



INTERPRETATION OF CHEMICAL COMPOSITION OF GEOTHERMAL FLUIDS FROM ÁRSKÓGSSTRÖND, DALVÍK, AND HRÍSEY, N-ICELAND AND IN THE KHANGAI AREA, MONGOLIA

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ABSTRACT

Data from four production wells in three low-temperature geothermal fields, Árskógsströnd, Dalvík and Hrísey in N-Iceland and 7 hot springs in the Khangai low-temperature geothermal area in Mongolia are presented. The chemical composition of these fluids was analysed by standard methods. The ternary Cl-SO₄-HCO₃ scheme was used to classify the geothermal fluids with respect to major anion composition and the ternary Na-K-Mg diagram was used to classify waters according to the state of equilibrium at given temperatures. The geothermal fluids of Árskógsströnd and Dalvík are of bicarbonate-sodium water type and the geothermal fluid of Hrísey is of sodium-chloride water type. The hot springs in the Khangai area are of bicarbonate-sodium water type except the Shivert hot spring, which is of sulfate-sodium type. Chemical geothermometers were used to estimate reservoir temperatures. Values for chalcedony geothermometer and Na/K geothermometer values are quite different. Mixing models were used to try to define the source temperature of the geothermal fluid component. The Ryznar stability index (RSI) and WATCH programs were used to interpret the equilibrium state of the reservoirs and to predict calcite scaling tendencies. The Hrísey geothermal fluid (HR-10) has a slight tendency for scaling, but this is not the case for the other geothermal fluids considered here.

1. INTRODUCTION

In this report, chemical analyses for geothermal fluids from the Árskógsströnd, Dalvík, and Hrísey geothermal fields, N-Iceland (Figure 1), and selected hot springs from the Khangai area in central Mongolia, were used to evaluate the probable existing chemical equilibria and to estimate subsurface temperatures in the geothermal systems. The temperatures of the geothermal fluids from fields in N-Iceland are in the range 64- 78°C and the temperatures of the Mongolian hot springs range between 25 and 88°C. The mixing processes in the up-flow zones were assessed using two mixing models. The subsurface temperatures predicted by various geothermometers are evaluated by comparison with measured (downhole) temperatures. Most calculations were performed with the SOLVEQ (Reed and Spycher, 1984) and WATCH (Arnórsson et al., 1983a; Bjarnason, 1994) programs. Finally, the WATCH program and Ryznar stability index (RSI) are used to evaluate calcium carbonate scaling tendency in the

hot springs in Mongolia and the Icelandic wells.

The Hamar at Dalvík (wells HA-10 and HA-11), the Árskógsströnd (well ÁRS-29) and Hrísey (well HR-10) geothermal systems are three separate systems, each with its own characteristics. Árskógsströnd is a small rural community of 300 people, on the western side of Eyjafjörður, a long fjord in N-Iceland, Dalvík is a small town of 2,000 inhabitants at the western coast of Eyjafjörður, Hrísey is an island off the coast of Eyjafjörður with nearly 300 inhabitants.

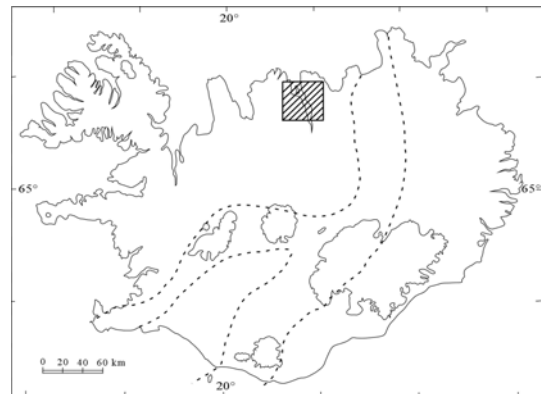


FIGURE 1: Map of Iceland and the study area in N-Iceland

Mongolia is located in the northern part of central Asia, far from the oceans, on a high plateau surrounded by mountain ridges. It shares a 3,005 km long border with Russia in the north and a 4,673 km long border with China in the south. Mongolia covers a vast territory of over 1.5 million km² and has a population of 2.5 million. Mongolia has 43 hot springs, with measured surface temperatures ranging from 20 to 92°C, mainly distributed in the central and western provinces (Popov, 1949). Mongolian hot springs are divided into the following three areas by distribution and hydrogeological characteristics: 1) Altai Soyonii area with four hot springs of sulfate-bicarbonate-sodium type, with measured surface temperatures of 25-32°C and flow rates of 0.5-3 l/s; 2) Khangai area with 36 hot springs of bicarbonate-sulfate-sodium type; and 3) Khentii area with 3 hot springs of bicarbonate-sulfate-sodium type and temperatures in the range 67-88°C (see Figure 2).

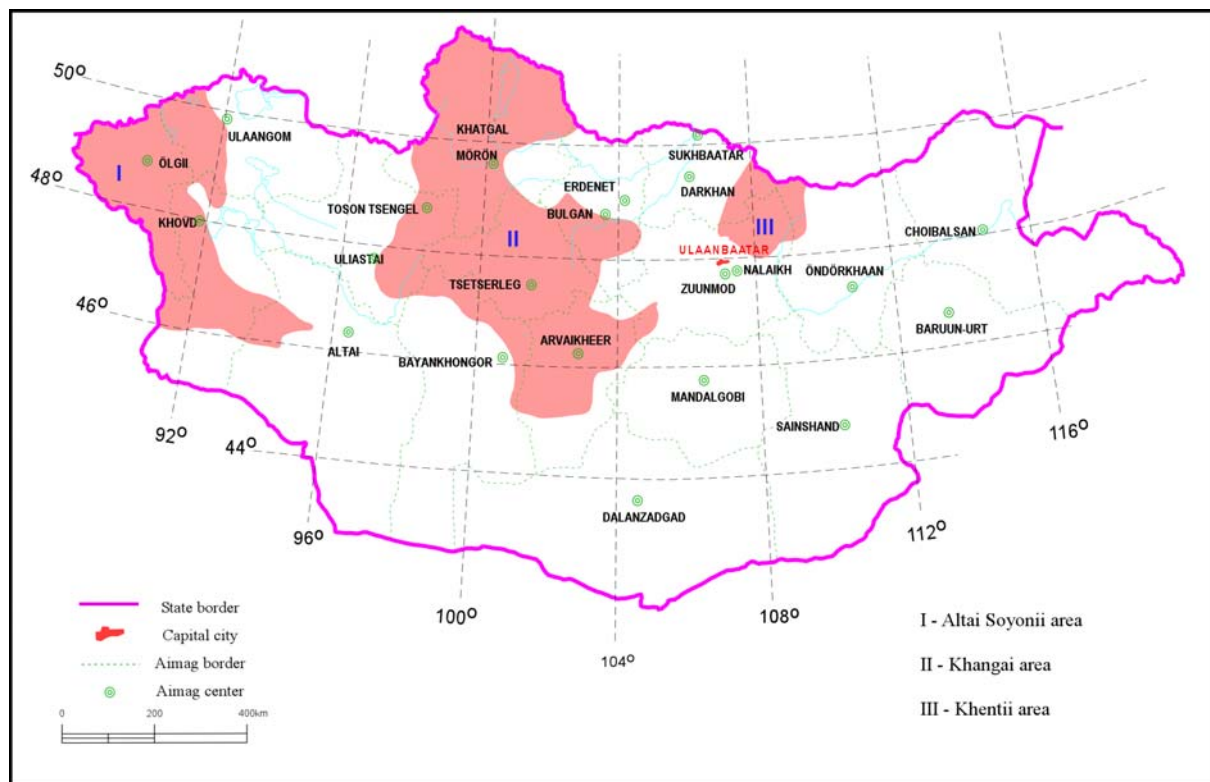


FIGURE 2: The classification of hot spring areas in Mongolia (Namnandorj, 1960)

2. SAMPLING AND ANALYSIS OF GEOTHERMAL FLUIDS

In the current study, data from 32 geothermal fluid samples from N-Iceland were used and 15 samples from the seven main hot springs of the Khangai area. The Icelandic samples were collected between 1977 and 2001 and the chemical data were taken from the ISOR (formerly Orkustofnun) database. The Mongolian samples were collected in August 2002. The samples were analysed by field and laboratory methods summarized in Tables 1 and 2.

The chemical composition of geothermal waters of Árskógsströnd, Dalvík, and Hrísey were determined in the chemical laboratory of ISOR, Iceland; and chemical compositions of hot springs of Khangai were determined in the hydrological laboratory of the Institute of Chemistry and Chemical Technology, Mongolia. Description of methods is mostly based on Greenberg et al. (1980).

TABLE 1: Sample volumes, sample treatment and analytical methods for geothermal water samples in the Khangai area, Mongolia (Mongolian National Standard, 1984)

Constituent	Sample volume (ml)	Sample treatment	Analytical method
pH	200	None	pH-meter
H ₂ S	100	None	Titration on site with 0.001M (CH ₃ COO) ₂ Hg solution using dithizone to indicate end point
CO ₂	200	Add 0.2 mg salt of segnetov to 200 ml	Titration with 0.1N NaOH solution using 1ml indicator phenolphthalein to indicate end point
SiO ₂	200	Add 2 ml concentration HCl and 3 ml molybdate solution to 100 ml	Spectrophotometer
SO ₄	200	Add 2 ml HCl (1:1) solution, methyl orange indicator solution and 5 ml 5% barium chloride solution to 200 ml	Precipitated sediment will be analyzed by gravimetric method
Na, K	200	Filtered, acidified with 0.2 ml concentration HCl to 200 ml sample	Flame emission photometric method
Ca	200	Add 2N 2ml NaOH solution to 100 ml	Titration with 0.1N Trilon-B solution using indicator murexide to indicate end point
Mg	200	Add 5ml ammonia buffer solution to 100 ml	Titration with 0.1N Trilon-B solution using indicator eriochrome to indicate end point
Fe	200	Add 5-7 drops HCl (1:5), solution to 100 ml	Titration with 0.1N Trilon-B solution using sulfasalazine acid to indicate end point
Cl	200	Add 0.5ml 10% potassium chromate indicator solution to 100 ml	Titration with standard silver nitrate titrant to indicate end point
F	200	Add 5ml indicator alzarín and 5 ml zirconyl acid solution to 100 ml	Spectrophotometer

3. PRACTICAL APPLICATIONS OF GEOCHEMICAL STUDIES OF GEOTHERMAL FLUIDS

3.1 Silica geothermometry

The earliest attempt to use water chemistry to evaluate subsurface temperature conditions in geothermal systems dates back to about 1960. The first geothermometer ever used for temperature prediction was entirely empirical. Bödvarsson and Pálmason (1961) detected that there was a good relation between silica content and reservoir temperature. Later, several calibrations were proposed for the quartz and chalcedony silica geothermometers. Table 3 shows the equations of quartz and chalcedony geothermometers used in the report.

