



MODELLING GEOCHEMICAL EFFECTS OF GEOTHERMAL REINJECTION IN THE TANGGU LOW-TEMPERATURE FIELD, CHINA

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ABSTRACT

As a strategy for geothermal production and a method to deal with the problem of pressure decline due to reservoir depletion, geothermal reinjection is going to be carried out in the Tanggu geothermal field, China. In the early stages of the project, it is necessary to obtain information about the reservoir response to the operation. Because of limited data at this stage, geochemical modelling is the only approach that can be used to understand the effects of reinjection on reservoir properties. Two different geochemical modelling approaches have been employed to simulate the geochemical effects of reinjection. In one, water-rock interaction is considered while in the other, account is taken of the migration of silica in the reservoir during reinjection. The results show that reinjection has a positive effect on the reservoir porosity on account of the volumetric balance of dissolution of rock and precipitation of secondary minerals. The effect is important at the beginning of reinjection in an area of about 50 m radius around the reinjection well. As a result of water-rock interaction, the secondary mineral assemblage is composed of quartz (chalcedony), calcite, muscovite, kaolinite and clinocllore. The composition of the reinjected water is the same as that of the geothermal water in the reservoir.

1. INTRODUCTION

In recent years, direct utilization of low-temperature geothermal energy is getting more and more important. Many countries, including Japan, Iceland, France, Italy, USA, and China, have benefitted from using low-temperature energy and accumulated a wealth of experience of geothermal utilization. They provide good examples of direct uses of low-temperature energy and show that geothermal energy is economic and competitive.

The main problems encountered during the exploitation of low-temperature geothermal fields are reservoir temperature and pressure decline with increasing time and production rate. To deal with these problems, it has been shown that reinjection is a good way to prevent pressure decline. Reinjection has primarily been used to dispose of waste water for environmental reasons. But now it has been shown

that not only the pressure can be maintained but also that extra energy can be extracted from the reservoir due to cool water heated by the reservoir rock. Two problems concerning reinjection are early thermal breakthroughs in the production well and injectivity decline in the injection well. Both problems affect the operation of a reinjection system. Scientists and reservoir engineers use different approaches to study such problems. Geochemical studies on reinjection have focused on silica precipitation and its effects on the reservoir properties. Chemical tracers have also been used to study reservoir properties and thermal breakthrough during reinjection. Geochemical studies have also been conducted for hot dry rock test systems and flooding of oil reservoirs. Recently, with geothermal reinjection regarded as the optimal method to extract energy, studies on geochemical processes taking place in the reservoir during reinjection are becoming more and more important.

Tanggu, a harbour and industrial area in Tianjin, China, has benefitted from the use of low-temperature geothermal energy since 1987. There is an area of 620,000 m² heated by geothermal water, and it serves 100,000 inhabitants. Now Tanggu has become a model of district heating with geothermal water and provides experience that can be drawn on later for other areas in China. However, as has happened in most geothermal fields responding to production, the geothermal water level has been dropping, at a rate of 3 m per year. For the purpose of production, the pump has been placed progressively deeper in the well, and the changes directly affect the production of geothermal water. Therefore, reinjection is being planned. But before its commencement, it is necessary to understand how the reservoir will respond to reinjection.

For the above purpose, the objective of the present study is to study the geochemical effects on reservoir properties by modelling geochemical processes during reinjection, and to aid in determining how to carry out and manage the reinjection system. The study is presented in three main parts:

1. Geothermal resources in Tanggu;
2. Geochemistry of the geothermal water;
3. Modelling geochemical effects of geothermal reinjection.

In the first part, information about the Tanggu geothermal field is summarized. In the second part, some new results from the present study obtained from geochemistry of geothermal fluids are presented and it is shown how they can be used for the modelling studies. Finally, two different approaches are used to model the geochemical effects on reservoir properties. The first uses water-rock interaction at different temperatures and reservoir pressure. The second employs modelling of silica transport in the reservoir taking the chemical kinetics of the reaction into account. Both approaches are employed to understand the geochemical effects on reservoir properties.

2. GEOTHERMAL RESOURCES IN TANGGU, TIANJIN

2.1 Geology and tectonics

Tanggu, a harbour and industrial area and an economic development area of Tianjin, is located in the east part of the North China plain, on the west coast of the Bohai Sea, about 60 km east of Tianjin. It covers an area of 859 km², with a population of 430,000.

The Tanggu geothermal field is situated in the Beitang depression, in the north part of the Huang-hua downwarp in the North China subsidence zone. The boundaries of this area are characterized by the Han-gu fault in the north, the Cang-dong fault in the west and the Da-gang fault in the south, while the Haihe fault crosses the area through the centre (Figure 1). The main faults in this area have two different

directions, NE or NNE, and NW. The two different faults intersect and result in a block structure.

The strata of Tanggu are characterized as argillis-arenaceous sediments of Cenozoic time. The basement is composed of Middle and Upper Proterozoic limestone. The Quaternary formation is composed of clay, silt and sand with a thickness of 550 m. The Cenozoic formation is about 2000-4000 m in thickness and becomes thicker from west to east, and from north to south. The geothermal gradient is about 25-35°C/km.

2.2 Geothermal reservoir

The Tanggu geothermal field has no surface manifestations. The geothermal water is stored in the sedimentary basin that is the main reservoir for oil and geothermal water in North China. The geothermal reservoir is in the Upper Tertiary Guantao group, which is composed of red argillite and sandstone. This group is distributed widely to a depth of 1600-2100 m. Its total thickness is about 300-500 m, and it can be divided into three lithological sections (Figure 2).

Guantao I aquifer (Ng I)

This aquifer is in the upper part of the Guantao group, whose thickness is 80-120 m, and is mainly dominated by a medium-fine sandstone with the porosity of 30-32%, and water temperature of 55-65°C. This reservoir has not yet been developed.

Guantao II aquiclude (Ng II)

This formation, which is mainly argillite with a thickness of 80-120 m, is a stable aquiclude between Guantao I and Guantao III.

Guantao III aquifer (Ng III)

This aquifer is composed of two layers. The upper is sandstone, about 100 m thick with porosity of 28-30%, geothermal water temperature of about 60-70°C and permeability about 2,000-2,300 mm². The lower is composed of sandstone and conglomerate. The porosity and permeability are 20% and 2,000-2,500 mm² respectively. The geothermal water is at a temperature of 65-78°C. This part is the main reservoir presently exploited. Now, 14 geothermal wells have been drilled into this aquifer with depths of 1900-2200 m. The average single well flow rate is 120-130 m³/h. They serve an area of 620,000 m² for space heating.

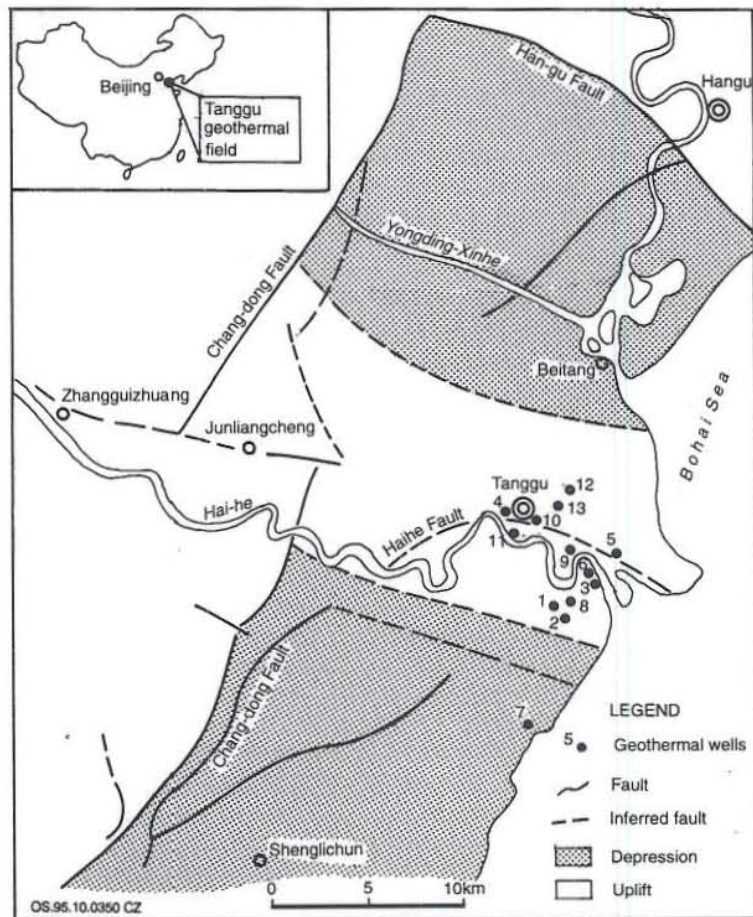


FIGURE 1: The geological structures and tectonics of the Tanggu geothermal field

FIGURE 2: The lithological sections of the Tanggu geothermal field

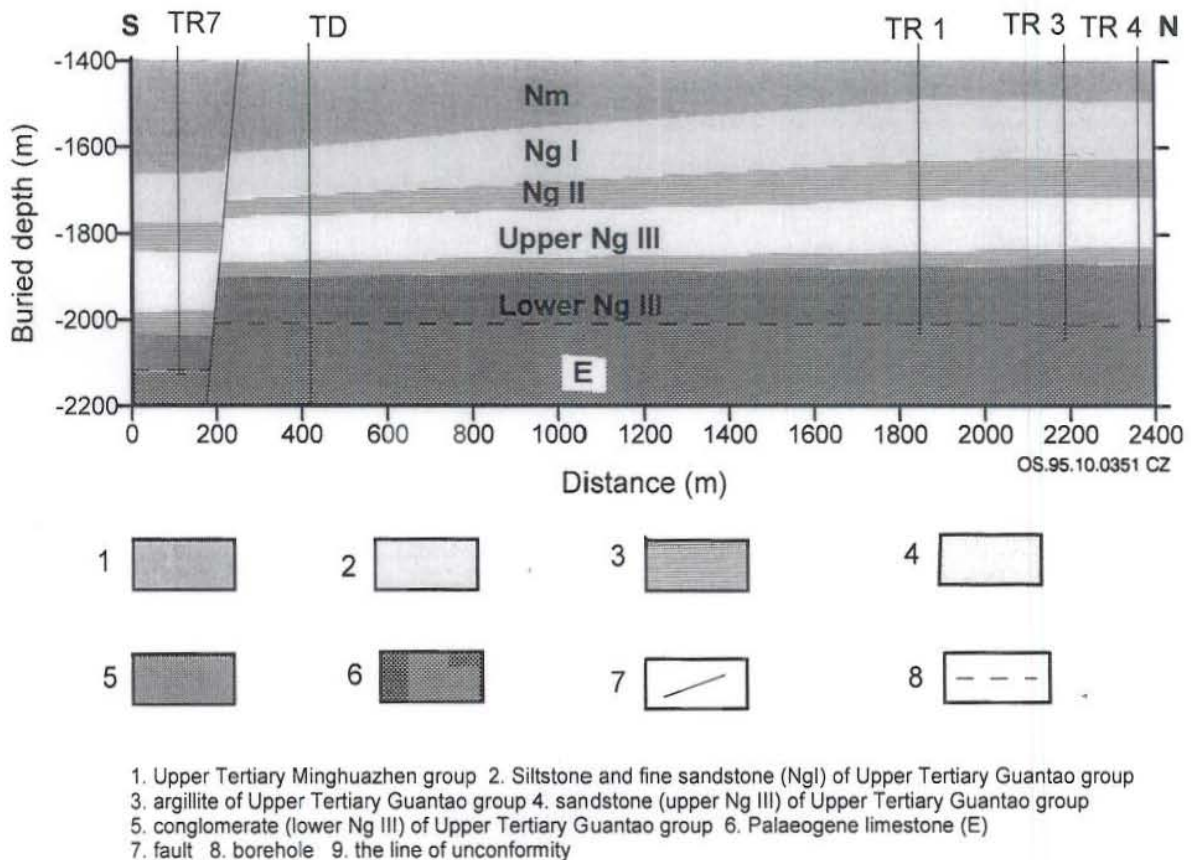


FIGURE 2: Cross-section of the classified aquifers in the Tanggu geothermal field

2.3 Production history and present status of geothermal utilization

The first geothermal well was drilled into the Guantao group reservoir in 1987 to a depth of 2050 m. Its flow rate is 130 m³/h and the water temperature is 74°C. This well was the first one to serve for space heating in an area of 63,000 m². In 1989, there were 8 geothermal wells in Tanggu, and 6 of them were used for heating an area of 318,300 m² with geothermal water production amounting to 2,100,000 m³. In 1991, 11 geothermal wells produced 3,100,000 m³ of geothermal water and served for heating an area of 620,000 m². Up to now, 14 geothermal wells have been used for space heating. Some data on geothermal wells is listed in Table 1. The total annual production is 3,636,000 m³ for heating an area of 620,000 m² (Cao, 1992).

