it was necessary to estimate the distance between the bottom-hole locations for interpretation of the interference test data. The distances between wells correspond to those between pairs of feed zones (Fig. 1).

Data on the 11-day interference test are presented in Table 2. Hot water production rate, measured using a weir, ranged from 80,000 to 83,500 kg/hr during the test. The total fluid (water + steam) production rate was calculated from the hot water production rate using energy-balance considerations for flashing water. For the 11-day interference test, the wellhead pressure, water production rate and total fluid production rate of flowing well 16T stabilized at 3.59 bars, 80,000 kg/hr, and 105,000 kg/hr, respectively. Wellhead pressures were monitored at all the observation wells, however data from wells 5T and 13T do not appear to be reliable because of equipment malfunction.

Fan et al. (2005) developed a conceptual linear flow model of the Chingshui geothermal reservoir based on geological data of the area. Fig. 7 shows a schematic drawing of the linear flow model; the geothermal reservoir is represented by a parallelepiped. Fluid flow is parallel to the main strike of the joints and the lateral boundaries of the prism. The cross-section of the parallelepiped is assumed to be a rectangle with a height h and a width b. The production well is represented by a planar source. The diffusivity equation governing fluid flow in an infinite linear reservoir for the constant-rate case is given by,

$$\frac{\partial p}{\partial t} = a \frac{\partial^2 p}{\partial x^2} \quad (6)$$

subject to the following initial and boundary conditions:

$$t = 0, \quad p = p_i \quad for \;\; x \geq 0$$

$$t > 0, \quad \frac{\partial p}{\partial x} = \frac{q\mu}{2bhk} \qquad for \;\; x = 0 \quad and$$

$$\lim_{x \to \infty} p(x,t) = p_i \quad for \;\; x \to \infty$$

where $a = k/\phi\mu c_t$, the formation diffusivity.

Miller (1960) investigated unsteady influx of water in linear reservoirs for the constant-rate case and an infinite-acting reservoir with the same equations as above. This author adapted Carslaw-Jeager (1959; page 75) solution of heat conduction to pressure drawdown in linear reservoirs as follows,

$$p(x,t) = p_i - \frac{q\mu}{2khb}\left[2\sqrt{\frac{at}{\pi}}\exp\left(-\frac{x^2}{4at}\right) - x\,erfc\left(\frac{x}{2\sqrt{at}}\right)\right] \quad (7)$$

Miller's solution, Equation (7), is similar to that developed by Jenkins and Prentice (1982) to analyze aquifer tests in fractured rocks assuming linear flow conditions. The Miller solution was previously applied to a steam reservoir for interference analysis (Ehlig-Economides et al., 1980). To apply Miller's solution to a fractured hot-water reservoir using SI units, the following dimensionless variables are defined,

$$P_{Dl} = \frac{2\ kh\ \Delta p}{q\mu} \quad (8)$$

$$x_D = x / b \quad (9)$$

$$t_{Db} = \frac{kht}{\phi h \mu c_t b^2} \quad (10)$$

where k is permeability, m^2; h is formation thickness, m; Δp is pressure change, Pa; q is volumetric well flow-rate at reservoir conditions, m^3/s; μ is viscosity, Pa-s; x is distance between the production and observation wells, m; b is width of the fractured reservoir, m; ϕ is porosity; c_t is compressibility of fluid, $(Pa)^{-1}$; and t is time, s. The well volumetric flow-rate at reservoir conditions can be calculated by multiplying the well mass flow-rate with the specific volume of geothermal fluid at reservoir conditions, or, $q = q_m v$ where q_m is well mass flow-rate, kg/s; and v is specific volume of geothermal fluid at reservoir conditions, m^3/kg.

Then, Equation (7) can be written in terms of the dimensionless variables defined above

$$\frac{P_{Dl}}{x_D} = 2\sqrt{\frac{t_{Db}}{\pi x_D^{\ 2}}} \exp\left(-\frac{x_D^{\ 2}}{4t_{Db}}\right) - erfc\left(\frac{x_D}{2\sqrt{t_{Db}}}\right) \quad (11)$$

Equation (11) can be used to calculate a log-log type-curve, P_{Dl}/x_D versus $t_{Db}/x_D^{\ 2}$ for linear flow.

If practical units are used instead (i.e., Darcies, bars, hours) of SI units, Equations (8) and (10) become,

$$P_{Dl} = \frac{0.0007106\ kh\ \Delta p}{q\mu} \quad (8a)$$

$$t_{Db} = \frac{0.0003553\ kht}{\phi h \mu c_t b^2} \quad (10a)$$

The Carslaw-Jaeger-Miller solution assumes that one half of the produced fluid comes from each side of the production plane/well; for this reason, Fig. 7 shows only one side for a production well along the direction of fluid flow and the main strike of the joints.