The silica geothermometer is based on the solubility of quartz and chalcedony, and is widely used to estimate subsurface temperature. Solubility of these minerals generally changes as a function of

TABLE 2: Analytical methods used for the constituents determined

Constituent	ISOR, Iceland		ICCT, Mongolia
	Fraction	Method	Method
pH	Ru	pH-meter	pH-meter
CO ₂	Ru	Electrometric titration	Titrimetric method
H ₂ S	Ru	Titrimetric method	Titrimetric method
SiO ₂	Rd	Spectrophotometer	Titrimetric method
F	Fu	Selective electrode	Spectrophotometer
Cl	Fu	Selective electrode	Argentometric method
SO ₄	Fu/Fp	Selective electrode	Gravimetric method
B	Fu	Spectrophotometer	none
Na	Fa	AAS DA	Flame emission photometric method
K	Fa	AAS DA	Flame emission photometric method
Mg	Fa	AAS DA	Titrimetric method
Ca	Fa	AAS DA	Titrimetric method
Al	Fa	AAS GF	none
Fe	Fa	AAS GF	Titrimetric method
TDS	Fu	Gravimetric	Gravimetric

Fa - Filtered, (0.2µm; 0.45µm) acidified

Fu - Filtered, untreated

Fp - Filtered, precipitated

Ru - Raw, untreated

Rd - Raw, diluted

AAS - Atomic absorption spectroscopy

DA - Direct aspiration

GF - Graphite furnace

ICCT - Institute of Chemistry and Chemical Technology

ISOR - Íslenskar orkurannsóknir (Iceland Geosurvey)

TABLE 3: Temperature equation (in °C) for the silica geothermometer (concentrations in ppm); S stands for the concentration of SiO₂

Geothermometer	Equation	Range (°C)	Source
Quartz	$-42.2 + 0.28831S - 3.6686 \times 10^{-4}S^2 + 0.1772 \times 10^{-7}S^3 + 77.0341 \log S$	25-330	Fournier and Potter (1982)
Quartz	$\frac{1309}{5.19 - \log S} - 273.15$	25-250	Fournier (1977)
Quartz	$-55.3 + 0.3659 S - 5.3954 \times 10^{-4}S^2 + 5.5132 \times 10^{-7}S^3 + 74.3601 \log S$	0-350	Arnórsson et al. (1983b)
Chalcedony	$\frac{1032}{4.69 - \log S} - 273.15$	0-250	Fournier (1977)
Chalcedony	$\frac{1112}{4.91 - \log S} - 273.15$	-	Arnórsson et al. (1983b)

temperature and pressure. At temperatures below 340°C, the solubility of the silica minerals decreases drastically as the temperature decreases.

At temperatures below 300°C and at a depth generally attained by commercial drilling for geothermal resources, variation in pressure has little effect on the solubilities of quartz and other silica polymorphs. Generally speaking, the solubility of quartz appears to control dissolved silica in a geothermal reservoir at temperatures higher than 120-180°C, so the quartz geothermometer probably works best in the temperature ranges of 150-250°C. At lower temperatures, other silica phases such as chalcedony, may

control the concentration of dissolved silica (Fournier, 1977). The reason for this is that even though chalcedony is thermodynamically unstable relative to quartz at any temperature and pressure it forms metastably at low temperatures because of faster reaction kinetics of chalcedony precipitation relative to quartz. In some cases, where water has been in a contact with rock at a given temperature for a relatively long time, quartz may control dissolved silica at temperatures less than 100°C. At lower temperatures this reaction will not occur due to slow kinetics although it is thermodynamically favourable. In other places, chalcedony may control dissolved silica at temperatures as high as 180°C (Arnórsson, 2000).

3.2 Cation geothermometry

Cation geothermometers are widely used to calculate subsurface temperatures of waters collected from hot springs and wells. There are many different geothermometers as shown in Table 4, and it is rare that they all give similar results, especially when applied to thermal fluids. They are based on ion exchange reactions with temperature-dependent equilibrium constants, as for example:



The assumption is that the activities of the solid phases (in this case albite and K-feldspar) are unity and the activities of the dissolved species are equal to their molal concentrations in the aqueous solution. The equilibrium constant K_{eq} is expressed by the ratio of the molalities of the respective ions, such as Na/K. If an exchange reaction involves divalent and monovalent ions, for example K^+ and Mg^{2+} , the equilibrium constant is approximately equal to K/\sqrt{Mg} .

TABLE 4: Temperature equations (in °C) for cation geothermometers

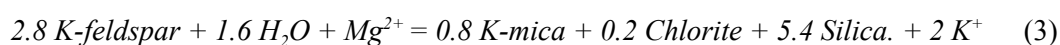
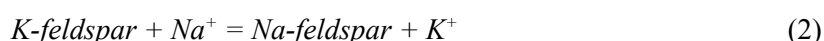
Geothermo- meter	Equation	Range (°C)	Source
Na-K	$\frac{933}{0.993 + \log(Na / K)} - 273.15$	25-250	Arnórsson et al. (1983b)
Na-K	$\frac{856}{0.857 + \log(Na / K)} - 273.15$	100-275	Truesdell (1976)
Na-K	$\frac{1217}{1.438 + \log(Na / K)} - 273.15$	-	Fournier (1979)
Na-K	$\frac{1390}{1.750 + \log(Na / K)} - 273.15$	-	Giggenbach (1988)
Na-K-Ca ^a	$\frac{1647}{\log(Na / K) + \beta [\log(\sqrt{Ca / Na}) + 2.06]} + 2.47 - 273.15$	-	Fournier and Truesdell (1973)

Notes: a) $\beta = 4/3$ for $t < 100^\circ\text{C}$; $\beta = 1/3$ for $t > 100^\circ\text{C}$ or for $\log \sqrt{Ca / Na} < 0$

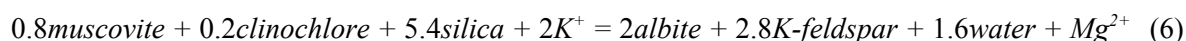
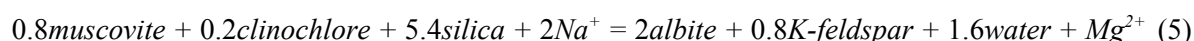
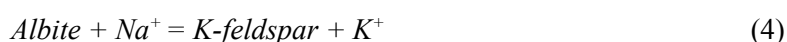
The Na-K-Ca geothermometer (Fournier and Truesdell, 1973) has proved to be more reliable than the Na-K geothermometer for low-temperature waters. It is completely empirical, and based on more than one exchange reaction. It takes into account reactions involving the exchange of Na^+ , K^+ and Ca^{2+} with mineral solid solution. The Na-K-Ca geothermometer appears to give excellent results for most waters above approximately 200°C, but erratic results are obtained for waters from reservoirs at less than 200°C. These erratic results are connected to samples with high CO_2 partial pressure (Paces, 1975) and probably with exchange reaction between Na, K and Ca with additional ions, such as Mg. Precipitation of calcium as carbonate after boiling of solution on the way from reservoir to the surface is another likely source of

error. To avoid the influence of the partial pressure of CO₂ on the Na-K-Ca geothermometer, Paces (1975) recommended a correction. A correction is also suggested for Mg, when this geothermometer is applied to Mg-rich waters (Fournier and Potter, 1979). Generally the Mg concentration in geothermal fluids decreases sharply as the temperature increases, and all Mg-rich fluids found in nature have undergone water-rock reaction at a relatively low temperature.

Ionic solute geothermometers, based on Na-K, Na-K-Ca and K-Mg content, are useful tools for the evaluation of subsurface temperature. Most of the problems in their use arise from their application to unsuitable samples. Recently a “self-policing” technique was devised giving an automatic indication as to the suitability of a given water for the application of ionic solute geothermometers. It is essentially based on the temperature-dependence of the two reactions:



Both equations involve minerals of the full equilibrium assemblage expected to form after the isochemical recrystallization of an average crustal rock under condition of geothermal interest. Na, K and Mg contents of water in equilibrium with this assemblage are accessible to rigorous evaluation. Giggenbach (1988) suggested that a triangular diagram with Na/1000, K/100 and \sqrt{Mg} at the apices could be used to classify waters according to their state of equilibrium at given temperatures (see below). He adopted the Na/K and slightly modified K/Mg geothermometer equations given by Giggenbach et al. (1983). The full equilibrium curve is for reservoir water composition corrected for loss of steam owing to decompressional boiling. Uncorrected boiled water will plot slightly above the full equilibrium line. This overcomes some of the disadvantages of the direct use of Na-K and Na-K-Ca geothermometers where e.g. mixing with an immature water or incomplete equilibration interferes with their use but the magnesium concentration (low in equilibrated geothermal waters) can help in weeding out unsuitable samples. The magnesium correction suggested by Fournier and Potter (1979), has the same aim but is more cumbersome and less successful. The use of the triangular diagram is based on the temperature dependence of three reactions:



They involve minerals of the full equilibrium assemblage after isochemical re-crystallization of an average crustal rock under conditions of geothermal interest. Na, K and Mg concentrations of waters in equilibrium with this assemblage are accessible to rigorous evaluation. The coordinates of a point on the diagram are calculated by

$$S = C_{Na} / 1000 + C_K / 100 + \sqrt{C_{Mg}} \quad (7)$$

$$\%Na = C_{Na} / 10S ; \quad \%K = C_K / S ; \quad \%Mg = \sqrt{C_{Mg}} / S \quad (8)$$

When the Cl-SO₄-HCO₃ ternary diagram (Giggenbach, 1991) is used, geothermal water is classified on the basis of major anions, i.e. Cl, SO₄ and HCO₃. The source of Cl is seawater and rock dissolution; for SO₄, a small amount comes from seawater; but a large amount is usually derived from the oxidation of sulfide in volcanic steam. Carbonate is present at the periphery of high-temperature systems. CO₂ is obtained by the degassing of volcanic formations like deep-seated intrusions. In cold water, carbonate originates from biological sources. High concentrations of sulfate and carbonate, thus obtained, may give erroneous cation geothermometer results. In some old magmatic systems, equilibrium with alteration minerals is established in the presence of excess sulfate in the water and water acidity is not increased. The Giggenbach diagram is obtained by the summation of the concentration of all the three constituents:

$$S = C_{Cl} + C_{SO_4} + C_{HCO_3} \quad (9)$$

and calculating the percentage of individual constituents:

$$\%Cl = 100 C_{Cl} / S; \quad \%SO_4 = 100 C_{SO_4} / S; \quad \%HCO_3 = 100 C_{HCO_3} / S \quad (10)$$

Here, the silica and cation geothermometers and Na-K-Mg and Cl-SO₄-HCO₃ ternary diagrams are used to study reservoir temperatures and reservoir water composition of selected wells and hot springs.

