

## APPLICATIONS OF GEOTHERMOMETRY

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### INTRODUCTION

Applications of geothermometry constitute one of the major tools for the exploration and development of geothermal resources. Geothermometry is simply the estimation of the reservoir (subsurface) temperatures using the chemical and isotopic composition of surface discharges from wells and/or natural springs/fumaroles.

Geothermometers are classified into two groups: *Chemical Geothermometers* and *Isotopic Geothermometers*. The former utilizes the chemical composition (essentially the silica and major cation contents of water discharges, and the gas concentrations or relative abundances of gaseous components in steam discharges), whereas the latter is based on isotope exchange reactions between various phases (water, gas, mineral) in geothermal systems.

This course is designed to focus on the chemical geothermometers as applied to water discharge. The course consists of two major parts: the first part summarizes the basic principles and types of chemical geothermometers along with the relevant equations applied; the second part is devoted to examples/problems concerned with the application of chemical geothermometers to geothermal systems.

### 1. CHEMICAL GEOTHERMOMETERS : BASIC PRINCIPLES AND TYPES

#### 1.1. BASIC PRINCIPLES

Chemical geothermometers are developed on the basis of temperature-dependent chemical equilibrium between water and minerals under deep reservoir conditions. The use of chemical geothermometers is based on the assumption that the water preserves its chemical composition during its ascent from the reservoir to the surface. This assumption of equilibrium is not necessarily always true. However, the studies of the chemistry of well discharge and the alteration mineralogy have demonstrated the presence of equilibrium in several geothermal fields and this is taken to indicate that the assumption of equilibrium is valid.

The assumption of the preservation of water chemistry may not always hold. This is because the water composition may be affected by processes such as cooling, and mixing with waters from different reservoirs.

During their ascent from a deep reservoir to the surface, geothermal waters may cool by conductive heat loss while migrating through cooler rocks, or by boiling because of

decreasing hydrostatic head. Conductive cooling does not by itself change the composition of the water, but may affect the degree of saturation with respect to several minerals. Thus, it may bring about a modification in the chemical composition of the water by mineral dissolution or precipitation. Cooling by boiling (adiabatic cooling), on the other hand, causes changes in the composition of ascending water; these changes include degassing and the increase in the solute content as a result of steam loss.

The composition of geothermal waters can also be affected by the process of mixing with cold groundwater on the way to the surface. Since the solubility of most of the compounds in waters increase with increasing temperature, mixing with cold groundwater results in the dilution of geothermal water.

It has to be emphasized here that applications of geothermometry are not simply inserting values into specific geothermometry equations. Interpretation of temperatures obtained from geothermometry equations requires a sound understanding of the chemical processes involved in geothermal systems. The main task of a geochemist is to verify or disprove the validity of assumptions made in using specific geothermometers in specific fields.

## 1.2. TYPES OF CHEMICAL GEOTHERMOMETERS

Chemical geothermometers are divided roughly into two groups:

- silica geothermometers
- cation geothermometers (alkali geothermometers)

The temperature equations related to these geothermometers are given in Tables 1 and 2.

### *1.2.1. Silica Geothermometers*

Silica geothermometers are based on the experimentally determined, temperature-dependent variation of the solubility of silica in water. Because silica occurs in various forms in geothermal fields (such as quartz, cristobalite, chalcedony, amorphous silica) different silica geothermometers have been developed by different workers (Table 1). Different geothermometers are valid in different temperature ranges. This is because they equilibrate at different rates and respond differently to cooling in upflow zones. The quartz geothermometer, for example, works best for well waters with subsurface temperatures above 150°C.

The following should be considered when using silica geothermometers :

- the temperature range in which the equations in Table 1 are valid ,
- effects of steam separation,
- possible precipitation of silica before sample collection (due to silica oversaturation during the ascent of the fluid to the surface) or after sample collection (due to improper preservation of sample),

- the effect of pH on silica solubility,
- possible mixing of hot water with cold water before the thermal water reaches the surface.

### Temperature range of equations

The equations for silica geothermometers are valid for temperature ranges up to 250 °C. Above 250°C, the equations depart drastically from the experimentally determined solubility curves. As can be seen from Fig. 1, amorphous silica is more soluble than quartz at low temperatures, and quartz geothermometer works best for well waters with subsurface temperatures above 150°C.

### Effects of steam separation

As water boils, the silica concentration in the residual liquid increases in proportion to the amount of separated steam. This results in high temperature estimates (overestimated temperatures) during geothermometric applications (note that in the equations in Table 1, greater amount of “C” - due to steam separation - results in smaller denominator, and hence in higher temperature estimates). Therefore, for boiling springs, boiling-corrected geothermometers are used (i.e. Quartz-max. steam loss equation on Table 1, or curve B in Fig.1)

### Precipitation of silica prior to or after sample collection

Quartz solubility increases with increasing temperature (Fig.1). When water ascends from the reservoir to the surface, it may cool and some silica precipitation may occur on the way to the surface. This precipitation results in lower temperature estimates (note that in the equation in Table 1, silica loss via precipitation causes lower “C” values and hence larger denominators, which results in lower temperature estimates)

When a cooling solution becomes saturated with respect to silica (e.g. during sample collection), some of *monomeric silica* (silica with single molecules) usually converts into *polymerized silica* (silica consisting of a number of combined molecules) with silica precipitation. To prevent silica polymerization and precipitation, sample solution must be (5 to 10 times) diluted by silica-free water immediately after sample collection.

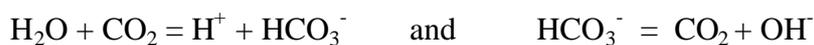
### Effect of pH

The effect of pH on quartz solubility is shown in Fig. 2. As can be seen from the figure, solubilities slightly increase with increasing pH (alkalinity). Also shown by a dashed line, is the pH required to achieve a 10% increase in quartz solubility (at different temperatures), compared to water with pH value of 7. The most pronounced effect is at 175 °C, where a pH of 7.6 will cause 10% increase in silica solubility. However, pH

**Table 1.** Temperature equations for silica geothermometers (C is SiO<sub>2</sub> concentration in mg/l; temperature (T) is in °C).