The initial hydrostatic level was at 6 m depth when the first well was drilled in 1987. With the addition of more geothermal wells and then increased production rate, the hydrostatic level dropped progressively. The rate of decline is about 3 m per year. The hydrodynamic level has also dropped with the decline of the hydrostatic level. In most geothermal wells, the dynamic water level was at 49-51 m depth in 1989. It declined to 55-60 m depth in 1990 and in 1992, it was at 69-72 m depth (Cao, 1992).

There are two types of geothermal heating systems in Tanggu (Zheng and Cao, 1995). The most important one is direct geothermal heating with a peak boiler. Generally, geothermal water is pumped into a system passing directly through circulating pumps. Then a part of the spent water is mixed with geothermal water and circulated again in the system, while the remainder is pumped into a separate pipeline for other purposes. A steam boiler is used to heat the circulating geothermal water to improve heating efficiency when the weather is coldest. The other type is indirect geothermal heating using a heat exchanger.

TABLE 1: The parameters of geothermal wells in the Tanggu geothermal field

Well No.	Depth (m)	Location of liner (m)	Wellhead temperature (°C)	Bottom temperature (°C)	Flow rate (m ³ /h)	Date of well completion
TR1	2034	1951-2006	74	76	135	1987.08
TR2	2049	1973-2033	73	76	135	1987.11
TR3	2041	1903-1975	67	73	145	1988.05
TR4	1935	1832-1900	65	69.5	158	1988.07
TR5	2025	1927-1988	72	76	174	1988.01
TR6	2180		66	73	80	1989.04
TR7	2150	2054-2119	78	81.7	160	1990.01
TR8	2070	1950-2033	71	76	130	1989
TR9			65.5	75.9		1989
TR10	2004			73		1990.12
TR11	1912	1815-1880	67	72		1991.05
TR13	2025	1952-1973	69	74		1991.01
TR12	2070		67	74		1991.07

3. GEOCHEMISTRY OF THE GEOTHERMAL WATER

3.1 Isotope data on geothermal water

Studies of stable isotopes in geothermal water of the Tianjin area have been carried out by Zhang (1989) and Liao (1991). Zhang (1989) suggested that the geothermal water had been formed at the end of the Tertiary or the beginning of the Quaternary period, but has been replaced by atmospheric precipitation. He also found out that the recharge area is at an altitude between 700-900 m to the north of the region.

TABLE 2: The δD and $\delta^{18}O$ values of different waters in Tanggu

Water type	Geothermal water	Atmospheric precipitation	Cold groundwater
$\delta D\text{‰}$	-63.1 to -69.6	-66.0 to -83.3	-66.1 to -78.3
$\delta^{18}O\text{‰}$	-8.8 to -9.9	-9.8 to -11.6	-9.7 to -9.8

The δD and $\delta^{18}O$ values for different waters in the Tanggu geothermal field, are listed in Table 2. The δD and $\delta^{18}O$ values for the geothermal water fall close to the world meteoric line (Figure 3). As compared with cold water, the geothermal water $\delta^{18}O$ value is high and an oxygen shift is observed. Liao (1991) suggested that the high $\delta^{18}O$ value in the Tertiary reservoir is consistent with that observed for precipitation from the glacial epoch. The helium-argon age of geothermal water below Tianjin

determined by Zhang (1989), is 1-3.5 Ma and that is the relative age of geothermal water in this area. An absolute age for the geothermal water can be determined using the carbon 14 dating method. The age of geothermal water determined using carbon 14 values for samples from 3 wells in Tanggu in 1995, corrected with reference to their carbon 13 values is 23.99 to 26.04 thousand years, i.e. late Pleistocene.

The isotope studies suggest that the geothermal water originates from meteoric water which has infiltrated to depth through fractures in the area of rock, and is heated during deep circulation. $\delta^{18}\text{O}$ exchanges between geothermal water and reservoir rock have taken place. The geothermal water is younger than the reservoir rock. The geothermal water was present in the reservoir when it was formed and has subsequently been gradually replaced by meteoric water. The younger water overlies the Tertiary aquifer, supplies the Tertiary reservoir through the fault and mixes with the geothermal water.

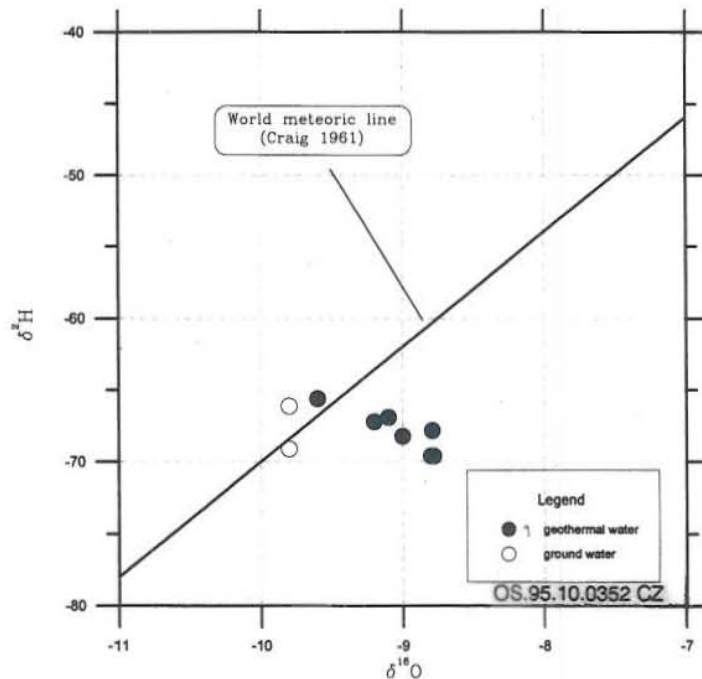


FIGURE 3: δD and $\delta^{18}\text{O}$ in Tanggu water

3.2 Some information on the reservoir obtained from the geochemistry of the geothermal water

The chemical composition of geothermal water in the Tanggu field has been determined at the Tianjin Geothermal Exploration and Development Institute. The results are listed in Table 3. The purpose of studying geochemistry of thermal water is, besides getting information on the properties of the fluid, to locate recharge areas, to estimate subsurface temperatures (geothermometers) and to obtain information about reservoir processes (boiling, cooling and mixing). Geothermometers have been widely used for the estimation of reservoir temperature. Two basic assumptions for the geothermometers are that a temperature-dependent equilibrium is attained in the reservoir and that the specific solute concentrations are not affected by chemical reactions during the upflow. The use of different geothermometers depends on data quality, temperature range (different geothermometers have different favorable temperature ranges), and rock type (equilibrium type different between different rock types).

Giggenbach (1988) has presented a $\text{Cl-SO}_4\text{-HCO}_3$ diagram to specify types of geothermal waters and to weed out waters that are not suitable for cation solute geothermometers. He suggested that most cation solute geothermometers "work" only if used with close to neutral water containing chloride as the major anion ("mature water"). The $\text{Cl-SO}_4\text{-HCO}_3$ triangular diagram (Figure 4) for 10 samples shows that the geothermal water is not a "mature water" that is suitable to use for cation solute geothermometers. For such cases, Giggenbach (1988,1991) recommended applying "geoindicators" with cations. He has developed a graphic technique to evaluate the water-rock equilibrium and to estimate the temperature using Na, K, Mg and Ca concentrations of discharge water. He presented temperature functions for cation-geothermometers based on Na/K and $\text{K}/\sqrt{\text{Mg}}$ ratios and a technique for estimating the partial pressure of CO_2 using the $\text{K}/\sqrt{\text{Ca}}$ ratio. For Na/K geothermometers, numerous empirical lines of correlation between $a_{\text{Na}}/a_{\text{K}}$ ratio and temperature have been proposed. The one provided by

TABLE 3: Chemical composition of thermal water (ppm) from the Guantao layers' reservoir (from The Tianjin Geothermal Exploration and Development Institute)

No.	Date	Na ⁺	K ⁺	Mg ²⁺	Ca ²⁺	SiO ₂	Fe ²⁺	Al ³⁺	Cl ⁻	SO ₄ ²⁻	HCO ₃	CO ₃ ²⁻	F ⁻	HBO ₂	S ²⁻	CO ₂ free	pH	TDS
TR1	28-12-1991	570	7.4	0.6	11.0	52.5	0.12	0.1	402.4	257.0	619.4	0.0	7.10	27.5	0.06	4.4	8.20	1920
	08-12-1993	590	6.5	0.0	10.0	52.0	(0.08)		395.3	249.8	601.0	3.0	6.50	31.1	0.02	0.0	8.36	1908
	17-12-1994	575	6.3	0.6	10.0	44.0	(0.08)		405.9	242.6	601.0	0.0	7.7	26.8	0.05	2.2	8.24	1885
	07-04-1995	580	7.3	1.2	11.0	45.5	0.56	(0.2)	391.7	230.5	582.7	6.0	6.7	29.1	0.1	0.0	8.38	1856
TR11	17-12-1994	575	5.9	0.6	13.0	40.5	(0.08)		390.0	247.4	567.5	0.0	8.00	30.8	0.07	8.8	8.02	1840
	11-03-1995	545	7.5	1.2	13.0	36.0	0.14	(0.2)	386.4	237.7	567.5	0.0	7.22	29.4		11.0	8.00	1794
TR13	17-12-1994	550	5.3	0.6	10.0	42.0	(0.08)		354.5	237.7	524.8	12.0	8.30	28.9	0.05	0.0	8.42	1737
	07-04-1995	510	6.5	1.2	10.0	43.0	(0.08)	(0.2)	342.1	230.5	524.8	6.0	7.50	28.8	0.07	0.0	8.42	1674
TR2	1989	562		1.8	9	44			386.4	235.3	501.3		8.4				8.7	1793
TR3	1989	544		0.6	11	44			365.1	230.5	552.2		8.4			2.2	8.2	1754
TR4	1989	528		1.8	9	39			382	245	533.9		6.8				8.4	1746
TR5	1989	504		1.2	10	46			317.3	220.9	549.2		7.2			8.8	8.1	1653
TR7	1990	624		0.6	10	51			422.2	243	579.7		7.4				8.3	1942
TR8	1989	570		1.2	10	45			390	201.7	506.2		2.5				8.8	1798
TR9	1989	520		0	10	49			360	235	522		3.1				8.5	1717