3.3 Mineral equilibrium geothermometry

Reed and Spycher (1984) have proposed that the best estimate of reservoir temperature was attained by considering simultaneously, the state of equilibrium between a specific water, and many hydrothermal minerals as a function of temperature. Equilibrium constants are both temperature and pressure dependent. However, pressure in the range occurring in geothermal systems (0-200 bar) has a very limited effect. On the other hand, equilibrium constants for mineral dissolution often vary strongly with temperature. Therefore, if a group of minerals converges to equilibrium at a particular temperature, this temperature corresponds to the most likely reservoir temperature, or at least the temperature of the source aquifer, for the particular water considered. Reed and Spycher (1984) consider their procedure capable of distinguishing between equilibrated geothermal waters and waters that have departed from equilibrium due to boiling or mixing with shallow colder waters. For the dissolution of a mineral, *m*, the activity quotient, Q_m , can be written

$$Q_m = \prod \frac{a_{i,m}^{v_{i,m}}}{a_m} \quad (11)$$

where $a_{i,m}$ is the activity and $v_{i,m}$ is the stoichiometric coefficient of a species in the mineral *m*, written with the mineral on the left hand side of the reaction equation, and the aqueous components on the right hand side, the activity of mineral *m*, a_m , being equal to one for pure minerals.

The free energy, ΔG , of chemical reaction including mineral dissolution is given by

$$\Delta G_m = \Delta G_{m,r}^{\circ} + RT \ln Q_m \quad (12)$$

and the stoichiometric coefficients $v_{i,m}$ are positive for those species that appear on the right and negative for those on the left side of the equilibrium expression. At equilibrium $\Delta G = 0$.

The relationship between the standard partial molal free energy of reaction, $\Delta G_{m,r}^{\circ}$, and the equilibrium constant, K_m , is

$$\ln K_m = \frac{\Delta G_{m,r}^{\circ}}{RT} \quad (13)$$

Therefore, $Q_m = K_m$, and $\log(Q_m/K_m) = 0$, at equilibrium. In a plot of $\log(Q_m/K_m)$ vs. temperature, the values for all minerals that are in contact with the solution will converge to a value of 0 at the same temperature. When geothermal fluids at equilibrium with hydrothermal minerals mix with solutions of different composition, the equilibrium is disturbed causing: a shift in the position in which the minerals are apparently at equilibrium to lower $\log(Q_m/K_m)$ values if the solutions that mix with the geothermal fluid are very dilute; or a complete lack of any identifiable equilibrium temperature if the solution mixing with the geothermal fluid contains solutes in proportions very different from the geothermal water. Furthermore, there will be a dispersion of the $\log(Q_m/K_m)$ plots, if the geothermal fluid has boiled before it is sampled, since the formation of the residual aqueous solution and a simultaneous increase in pH, both tend to cause supersaturation of certain mineral phases (e.g calcite and pyrite). These factors will lead to complex changes in the various apparent equilibrium mineral/solution temperatures.

3.4 Mixing models

Water ascending from a geothermal reservoir to the surface and emerging in hot springs may cool on the way, by conduction, boiling, or mixing with shallow cold water or by any combination of these three processes. Fluid geothermometry can provide evidence of the location and mechanism of these processes. Water that ascends relatively rapidly and directly from the reservoir with little conductive cooling is likely to have chemical compositions that reflect rock-water equilibrium at the reservoir temperature. Other thermal water has ascended from deep reservoirs so slowly that much of their initial chemistry has been altered by equilibration at lower temperatures. Where the aquifer is at temperatures above the boiling point at atmospheric conditions the water will cool by boiling on its way to the surface whereas waters from reservoirs at temperatures below atmospheric boiling can reach the surface at about the temperature of the aquifer. The chemical composition of mixed water can be used for temperature estimation applying a mixing model (Fournier and Truesdell, 1973). The model can be applied if there is an independent indication of mixing. The mixing models should be applied with great care in geothermometry as the model results are very sensitive to the estimated mixing proportions. The chemical characteristics of mixed water which should be used as evidence for mixing are discussed by Fournier (1981) and Arnórsson (1985). These include large discrepancies between the temperatures indicated by the silica and the cation geothermometers; cold springs with large mass flow rates and relatively high estimated subsurface temperatures by chemical geothermometers; low pH relative to the water salinity; variation in oxygen and hydrogen isotopes; the tendency of calcite undersaturation and low calcium/proton activity ratio compared to geothermal water; variation in chloride concentrations; or systematic variation in temperature and composition of water in springs from the same area. Mixing models have been developed to allow estimation of the hot water component in mixed waters emerging in springs or discharged from shallow wells. There are essentially three kinds of mixing models: 1) the chloride-enthalpy mixing model; 2) the silica-enthalpy warm spring mixing model; 3) the silica-carbonate mixing model.

In this report, the silica-enthalpy and the silica-carbonate mixing models were used. These mixing models are used to study both mixing and boiling processes.

3.5 Scaling in geothermal systems

Scaling and corrosion have presented problems in the exploitation of many geothermal systems. Dissolved materials in geothermal waters can exhibit aggressive corrosive properties or have the tendency to deposit large amounts of mineral scales. Either property can seriously shorten the life of pipes in production wells or reinjection wells. Scaling and corrosion, thus, constitute technical barriers to the utilization of geothermal resources. These are two of the more important geothermal utilization problems that require the close attention of chemists and geologists. The composition of the geothermal fluid results from the environment from which it is extracted (Miller, 1980). The history of the fluid as it comes into contact with various minerals is the key to understanding the tendency of the fluid to promote scaling or corrosion and, as a consequence, making rational material selections for construction. The three major classes of geothermal scales are a) silica phases and silicates, b) sulphides of iron and heavy minerals, and c) carbonates of calcium and sometimes iron. Calcium carbonate scales are the most common of all geothermal scale deposits. Almost all geothermal systems contain dissolved carbon dioxide. The most important facts regarding calcium carbonate deposition problems are:

- 1) Deposits can be readily dissolved, transported or redeposited in water;
- 2) Carbon dioxide and water play major roles in the dissolution, transport and redeposition cycles;
- 3) Calcium carbonate scale deposits can be prevented by pH adjustment or pressure adjustment.

Deposits of calcium carbonate can be readily dissolved and precipitated by water in a variety of conditions. Simple contact of water with air will expose the dissolved calcium to varying concentrations of carbon dioxide. Usually, geothermal systems in which calcium carbonate tends to form have the

characteristics that their waters are of the bicarbonate type and weakly alkaline. If such geothermal fields are to be developed, the calcite scaling potential has to be estimated. Generally, the deposition potential can be estimated from the concentration of chemical components involved in deposition from the water, i.e., when the concentration is higher than the soluble concentration of the chemical compound, precipitation (deposition) of the chemical compound will occur in the water. Two methods to predict calcite scale formation from the chemical components of the water are considered below.

Saturation index (activity product).

This method involves the comparison of the calcium carbonate ion activity product of the solution to the thermodynamic equilibrium constant. An expression of this, the saturation index (SI), is sometimes used to predict scale formation:

$$SI = \log (Q/K) \quad (14)$$

If SI is higher than 0, the water can deposit calcite; if $SI = 0$ the solution is in equilibrium with calcite and will neither deposit nor dissolve calcite; and if SI is lower than 0, the water does not have a calcite deposition potential (is aggressive). The potential scaling problems when utilizing geothermal water depends on the type of water as well as the temperature and the changes the water undergoes in the installations. Therefore, a reliable analysis of the water and a simulation of the changes occurring during the utilization are needed to predict possible scaling. Here, the WATCH program was used to calculate the concentrations of resulting species, activity products and solubility products when the equilibrated fluid is allowed to cool conductively from the reference temperature to some lower temperatures. The scaling potential is estimated by calculating SI during simulated changes in conditions that correspond to changes that the fluid might experience during utilization.

Ryznar stability index (RSI).

This is a method to predict scale formation by comparing the measured pH of a discharged geothermal water with the calculated pH of water equilibrated with calcium carbonate, taking into account calcium ion concentration, alkalinity, temperature and total dissolved solids (Shimada, 1987):

$$RSI = 2 pH_s - pH \quad (15)$$

The Ryznar stability index (Table 5) produces a slightly different value numerically but is interpreted in a similar fashion. It is important to point out that the accuracy of the RSI is greater as a predictor of scaling than of corrosion. This results from the fact that this method is based upon the saturation of calcium carbonate. Calculation of the value for pH_s can be carried out by a graphical method (see e.g. Edstrom, 1998) or through the use of the following equation:

$$pH_s = 9.3 + [(\log (TDS) - 1)/10] + [-13.12 \log (^\circ K)] - \log (\text{calcium hardness}) + \log (\text{alkalinity}) \quad (16)$$

TABLE 5: Interpretation of the Ryznar Stability Index (Carrier, 1965)

RSI-index value	Indication
4.0-5.0	Heavy scale
5.0-6.0	Light scale
6.0-7.0	Little scale
7.0-7.5	No scale and little
7.5-9.0	Light corrosion
>9.0	Heavy corrosion

4. APPLICATION OF SPECIATION PROGRAMS

4.1 The computer program WATCH

The computer program WATCH (Arnórsson et al., 1983a; Bjarnason, 1994), as an aqueous speciation program, is a useful tool for interpreting the chemical composition of geothermal fluids. It may also be used for calculating different geothermometer temperatures. It has been developed to calculate the composition and aqueous speciation of geothermal reservoir waters. The input to the program is a component analysis of each phase (liquid, gas, steam) of the geothermal fluid at the surface, including the water pH and temperature at which it was measured, and specification of the reference temperature at which the aqueous speciation is calculated. The program calculations include the quartz, chalcedony, and Na-K geothermometer temperature. All the chemical components that occur in major concentrations in geothermal waters and/or rocks commonly found in the geothermal systems, are included. WATCH computes the concentration and activity coefficient for each of the 69 aqueous species that are considered in the program assuming homogeneous equilibrium in the solution.

Two sets of equations are solved simultaneously by an iterative procedure, which is carried out a few times during each run of the program. An initial estimate of the ionic strength of the water is obtained using only the concentrations of the major cations. This value is used to calculate all activity coefficients at the temperature of the pH measurement, using the extended Debye-Huckel formula. The homogeneous equilibria and mass balance equations are then solved simultaneously at the given temperature to yield a distribution of species. From the distribution of species, a new value for the ionic strength is computed. This value is used to recompute the activity coefficients and species concentrations, still at the temperature of the pH measurement. Activity coefficients of non-charged aqueous species are taken to be equal to unity. The pH-calculation, the calculation of the activity coefficients, the mass balance iteration, and the ionic strength calculation are repeated, in turn, a couple of times. The equilibrium gas pressures are then computed from the final equilibrium species concentration.