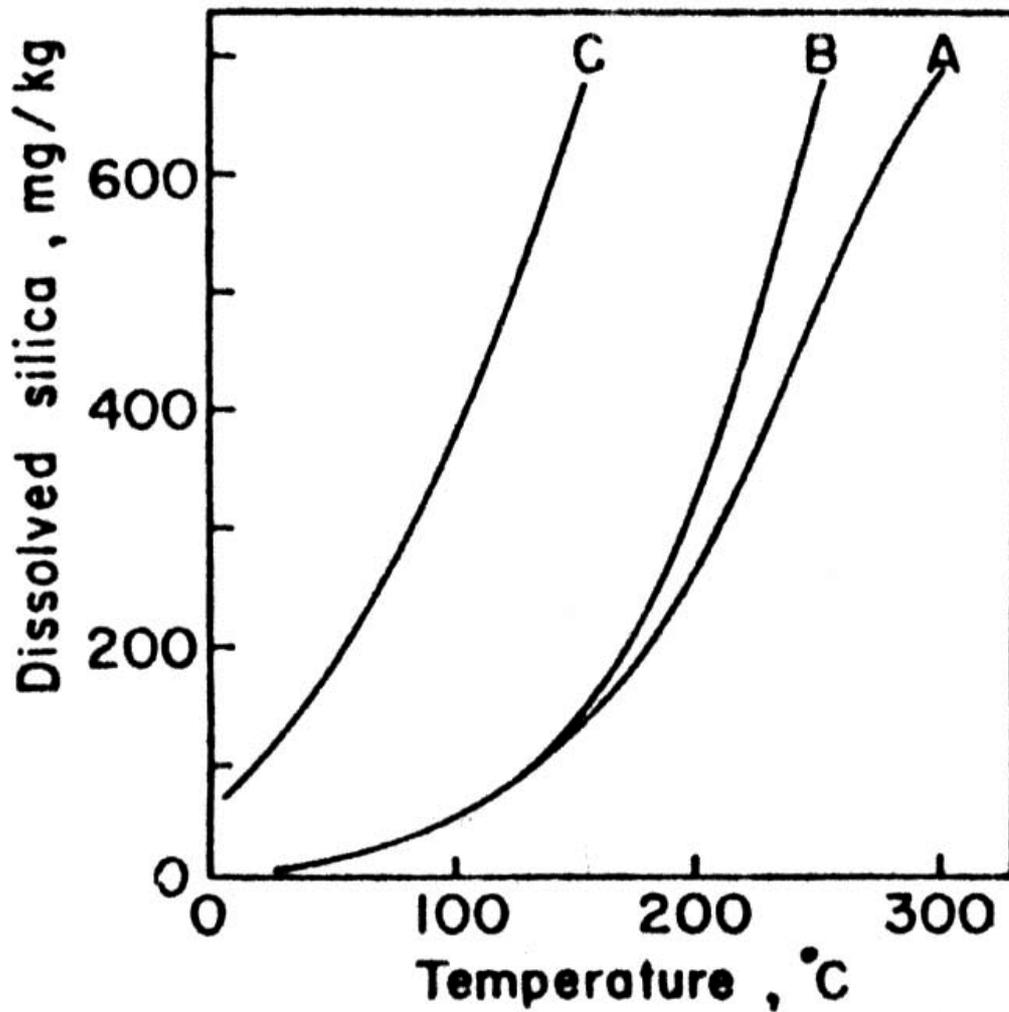
Geothermometer	Equation	Reference
Quartz-no steam loss	$T = 1309 / (5.19 - \log C) - 273.15$	Fournier (1977)
Quartz-maximum steam loss at 100 °C	$T = 1522 / (5.75 - \log C) - 273.15$	Fournier (1977)
Quartz	$T = 42.198 + 0.28831C - 3.6686 \times 10^{-4} C^2 + 3.1665 \times 10^{-7} C^3 + 77.034 \log C$	Fournier and Potter (1982)
Quartz	$T = 53.500 + 0.11236C - 0.5559 \times 10^{-4} C^2 + 0.1772 \times 10^{-7} C^3 + 88.390 \log C$	Arnorsson (1985) based on Fournier and Potter (1982)
Chalcedony	$T = 1032 / (4.69 - \log C) - 273.15$	Fournier (1977)
Chalcedony	$T = 1112 / (4.91 - \log C) - 273.15$	Arnorsson et al. (1983)
Alpha-Cristobalite	$T = (1000 / 4.78 - \log C) - 273.15$	Fournier (1977)
Opal-CT (Beta-Cristobalite)	$T = 781 / (4.51 - \log C) - 273.15$	Fournier (1977)
Amorphous silica	$T = 731 / (4.52 - \log C) - 273.15$	Fournier (1977)

values of most geothermal waters are less than 7.5. Values of pH greater than 7.5 are caused by the loss of CO<sub>2</sub> as shown below:

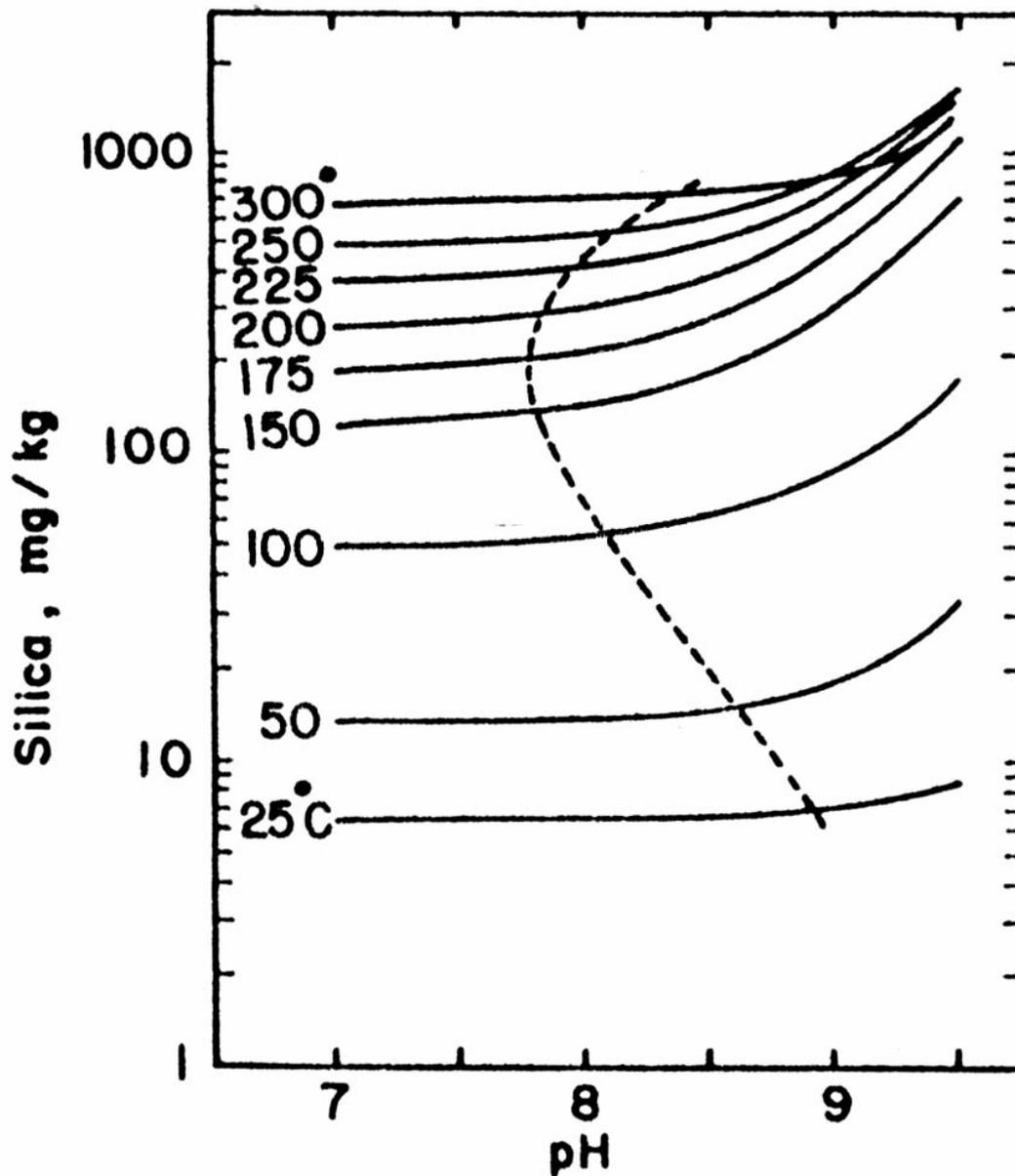


(when CO<sub>2</sub> dissolves in water it forms HCO<sub>3</sub><sup>-</sup> and leaves free H<sup>+</sup> which decreases pH or increases acidity ; when CO<sub>2</sub> is lost, pH of solution increases)