Giggenbach (1988) is very similar to the theoretical line calculated using thermodynamic data. Fournier (1991) listed seven Na-K geothermometers and suggested using these only for temperatures greater than 150°C. Thus, different Na-K geothermometers gave different temperatures and may be unreliable at low temperatures. The reason for this is probably different rock types that have been used when obtaining either empirical or experimental equations. Giggenbach's Na/K geothermometer tends to give the highest result of all at this low temperature, probably because he has considered relatively old

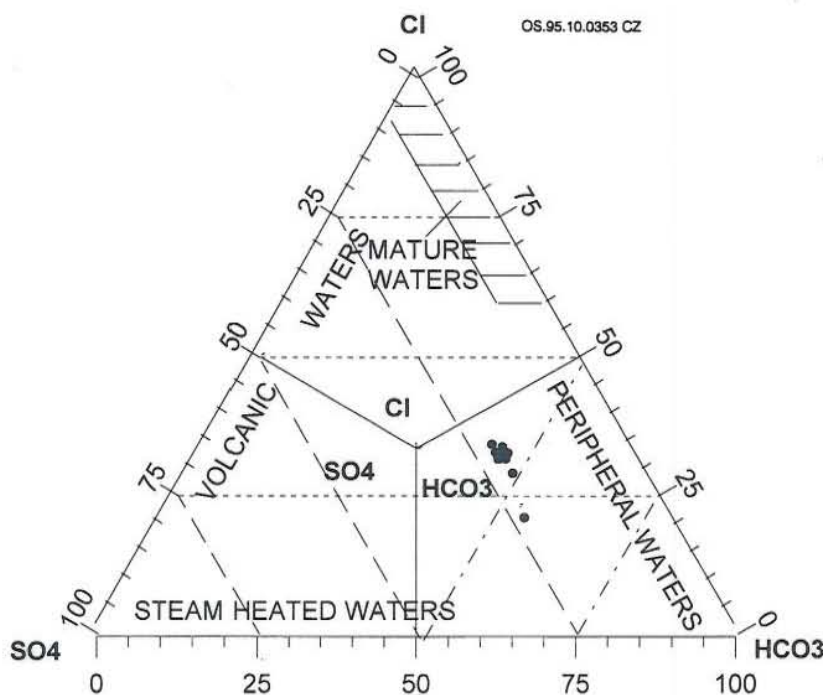


FIGURE 4: The Cl-SO₄-HCO₃ diagram for geothermal water in Tanggu

rocks that have taken a long time to reach equilibrium. The age of the geothermal water in the Tanggu field is about 24-26,000 years and the K-feldspar exchange reaction is little affected by changes in physical environment, therefore it should attain equilibrium in the reservoir even though the water does not qualify as a "mature water" on the Cl-SO₄-HCO₃ diagram. Thus, this Na/K geothermometer is

expected to be reliable for the estimation of reservoir temperature. The K/Mg geothermometer (Giggenbach, 1988) based on the equilibrium between water and the mineral assemblage K-feldspar, K-mica and chlorite is found to respond fast to changes in the physical environment and usually gives a relatively low temperature. This geothermometer is also reasonable for the estimation of reservoir temperature.

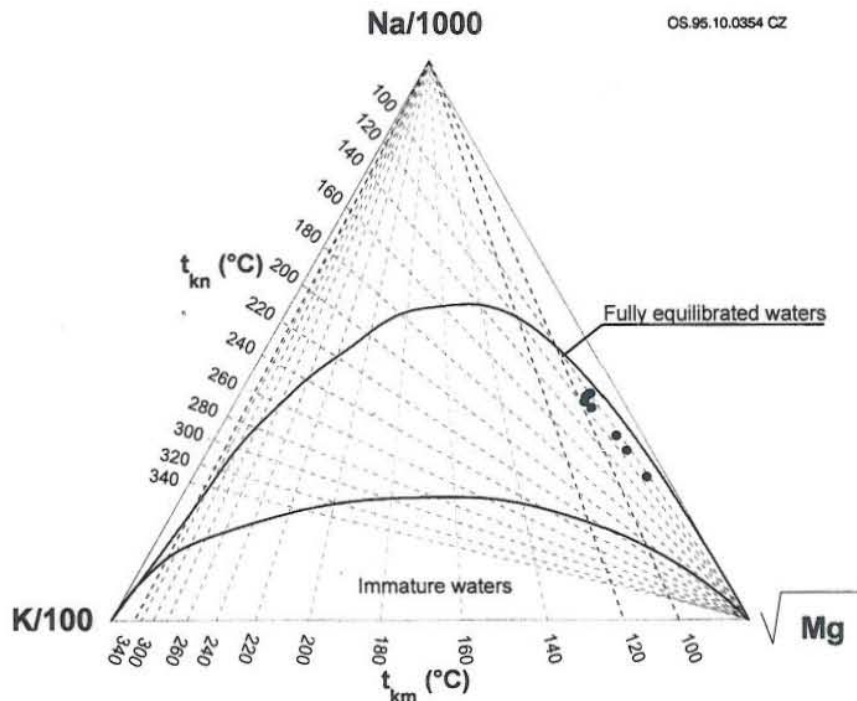


FIGURE 5: The Na-K-Mg diagram for geothermal water in Tanggu

The Na-K-Mg plot (Giggenbach, 1988) is employed here to determine whether the fluid is in full equilibrium with certain hydrothermal minerals. This diagram was constructed by adopting the Na/K and K/Mg geothermometer equations given by Giggenbach (1988). All the data from Tanggu were plotted on the diagram (Figure 5). It shows that the shallow geothermal waters plot close to the equilibrium curve. Thus the K/Mg and the Na/K geothermometers can be considered as reliable cation geothermometers.

Considering the high calcium concentration, the fluid may be approaching equilibrium described in terms of a reaction consisting of the deposition of calcite and the conversion of feldspar to layer silicates. Giggenbach (1986) has suggested the following relationship linking the K^+ and Ca^{2+} concentrations of the water to the partial pressure of CO_2 in the reservoir:

$$\log P_{CO_2} = \log (K^2 / Ca) - 3.0 \quad (1)$$

Fournier (1991) reported various silica geothermometer equations based on equilibria for different crystalline forms of silica and discussed the properties of dissolved silica in thermal water including equilibrium, mixing models, salinity and pH effects. It is thought that chalcedony often controls dissolved silica concentration at temperatures below about 180°C because its small crystals dissolve quickly. But as Fournier (1991) pointed out, in some places where water has been in contact with rock at a given temperature for a long time, such as in deep sedimentary basins, quartz may control the dissolved silica at temperature down to 100°C. Since the geothermal water in Tanggu is contained in the Upper Tertiary sandstone reservoir and its age is late Pleistocene according to ^{14}C dating, the quartz may control the dissolved silica concentration. From the results computed by the programme WATCH, chalcedony seems to equilibrate with the geothermal water at 40-90°C, therefore, both the quartz and chalcedony concentrations were used as silica geothermometers and calculated accordingly.

The reliable geothermometers applied to the Tanggu geothermal field are as follows:

K/Mg (Giggenbach, 1988):

$$t(^{\circ}\text{C}) = 4410 / (13.95 - \log(K/\sqrt{Mg})) - 273.15 \quad (2)$$

Na/K (Giggenbach, 1988):

$$t(^{\circ}\text{C}) = 1390 / (1.75 - \log(K/Na)) - 273.15 \quad (3)$$

Quartz (Fournier and Potter, 1982):

$$t(^{\circ}\text{C}) = -42.198 + 0.28831 SiO_2 - 3.6686 \times 10^{-4} (SiO_2)^2 + 3.1665 \times 10^{-7} (SiO_2)^3 + 77.034 \log SiO_2 \quad (4)$$

Chalcedony (Arnórsson et al., 1983):

$$t(^{\circ}\text{C}) = 1112 / (4.91 - \log(SiO_2)) - 273.15 \quad (5)$$

TABLE 4 : The reservoir temperature calculated by geothermometers

No.	Sampling date	T _{meas} ($^{\circ}\text{C}$)	Na ⁺ (ppm)	K ⁺ (ppm)	Mg ²⁺ (ppm)	Ca ²⁺ (ppm)	SiO ₂ (ppm)	T _{Na/K} ($^{\circ}\text{C}$)	T _{K/Mg} ($^{\circ}\text{C}$)	T _{quartz} ($^{\circ}\text{C}$)	T _{mean1} ($^{\circ}\text{C}$)	T _{chal.} ($^{\circ}\text{C}$)	T _{mean2} ($^{\circ}\text{C}$)	PCO ₂ K-Ca (bar)
TR1	28-12-1991	73	570.0	7.4	0.6	11.0	52.5	109.22	94.67	97.6	100.5	75.46	94.24	0.004978
	08-12-1993	72	590.0	6.5	0.0	10.0	52.0	101.87		97.17	99.52	75.00	91.35	0.004225
	17-12-1994	71	575.0	6.3	0.6	10.0	44.0	101.63	90.43	89.83	93.96	67.27	87.30	0.003969
	07-04-1995	70	580.0	7.3	1.2	11.0	45.5	107.81	85.31	91.28	94.80	68.79	89.30	0.004845
TR11	17-12-1994	66.5	575.0	5.9	0.6	13.0	40.5	98.78	88.72	86.30	91.27	63.56	84.24	0.002678
	11-03-1995	70	545.0	7.5	1.2	13.0	36.0	111.89	86.0	81.38	93.09	58.42	84.42	0.004327
TR13	17-12-1994	66	550.0	5.3	0.6	10.0	42.0	96.08	85.98	87.84	89.97	65.18	83.77	0.002809
	07-04-1995	69	510.0	6.5	1.2	10.0	43.0	108.38	82.40	88.84	93.21	66.23	86.46	0.004225
TR2	1989	73					43.6			89.44		66.86	78.15	
TR3	1989	67					44.4			89.83		67.27	78.55	
TR4	1989	65					38.8			84.49		61.27	73.08	
TR5	1989	72					46.1			91.85		69.39	80.62	
TR7	1990	74					50.6			95.95		73.72	84.84	
TR8	1989	71					45.2			90.99		68.49	79.74	
TR9	1989						48.5			94.07		71.74	82.91	

Note: T_{Mean1} is the average temperature of the Na/K, K/Mg and Quartz geothermometers, while T_{Mean2} is the average of the Na/K, K/Mg, Quartz and Chalcedony geothermometers

The calculated geothermometer temperatures are listed in Table 4. Because of lack of analytical results for K⁺, the Na/K and K/Mg geothermometer could only be used for 3 wells, while quartz and chalcedony geothermometers were used for all wells. The Na/K geothermometer gives a higher temperature than the K/Mg and quartz geothermometers. This geothermometer has the advantage of not being based on very fast reactions and therefore it will show better a relatively high temperature that has been present

in the system recently or is still present at some distance from the sampling point. It can be regarded as the hotter water temperature in the feed zone of wells. The K/Mg geothermometer gives a lower temperature. Because it is based on a very fast reaction, it only tends to inform on the very latest situation in the system and often tells us more about the temperature of sampling than the temperature in the system itself. These results suggesting two different temperatures are supported by the results of calculations using the WATCH speciation programme which are discussed below. The quartz temperature is intermediate between the Na/K and K/Mg temperatures. The average temperature of the three geothermometers is 80-100°C for the different wells. The calculated results for well TR1 show a slight cooling of the reservoir after high production rate in the past few years. This is probably due to cool water recharging the reservoir in response to an increased production rate from 1988 to 1992. But there is no isotopic evidence to show possible mixing of different waters because the isotopic data only extend over two years. Meanwhile, the CO₂ partial pressure in the reservoir has also decreased with time. This is expected to correspond to the reservoir pressure decline due to increased production rate. In order to learn how the reservoir responds to production, it is necessary to carry out further monitoring studies on the geochemistry of the geothermal water.