The output from the program lists the components and species concentrations at the reference temperature, as well as the activity coefficients. Ionic balance, geothermometer temperatures, partial pressures of gases, and redox potentials are also calculated. Finally, the ion activity products $\log Q$ and solubility products $\log K$ of 29 selected minerals are computed, from which it is straight forward to compute the corresponding saturation indices, $\log(Q/K)$. The WATCH program may also be used to study chemical changes accompanying adiabatic boiling and conductive cooling of waters in geothermal reservoirs, to compute the resulting species concentrations, activity coefficients, and activity and solubility products from the reference temperature to some lower temperature (Bjarnason, 1994).

In this report, the WATCH program is used to study the chemical equilibria between minerals and solutions as well as to calculate the subsurface chemical equilibrium temperature of 47 chemical analyses of geothermal fluids at Árskógsströnd, Dalvík, and Hrísey, N-Iceland, and in the Khangai area, Mongolia.

4.2 The computer program SOLVEQ

The program SOLVEQ was the first in a series of programs that have been developed since the mid 1970s for calculating equilibria in aqueous systems. SOLVEQ can be used for computing multicomponent homogeneous and partial heterogeneous chemical equilibrium, where equilibration of a given water with specified minerals or gas fugacities can be forced. The forced equilibria are then used to fix the total concentration of specified component species. SOLVEQ's forced mineral equilibrium capability and its ability to calculate pH at high temperatures from a low-temperature pH measurement (Reed and Spycher, 1984) make it useful for geothermometry and other studies of geothermal and sedimentary formation waters and for processing hydrothermal experimental run products to derive thermodynamic data or evaluate mineral equilibria. SOLVEQ solves a system of mass-balance and mass-action equations by the Newton-Rahson numerical technique (Reed and Spycher, 2001). In SOLVEQ, the chemical system is defined in terms of component species. Other species, complexes or redox species are expressed in terms

of these component species. Minerals and gases are also expressed in terms of component species. The total chemical composition of the system is, thus, given as the total molar amount of each component species in that system.

Here, the SOLVEQ program was used to enlarge the number of minerals which can be equilibrated with water at a given temperature, and to study the mineral equilibria for four samples from Iceland, and 7 samples from Mongolia.

5. THE ÁRSKÓGSSTRÖND, DALVÍK, AND HRÍSEY LOW-TEMPERATURE GEOTHERMAL FIELDS, N-ICELAND

5.1 General background

5.1.1 Árskógsströnd

Árskógsströnd is a small community of only a few hundred people on the western side of Eyjafjörður, (Figure 3). Árskógsströnd is approximately 60 km west of the present spreading axis of N-Iceland and the age of the crust is 6-10 m.y. The crust consists of sequences of flood basalts which were originally formed close to the spreading axis but have now moved to the present location through the processes of ocean floor spreading. Close to the study area, a NNE-SSW trending anticline axis is found. West of the anticline, the dip of the lava pile is 3-4° to the southwest, but east of the anticline, in the research area, the dip is 6-8° to the southeast (Saemundsson, 1979). The basalt lavas are of tholeiitic composition and some are porphyritic. The lava thickness varies from a few metres to about 30 m and the flows are interbedded by very thin layers of scoria and sediments. The lava pile is intersected by numerous near-vertical dykes and normal faults. Glacial erosion of the Ice Age has removed about one kilometre of the original lava pile at Árskógsströnd, and brought to the surface lavas

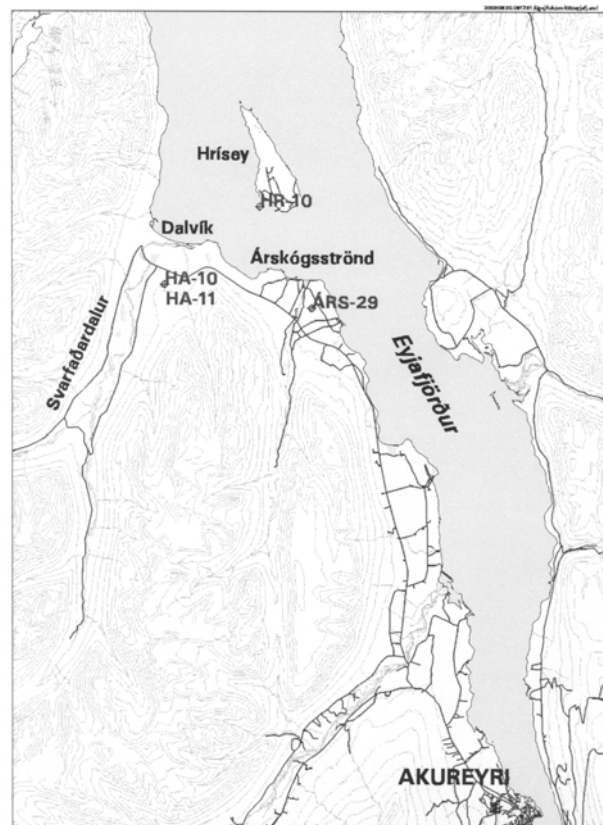


FIGURE 3: The location of wells ÁRS-29, HA-10, HA-11 and HR-10, at Árskógsströnd, Dalvík and in Hrísey

that had been buried to that depth and reheated to about 100°C during the spreading process. During this process, the lavas mineral alteration produced zeolites with mesolite and scolecite as the dominant secondary minerals. As a result, the permeability of the rocks is quite low as the pores and fractures, which comprise the primary permeability, are full of secondary minerals. However, recent tectonic activity, possibly related to postglacial rebound or the present seismic zones, has created fractures leading to a secondary permeability and, thus, opening the way for the formation of convective geothermal fields.

At the end of the year 1998, production started from the Brimnesborgir geothermal system, from well ÁRS-29 (Figure 3). The well was drilled to a depth of 440 m. Only small feedzones were observed until a major one was struck at 429 m depth. The initial pressure within the aquifer was low. The water level in the open hole was at 25 m depth, with the wellhead at 50 m elevation above sea level. During drilling, temperature logs were run during stops. Figure 4 shows temperature logs from well ÁRS-29 and also from wells HA-10, HA-11 and HR-10.

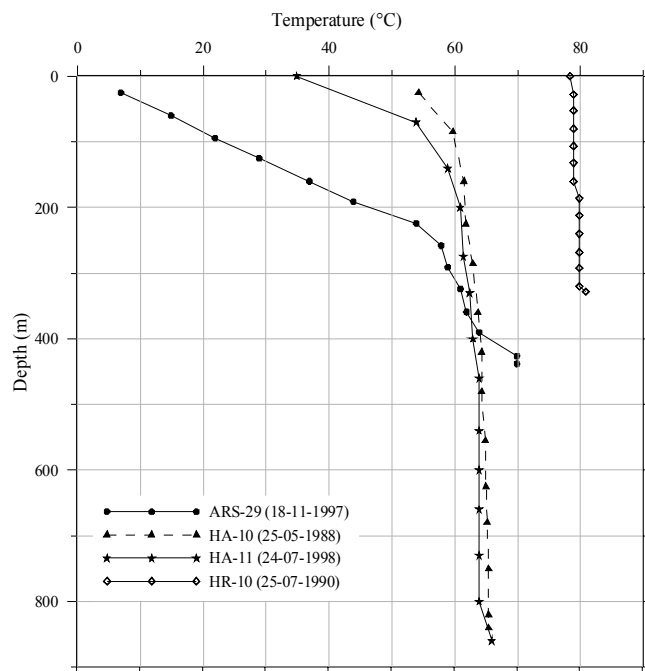


FIGURE 4: Temperature profiles for wells ARS-29, HA-10, HA-11 and HR-10

heat flow to the surface is about 150 mW/m^2 (Flóvenz and Saemundsson, 1993). It reflects the heat conducted through the crust from the underlying mantle.

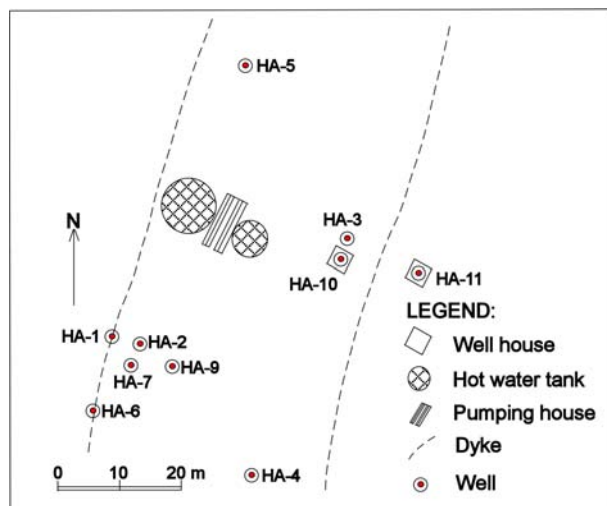


FIGURE 5: Location map of the production field of the Hamar geothermal field in Dalvík, N-Iceland (modified after Karlsdóttir and Axelsson, 1986)

Since 1966, eleven wells have been drilled in the Hamar field (Figure 5). Of these, only four wells have produced, wells HA-2, 9, 10 and 11. HA-2 was in production in 1970-1975; HA-9 in 1975-1977; HA-10 from 1977-1988 and from 1998 to present, and HA-11 in 1988-1998. The last one is also the deepest at 860 m. The production wells have feedzones at 500-800 m depths (Hjartarson and Hardardóttir, 2001), with water temperature of 64°C . Figure 5 shows the location of the wells in the Hamar geothermal field. During drilling, free-flow appeared in wells HA-1, 3, 5 and 8 in which wellhead pressures were estimated to be 1-1.5 bar. In April - September 1990, wells HA-1, 2 and 6 were filled with cement; wells HA-3, 4, and 5 were cased and cemented to the bottom; and HA-7 and 9 were cased with a perforated casing at the feed zones, which means they show true water level. Table 6 shows main characteristics of the wells at Hamar. Well HA-10 is 838 m deep and was drilled in 1977. Only one temperature log is recorded. Well HA-11 was drilled in 1987, and is the deepest in the area, 860 m. Seven temperature logs were carried out from 1987 to 1998.