An increase in dissolved silica content, due to increase in pH, will cause overestimates in geothermometry calculations.



**Fig.1.** Solubility of quartz (curve A) and amorphous silica (curve C) as a function of temperature. Curve B shows the amount of silica that would be in solution after an initially quartz-saturated solution cooled adiabatically to 100 °C without any precipitation of silica (from Fournier and Rowe, 1966, and Truesdell and Fournier, 1976).



**Fig. 2.** Calculated effect of pH upon the solubility of quartz at various temperatures from 25 °C to 350 °C , using experimental data of Seward (1974). The dashed curve shows the pH required at various temperatures to achieve a 10% increase in quartz solubility compared to the solubility at pH=7.0 (from Fournier, 1981).

### Effects of dilution (mixing)

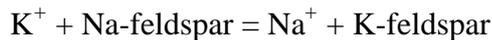
Owing to the fact that silica solubility increases with increasing temperature, hot waters have higher silica concentrations compared to cold waters. If (ascending) hot water mixes with cold groundwater, silica concentration decreases (denominator of equation in Table 1 will be higher) and results in lower temperature estimates (underestimated temperatures).

### *1.2.2. Cation Geothermometers*

Cation geothermometers are based on the partitioning of alkalis between solid and liquid phases, and are also called as *alkali geothermometers*. The majority of these geothermometers are empirically developed geothermometers. The temperature equations are given in Table 2.

### Na/K Geothermometer

The Na/K geothermometer is based on the temperature dependent partitioning of sodium and potassium between aluminosilicates and solutions:



The equilibrium constant,  $K_{\text{eq}}$ , for this partitioning is expressed as :

$$K_{\text{eq}} = [\text{Na}^+] / [\text{K}^+]$$

Variation of the equilibrium constant is temperature-dependent and is expressed by the Van't Hoff equation:

$$d \log K_{\text{eq}} / d(1/T) = (-\Delta H^\circ_T) / R$$

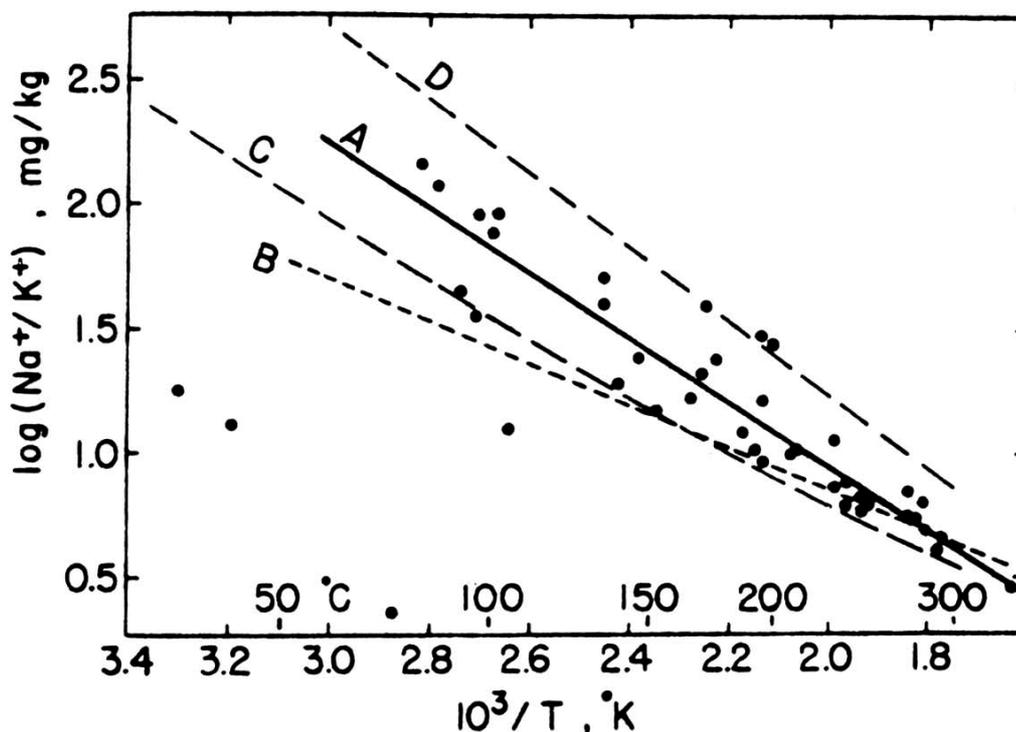
A plot of  $\log [\text{Na}^+] / [\text{K}^+]$  vs.  $1/T$  is given in Fig. 3 together with several empirical curves developed by using the analytical data from geothermal wells throughout the world. The equation of curve A in Fig. 3, is equation no. 5, and that of B is equation no.1 in Table 2.

Compared to the silica geothermometers, the Na/K geothermometer is less affected by processes like boiling and mixing; this is because ratios are used in this geothermometer (rather than absolute abundances). The Na/K geothermometer,

- usually gives good results for reservoir temperatures above 180°C,
- yields erroneous estimates for low-temperature waters, because temperature-dependent exchange equilibrium between feldspars and geothermal waters is not attained at low temperatures and the Na/K ratio in these waters are governed by leaching rather than chemical equilibrium,
- yields unusually high estimates for waters having high calcium contents; For Ca-rich waters (can be guessed from the travertine deposition around), the Na-K-Ca geothermometer is developed.