In contrast to using individual geothermometers, Reed and Spycher (1984) have presented a different approach to geothermometry. Their method is not based on the assumption of predetermined mineral equilibria or the use of empirically calibrated geothermometers. It involves evaluating the saturation state of water of a specific composition with a large number of minerals as a function of temperature. If a group of minerals is close to equilibrium at a particular temperature, the water has equilibrated with this group of minerals and the temperature represents the reservoir temperature. The procedure for using this method requires the calculation of the saturation index ($\log Q/K$) at different temperatures and the construction of a diagram of $\log Q/K$ versus temperatures. When an aqueous solution has reached equilibrium with respect to a mineral, the ion activity product, Q , calculated from analytical results from the water sample should be the same as the solubility product, K , i.e. $Q/K=1$. Because Q values may vary by orders of magnitude, it is more convenient to take the logarithm of the ratio, commonly called the saturation index ($\log Q/K$), in which case an index value of zero denotes the equilibrium condition.

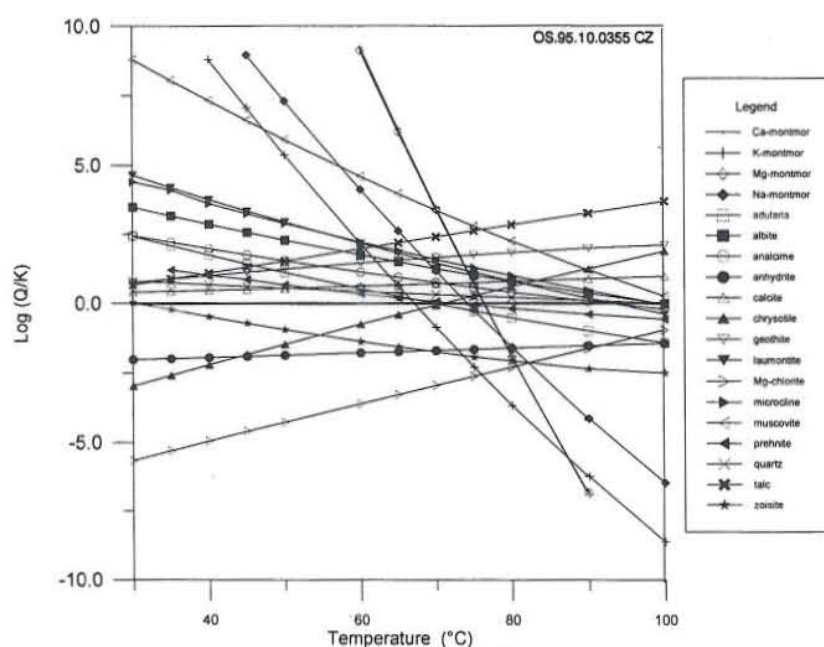


FIGURE 6: The $\log Q/K$ values for minerals in Tanggu geothermal water

If a water is supersaturated, then the index is positive and the mineral has a tendency to precipitate. If the water is undersaturated, then the index is negative, and the mineral has a tendency to dissolve. Now, if it is possible to calculate activities of free ions from a given water analysis, then it is a relatively simple matter to calculate the index quantitatively. The geochemical speciation program WATCH which was developed in Iceland, makes it possible to compute the aqueous speciation, gas pressure, activity and solubility products at any temperature from 0-300°C. It can also be used to

compute the species concentrations, activity coefficients, activity and solubility products when the equilibrated fluids are allowed to cool conductively or boil adiabatically from the reference temperature to a lower temperature. Conductive cooling of well water (from TR1) at 10°C intervals from the reservoir temperature to 35°C was calculated using WATCH. The indices ($\log Q/K$) for 29 minerals were calculated. The water may be undersaturated with respect to certain minerals because of their limited availability and kinetic effects may cause it to be supersaturated with respect to others, so all minerals can not attain equilibrium in this temperature range. The non-equilibrium minerals were weeded out from the results. The resulting saturation indices, $\log Q/K$, for chosen minerals at each temperature were plotted in Figure 6. The results for TR1, suggest that it is in equilibrium with the greatest number of minerals at 75 and 100°C, respectively, i.e. that there are two equilibrium convergence points for minerals. These two points suggest two sets of equilibrium temperatures for the water. One is about 75°C and represents the most recent equilibrium close to the K/Mg temperature. The other one is about 100°C and represents the temperature of a hotter water in the reservoir itself. This temperature is very close to that of the Na/K temperature which is based on a relatively slow reaction and is little affected by physical changes in the reservoir.

4. MODELLING THE GEOCHEMICAL EFFECTS OF GEOTHERMAL REINJECTION

4.1 Ways of modelling

The geochemical effects of reinjection were modelled in two different ways, firstly by modelling injected water-rock interactions involving multiple simultaneous reactions, and secondly by modelling silica transport in the reservoir with respect to dissolution and precipitation. Both modelling processes were used in an attempt to understand the geochemical effects on the reservoir properties due to reinjection.

4.2 Modelling water-rock interaction

4.2.1 Modelling approach and processes

The geochemical processes taking place during geothermal reinjection are very complicated. It is impossible to understand or predict them easily without geochemical modelling. The geochemical programmes SOLVEQ and CHILLER were employed to model the geochemical process of reinjection. The CHILLER is a reaction path programme for simulating water-rock interaction at a given temperature and pressure, calculating aqueous phase cooling and heating, computing mixing of fluids at different temperatures and modelling boiling and condensation processes. The principles and detailed information about these two programmes were given by Reed (1982), Reed and Spycher (1990) and Spycher and Reed (1990, 1992).

Two modelling approaches were used. One is modelling reinjected water-rock interaction at different temperatures, in an attempt to understand processes affecting the vicinity of the injection well due to reservoir cooling while the cooled water continues to be injected. Another is modelling the successive interaction between injected water and reservoir rock with increasing temperature, which is used to study processes that take place as the injected water flows to the production well and is heated by the reservoir. On modelling the successive interactions, two aspects of mass transfer were taken into account, i.e. mass transfer due to the dissolution of undersaturated minerals and precipitation of secondary minerals at a given temperature, and mass transfer due to mineral dissolution and precipitation with increasing temperature. These two processes were simulated by modelling water-rock interaction at a fixed temperature and upon heating, respectively.

The modelling of successive water-rock interactions is considered as the combination of heating and water-rock interaction. The water-rock interaction is considered at a fixed temperature. The new composition is defined at the end of this simulation. The resulting solution is then heated to a new temperature and a new composition due to the heating effect is computed at this temperature. The water-rock interaction is again modelled at this temperature using the last new composition as initial composition. Both modelling processes are repeated again and again from 35 to 95°C at 10°C intervals. The increment of temperature is 1°C during the modelling of the heating process. At each step of modelling, the calculation considers all information from the former step. The approach of combining the two processes instead of coupling hydrogeology and hydrochemistry is, of course, an improved way to model the geochemical process when there is a lack of hydrological parameters of reinjection. A very similar method has been used by Ben Baccar and Fritz (1993) to study sandstone diagenesis and evolution of porosity.

It is expected that reinjection will be conducted using spent geothermal water. As we do not have the chemical composition of spent water, we consider the chemical composition of geothermal water to have been changed by cooling. Such changes were calculated by CHILLER and the result was taken as the initial composition of reinjected water. But in a real system, because of kinetics, silica dissolution is faster upon heating than its deposition upon cooling. The silica concentration of the spent geothermal water is not likely to have reached equilibrium by being cooled down to 35°C, so that the silica concentration in spent geothermal water will be higher than that used here. This suggests that the amount of dissolved silica from the rock in the modelling is likely to be larger than that in the real case during reinjection, and its contribution to the effects of reservoir porosity may also be greater than in the real case. On the other hand, heating groundwater to 35°C and reinjecting it would in reality have given a similar result to the modelled one. In the case of a lack of data on the chemical composition of spent water, the initial composition of spent geothermal water used here is only an approximation. The chemical composition of reservoir rock used in the modelling is the average composition of sedimentary rock in that area given by Li (1989) and presented in Table 5. The chemical characteristics of this rock are high CaO content and low Na₂O content. In the modelling, Fe was not considered as a reaction component because Fe is sensitive to redox processes and we have no data about such processes. Another reason is that redox reactions reach equilibrium very slowly even at high temperatures and the iron minerals in geothermal systems are not well documented (Michard, 1991). Thus, in considering the assemblage of secondary minerals Fe-minerals are left out. We considered a rock consisting of SiO₂, Al₂O₃, CaO, MgO, Na₂O, K₂O, H₂O+ and CO₂. The total percentage composition is 94.87%. The temperature of reinjected water is considered to be 35°C. The reservoir temperature is regarded as 95°C, that is close to the temperature given by geothermometers. The pressure of the reservoir calculated from the hydrostatic pressure is about 150 bar. That is thought to be slightly lower than the actual reservoir pressure. At each step of modelling, 1 kg of solution was successively reacted with 1 g rock, and the sum of the dissolved rock is 7 g in 1 kg solution at the end of the whole modelling process. It means that the same mass and volumetric balance resulted in each step. Also, volumetric changes were emphasized to be due to chemical reactions and the effect of compaction was not considered. Meanwhile, the evolution of the fluid chemistry and the resulting minerals was modelled.

TABLE 5: The average chemical composition of the sedimentary rock in % (Li, 1989)

SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	H ₂ O+	P ₂ O ₅	CO ₂
57.61	0.44	9.61	2.37	2.13	0.03	2.27	10.43	0.30	2.00	2.96	0.16	9.69

4.2.2 The results of modelling

Modelling geothermal water during cooling

As mentioned above, the method of geothermal utilization in Tanggu involves pumping the water directly into the pipeline and piping it to the user for space heating. At the end of energy extraction, the tail water temperature is about 35-40°C. From the calculation of water-mineral equilibrium with WATCH, it was found that some minerals can be expected to be deposited during cooling. A qualitative study of this process was carried out using CHILLER for cooling from 80 to 35°C with a temperature increment of 5°C. A closed system cooling at a pressure of 1 bar was modelled. In real systems, the pressure may be higher than 1 bar. Since many studies show that the pressure effects on equilibrium are not very important, we assumed a pressure of 1 bar in the system. The geothermal water in Tanggu is very dilute. Mineral deposition in the pipelines is not a serious problem. The results of modelling the cooling process showed that the scale should be composed of calcite, clc14/2, dolomite, pyrite and quartz. Many studies have shown that the rate of crystallization of quartz is so slow that amorphous silica, a metastable phase should be responsible for silica deposition. As a result of cooling, the silica deposition product will be amorphous silica rather than quartz. Cooling to 45°C, appeared to produce about 0.058 g solids in 1 litre of solution, which was composed of 68.84% SiO₂, 0.45% pyrite, 1.32% dolomite, 1.78% clc14/2 and 27.62% calcite. At 35°C, the weight of the deposited minerals would be 0.059 g, composed of 72.72% SiO₂, 0.43% pyrite, 0.40% dolomite, 1.72% clc14/2 and 24.72% calcite. In response to mineral deposition, the concentrations of chemical components in the solution changed, mainly those of Ca²⁺, Mg²⁺, SiO₂, Al³⁺, Fe²⁺ and HCO₃⁻. The composition of the geothermal water and the calculated results from this simulation are given in Table 6. The mineral assemblage basically corresponds to the scale found in the real systems which is composed of silicate and carbonate.