5.1.2 Dalvík town

Dalvík is a small town of 2,000 inhabitants located at the western side of Eyjafjörður (see Figure 3). A small geothermal field, Hamar, is utilized by a district heating service that serves Dalvík and is approximately 5 km south of the town in the slopes of a hill 60 m a.s.l., and about 65 km from the volcanic rift zone in N-Iceland in a 6-10 m.y. old formation. The main stratigraphic layers consist of porphyritic basalt and tholeiitic basalt. There are some thin layers of scoria and sediments between the basaltic layers. The layers generally dip $3-4^\circ$ towards the southwest. The axis of an anticline is found some 4 km to the northeast of the geothermal field. The basaltic lava pile is intersected by numerous near-vertical dykes and normal faults which probably are a part of a fissure swarm belonging to an extinct central volcano 10-20 km away from Hamar. The regional temperature gradient outside the geothermal area is close to 60°C/km , and the

There are indications that some of the low-temperature geothermal fields in Iceland were formed by crustal movements during the last deglaciation, and are therefore about 10,000 years old (Bödvarsson, 1982). It has been proposed that tectonic movement, that followed deglaciation, formed macroscopic fractures in which convection started and formed geothermal systems. Once the convection is initiated it is a self-sustainable process. Contraction in the deepest part of a fracture due to heat mining will extend the fracture to greater depth and the connecting liquid comes continuously in contact with new hot rock (Flóvenz and Saemundsson, 1993).

TABLE 6: Main characteristics of the wells at Hamar

Well no.	Drilling completed	Depth (m)	Casing		Production period
			Width (")	Length (m)	
1*	Dec.1966	101	4	0-2.4	Not used
2*	Jan.1969	300	8	0-38.0	1970-1975
3**	Apr.1969	504	10; 8	0-5.3; 5.3-39.4	Not used
4**	Jun.1969	303	8	0-28.0	Not used
5**	Feb.1971	587	7	0-60.0	Not used
6*	Mar.1971	373	12	0-3.0	Not used
7***	Jul.1971	302	8	0-78.0	Not used
8	Jul.1974	108	4	0-119.0	Not used
9***	Sep.1975	253	10	0-122.2; 122.2-174.6	1975-1977
10	Sep.1977	838	10 ³ / ₄ ; 9 ⁵ / ₈	0-253.6	1977-1988, 1998-pres.
11	Aug.1987	860	11 ³ / ₄		1988-1998

*Filled with cement;

**Cased and cemented to the bottom;

***Cased with a perforated casing at the feed zones.

5.1.3 Hrísey island

Hrísey is an island about 6.5 km long in a northwesterly direction and just over 2 km wide at most. The island is flat but rises to its maximum altitude, about 100 m at the north end. The island's area is close to 7 km² (Fridleifsson, 1989). The bedrock in Hrísey is Tertiary basaltic lava layers and relatively thin intermediate layers. The inclination of the geological strata is about 3-4° to the southeast. The anticlinal axis is a short distance west of Hrísey and to the south along the Tröllaskagi peninsula. The geological strata are inclined in the direction away from the anticlinal axis to the southwest west of the anticline but to the southeast east of it. A great number of dykes are found in Hrísey, and their direction is north-northeast. Their inclination is 86-87° from horizontal to the northwest, or close to right angle to the geological strata. The rocks are about 10 m.y. old. Natural geothermal manifestations have been observed at two locations in Hrísey, by Laugarkambur on the seashore (Figure 6); and about 1 km to the north of the village there was a hot pool on the shore but it has disappeared now (Einarsson et al., 1979). Well HR-10 has been the main production well of the Hrísey Space Heating Company since December 1987. There are horizontal permeable layers at 40 and 70-80 m depth and at 95-115 m depth (Björnsson and Kristmannsdóttir, 1991). Figure 6 shows location of the wells in the Hrísey field. A slight seawater inflow into the geothermal system affects the chemical composition so that the chloride concentration is relatively high and may cause corrosion in heaters if oxygen is taken up by the water (Hauksdóttir and Björnsson, 2002). Geothermal development started in 1966. A total of 11 wells have been drilled, of which 4 are productive, 5 are used for observation, and 2 for reserve (Table 7). One well is used today for the Hrísey district heating system.

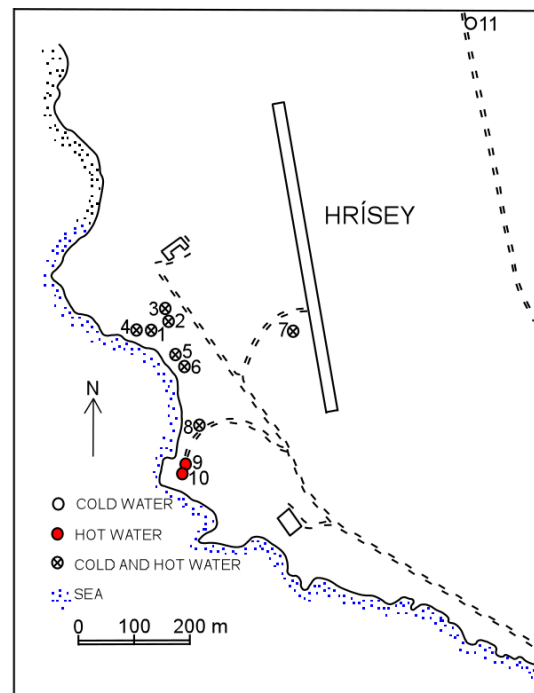


FIGURE 6: The location of the wells in the Hrísey field (see Figure 3) (modified after Björnsson and Flóvenz, 1985)

5.2 Properties of geothermal fluids

Here, chemical composition, the Na-K-Mg and Cl-SO₄-HCO₃ ternary diagrams, various geothermometers, log(Q/K) diagrams, silica-enthalpy and carbonate-silica mixing models and scaling tendency of selected geothermal fluids, are presented and discussed.

TABLE 7: Overview of the Hrísey geothermal wells

Well no.	Type of well	Production history	Depth of well (m)	Depth of casing (m)
1	Observation well		99	2
2	Production well	-1977	132	53
3	Observation well		637	4.4
4	Observation well		322	2.7
5	Production well	1977-1988	320/1055	92
6	Reserve well for well-5		145	1.8
7	Observation well		156	6.5
8	Observation well		176	5
9	Reserve well for well-10		224	2
10	Production well	1988 - until present	216	143
11	Production well	1988 - until present	130	16.4

5.2.1 Chemical composition

Analyses of 4 samples from well ÁRS-29 of Árskógsströnd, 9 samples from well HA-10 and 11 samples from well HA-11 of Dalvík, and 8 samples from well HR-10 of Hrísey, were chosen. Tables 8-10 show chemical composition of waters for wells ÁRS-29, HA-10 and HA-11, and HR-10, respectively.

TABLE 8: Chemical composition of waters (in ppm) from well ÁRS-29, Árskógsströnd, N-Iceland

Date No.	21.01.1998 1998-0010	22.11.1999 1999-0501	5.10.2000 2000-0386	13.11.2001 2001-0384
T _{meas} (°C)	73.5	73.4	73.0	73.5
pH	10.15	10.04	10.05	10.12
Silica (SiO ₂)	104.9	106	106.5	104
Sodium (Na)	55.4	54.8	56.2	55.0
Potassium (K)	0.93	0.81	0.79	0.83
Calcium (Ca)	2.11	2.09	2.02	2.1
Magnesium (Mg)	0.011	0.004	0.001	0.006
Carbonate (CO ₂)	15.9	16.1	17.0	16.6
Sulfate (SO ₄)	15.7	15.8	15.5	16.0
Hydrogen sulfide (H ₂ S)	0.18	0.16	0.16	0.15
Chloride (Cl)	13.7	14.5	13.9	14.0
Fluoride (F)	0.86	1.01	0.92	1.0
Iron (Fe)	0.0034	0.0007	0.0019	0.0016
Manganese (Mn)	0.0026	0.0001	0	0
Boron (B)	0.15	0.16	0.16	0.16
Aluminium (Al)	0.0073	0.074	0.068	0.065
Ionic balance (%)	0.93	2.47	4.28	-0.6
TDS	150	185	225	232

TABLE 9: Chemical composition (in ppm) of waters from wells HA-10 and HA-11, Dalvík, N-Iceland

	HSA-10										HA-11																			
	1977-0148	1979-3042	15.5.1981	1910.1985	1985-0268	26.10.1987	1987-0164	18.10.1988	1988-0180	22.11.1999	5.10.2000	2001-0383	17.10.1988	1988-0175	6.10.1989	23.11.1990	1990-0282	23.10.1991	1991-0198	31.10.1992	1992-0267	13.10.1993	15.11.1994	1994-0345	14.11.1995	1995-0315	6.11.1996	1996-0347	17.11.1997	1997-0719
T _{meas} (°C)	64.2	64	64.9	64.3	64.1	63.5	64.2	64.2	64.2	64.0	65.2	64.4	64	64.0	65.2	64.1	64.8	64.1	64.1	64	64	63.5	64.1	64.1	64.8	64.1	64.1	63.7	64	
pH	10.3	10.1	10.2	10.2	10.2	10.0	10.2	10.2	10.2	10.3	10.31	10.3	10.3	10.3	10.31	10.4	10.3	10.3	10.38	10.38	10.3	10.32	10.4	10.38	10.3	10.3	10.38	1	10.27	
Silica (SiO ₂)	95	95.5	92.3	88.9	89.3	88.4	88.5	88.5	88.5	88.2	89.4	90.4	89.2	88.2	89.4	90.8	90.1	89.8	89.8	89.8	89.2	88.9	90.8	90.8	90.1	89.8	89.8	88.3	90.2	
Sodium (Na)	46.7	46.4	48.6	46.5	46.0	47.9	48.8	48.8	48.8	48.2	48.45	48.2	48.2	48.2	48.45	49.3	49.2	48.9	48.9	48.9	48.2	49.5	49.3	49.3	49.2	48.9	48.9	49.3	49.3	
Potassium	0.53	0.67	0.69	0.74	0.55	0.7	0.54	0.54	0.54	0.6	0.62	0.46	0.55	0.55	0.62	0.51	0.54	0.51	0.51	0.51	0.55	0.53	0.51	0.51	0.54	0.51	0.53	0.53	0.55	
Calcium (Ca)	2.19	1.98	1.82	2.04	2.1	2.01	2.1	2.1	2.1	1.94	2.23	2.12	2.14	2.03	2.23	2.17	2.08	2.11	2.11	2.14	2.08	2.08	2.17	2.08	2.08	2.11	2.09	2.15		
Magnesium	0.01	0.02	0.01	0.00	0.01	0.00	0.004	0.004	0.004	0.000	0.000	0.00	0.00	0.00	0.000	0.00	0.00	0.002	0.002	0.00	0.00	0.003	0.00	0.00	0.00	0.002	0.002	0.003	0.004	
Carbonate	13.5	14.0	11.9	14.2	15.1	16.8	14.9	14.9	16.4	13.3	16.7	15.0	15.7	12.0	13.3	15.6	16.5	14.1	14.1	15.6	16.3	12.8	13.0	12.5	16.5	14.1	14.9	15.8		
Sulfate (SO ₄)	15.5	14.1	14.2	13.3	12.9	13.4	13.3	13.3	12.9	13.31	13.4	13.2	13.5	13.4	13.4	13.0	12.5	13.4	13.4	13.0	13.5	12.8	13.0	12.5	12.5	13.4	12.7	13.1		
Hydrogen	0	0.06	0	0	0	0.02	0.04	0.04	0.03	0	0	0	0	0.02	0	0	0	0.03	0	0	0	0.038	0	0	0	0.03	0.04	0.03		
Chloride (Cl)	10.1	10.2	11.0	9.54	9.36	9.13	9.35	9.35	9.22	9.25	9.0	9.37	9.13	9.19	9.25	8.75	8.94	9.2	9.2	8.75	9.13	8.89	8.75	8.94	8.94	9.2	8.42	9.19		
Fluoride (F)	0.5	0.52	0.5	0.52	0.48	0.5	0.51	0.51	0.48	0.494	0.482	0.49	0.48	0.49	0.494	0.5	0.5	0.49	0.49	0.5	0.48	0.46	0.5	0.5	0.5	0.49	0.44	0.5		
Iron (Fe)	-	0.02	-	-	0	-	0.003	0.003	0.003	0	-	-	-	-	0	-	0.00	0.003	0.003	-	-	0.003	-	0.00	0.00	0.003	0.001	0.002		
Manganese	-	-	-	-	-	-	0.000	0.000	0	-	-	-	-	-	-	-	0	0.001	0.001	-	-	0.000	-	0	0	0.001	0.000	0.000		
Boron (B)	-	0.08	0.07	-	-	-	0.08	0.08	0.1	0.09	0.09	0.1	0.09	-	0.09	0.08	0.08	0.07	0.07	0.08	0.09	0.08	0.08	0.08	0.08	0.07	0.07	0.08		
Aluminium	-	0.12	-	-	-	-	0.083	0.083	0.075	-	0.088	0.08	0.08	-	0.088	-	0.08	0.082	0.082	-	0.08	-	-	0.08	0.08	0.082	0.075	0.068		
Ionic balance	-	1.83	-0.3	-5.27	0.52	11.5	2.13	2.13	-2.6	4.06	-2.81	-5.59	-4.9	25.2	207.6	-3.49	-2.14	-0.11	-0.11	-	1.2	-	1.2	-3.49	-2.14	-0.11	3.26	-1.36		
TDS	220	226	220.	203.	178.	227	214	208	199	205.4	207.6	211.	211	225	205.4	216	197	228	228	216	211	183	216	197	197	228	188	163		