Fig. 8 is a match of the well 4T pressure versus time data against the Miller linear flow solution. The linear flow type-curve matching results for all pairs are given in Table 3. As can be seen from Table 3, porosity-thickness products obtained from the linear flow model was from 279 to 956 m. To apply the linear flow model, the width of the Chingshui geothermal reservoir was estimated to be around 300 m based on the bottom-hole locations of the production zones of seven wells (4T, 5T, 9T, 12T, 13T, 14T, and 16T) shown on Fig. 1.

The main importance of the porosity-thickness product obtained from interference testing is that it can provide an estimate of the fluid-in-place. Based on the isotherms map (Fig. 1), the area of the Chingshui geothermal reservoir is estimated to be around 0.4 km^2. The fluid-in-place can be calculated by a volumetric method using the following equation,

$$FIP = \phi h A \quad (12)$$

where FIP is fluid-in-place, m^3; ϕh is porosity-thickness product, m; and A is area, m^2. Therefore, the fluid-in-place for Chingshui geothermal reservoir is slightly higher than $2\times10^8\ m^3$ (i.e., 525 m x 0.4 km^2).

DISCUSSIONS

The annual natural recharge can be estimated using the mean residence-time of Chingshui geothermal water and the fluid-in-place of Chingshui reservoir. The age or the mean residence-time of Chingshui geothermal water was determined with tritium data. Estimate for the mean residence time of Chingshui geothermal water is 15 years. An interference test in Chingshui geothermal reservoir was analyzed to determine aquifer porosity-thickness product which was then used to calculate the fluid-in-place by volumetric method. Estimate for fluid-in-place for Chingshui geothermal reservoir is $2\times10^8\ m^3$. The natural recharge for Chingshui geothermal reservoir can be calculated using the following equation,

$$R = (FIP) / (\Delta t) \quad (13)$$

where R is the natural recharge for Chingshui geothermal reservoir; Δt is the mean residence time of Chingshui geothermal water, day; and FIP is the fluid-in-place for Chingshui geothermal reservoir, m^3. Therefore, the natural recharge for Chingshui geothermal reservoir is $1.3\times10^7\ m^3$ / year.

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Table 1 Tritium concentration in Chingshui geothermal water

Well number	Tritium concentration (pCi / L)*							
	5-82	10-82	1-84	5-84	8-84	11-84	3-85	5-85
4T	116	72	35	29	55	27	27	18
5T	103	62	33	49	58	27	21	39
9T	112	71	44		23	54	30	49
12T	91	78	28		16	36		
13T	116	101		38		18	30	47
14T		98	27	30		32	17	14
16T	113	56	30	18	61	55	33	33
Mean	109	77	33	33	43	36	26	33
Std. Dev.	10	17	6	12	21	14	6	16

*1 TU = 3.238 pCi / L

Table 2 Interference test in Chingshui geothermal field (Chang and Ramey, 1979)

Time	Observation Wells								Flowing Well		
	4 T		9 T		12 T		14 T		16 T		
(hr)	WHP* (bar)	Δp** (bar)	WHP (bar)	Δp (bar)	WHP (bar)	Δp (bar)	WHP (bar)	Δp (bar)	WHP (bar)	Hot Water Rate ($\times 10^3$ kg/hr)	Total Fluid Rate ($\times 10^3$ kg/hr)
0.0	11.86	0.00	9.51	0.00	12.89	0.00	9.17	0.00	17.79	0.0	0.0
18.5	11.79	0.07	9.45	0.07	12.76	0.14	9.17	0.00	4.76	24.0	30.8
42.5	11.58	0.28	9.31	0.21	11.17	0.34	8.96	0.21	4.00	83.5	108.7
66.5	11.45	0.41	9.17	0.34	12.55	0.34	8.62	0.55	3.86	83.1	108.4
90.5	11.45	0.41	8.96	0.55	12.41	0.48	8.62	0.55	3.86	83.1	108.4
114.5	11.38	0.48	8.96	0.55	12.34	0.55	8.48	0.69	3.86	82.0	107.0
138.5	11.31	0.55	8.96	0.55	12.27	0.62	8.34	0.83	3.86	82.4	107.5
162.5	11.31	0.55	8.89	0.62	12.20	0.69	8.27	0.90	3.72	82.4	107.8
186.5	11.24	0.62	8.83	0.69	12.13	0.76	8.20	0.97	3.72	81.0	106.0
210.5	11.17	0.69	8.76	0.76	12.07	0.83	8.20	1.03	3.65	80.0	104.8
234.5	11.17	0.69	8.76	0.76	12.07	0.83	8.07	1.10	3.59	80.0	105.0
258.5	11.10	0.76	8.69	0.83	12.07	0.83	7.93	1.24	3.59	80.0	105.0

* WHP: Wellhead pressure.