TABLE 6: The chemical composition of geothermal water and modelled water (ppm)

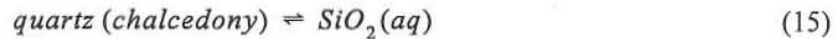
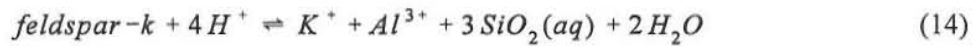
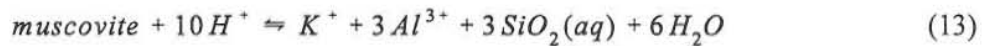
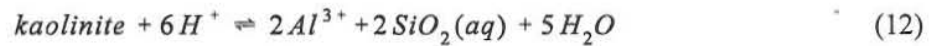
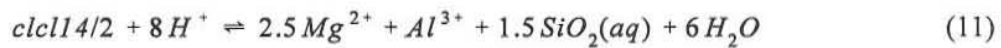
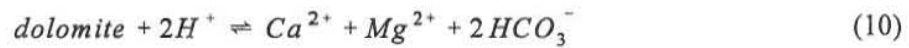
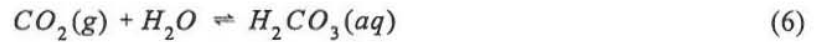
No.	K ⁺	Na ⁺	Ca ²⁺	Mg ²⁺	Fe ²⁺	Al ³⁺	Cl ⁻	SO ₄ ²⁻	HCO ₃ ⁻	S ²⁻	SiO ₂	pH
1	7.4	570.0	11.0	0.6	0.12	0.1	402.4	257.0	619.4	0.06	52.5	8.20
2	7.4	570.0	5.06	0.345	0.0	0.00047	402.4	257.0	610.3	0.0	8.91	7.87
3	3.2	585.5	7.7	0.1	0.0	0.12	402.4	256.9	751.2	0.0	44.4	6.95

Note: No. 1 is the chemical composition of geothermal water sampled from well TR1 on 28-12-1991; No. 2 is the calculated results for tail water composition from cooling geothermal water; No. 3 is the result of modelling successive water-rock interaction to 95°C.

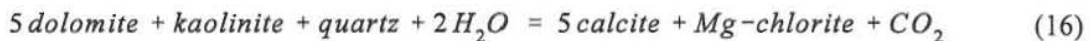
Modelling water-rock interaction during reinjection

The saturation state of cooled water was tested by SOLVEQ before modelling the water-rock interaction. The results show that the water is undersaturated with all the minerals considered at temperatures above 35°C. Thus, these minerals will dissolve and approach equilibrium with water when this water is injected into the reservoir and heated.

The results of modelling water-rock interaction at different temperatures show that the secondary mineral assemblage is composed of calcite, clinocllore, kaolinite, muscovite and quartz (chalcedony). But the amount of the secondary minerals decreases with temperature due to the high solubility of most minerals at high temperature. The simulation of successive water-rock interaction was carried out by combining water-rock interaction and heating. Here, 0.1 g rock portions were titrated until a total of 1 g rock had been titrated with 1 litre water and the solution heated to next temperature and the rock titrated again. The secondary mineral assemblage consists of calcite, clinocllore, kaolinite, muscovite and quartz (chalcedony). The system can simply be written in terms of the following equations:



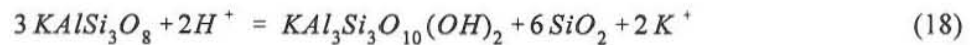
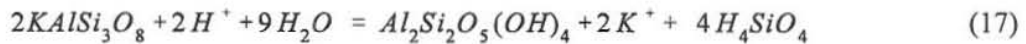
Feldspar, the major mineral formed in the sandstone, will dissolve with other similar aluminosilicates, release SiO_2 and consume H^+ according to Reaction 14, when the water is reinjected into the reservoir because it is undersaturated with respect to it. As the water is saturated with calcite, dolomite, clinocllore and quartz (chalcedony) at $35^\circ C$, the increase of silica and pH in the water make Reactions 9, 10, 11, 12 and 13 proceed to the left to precipitate dolomite, calcite, clinocllore, kaolinite, and muscovite. Relative to the reinjected water, the sequences of water-rock interaction at $35^\circ C$ consumed Ca^{2+} , K^+ , Mg^{2+} , HCO_3^- and H^+ from the solution, but increased its SiO_2 and Al^{3+} concentrations. But in the secondary mineral assemblage, kaolinite and dolomite are not precipitated at $35^\circ C$. The possible reason is the predominance of the carbonate and silicate mineral reactions. Zen (1959) suggested that the carbonate-silicate reaction is



This reaction will proceed to the right due to CO_2 consumption by feldspar dissolution. Hutcheon et al. (1980) used this reaction to account for diagenetic transformation from dolomite and kaolinite to chlorite and calcite found in sandstone. However, many other experimental studies showed Mg-smectite forming instead of chlorite using the same reactants and given a similar reaction. And in Fe- and Ca-bearing

systems, Fe-chlorite and Ca-smectite will form instead of Mg-chlorite. Wu-Liang and Longo (1994) found that the concentration of CO₂ in the reaction fluid controls which clay type is formed in experiments, in which high CO₂ concentration led to chlorite formation and low CO₂ concentration led to smectite formation. In our case, we have not found smectite replaced by chlorite. This is due to the high CO₂ concentration of the reaction solution and rock.

Upon completion of water-rock interaction at 35°C, the water was heated to 45°C and reacted with rock at this temperature. Because the reactions at 35°C tend to remove the K⁺, Ca²⁺ and Mg²⁺ from the solution, the water dissolves the rock and the concentrations of these components increase when the temperatures increase to 45°C. H⁺, Ca²⁺ and Mg²⁺ increase in the solution from 35 to 45°C. Kaolinite precipitates at 45°C. Because kaolinite, muscovite and clinochlore form from feldspar dissolution, the K⁺ concentration increases and the Al³⁺ concentration decreases in the solution, which is close to a new equilibrium with the minerals. The same result was found in the studies of diagenetic processes in sandstone by Ben Baccar and Fritz (1993). The kaolinite and muscovite formation is due to the dissolution of feldspar from the rock. Kaolinite is the chief alteration product of aggressive water reacting with silicate minerals. The following reaction equations can be used to explain these processes:



It has been recognized that the K-feldspar dissolves at 25 to 80°C (Ben Baccar and Fritz, 1993). They have confirmed that feldspar end-members are unstable at low temperatures, but may become stable and precipitate as secondary phases at high temperatures. They also found that kaolinite precipitated at 40 and 80°C, while illite precipitated during the whole diagenetic processes. Their results suggest that kaolinite precipitates over the whole temperature range.

At temperatures above 45°C, the dissolution of rock increases the SiO₂, Al³⁺, CO₂ and H⁺ concentrations of the water due to the large solubility of some minerals at elevated temperatures. The saturated minerals continually precipitate with temperature increase and rock addition. Calcite precipitates and the Ca²⁺ concentration in the solution decreases at high temperatures because the solubility of calcite and other carbonate minerals decreases with increased temperature. Most minerals including calcite, clinochlore, kaolinite and muscovite are precipitated because feldspar minerals tend to become more soluble as PCO₂ increases and their amount increases with increased addition of rock and high temperatures. Quartz (chalcedony) is supersaturated at all temperatures. Secondary minerals formed at different temperatures are listed in Table 7.

The water composition changes with the dissolution of rock and precipitation of secondary minerals. The general trend of evolution in chemical composition is close to that of the geothermal water in the reservoir. This suggests that the condition of water-mineral equilibrium at the end of our modelling is close to that in the geothermal system. Relative to the composition of the reinjection water, the K⁺ and Mg²⁺ concentrations decrease but the Ca²⁺, Al³⁺ and SiO₂ concentrations increase in the water. This is due to dissolution of feldspar and formation of clay minerals, such as kaolinite and clinochlore. The same findings have been reported from experimental studies of hot dry rock geothermal system at Rosemanowes, Cornwall, UK (Richards et al., 1992; Savage et al., 1992). Figure 7 shows the concentrations of major components in the solution at different temperatures used in the modelling. Water-rock interaction at 35°C tends to decrease the concentrations of HCO₃⁻, Ca²⁺, Mg²⁺ and K⁺ concentrations in the reinjected water due to saturation of many minerals. With increased temperature, some minerals become undersaturated and dissolve from the rock.

TABLE 7: Secondary minerals formed at different temperatures and volumetric changes as a result of water-rock interaction at different temperatures

Temperature (°C)	Secondary minerals (percentage of deposition)	Volumetric changes Vd-Vp (cm ³)	Porosity effects
35	calcite (21.14%), muscovite (23.28%), clinochlore (6.87%), quartz (48.32%).	0.0573989	increase
45	calcite (19.5%), muscovite (20.95%), clinochlore (6.86%), kaolinite (3.03%), quartz (49.66%).	0.065999	increase
55	calcite (19.73%), muscovite (19.96%), clinochlore (6.85%), kaolinite (3.97%), quartz (49.49%)	0.0577748	increase
95	calcite (20.18%), muscovite (19.09%), clinochlore (6.84%), kaolinite (4.76%), quartz (49.13%).	0.058178	increase
65	calcite (19.89%), muscovite (19.51%), clinochlore (6.85%), kaolinite (4.38%), quartz (49.37%).	0.057559	increase
75	calcite (20.01%), muscovite (19.28%), clinochlore (6.84%), kaolinite (4.59%), quartz (49.28%).	0.05757	increase
85	calcite (20.10%), muscovite (19.16%), clinochlore (6.84%), kaolinite (4.70%), quartz (49.2%).	0.057784	increase

The concentrations of Ca²⁺ and Mg²⁺ increase to a maximum at 45°C due to Ca²⁺ and Mg²⁺ removal from the solution at 35°C and then decrease with the onset of deposition of calcite and chlorite, corresponding to these, the HCO₃⁻ moles in the solid (Figure 8) decrease from 35 to 45°C and then increase. The concentration of K⁺ in the solution increases with the reaction process to 70°C due to the formation of kaolinite. The concentration of SiO₂ increases with temperature. This is due to increased SiO₂ solubility with temperature. Another contribution is the dissolution of feldspar, the major rock-forming mineral in sandstone. The concentration of Al³⁺ decreases from 35 to 45°C and then increases with the formation of kaolinite above 45°C. The concentration of HCO₃⁻ increases with temperature due to addition of CO₂ from the rock, and as a result of this, the pH decreases. A high P_{CO2} value in the reservoir is consistent with high alkalinity in formation water and may be explained with reference to dissolution of carbonates suggested as a CO₂ source in sedimentary basins. A high P_{CO2} value is likely to affect greatly pH which in turn controls the aluminosilicate equilibria. The P_{CO2} values calculated by modelling are much higher than those calculated by the K-Ca indicator given by Giggenbach (1988). This is due to the high content of CO₂ in the reaction rock.