**Table 2.** Temperature equations for cation geothermometers (concentrations are in mg/l, temperature (T) is in °C).

Geothermometer	Equations	Reference
1. Na/K	$T=[855.6/(0.857+\log(\text{Na/K}))]-273.15$	Truesdell (1976)
2. Na/K	$T=[833/(0.780+\log(\text{Na/K}))]-273.15$	Tonani (1980)
3. Na/K	$T=[933/(0.993+\log(\text{Na/K}))]-273.15$ (25-250 °C)	Arnorsson et al. (1983)
4. Na/K	$T=[1319/(1.699+\log(\text{Na/K}))]-273.15$ (250-350 °C)	Arnorsson et al. (1983)
5. Na/K	$T=[1217/(1.483+\log(\text{Na/K}))]-273.15$	Fournier (1979)
6. Na/K	$T=[1178/(1.470+\log(\text{Na/K}))]-273.15$	Nieva and Nieva (1987)
7. Na/K	$T=[1390/(1.750+\log(\text{Na/K}))]-273.15$	Giggenbach (1988)
8. K-Mg	$T=[4410/(14.00-\log(\text{K}/\sqrt{\text{Mg}}))]-273.15$	Giggenbach (1988)
9. Na-K-Ca	$T=[1647/(\log(\text{Na/K})+\beta(\log(\sqrt{\text{Ca}}/\text{Na})+2.06)+2.47)]-273.15$ a) if $(\log(\sqrt{\text{Ca}}/\text{Na})+2.06) < 0$ , use $\beta=1/3$ and calculate T b) if $(\log(\sqrt{\text{Ca}}/\text{Na})+2.06) > 0$ , use $\beta=4/3$ and calculate T c) if calculated $T > 100^\circ\text{C}$ in (b), recalculate T using $\beta=1/3$	Fournier and Truesdell (1973)
10. Na-K-Ca-Mg	$T = T_{\text{Na-K-Ca}} - \Delta t_{\text{MG}}^\circ\text{C}$ $R = (\text{Mg} / \text{Mg} + 0.61\text{Ca} + 0.31\text{K}) \times 100$ <b>if R from 1.5 to 5</b> $\Delta t_{\text{MG}}^\circ\text{C} = -1.03 + 59.971 \log R + 145.05 (\log R)^2 - 36711 (\log R)^2 / T - 1.67 \times 10^7 \log R / T^2$ <b>if R from 5 to 50</b> $\Delta t_{\text{MG}}^\circ\text{C} = 10.66 - 4.7415 \log R + 325.87 (\log R)^2 - 1.032 \times 10^5 (\log R)^2 / T - 1.968 \times 10^7 (\log R)^3 / T^2$ <b>Note:</b> Do not apply a Mg correction if $\Delta t_{\text{MG}}$ is negative or $R < 1.5$ . If $R > 50$ , assume a temperature = measured spring temperature. T is Na-K-Ca geothermometer temperature is Kelvin.	Fournier and Potter (1979)



**Fig. 3.** Na/K atomic ratios of well discharges vs reciprocal measured downhole temperatures. Curve A is the least square fit of the data points above 80 °C. Curve B is another empirical curve (from Truesdell, 1976). Curves C and D show the approximate locations of the low albite-microcline and high albite-sanidine lines derived from thermodynamic data (from Fournier, 1981).

#### Na-K-Ca Geothermometer

Na-K-Ca geothermometer reflects the equilibria between feldspars (Na & K-feldspars), calcite or Ca-bearing minerals and geothermal waters.

The procedure to use this geothermometer is as follows:

- using mg/l values, calculate  $[\log \sqrt{Ca}/Na) + 2.06]$
- if  $[\log \sqrt{Ca}/Na) + 2.06] < 0$ , use  $\beta = 1/3$  and calculate T°C
- if  $[\log \sqrt{Ca}/Na) + 2.06] > 0$ , use  $\beta = 4/3$  and calculate T°C
- if calculated T > 100°C, recalculate T°C using  $\beta = 1/3$

For CO<sub>2</sub> rich or Ca-rich environments, this geothermometer works well, provided that calcite is not deposited after the water left the reservoir (otherwise, Ca concentration used in the temperature equation - in Table 2 - will be less than the true concentration;

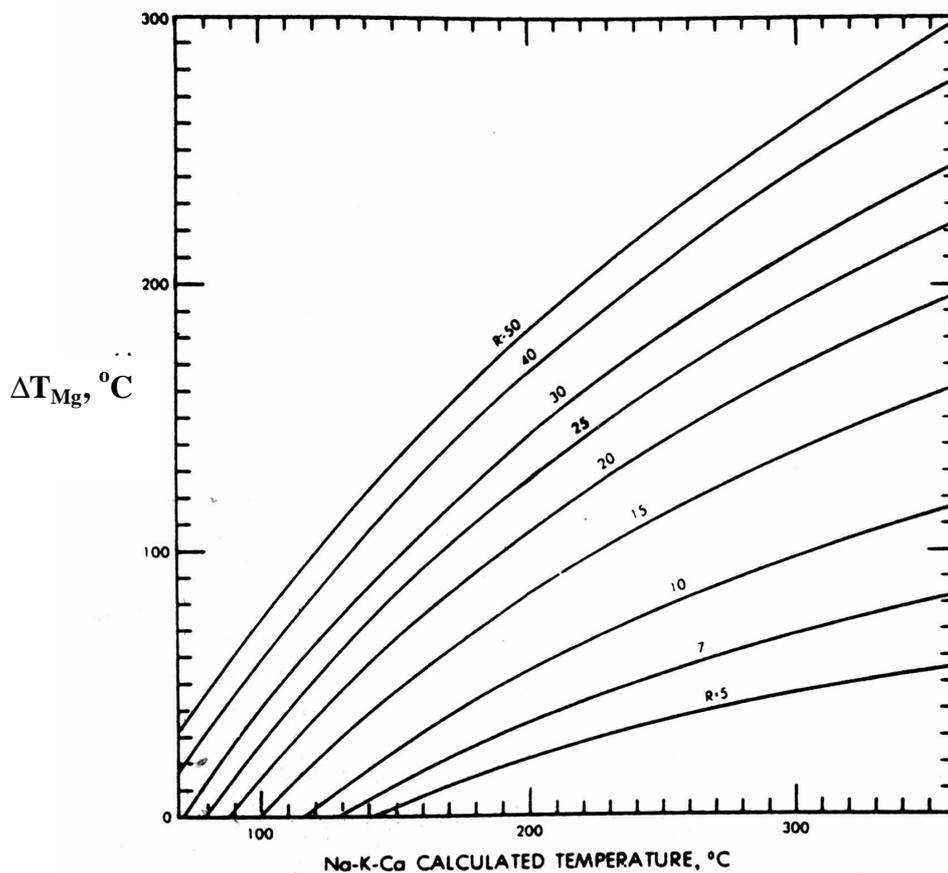
i.e. the denominator of the equation will decrease and hence the temperature will be overestimated).

For waters with high Mg contents, Na-K-Ca geothermometer yields erroneous results. For these waters, Mg correction is necessary and this is done using the Na-K-Ca-Mg geothermometer.

#### Na-K-Ca-Mg Geothermometer

The Na-K-Ca-Mg geothermometer is a version of the Na-K-Ca geothermometer which is corrected for high Mg contents.

The Mg correction to the Na-K-Ca-Mg geothermometer is done using Fig. 4 which gives the Mg-correction factor ( $\Delta T_{Mg}$ ) that should be subtracted from Na-K-Ca-Mg temperatures. Note that  $R=100Mg/(Mg+Ca+K)$  uses *concentrations* expressed in *molality / ionic charge*, i.e., *equivalents*.