** Δp : Pressure change.

Table 3 Type curve matching using Miller (1960) solution*

Linear flow model	Observation wells			
	4T	9T	12T	14T
Match point				
Δp = 0.6895 bars,				
p_{Dl}/x_D =	4.8	1.6	2.9	1.4
t =100 hrs,				
$t_{Db}/x_D^{\,2}$ =	13	3	7.5	3.4
Distance, m	175	300	90	330
kh, Darcy-meter	85.5	48.9	26.6	47.1
ϕh, m	468	396	956	279

$^*q_m = 105{,}000\ kg/hr \qquad v = 1.188\times10^{-3}\ m^3/kg$

$\mu = 0.12\times10^{-3}\ Pa{-}\sec \qquad c_t = 1.45\times10^{-4}(bar)^{-1}$

$b = 300\ \ m$

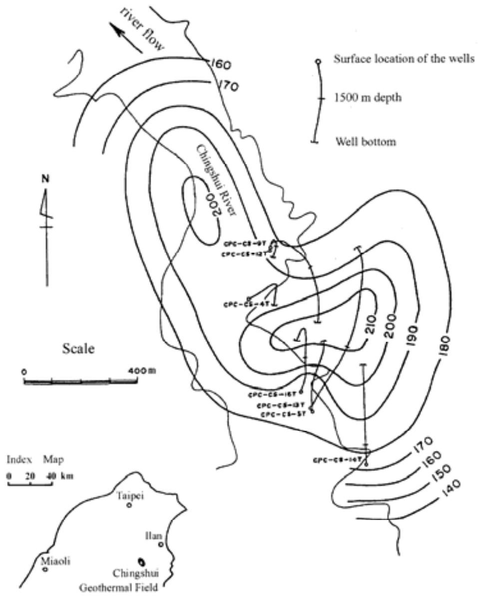

Fig. 1 Well locations and inferred temperature distribution (in ºC) at 1500 m depth in the Chingshui geothermal area (from Chang and Ramey, 1979).

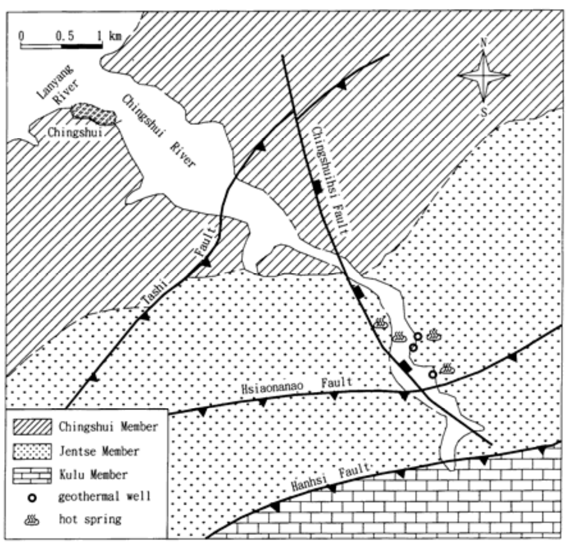

Fig. 2 Geological map of the Chingshui geothermal area showing the Chingshuihu, Jentse, and Kulu members of the Miocene Lushan Formation. Triangles and rectangles indicate the up-dipped sides of the reverse faults and the direction of dip of the normal fault, respectively.

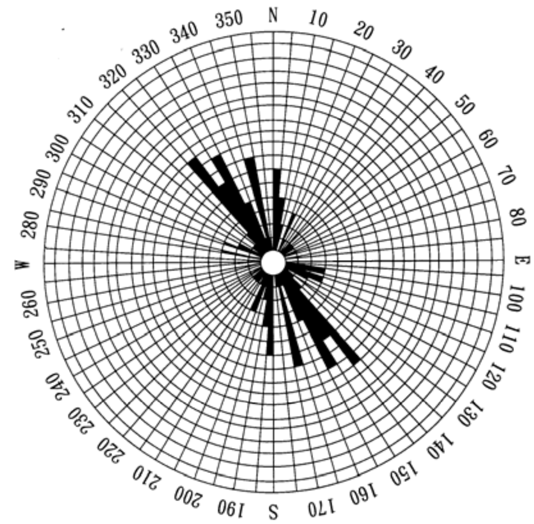

Fig. 3 Rose diagram for 67 joints in the Chingshui geothermal area (Tseng 1978).