The porosity effects caused by water-rock interaction in the reservoir are modelled by CHILLER. Only chemical transformations (precipitation and dissolution) are considered to affect volume balance which is used to evaluate the changes in porosity. The porosity caused by water-rock interaction is defined as

$$Porosity = \frac{100 \times (Volume\ of\ dissolved\ rock - Volume\ of\ precipitated\ minerals)}{Volume\ of\ dissolved\ rock} \quad (19)$$

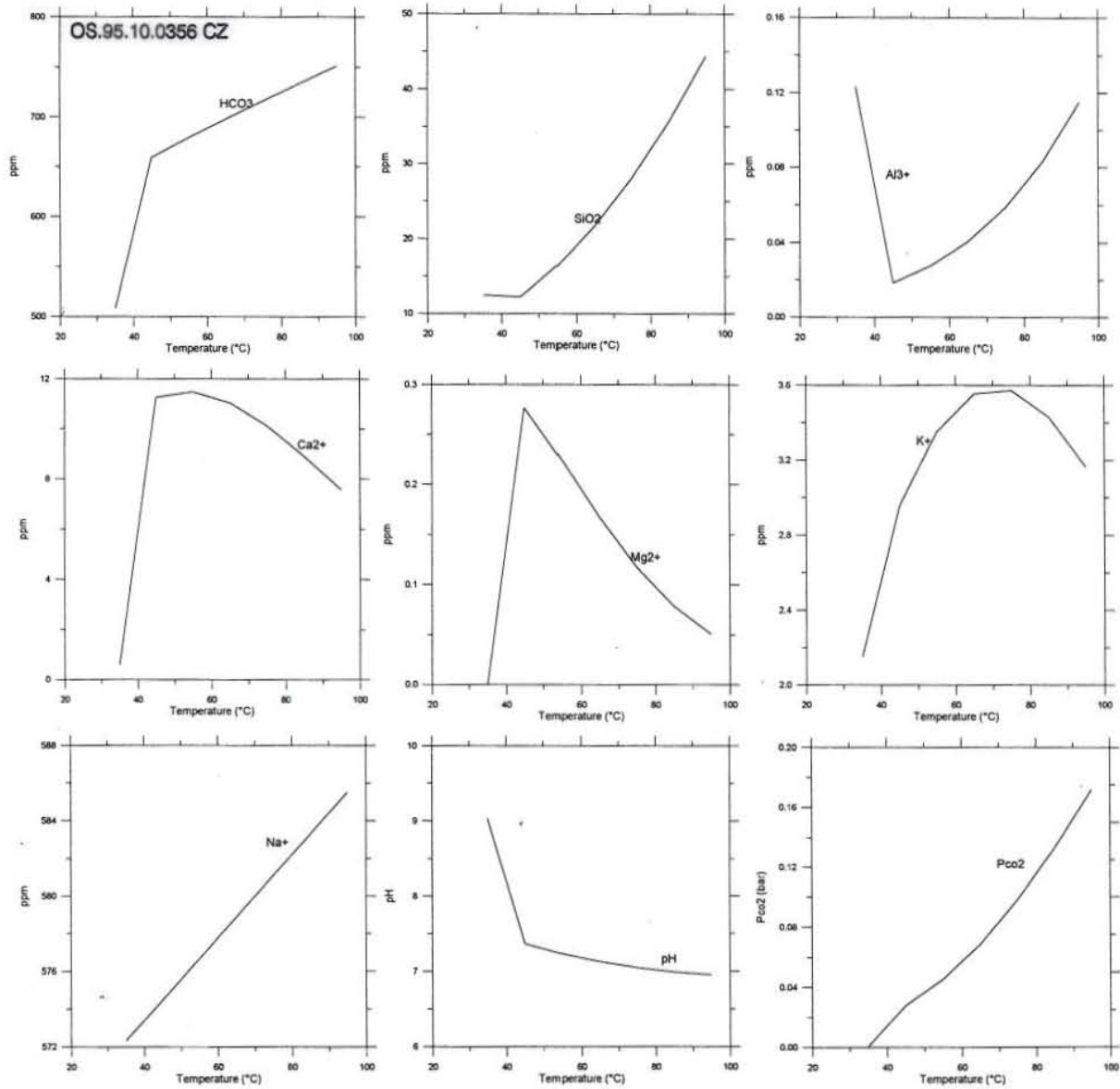
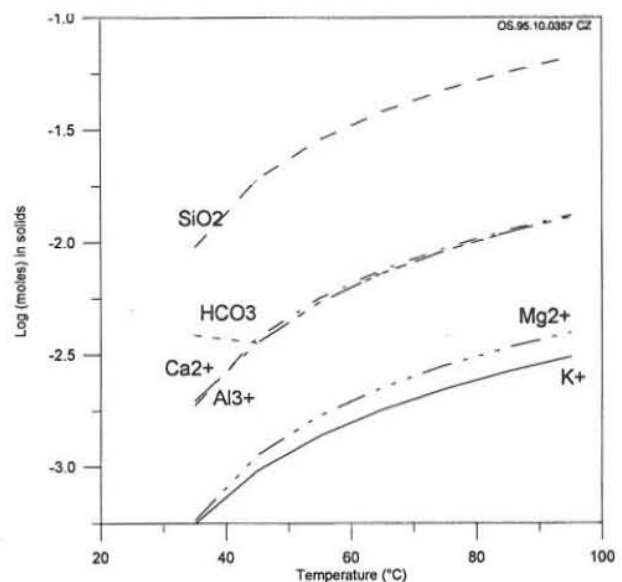


FIGURE 7: The concentrations of chemical components in the solution at different temperatures calculated by modelling

The results of simulations of porosity suggest positive porosity effects. The net porosity caused by water-rock interaction is about 12 - 14%. The temperature decreases around the reinjection well with time and the chemical effects on porosity are not important. Along the flow path to the production well, the movement of the fluid allows transport of chemical elements from one place (dissolution) to another (precipitation) and the porosity is decreased due to secondary mineral precipitation.

FIGURE 8: The moles of individual components in solids calculated by modelling successive water-rock interaction



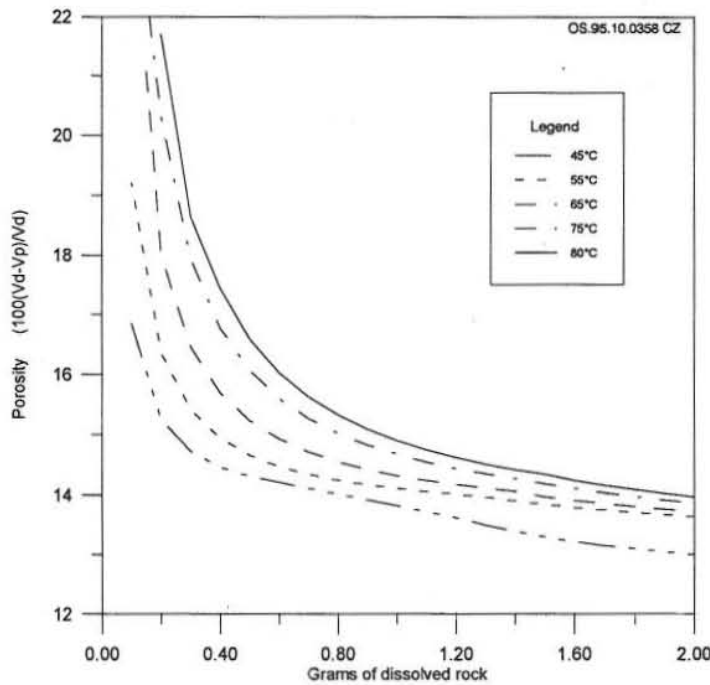


FIGURE 9: The porosity calculated by modelling water-rock interaction at different temperatures

suggests that the effects of porosity increase with temperature. Thus, at the beginning of reinjection, the porosity caused by reaction is large because the reinjected water is heated up quickly by the reservoir and the dominant chemical reaction is dissolution. It seems that reinjection is favourable with respect to porosity at the beginning. Similar effects have been reported for many geothermal fields, in which

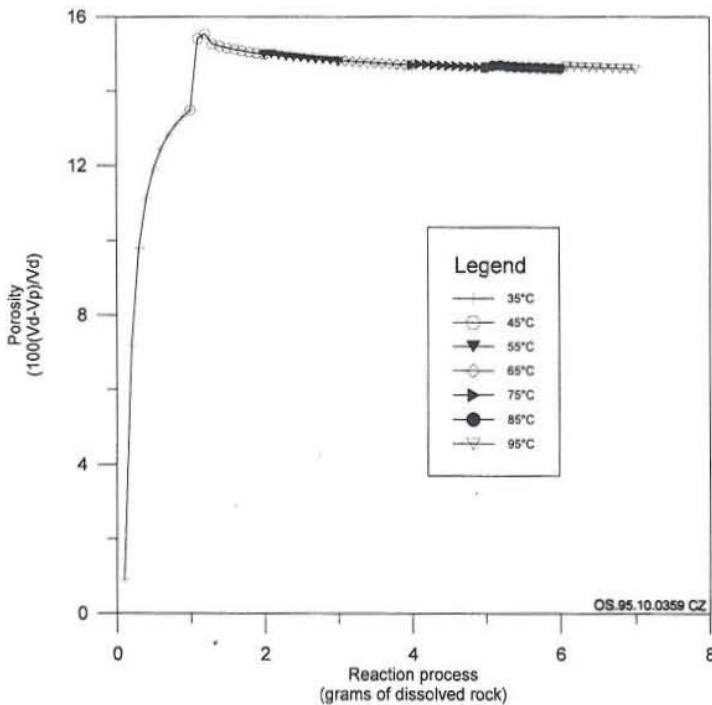


FIGURE 10: The porosity caused by successive water-rock interaction

the short term injectivity has been found to increase with time. This has been observed in several fields in Iceland, New Zealand and Mexico. It has been explained as a result of thermal contraction of reservoir rock. With continued reinjection, the situation around the well can be considered as isothermal, and the porosity caused by water-rock interaction is not significant due to the saturation of most minerals at the reinjection temperature. In the isothermal case, the high reinjection rate is accompanied by a large and the low reinjection rate by a small porosity effect.

Figure 10 shows the results of modelling water-rock interaction successively. The purpose is to study its effects on the porosity with respect to water flow in the reservoir. Reinjection time and distance from

the reinjection well increase to the right. The result shows that porosity increases fast at the beginning and near the reinjection well and then reaches a maximum. This is because the water is undersaturated with minerals at the beginning and when the temperature increases the water-rock interaction is dominated by dissolution. This also shows that reinjection is favourable for porosity at the beginning. When the water flows to the production well and the temperature rises, it approaches equilibrium with certain minerals. The water-rock interaction is switched to the precipitation of secondary minerals and the porosity decreases, approaching a constant value of 14% at a temperature close to that of the reservoir. The small change in porosity is due to the large amount of deposition of secondary minerals and the small volume balance between dissolved rock and secondary minerals. The reaction is close to equilibrium and a mineral that dissolves near the well is deposited at this point.