TABLE 10: Chemical composition (in ppm) of waters from well HR-10 Hrísey, N-Iceland

Date Number	15.11.1994 1994-0344	15.11.1995 1995-0322	7.11.1996 1996-0352	17.11.1997 1997-0717	9.11.1998 1998-0559	22.11.1999 1999-0494	5.10.2000 2000-0384	12.11.2001 2001-0380
T _{meas} (°C)	77.3	77.0	77.0	76.8	76.9	77.7	77.9	76.8
pH	9.53	9.45	9.45	9.29	9.36	9.41	9.26	9.26
Silica (SiO ₂)	66.5	67.8	65.9	65.9	66.8	67.1	68.3	65.3
Sodium (Na)	248.7	246.7	252	254.9	252	250	257	256
Potassium (K)	4.27	4.17	4.13	4.13	4.22	4.22	4.52	4.25
Calcium (Ca)	81.3	81.1	90.3	92.8	88.9	84.7	88.1	97.5
Magnesium (Mg)	0.006	0.013	0.012	0.025	0.026	0.024	0.02	0.05
Carbonate (CO ₂)	5.31	5.8	3.4	3.7	5.2	3.3	4.16	3.77
Sulfate (SO ₄)	59.0	58.2	59.5	59.8	61.32	60.3	58.0	59.9
Hydrogen sulfide (H ₂ S)	0	0	0.06	0.05	0.05	0.05	0.05	0.05
Chloride (Cl)	471	478	471.4	518	490	487	498	510
Fluoride (F)	0.25	0.28	0.24	0.245	0.26	0.28	0.25	0.27
Iron (Fe)	-	0.002	0.0039	0.0027	0.0049	0.007	0.0026	0.002
Manganese (Mn)	-	0	0.0046	0.0007	0.0009	0.0038	0.0004	0.0006
Boron (B)	0.1	0.1	0.11	0.08	0.09	0.12	0.11	0.1
Aluminium (Al)	-	0.021	0.022	0.0018	0.016	0.0262	0.019	0.017
Ionic balance (%)	-0.06	-1.83	4.35	-2.52	-0.31	-0.95	0.35	0.72
TDS	988	965	1056	1042	670	900	942	1028

Water analyses from well ÁRS-29, Árskógsströnd, from 1998 to 2001, are shown in Table 8. The concentrations of dissolved solids in the geothermal waters of Árskógsströnd are generally low, and the electrical conductivity is in the range 250-260 $\mu\text{S}/\text{cm}$. The average temperature is 73.4°C, pH is 10.1, and the silica content is 105 ppm.

Water analyses from well HA-10, (1977-1988, and 1998 to present), and well HA-11, Dalvík (1988-1998), are shown in Table 9. The concentrations of dissolved solids in the geothermal waters from well HA-10 are generally low and the electrical conductivity ranges from 195 to 237 $\mu\text{S}/\text{cm}$. The average temperature is 64.2°C, pH is 10.2, and the silica content is 91 ppm. In well HA-11 the concentrations are generally low, and the electrical conductivity is in the range 224-248 $\mu\text{S}/\text{cm}$. The average temperature is 64.2°C, pH is 10.3, and silica content is 90 ppm.

Water analyses from well HR-10, Hrísey (1994-2001), are shown in Table 10. The concentrations of dissolved solids in the geothermal waters are generally low, although significantly higher than at Hamar and Árskógsströnd due to the sea water component. Hence the electrical conductivity is high, in the range 1584-1830 $\mu\text{S}/\text{cm}$. The average temperature is 77.2°C, pH is 9.4, and silica content is 67 ppm.

5.2.2 Ternary diagrams

The $\text{Cl-SO}_4\text{-HCO}_3$ ternary diagram is commonly used for classification of the thermal fluids. The diagram indicates several types of thermal fluids such as immature waters, peripheral waters, volcanic, and steam-heated waters. The degree of separation between data points for high chloride and bicarbonate waters gives an idea of the relative degrees of interaction of the CO_2 charged fluids at lower temperatures, and of the HCO_3 contents increasing with time and distance travelled underground. Figure 7 shows that most of the samples plot in the peripheral waters regions (samples from wells ÁRS-29, HA-10 and HA-11), but plots for water from well HR-10 (Hrísey) near the chloride corner. The four water samples from well HR-10, which are the hottest of those discussed (77°C), are close to mature waters, but a little high in sulfate (chloride water with a sulfate component). This reflects the seawater influence in the samples.

The Na-K-Mg ternary diagram (Giggenbach, 1988) is used for evaluating equilibrium between the hot waters and rocks at depth and to estimate reservoir temperature. Results for the wells discussed are shown in Figure 8.

5.2.3 Geothermometry

Geothermometer temperatures have been calculated by the computer program WATCH (Arnórsson et al., 1983a; Bjarnason, 1994) using the equations shown in Tables 3 and 4 using data from the selected wells in N-Iceland. The results are shown in Table 11. The quartz geothermometer temperatures of well ÁRS-29 are in the range 98-103°C (Fournier and Potter, 1982); 139-141°C (Fournier, 1977); and 127-128°C (Arnórsson et al., 1983b). The chalcedony geothermometer temperatures are 67-73°C (Fournier, 1977); and 111-113°C (Arnórsson et al., 1983b). The Na-K geothermometer temperatures are 57-65°C (Arnórsson et al., 1983b); 43-52°C (Truesdell, 1976); 99-106°C (Fournier, 1979); and 113-121°C (Giggenbach, 1988). Finally, the Na-K-Ca geothermometer temperature is 95-101°C (Fournier and Truesdell, 1973).

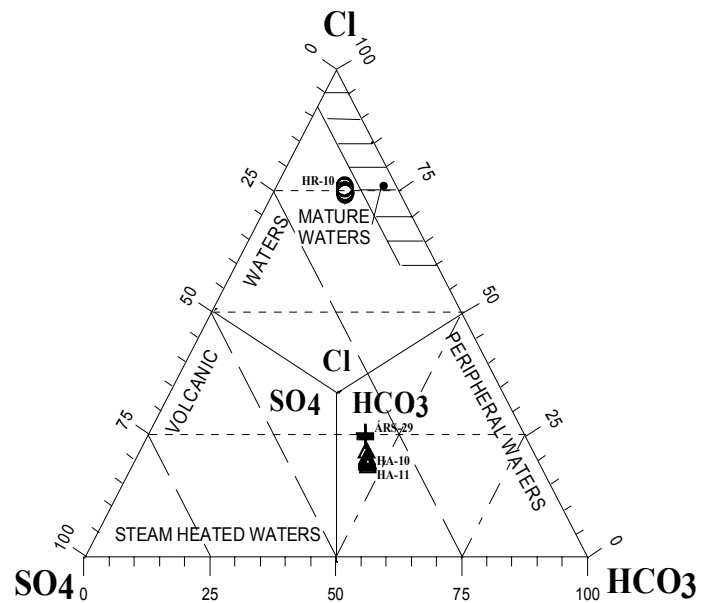


FIGURE 7: The Cl-SO₄-HCO₃ ternary diagram for waters from wells ÁRS-29, HA-10, HA-11 and HR-10