**Fig. 4.** Graph for estimating the magnesium temperature correction to be subtracted from Na-K-Ca calculated temperature.  $R = 100 Mg / (Mg + Ca + K)$  with concentrations expressed in equivalents (from Fournier, 1981).

### Other Cation Geothermometers

Various other cation geothermometers are suggested for use in some specific cases. Included in these are the Na/Li geothermometer (Fouillac and Michard, 1981), which is based on the cation exchange reactions with clay and zeolite minerals, rather than the equilibria between the solution and the Li-bearing discrete minerals. The K-Mg geothermometer (Giggenbach, 1988) is based on the equilibrium with the mineral assemblage K-feldspar, chlorite and muscovite.

#### *1.2.3. Silica-Enthalpy Mixing Model*

Under some circumstances the dissolved silica content of mixed waters can be used to determine the temperature of the hot-water component in mixture (Fournier, 1977). The simplest method is to plot dissolved silica against enthalpy of liquid water (Fig.5). Although temperature is the measured property, and enthalpy is a derived property, enthalpy is used as a coordinate rather than temperature. This is because the combined heat contents of two waters are conserved when those waters are mixed but the combined temperatures are not. The enthalpy values (at the corresponding discharge temperatures) are obtained from Steam Table (Henley et al., 1984).

In silica-enthalpy diagram, the reservoir temperature can be estimated from the intersection of the mixing line (joining cold water to hot water sample) with the quartz solubility curve (Fournier, 1977).

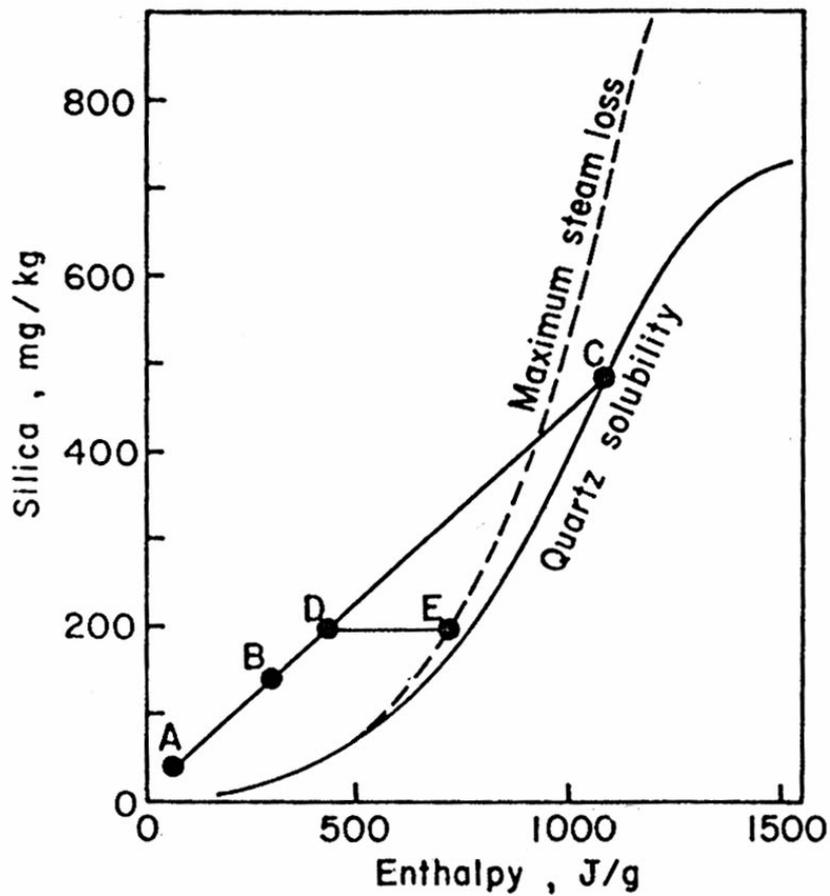
The procedure to estimate the reservoir temperature (using Fig. 5) is described below:

- The sample points (A, B, D) are plotted.
- A straight line is drawn from the point representing the non-thermal component of the mixed water (i.e. the point with the lowest temperature and the lowest silica content: point A in Fig. 5), through the mixed warm water springs (points B and D in Fig.5).
- The intersection of this line with the quartz solubility curve (point C in Fig.5) gives the enthalpy of the hot-water component (at reservoir conditions).
- From Steam Tables, the temperature corresponding to this enthalpy value is obtained as the reservoir temperature of the hot-water component.

This procedure assumes that the steam fraction did not separate (boiling did not occur) before mixing. If the steam separation (boiling) occurs before mixing, then the procedure to follow is as follows:

- The enthalpy at the boiling temperature (100°C) is obtained from the Steam Table (which is 419 J/g).
- A vertical line is drawn from the enthalpy value of 419 J/g.
- From the intersection point of this line with the mixing line (Line AD), a horizontal line (DE) is drawn.

- The intersection of line DE with the solubility curve for maximum steam loss (point E) gives the enthalpy of the hot-water component.
- From the Steam Table, the reservoir temperature of the hot-water component is determined.



**Fig. 5.** Dissolved silica-enthalpy diagram showing procedure for calculating the initial enthalpy (and hence the reservoir temperature) of a high-temperature water that has mixed with a low-temperature water (from Fournier, 1981).

In order to yield accurate results for the silica mixing model, it is vital that conductive cooling does not occur after mixing. If conductive cooling occurs after mixing, then the calculated temperatures would be too high (overestimated temperatures). This is because:

- The original points before conductive cooling would lie to the right of the line AD (i.e. towards the higher enthalpy values at the same silica concentrations, and conductive cooling only affects the temperatures, but not the silica contents).
- The intersection of mixing line with the quartz solubility curve yields lower enthalpy values (i.e. lower temperatures) than that obtained by conductive cooling. In other words, the temperatures obtained in case of conductive cooling are higher than the actual reservoir temperatures (i.e. if conductive cooling occurs after mixing, the temperatures are overestimated).

Another requirement for the use of enthalpy-silica model is that no silica deposition occurs before or after mixing. If silica deposition occurs, the temperatures are underestimated. This is because:

- The original points before silica deposition tend towards higher silica contents (at the same enthalpy values).
- The intersection of mixing line with the silica solubility curve is on higher enthalpy values (higher temperatures) than that obtained in case of silica deposition. In other words, the temperatures obtained in case of no silica deposition are higher than those in case of silica deposition (i.e. the temperatures are underestimated in case of silica deposition).