4.3 Modelling silica transport in the reservoir during reinjection

Two problems concerning reinjection are injectivity decline and early thermal breakthrough. Studies of water-rock interaction can be used for understanding the injectivity changes as discussed above, while studies of chemical component transport in the reservoir during reinjection can be used for predicting the thermal breakthrough in the production well. Usually, if a water different from the geothermal water, or separated geothermal water is injected into a reservoir, isotopic ratios and conservative components, such as δD , $\delta^{18}O$, 3H and Cl , are good candidates as natural chemical tracers. If the content of these components is the same in the reinjected and geothermal waters, they cannot be used as tracers. On the other hand, the breakthrough time for a conservative component represents the hydrodynamic breakthrough time because it does not react with rock during injection. It means that thermal breakthrough occurs a long time after the chemical tracer breakthrough. This time tells us more about reservoir permeability and the fraction of injected water returning to the production well than thermal breakthrough time does. The concentration changes of a reactive component are dependent on temperature changes during injection. The transport of a reactive component depends on the thermal gradient and hydrodynamic parameters, and the thermal breakthrough of the tracers comes after that. If we can determine the concentration changes for a reactive component, we can predict the thermal breakthrough time more accurately than by using tracers. Besides, from the study of its transport with respect to reaction, we can obtain information about reservoir property changes. Bödvarsson and Stefánsson (1989) have reviewed chemical aspects of reinjection using both theoretical and field studies.

As for geothermal reinjection in Tanggu, it will probably be carried out using spent geothermal water. Since there is no difference between the concentrations of conservative components in injected and geothermal water, a reactive component is taken into account. As has been discussed above, the water-rock interaction tends to change concentrations of Ca^{2+} , Mg^{2+} , K^+ and SiO_2 , and the K/Mg geothermometer has the advantage of a fast response to the temperature change (K^+ is taken up, Mg^{2+} is deposited with increased temperature). Thus, all the above components should be considered as good candidates for modelling solute transport during reinjection and used for monitoring the system. In fact, the K/Mg geothermometer can be used in the practical operation of reinjection. Because there are not well documented chemical kinetic equations for the reactions between the Mg^{2+} , K^+ , Ca^{2+} and minerals, the chemical process cannot be described in the mathematical model. Therefore, only SiO_2 is considered as a candidate for the modelling in this scale.

4.3.1 Mathematical model and approach

The reinjected water is not saturated with silica before it is heated by the reservoir. So silica will dissolve from the reservoir rock when the reinjected water is transported to the production well. Modelling this process can aid in understanding the process of reservoir property changes with regard

to hydrodynamic and kinetic properties. The modelling can be done by coupling the solute transport equation, the energy equation and chemical reaction kinetic equation, which represent conservation of mass, momentum and energy. The physical system used here is considered to be an idealized confined aquifer composed of a chemically inert material with a fine pore. The aquifer is homogeneous, of uniform thickness, and effectively of infinite horizontal extent. The aquifer is isotropic with hydraulic and thermal conductivity. Darcy's law is valid for fluid flow. The fluid and the rock are local and in instantaneous thermal equilibrium. The following equations describe the flow field, temperature field and chemical field in the physical system:

$$\bar{v} = -K\nabla h \quad (20)$$

$$\frac{\partial h}{\partial t} = \frac{Kb}{\mu} \nabla^2 h \quad (21)$$

$$\rho_a c_a \frac{\partial T}{\partial t} + \rho_l c_l \bar{v} \nabla T + \lambda \nabla^2 T = 0 \quad (22)$$

$$D \nabla^2 C - \bar{v} \nabla C = \frac{\partial C}{\partial t} + R_c \quad (23)$$

$$R_c = k^- \frac{A}{M} (C - C_s) \quad (24)$$

$$\log k^- = -0.707 - \frac{2598}{T_K} \quad (25)$$

$$h(x, y, 0) = h_0 ; \quad T(x, y, 0) = T_r ; \quad C(x, y, 0) = C_{s0} \quad (26)$$

$$\lim_{x, y \rightarrow \infty} T(x, y, t) = T_r \quad t > 0 \quad (27)$$

$$\lim_{x, y \rightarrow \infty} h(x, y, t) = h_0 \quad (28)$$

$$\lim_{x, y \rightarrow \infty} C(x, y, t) = C_{s0} \quad t > 0 \quad (29)$$

$$T(0, 0, t) = T_i \quad t > 0 \quad (30)$$

$$C(0, 0, t) = C_i \quad t > 0 \quad (31)$$

where

- A/M = Ratio of surface area to mass of water (m^2/kg)
- b = The effective thickness of aquifer (m)
- c_a = Effective aquifer heat capacity ($\text{J kg}^{-1}\text{C}^{-1}$)
- c_l = Fluid heat capacity ($\text{J kg}^{-1}\text{C}^{-1}$)
- C = Silica concentration (ppm)
- C_i = Silica concentration in reinjected water (ppm)
- C_s = Silica equilibrium concentration (ppm)

C_{s0}	= Silica equilibrium concentration at reservoir temperature (ppm)
D	= Coefficient of hydrodynamic dispersion
h	= Hydraulic head (m)
h_0	= Initial hydraulic head (m)
k	= Apparent rate constant (s^{-1})
K	= Hydraulic conductivity (m/s)
R_c	= Chemical reaction term
t	= Time (s)
T	= Temperature ($^{\circ}C$)
T_i	= Temperature of reinjected water ($^{\circ}C$)
T_K	= Temperature (K)
T_r	= Reservoir temperature ($^{\circ}C$)
v	= Darcy's velocity field (m/s)
λ	= Thermal conductivity of aquifer ($W\ m^{-1}^{\circ}C^{-1}$)
μ	= Storativity of aquifer (m^2)
ρ_a	= Effective aquifer density (kg/m^3)
ρ_l	= Fluid density (kg/m^3)
v	= The average particle velocity (m/s)

The two dimensional fixed solutions for the models can be obtained by using an integrated finite differential method. For the case of constant particle velocity and thermal velocity, if the effect of thermal conduction and chemical diffusion can be neglected, Malate and O'Sullivan (1992) gave an analytical solution as radial flow by using the method of characteristics

$$C(r, t) = C_{s1} + (C_i - C_{s1}) \exp\left(\frac{-k_1 r^2}{2V_1}\right) ; (r^2 < Ut ; T = T_i) \quad (32)$$

$$C(r, t) = C_{s0} + (C_{s1} - C_{s0}) \exp\left(\tau\left(\frac{r^2}{2} - Ut\right)\right) + (C_i - C_{s1}) \exp\left(\frac{(\sigma r^2/2) + \omega t}{\delta}\right) ; (r^2/2 < V_0 t, T = T_r) \quad (33)$$

with

$$\tau = \frac{k_0}{(U - v_0)} , \quad \sigma = (k_0 v_1 - k_1 U) , \quad \omega = U(k_1 v_0 - k_0 v_1) , \quad \delta = v_1(U - v_0)$$

and

$$C(r, t) = C_{s0} ; (r^2/2 > V_0 t, T = T_r) \quad (34)$$

where

- C_{s0}, C_{s1} = Silica equilibrium concentration at reservoir and reinjected water temperature, respectively
 k_0, k_1 = Chemical reaction rate constants at reservoir and reinjected water temperature, respectively, which are $k = k(C_r - C_s)$ (s^{-1})
 r = radial distance (m)

v_0, v_l = Particles velocity in the condition of reservoir temperature and reinjected temperature, respectively (m/s)
 U = Thermal front velocity (m/s)

This analytic solution is good for the radial volume around the reinjection well which is convection dominated. In the present modelling, this solution is used for silica transport in the reservoir. The velocity of the water particles and the thermal front is considered to be an average velocity. Generally, there are three methods for determination of the velocity of the water. The first includes all techniques that are directly dependent on use of the Darcy equation. The second involves the use of artificial tracers. The third consists of ground water age-dating methods using environmental isotopes such as tritium and carbon 14. Since no tracer test has been carried out in Tanggu, the first method is used to determine the velocity from the information on the hydraulic conductivity, hydraulic gradient, and porosity in the flow field. From the different flow rate well test, the effective radii of influence were obtained in Tanggu in 1989. The Darcy velocity of different rate can be approached according to the equation

$$v = \frac{Q}{2\pi b} \int_{r_w}^R \frac{1}{r} dr \quad (35)$$

where R and r_w are the effective radius of influence and the well radius, respectively in m and Q is the flow rate of the well in m^3/h .

The velocity of water particles is

$$\bar{v} = \frac{v}{\phi} \quad (36)$$

where ϕ is the porosity of the reservoir.

If the heat conduction term in Equation 3 is ignored, the thermal front velocity can be obtained as

$$U = \frac{\rho_l c_l}{\rho_a c_a} v \quad (37)$$

Three cases of different flow rates were used to model the silica transport process. Since the geothermal water is close to equilibrium with chalcedony, the equilibrium concentrations at reservoir and reinjected water temperature were calculated from the chalcedony geothermometer temperature. The reservoir temperature is considered to be 80°C . The initial surface area is difficult to determine, but it is thought that $200 \text{ m}^2/\text{kg}$ corresponding to a surface area of fine grained sediments is a reasonable approximation. The parameters used in modelling are listed in Table 8.

4.3.2 Results

Figures 11 and 12 show the results for silica transport in the reservoir during reinjection. Moving radially outwards from the injection well, three regions appear:

TABLE 8: Parameters used in the model for Tanggu

Parameters	Values
Well flow rate, Q (m ³ /h)	$Q_1 = 90, Q_2 = 133, Q_3 = 90$
Effective radius of influence, R (m)	$R_1 = 990, R_2 = 1500, R_3 = 2000$
Temperature of reinjected water, T_i (°C)	$T_i = 35$
Reservoir temperature, T_r (°C)	$T_r = 80$
Silica concentration in reinjected water, C_i (ppm)	$C_i = 8.9$
Silica equilibrium concentration at 35°C, C_{s1} (ppm)	$C_{s1} = 20.2$
Silica equilibrium concentration at 85°C, C_{s0} (ppm)	$C_{s0} = 57.7$
Surface area, A/M (m ² /kg)	$A/M = 200$
Aquifer thickness, b (m)	$b = 61$
Effective porosity of aquifer, ϕ	$\phi = 0.1998$
Aquifer heat capacity; $\rho_a c_a$ (MJ/m ³ K)	$\rho_a c_a = 2.475$
Heat capacity of fluid, $\rho_f c_f$ (MJ/m ³ K)	$\rho_f c_f = 4.18$
Well radius, r_w (m)	$r_w = 0.1$

1. An inner zone in which silica concentration is the same as that in the reinjection water;
2. A concentration transition zone termed the "front" where the silica concentration rises relatively abruptly;
3. An outer zone with a silica concentration close to the equilibrium concentration.

The reaction reaches equilibrium in regions 1 and 3. The porosity of the reservoir does not change in these two regions. The reaction takes place mainly in region 2 and the porosity of the reservoir increases with silica dissolution in this region. The front will move outwards from the injection well with injection time and the outermost fluid with an equilibrium concentration in this zone will reach the production well first. The silica front region is about 50 m wide in the reservoir, and moves to the production well. The velocity of the silica front is large when the reinjection flow rate is high. At the beginning of reinjection, the porosity increases within an area of 50 m radius around the well and the effect is important for the operation at this time.