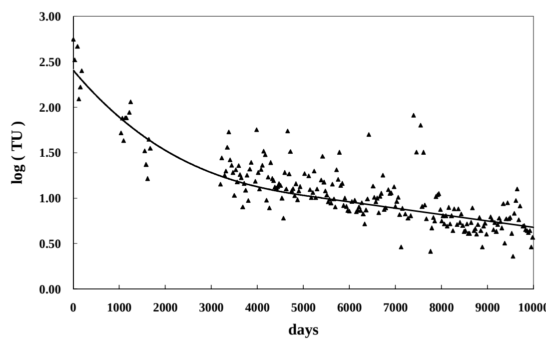

Fig. 4 Variations of tritium in precipitation at Hong Kong. (Time zero is set on April 1, 1963.)

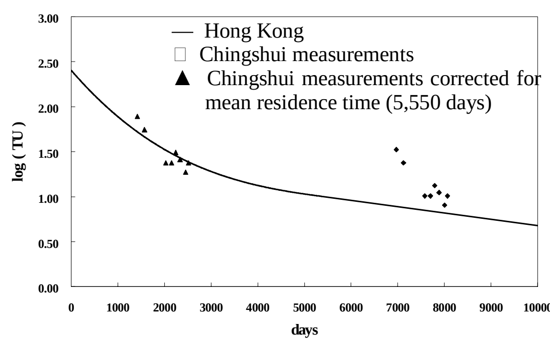

Fig. 5 Comparison of tritium content in Chingshui geothermal water with that in precipitation at Hong Kong. (Time zero is set on April 1, 1963.)

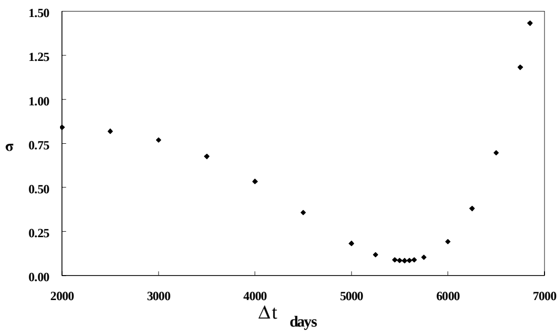

Fig. 6 Variation of sum of squares of residuals with the mean residence time of Chingshui geothermal water.

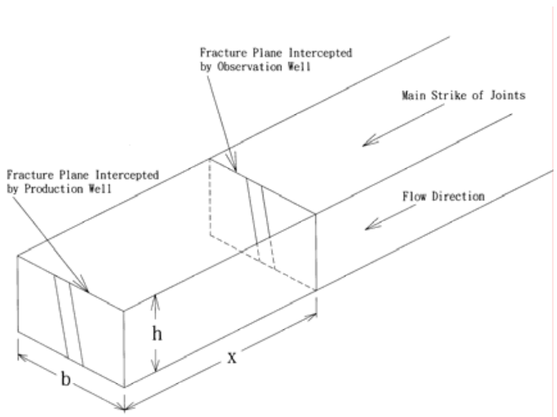

Fig. 7 Sketch of a linear flow model for a fractured reservoir.

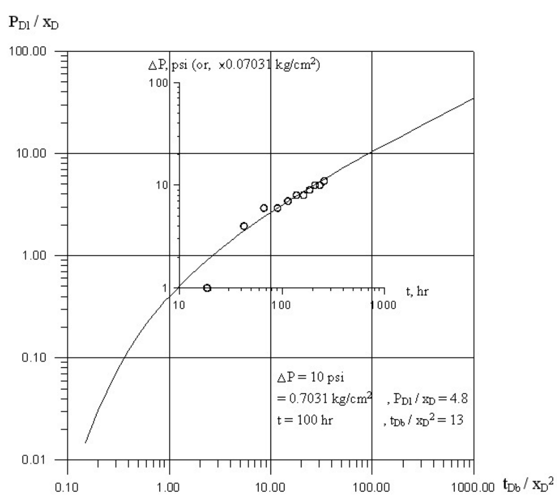

Fig. 8 Well 4T. Type-curve match using the linear flow model.