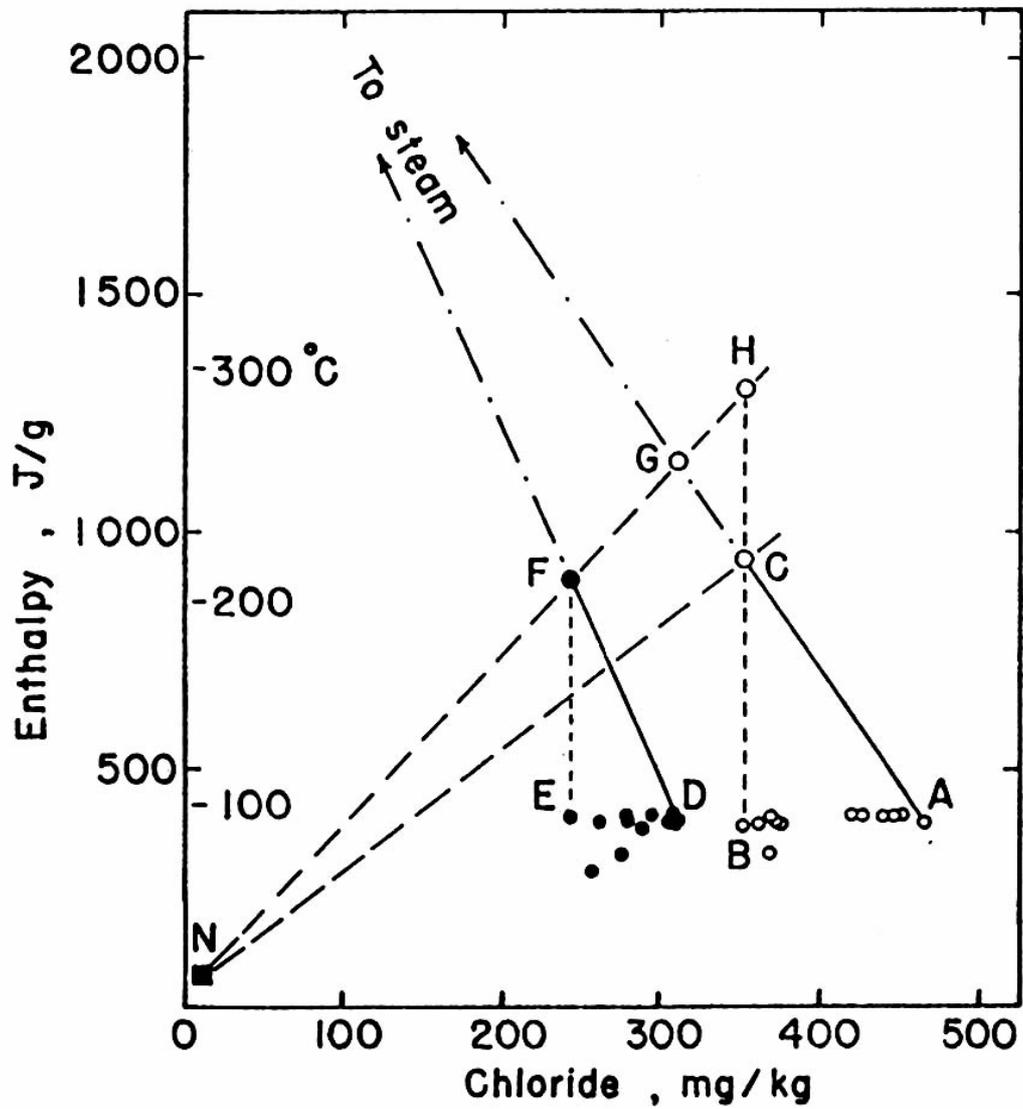
#### *1.2.4. Chloride-Enthalpy Mixing Model*

The enthalpy-chloride diagram can be used to estimate the reservoir temperatures when boiling yields a range of variations of Cl contents (Truesdell ve Fournier, 1975). Fig. 6 shows such a diagram for waters from Yellowstone National Park.

There are two groups of waters in Fig.6 : Geyser Hill-type and Black Sand-type. The enthalpy values used in the diagram are those corresponding to the measured temperatures of the waters.

The temperature of the reservoir feeding the Geyser Hill waters is determined by

- drawing a straight line from the water having maximum Cl content (point A) to the enthalpy of steam at 100°C,
- extending a vertical line from the water with the least Cl content (Point B),
- finding the intersection of these two lines (Point C) which gives the minimum enthalpy at the reservoir ( $936 \text{ J/g} = 218 \text{ }^\circ\text{C}$ ).



**Fig. 6.** Enthalpy-chloride diagram for waters from Yellowstone National Park. Small circles indicate Geyser Hill-type waters and small dots indicate Black Sand-type waters (From Fournier, 1981).

The temperature of the reservoir feeding the Black Sand waters (E to D) is determined using the same procedure described above for Geysir Hill waters (reservoir temperature for Black Sand waters is 209 °C, point F in Fig.6)

Chloride-enthalpy mixing model can also be used to assess the possible origins of waters. For example, concerning the relationship between Geysir Hill and Black Sand waters, the following points are worth to note:

- Although point F represents a more dilute and slightly cooler water than the water represented by point C, the former can not be derived from the latter by a mixing process between hot and cold water (points C and N, respectively), because any mixture would lie on or close to line CN.
- Waters C and F are probably both related to a still higher enthalpy water such as point G or H.
- Water F could be related either to water G or H by mixing in different proportions by water N.
- Water C could be related to water G by boiling (adiabatic cooling) (temperature/enthalpy decreases, whereas Cl concentration increases with boiling).
- Water C could also be related to H by conductive cooling of a slow moving solution (temperature decreases with no change in Cl content).

## 2. EXERCISE

### PROBLEM 1.

Given below are the chemical compositions (in mg/l) of well waters from Tuzla-Çanakkale (sample no.1) and Ömer-Gecek–Afyon (sample no.2) geothermal fields in western Anatolia.

Sample No.	T°C	pH <sub>lab</sub>	K	Na	Ca	Mg	HCO <sub>3</sub>	CO <sub>3</sub>	SO <sub>4</sub>	Cl	SiO <sub>2</sub>	B <sub>T</sub>
1	*171	7.40	1280	16640	2920	71	84	0	150	33600	94	22.8
2	+92	6.70	158	1750	78.8	10	1294	0	537	1862	198	10.0

\* Bottom-Hole temperature

+ Well-head temperature

- a) Estimate the reservoir temperatures for both fields by using the quartz, quartz-with max. steam loss, chalcedony, Na/K, and N-K-Ca geothermometers.

NOTE: use Fournier (1981) equations for quartz and quartz-with max. steam loss, Fournier (1977) equation for chalcedony, Arnorsson et al. (1983) equation for Na/K, and Fournier and Truesdell (1973) equation for Na-K-Ca geothermometers (see Tables 1 and 2).

- b) Which one(s) of the geothermometers is (are) likely to yield the most representative reservoir temperature for each geothermal field? Discuss your reasoning separately for each field (please note that widespread travertine deposition occurs in the Ömer-Gecek geothermal field of Afyon).

#### PROBLEM 2.

Given below are the silica concentrations and the discharge temperatures of the spring and well waters from the Seferihisar geothermal field in western Anatolia.