But with continued reinjection, the front moves to the production well and in the inner region the silica concentration is close to the equilibrium concentration at the reinjection temperature. Therefore, the porosity effect near the reinjection well is not so important during the later stages of reinjection. After the equilibrium concentration of silica in the outermost front reaches the production well, the silica concentration will decrease due to the front region breakthrough, because the silica front moves faster than the thermal front and more slowly than the hydrodynamic front. The silica concentration in

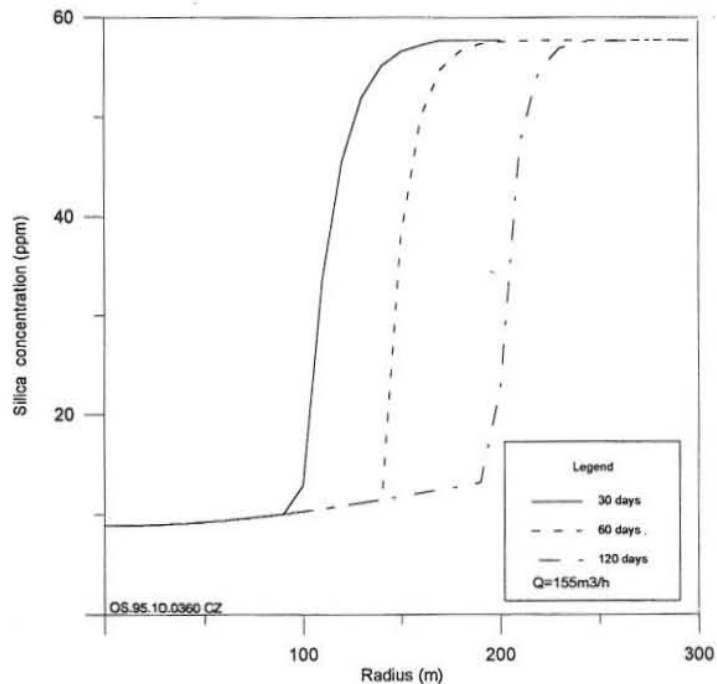


FIGURE 11: The extent of silica front migration with time

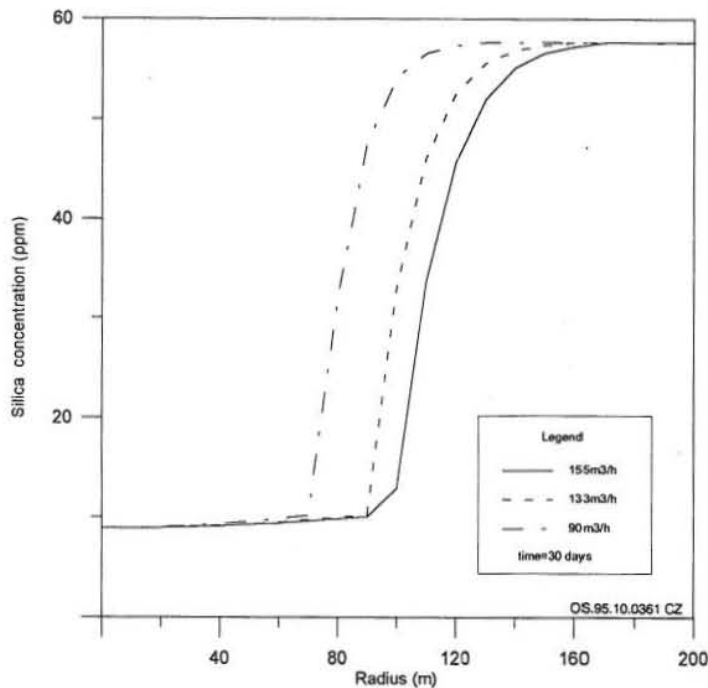


FIGURE 12: The extent of silica front migration at different reinjection rates

In the absence of analytical data on tail water, the result from the modelling can only be regarded as an approximation.

- a) The district heating system is very complicated, and the chemical composition of reinjected water, calculated by simple modelling of the cooling process, may not yield a good fit to the real system.
- b) The results for porosity by modelling water-rock interaction show a positive effect. The effects of mineral dissolution and precipitation are the only factors considered. In fact, many factors, such as plugging of pores by fine particles and gas bubbles can significantly damage the reservoir during reinjection. This has been found to be the case during a short injection test in the North China Oil Field. The effects of redox reactions and bacterial activity are not included in our model because of lack of data. But in many studies (e.g. De Hann et al., 1994) it has been found that redox and bacterial reactions are very important in determining reservoir porosity in low-temperature aquifers, especially in Fe-bearing sedimentary environments. The result may also deviate from the real situation because an average chemical composition of rock, which is probably different from that of the formation rock, was used in the modelling.
- c) The velocity of water particles used in the modelling of silica transport, has been estimated by Darcy's equation. But the parameters in Darcy-based equations have large inherent uncertainties that generally cannot be avoided. The hydraulic conductivity determinations often have large uncertainties. Errors in hydraulic conductivity measurements combined with errors associated with determinations of the gradient and the porosity result in a considerable error, associated with the computed velocity. Usually, better accuracy can be attained by the use of chemical tracers.

production water can be used to monitor the thermal breakthrough and aid in determining whether the operation should be continued.

4.4 Discussion

As a method for the study of geothermal reinjection, the modelling of geochemical effects provides significant information about the chemical effects caused by reinjection. The accuracy of the model was acceptable within the limits of the data, at least for the early part of the operation. The following factors may affect the accuracy of the model

5. CONCLUSIONS

1. Studies of the chemistry of the geothermal water in Tanggu show that the geothermal water originated from meteoric water. The age of the geothermal water is about 24-26 thousand years according to the results of ^{14}C dating. The average reservoir temperature calculated by geothermometers is 80-100°C. After high rate production, the chemical composition changed and the reservoir temperature based on the chemical geothermometer result shows a slight cooling of the reservoir due to cool water recharges. Two sets of equilibrium temperatures in the reservoir obtained from solute geothermometers and the use of $\log(Q/K)$ diagrams suggest that a higher temperature may exist in the reservoir. Further monitoring and studies of the chemical composition of the geothermal water should be carried out in order to obtain more reliable information about the reservoir.
2. In the early stages of reinjection, it is important to learn the reservoir response. Because of the limited data, it is difficult to obtain information on that without modelling. The geochemical effects of geothermal reinjection have been studied by modelling water-rock interaction and the movement of the chemical reaction front in the reservoir. The modelling has yielded the following results:
 - a) The chemical composition of the geothermal water changes because of mineral precipitation upon the lowering of temperature caused by energy extraction. When the tail water is reinjected into the reservoir, it will be heated by the reservoir and some minerals will dissolve. As a result of water-rock interaction, the secondary mineral assemblage is found to be mainly composed of kaolinite, clinocllore, calcite, and quartz (chalcedony). The chemical composition of the water approaches that of the geothermal water in the reservoir.
 - b) The porosity effect caused by water-rock interaction is positive. The effects take place most significantly in the thermal front region. At the beginning of reinjection, the porosity increases around the reinjected well and reinjection is favourable. With time, the effect around the reinjection well becomes less important. The net porosity contributed by water-rock interaction is about 12-14%. Reservoir damage caused by mineral precipitation is not a serious problem according to the modelling results.
 - c) The modelling of silica reaction front movement in the reservoir shows that the chemical front is about 50 m wide. It can be taken as the porosity effect width. The chemical front velocity is greater than that of the thermal front and slower than that of the hydrodynamic front. It suggests that silica can be used as a good indicator for monitoring thermal breakthrough during reinjection.

6. RECOMMENDATIONS REGARDING GEOTHERMAL REINJECTION IN TANGGU

As a strategy for geothermal production and a method to deal with the problem of pressure decline in Tanggu, there is no doubt that geothermal reinjection is a good choice and it is urgent that it is carried out sooner than later, considering the experience of reinjection in other geothermal fields. The following recommendations are made:

1. Theoretical studies should be conducted, which should involve the chemistry of geothermal fluid, water-rock interaction, geothermal reservoir engineering and mechanical engineering. The

geothermal engineering studies should focus on the movement of the thermal and chemical fronts. The present study is an initial study of water-rock interaction in this field, so further studies should be conducted, taking into account redox and bacterial reactions as significant chemical processes.

2. As some reinjection wells have been drilled in Tanggu, it is necessary to carry out a short time reinjection test with chemical tracers before the reinjection operation starts. The purpose is to obtain the parameters of the reservoir because its response to reinjection depends on the initial thermodynamic state of the reservoir. The reinjection test should be done for different injection rates and temperatures in order to design the optimal injection rate. Meanwhile, more detailed modelling studies using the test data should be carried out to learn the response to the reinjection.
3. It is necessary to do some theoretical and experimental studies on possible reservoir damage caused by fine particles plugging pores because of mechanical migration. The experience of reinjection in France and other European countries shows that sandstone reservoir damage is mainly caused by physical rather than chemical processes, especially in argillaceous sandstone reservoirs (Egberts et al., 1995).
4. Reinjection using produced water has been carried out in the North China Oil Field and a summary report based on the results has been finished. Some problems, such as the blocking of pores and deposition have been dealt with or prevented by washing the well with regular intervals after reinjection. This experience should be very helpful for the geothermal reinjection operation in Tanggu.
5. Setting up of a long-term monitoring programme. The monitored parameters include the production and injection rates, the temperature of the produced water and reinjected water, the water level in the production and observation wells inside and outside the area, the chemical composition of produced and reinjected water. The monitoring programme and the data collected depend on the internal parameters of the system and on the production and reinjection procedures applied in the field. The distribution of temperature, pressure and chemicals in the geothermal water can be disturbed. For chemicals in the geothermal water, the monitoring programme depends on the production and reinjection rate and the spacing of the wells. The procedure for monitoring production wells in Iceland, is to carry out a complete chemical analysis twice per year for each well and analysis for selected elements up to every month (Kristmannsdóttir et al., 1995). As found in Iceland, the sample frequency needs to be about once per month if the changes in chemical composition are to be correlated to seasonal changes in production. For an injection system, the frequency of sampling should be greater than that for a non-reinjection system. Dissolved O_2 , non-condensable gas, SiO_2 , K, Mg and pH in the geothermal water in both production and reinjection wells are suggested as monitoring parameters. The O_2 content of the reinjection water is the major agent that may cause corrosion. The changes of geothermometer temperatures for quartz and K/Mg in the production well can be used to determine the thermal breakthrough and aid in deciding whether reinjection will be continued or not.

ACKNOWLEDGEMENTS

I would like to express my gratitude to Dr. Ingvar Birgir Fridleifsson for his support and help throughout the training program. I sincerely thank Dr. Halldór Ármannsson, my advisor, for sharing his experience

and knowledge of geochemistry. My thanks go to Mr. Lúdvík S. Georgsson, Ms. Súsanna Westlund and Ms. Margrét Westlund for their efficient guidance. I wish to give my thanks to all the lecturers and staff members of the Chemical laboratory of Orkustofnun for their willingness to help.

I am also indebted to Prof. Li Feng-zhong and Prof. Cai Guan-ying from the Tianjin Bureau of Geology and Mineral Resources for providing me with the information used here and help during sampling of water.

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