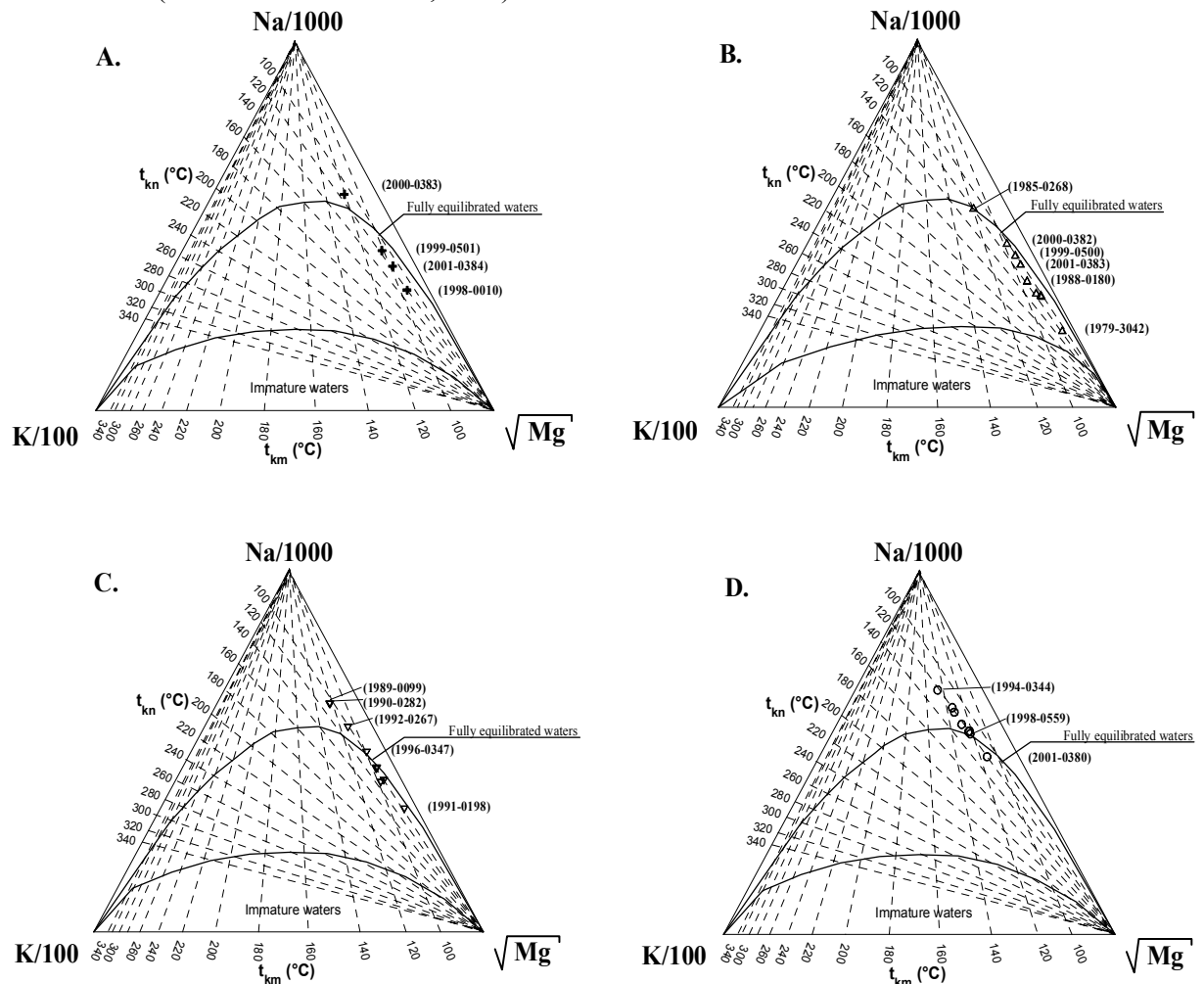


FIGURE 8: Classification of the geothermal waters according to Na-K-Mg equilibrium diagrams; A) ÁRS-29, Árskógsströnd; B) HA-10, Dalvík; C) HA-11, Dalvík; and D) HR-10, Hrísey

TABLE 11: Results of different geothermometers from Árskógsströnd, Dalvík and Hrísey

Well number	Sample number	T _{meas} (°C)	T _{qtz} ¹ (°C)	T _{qtz} ² (°C)	T _{qtz} ³ (°C)	T _{chal} ¹ (°C)	T _{chal} ³ (°C)	T _{NaK} ¹ (°C)	T _{NaK} ⁴ (°C)	T _{NaK} ⁵ (°C)	T _{NaK} ⁶ (°C)	T _{NaKCa} ⁷ (°C)
ÁRS-29	1998-0010	73.5	98	140	127	67	112	65	52	106	121	101
	1999-0501	73.4	103	141	128	73	112	59	45	99	115	96
	2000-0386	73.0	103	141	128	73	113	57	43	97	113	95
	2001-0384	73.5	98	139	127	68	111	60	46	100	116	97
HA-10	1977-0148	64.2	84	134	122	52	106	48	32	87	103	85
	1979-3042	64.0	98	135	122	67	106	58	44	98	114	94
	1981-0082	64.9	87	133	120	56	104	57	43	97	113	94
	1985-0268	64.3	82	131	118	51	102	62	49	103	119	97
	1987-0164	64.1	90	131	118	59	103	50	35	89	105	87
	1988-0180	63.5	97	130	118	67	102	58	45	99	114	94
	1999-0500	64.2	86	131	118	54	102	47	31	85	102	85
	2000-0382	64.2	83	131	117	51	103	47	32	86	103	85
	2001-0383	64.1	82	130	118	50	102	47	32	86	103	85
HA-11	1988-0175	64.0	101	131	119	71	103	55	41	95	111	92
	1989-0099	64.0	87	130	118	55	102	51	37	91	107	89
	1990-0282	65.2	84	131	118	52	103	53	38	92	108	89
	1991-0198	64.4	81	132	119	50	103	41	24	79	96	80
	1992-0267	64.0	81	131	119	49	103	48	33	87	103	85
	1993-0196	63.5	82	131	118	51	102	46	30	84	100	84
	1994-0345	64.1	81	132	117	49	104	44	28	82	99	82
	1995-0315	64.8	83	132	119	52	103	46	31	85	102	84
	1996-0347	64.1	84	131	118	52	103	45	28	83	99	83
	1997-0719	63.7	84	130	118	52	102	47	30	84	101	84
1998-0555	64.0	84	132	118	52	103	47	32	86	102	85	
HR-10	1994-0344	77.3	104	116	102	74	87	65	53	107	122	98
	1995-0322	77.0	106	117	103	76	88	64	53	106	122	97
	1996-0352	77.0	105	115	102	75	87	63	51	104	120	96
	1997-0717	76.8	106	115	102	76	87	62	50	104	119	95
	1998-0559	76.9	105	116	102	75	87	64	52	106	121	96
	1999-0494	77.7	107	116	103	77	88	64	52	106	121	97
	2000-0384	77.9	108	117	104	78	88	66	55	108	123	98
	2001-0380	76.8	106	115	101	76	86	64	51	105	120	96

1) WATCH program;

2) Fournier (1977);

3) Arnórsson et al. (1983b);

4) Truesdell (1976);

5) Fournier (1979);

6) Giggenbach (1988);

7) Fournier and Truesdell (1973).

For wells HA-10 and HA-11 the quartz geothermometer temperatures of are in the range 81-101°C (Fournier and Potter, 1982); 130-135°C (Fournier, 1977); and 117-122°C (Arnórsson et al., 1983b). The chalcedony geothermometer temperatures are 49-71°C (Fournier, 1977); and 102-106°C (Arnórsson et al., 1983b). The Na-K geothermometer temperatures are 44-58°C (Arnórsson et al., 1983); 24-45°C (Truesdell, 1976); 82-103°C (Fournier, 1979); and 96-119°C (Giggenbach, 1988). Finally the Na-K-Ca geothermometer temperature is 82-97°C (Fournier and Truesdell, 1973).

For well HR-10, Hrísey, the quartz geothermometer temperature varies in the range 104-108°C (Fournier and Potter, 1982); 115-117°C (Fournier, 1977); and 102-104°C (Arnórsson et al., 1983b). The chalcedony geothermometer temperatures are 74-78°C (Fournier, 1977); and 86-88°C (Arnórsson et al., 1983b). The Na-K geothermometer temperatures are 62-66°C (Arnórsson et al., 1983b); 50-55°C (Truesdell, 1976); 104-108°C (Fournier, 1979); and 119-123°C (Giggenbach, 1988). Finally, the Na-K-Ca geothermometer temperature is 95-98°C (Fournier and Truesdell, 1973).

5.2.4 Log (Q/K) diagrams

The log Q and log K can be computed by computer programs like WATCH and SOLVEQ at different temperatures. Here results from the SOLVEQ program are used. For a given temperature and total composition of homogeneous aqueous solution, SOLVEQ computes the activities of all species and saturation indices of solids and fugacities of gases. The central thermodynamic database of SOLVEQ is SOLTHERM (Reed and Spycher, 2001), which contains all thermodynamic data for aqueous, gas and mineral species. The main output file contains the data, the speciation, gas fugacities and mineral saturation indices. The plot file contains the list of saturation indices for minerals and gases at each temperature. This file can be used by the program SOLPLOT to plot the saturation indices versus temperature. Figure 9 is composed of four graphs representing samples 2001-0384, 2001-0383, 1998-0555, and 2001-0380, respectively showing SI (saturation index) in the range 20-160°C for the Árskógsströnd, Dalvík and Hrísey samples. The temperature, at which the SI curves intersect the zero line, determines the equilibrium temperature for each mineral. In all four graphs, there is no clear temperature-rock equilibrium reached between all the minerals and the thermal fluid, which may be influenced by the selected minerals. In some cases three or four minerals attain equilibrium at the same temperature. If the fluid is in equilibrium with calcite, we get four intersections in the region 60-80°C, for calcite, albite-low, muscovite and quartz.

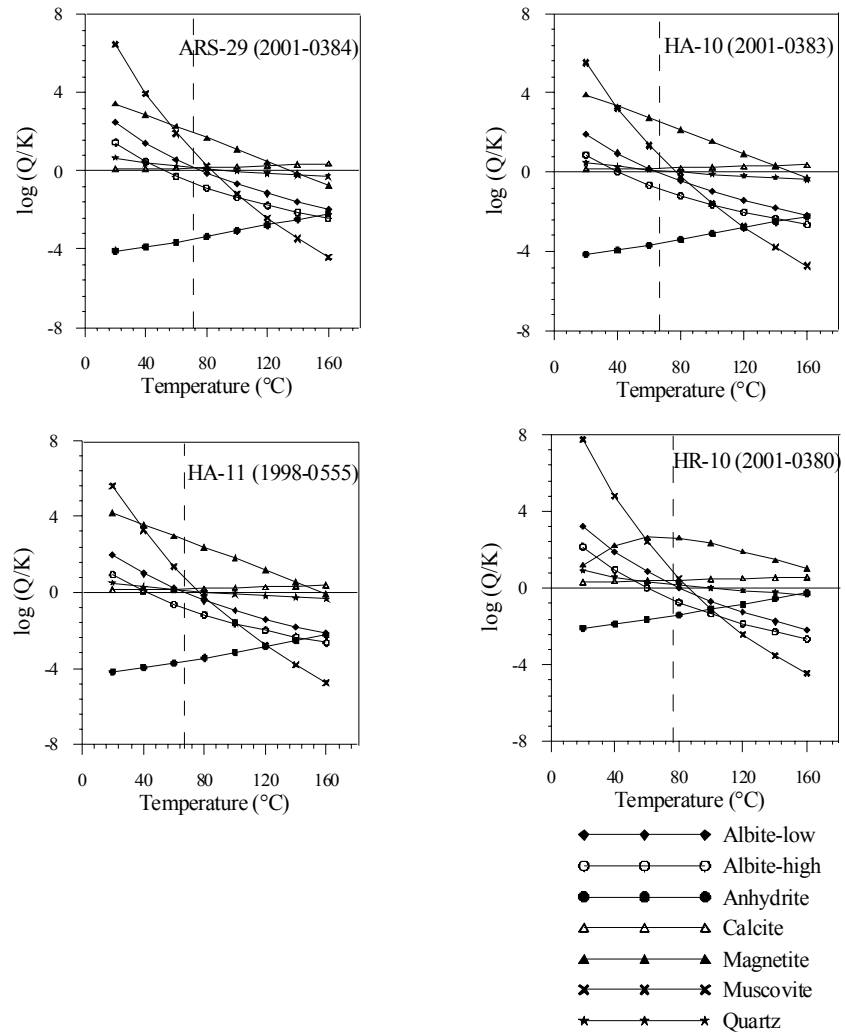


FIGURE 9: Mineral equilibrium diagrams for geothermal fluids from Árskógsströnd, Dalvík and Hrísey

5.2.5 Mixing models

The evident mixing processes diminish the reliability of the subsurface temperatures estimated by solute geothermometers. In order to eliminate such mixing effects, two mixing models have been applied to evaluate the subsurface temperatures at the Árskógsströnd, Dalvík and Hrísey low-temperature geothermal fields.