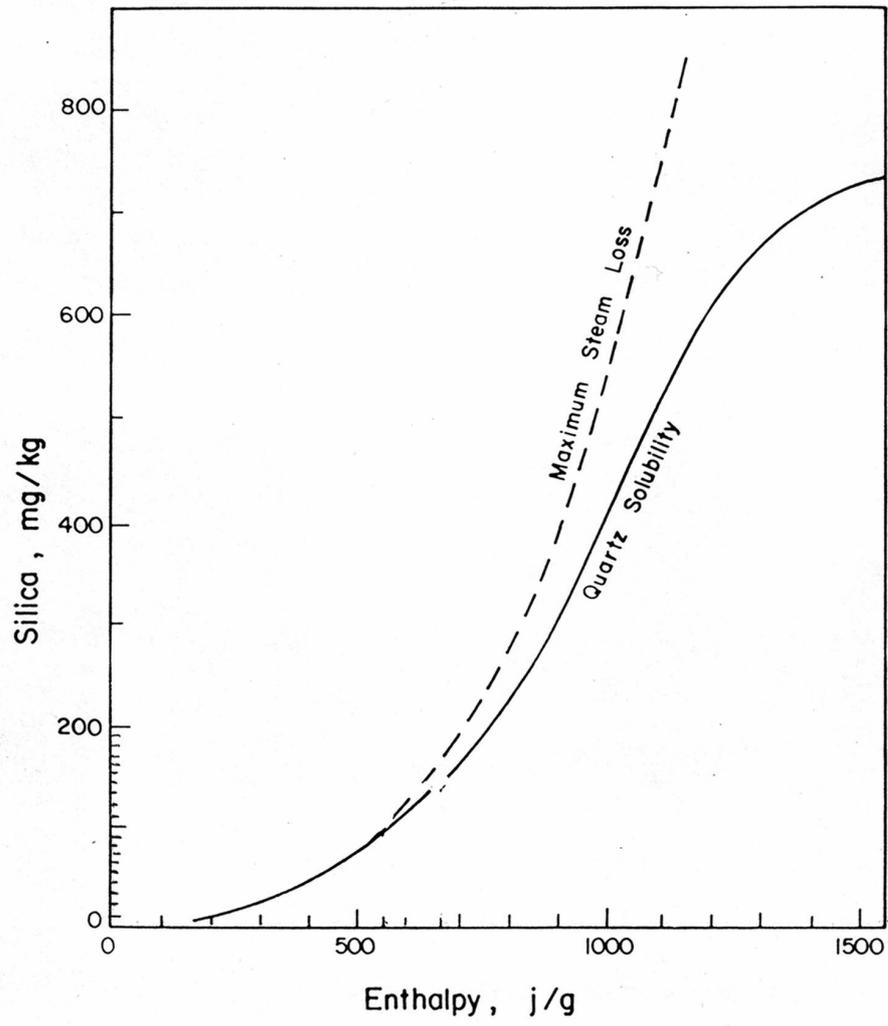
Sample (w: well, s: spring)	Discharge temperature (°C)	SiO <sub>2</sub> (mg/l)
S-2 (s)	7	23
S-5 (s)	36	73
S-10 (s)	72	105
S-12 (s)	60	72
TS-2 (w)	*92	123

\*well head temperature

- a) Estimate the reservoir temperature with the help of the silica-enthalpy model (use the attached base diagram containing the quartz solubility curves and the Steam Table in Henley et al., 1984).
- b) The application of the quartz and chalcedony geothermometers to these samples yields the following reservoir temperatures(°C):

Sample no	Qtz	Qtz – max.steam loss	chalcedony
S-2	68.78	73.68	36.92
S-5	120.33	118.44	91.94
S-10	139.94	135.02	113.54
S-12	119.63	117.84	91.17
TS-2	149.09	142.69	123.76

Discuss the reasons for the differences between estimates from silica geothermometers and the silica-enthalpy model.



**BASE DIAGRAM to be used for Problem 2**

**PROBLEM 3.**

Given below are the temperatures and the chloride contents of the spring and well waters from the Kızılcahamam geothermal field.

Sample	Big Bath (BB) (Spring)	MTA-1 (Well)	Small Bath (SB) (Spring)	Mineral Water (MW) (Spring)	Cold Spring (CS)	KHD-1 (Well)
*T°C	66	80	71	28	9	106
Cl (mg/l)	272	272	266	494	5	310

\*Temperatures are discharge temperatures for springs and bottom-hole temperatures for wells (at 180 m and 1556 m depths for MTA-1 and KHD-1 wells, respectively)

**a)** State what is the most likely process (boiling?, mixing?, conductive cooling?) responsible for the mineral water to have higher concentrations in almost all the chemical constituents yet with lower discharge temperature compared to other waters ? Explain your arguments.

**b)** According to your answer under (a) state, what sort of a (further ?) process might have been responsible for the low silica content of mineral water which have considerably high concentrations of other chemical constituents. Explain your arguments.

**c)** Using a graph paper plot your samples on a chloride-enthalpy model and then follow the procedure given below :

1. Assume that warm-spring water is produced by conductive cooling of hot-water at 100°C, show the position of the warm-spring water before conductive cooling in the diagram and mark this point as “A”,
2. Assume that the hot-water “A” (at 100°C) represents a residual liquid left after boiling of an even hotter “reservoir fluid”, draw a boiling line connecting “A” to the enthalpy of steam at 100°C,
3. Assuming that the hot-springs and the well water in this geothermal field are products of mixing between this “reservoir fluid” and the water from a “cold-spring”, draw the possible mixing line in the diagram,
4. Using the intersection of this mixing line with the boiling line, estimate the reservoir temperature,
5. Assume that the Kızılcahamam hot waters represent products of conductive cooling of mixed waters at 100 °C, estimate the range of possible reservoir temperatures (hint: project the data points upwards to the enthalpy values at 100 °C; draw possible mixing lines through the projected points; join these mixing lines to the boiling line)

**d)** Given that the geothermal gradient measured in KHD-1 well is 0.38 °C / 10 m, estimate the depth at which a reservoir with a temperature you have obtained in (4) is likely to be encountered.

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## SOLUTIONS FOR PROBLEMS

### PROBLEM 1.

a)

<i>Type of Geothermometer</i>	<i>Tuzla-Çanakkale (T°C)</i>	<i>Ömer-Gecek-Afyon (T°C)</i>
Quartz	133.77	179.27
Quartz-max. Steam loss	129.83	167.58
Chalcedony	106.70	158.05
Na/K	169.67	184.79
Na-K-Ca	186.40	195.50

b)

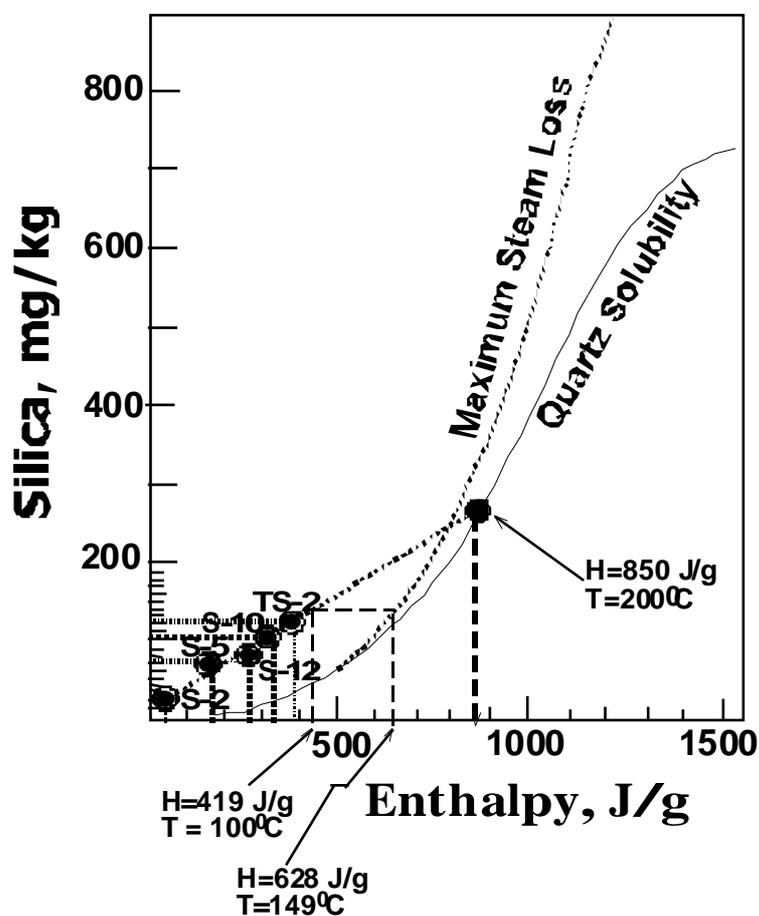
#### *Tuzla-Çanakkale*

Na/K and Na-K-Ca geothermometers are likely to yield the most representative reservoir temperatures as the others yield temperatures which are considerably below the measured bottom-hole temperature.

#### *Ömer-Gecek-Çanakkale*

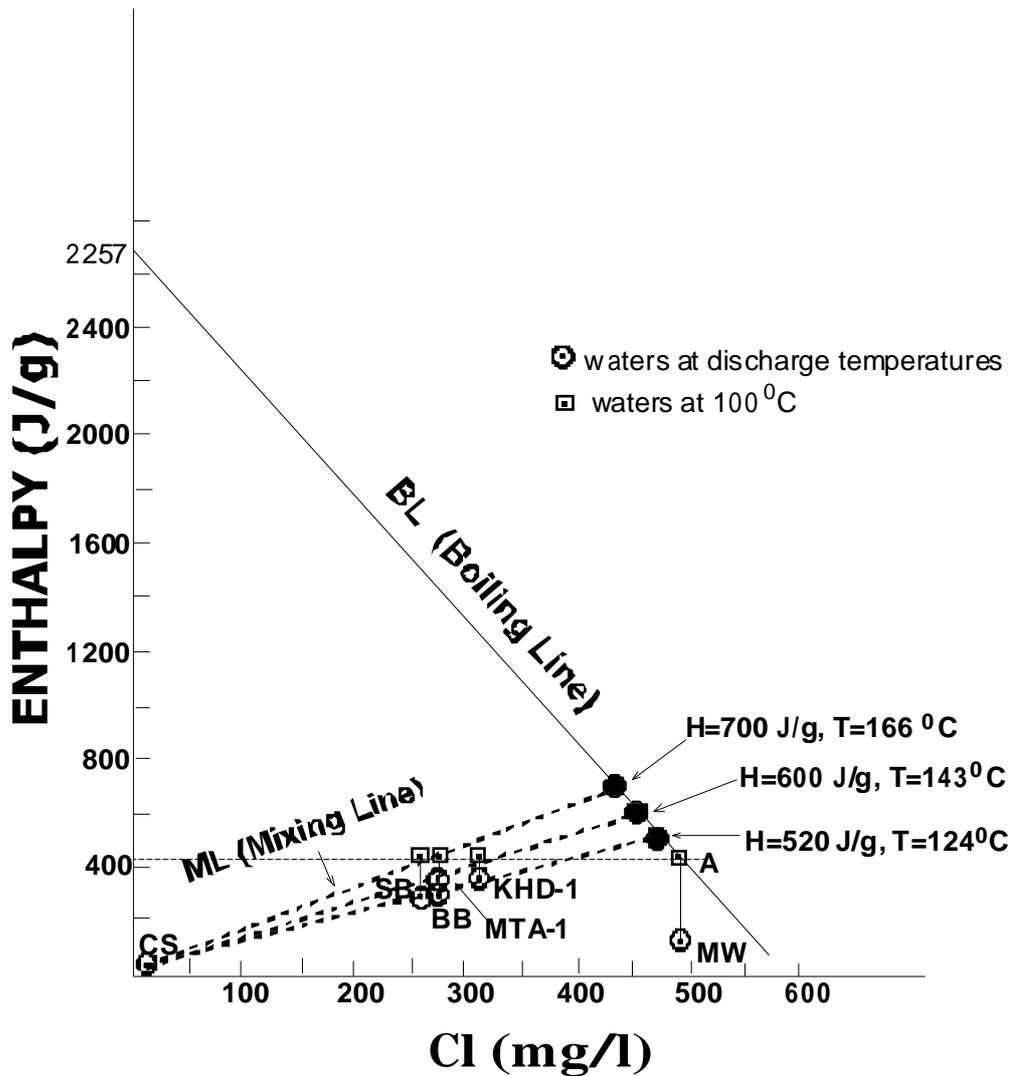
We know that calcite precipitation (travertine deposition) occurs in the field. Therefore, the Ca concentration used in the equation is supposed to be lower than the actual concentration. Hence, the temperature estimate by Na-K-Ca geothermometer is supposed to be an overestimate. In this respect, the real reservoir temperature should be below 195.5 °C. Na/K and silica geothermometers are, therefore, more reliable and suggest a temperature range of 158 to 185 °C for the reservoir.

## PROBLEM 2.



- Silica-enthalpy mixing model yields a reservoir temperature of 199-200 °C, if no steam separation before mixing is assumed, and of 149 °C, if steam separation is assumed to have occurred before mixing.
- Chalcedony and quartz geothermometers give lower temperature estimates than the silica-enthalpy mixing model. The reason for this is that these waters are mixed waters (i.e. the silica concentration is less than the concentration in reservoir conditions). Hence, silica geothermometers yield underestimated temperatures.

## PROBLEM 3.



a) The most likely process is boiling. In this way, the mineral water (representing the residual liquid after steam separation) becomes concentrated and cooled adiabatically.

b) The mineral water, after steam separation, might have lost its (enriched) silica by silica precipitation on the way to surface (as a result of adiabatic enrichment, water may have been oversaturated with respect to silica and – given adiabatic cooling which decreased silica solubility – the excess silica may have been deposited on the migration path of the mineral water.

c)

1. See the diagram.
2. See the diagram.
3. See the diagram.
4. 124 °C
5. 143 – 166 °C
6. gradient = 0.38 °C / 10 m

$$124 - 106 = 18 \text{ °C}$$

$$0.38 \text{ °C} \dots\dots\dots 10 \text{ m}$$

$$18 \text{ °C} \dots\dots\dots 421 \text{ m}$$

$$421 + 1556 = 1977 \text{ m.}$$