Silica-enthalpy mixing model: Dissolved silica concentration versus enthalpy can be used to determine the temperature of the hot water component of mixed water. The model is based on the assumption that no conductive cooling has occurred after mixing. If the mixed water has cooled conductively after mixing,

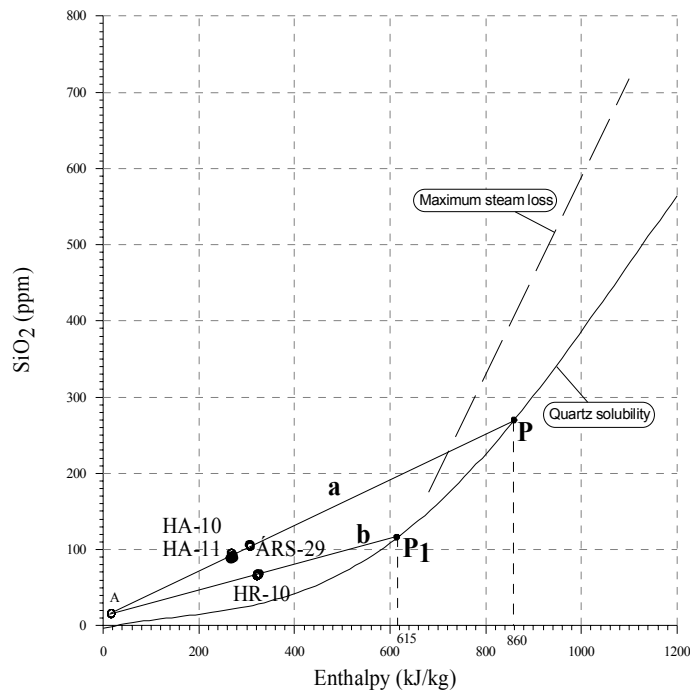


FIGURE 10: The silica-enthalpy mixing model for samples from wells ÁRS-29, HA-10, HA-11 and HR-10

the calculated temperature of the hot water component will be too high. Here, it is assumed that no silica deposition has taken place before or after mixing, and also that quartz but not chalcedony controls the solubility of silica in the hot water component. Figure 10 shows silica-enthalpy diagrams for the samples from the Árskógsströnd, Dalvík and Hrísey geothermal fields. For the 4 wells, two mixing lines, a and b, are possible. Line a intersects the quartz solubility curve at point P, with an enthalpy of 860 kJ/kg corresponding to the temperature 202°C. It connects point A (ground water), and the wells HA-10, HA-11, and ÁRS-29 with P. Line b connects point A with well HR-10 and P₁. The intersection with the quartz solubility curve at point P₁, with an enthalpy of 615 kJ/kg, corresponds to the temperature 146°C. The observed discrepancy between measured temperatures at the geothermal systems in Eyjafjörður and temperatures predicted by the mixing model is probably the result of the choice

of silica phase in the model. If chalcedony had been used as a controlling phase in the models instead of quartz the model results would have indicated both lower reservoir temperatures and less mixing.

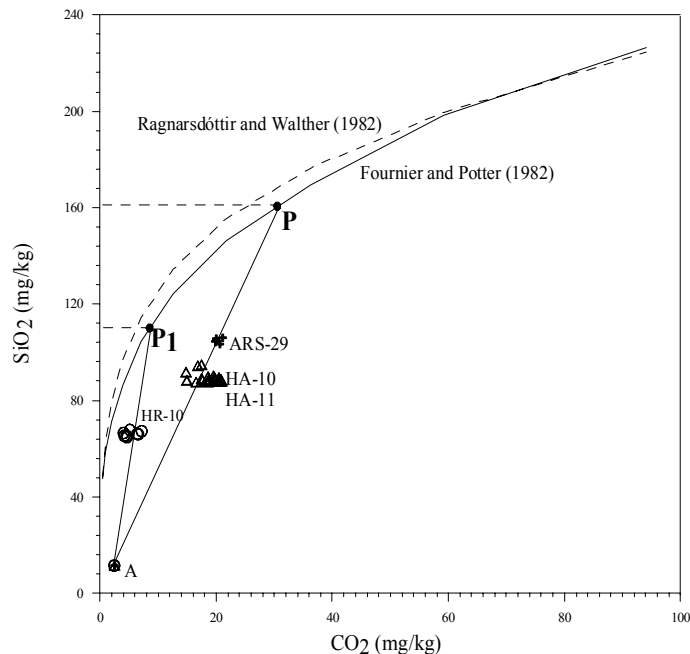


FIGURE 11: The carbonate-silica mixing model for the selected wells

The carbonate-silica mixing model. This model is based on the relationship between silica and the total carbonate concentrations in geothermal waters to estimate subsurface temperatures. The model assumes that both aqueous silica and CO₂ concentrations are fixed by temperature dependent solute/mineral equilibrium in the reservoir, that most of the dissolved total carbonate is in the form of carbon dioxide, and that temperature dependence of silica is controlled by quartz at temperatures above approx 200°C. Figure 11 shows the carbonate-silica mixing model for water samples from wells ÁRS-29, HA-10, HA-11 and HR-10. An extrapolation of a line from A through the data points for Hrísey (point P₁) indicates that the temperature of the hot water component is about 143°C. Well HR-10 water contains relatively little total carbonate (3.3-5.8 ppm). For the Árskógsströnd and Dalvík geothermal fields, a linear relationship (line between

plot A and P) between silica and carbonate is observed for ground water (point A) and hot water (points for samples from wells HA-10, HA-11 and ÁRS-29), which indicates that the hot water component of the mixed water has not boiled. An extrapolation of a line through the data points (to point P) indicates that the temperature of the hot water component is about 166°C.

5.2.6 Scaling tendency

In order to study changes during the production period, available chemical analyses (Tables 8-10) were entered into the WATCH program, and the results used to compute the saturation index and the Ryznar stability index by Equation 15. Figure 12 shows calcite saturation as a function of time. The calcite saturation index boundary between scaling and no scaling is approx. 0.36-0.5. In well ÁRS-29 water, SI is between 0.1 and 0.14. In well HA-10 water, it is about 0.014-0.195 and in well HA-11 water, it is lower than 0.189. For well HR-10, SI is higher than for waters from wells ÁRS-29, HA-10 and HA-11.

Figure 13 shows how chloride concentration affects the calcite saturation index in the well waters. SI appears to increase with the chloride content. The behaviour of the Ryznar stability index with time is shown in Figure 14. The RSI gives a qualitative estimate of the calcium carbonate scaling tendency of the water. Comparison of Figures 12 and 14 illustrates that the two methods agree reasonably well; both predict that no scaling will occur from fluids from Hamar and Árskógsströnd, whereas scaling might possibly become a problem in utilization of the geothermal fluid in Hrísey.

5.3 Summary

One of the main aims of this report is to predict subsurface temperatures and to categorise the fluids from the geothermal wells in the Árskógsströnd, Dalvík and Hrísey fields in N-Iceland.

Temperatures and chemical composition of thermal fluids from the wells suggest that their sources may be different. Using the

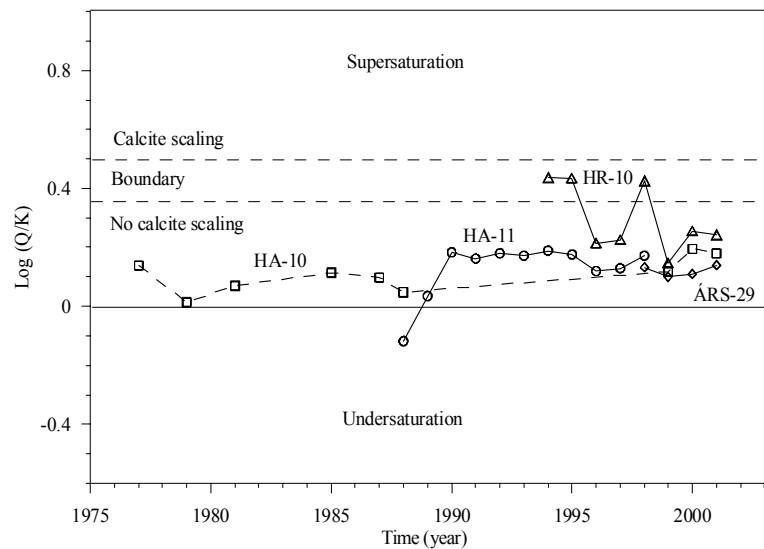


FIGURE 12: Saturation index (SI) of calcite for wells ÁRS-29, HA-10, HA-11 and HR-10 water, N-Iceland, from 1977 to 2001

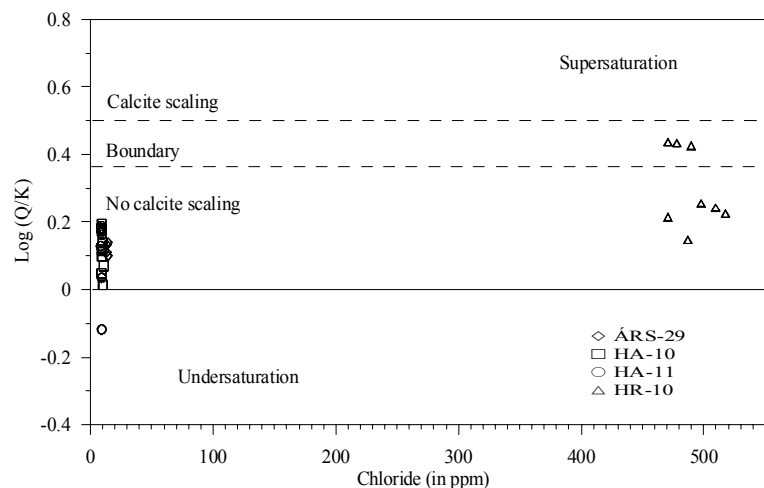


FIGURE 13: CaCO₃ saturation state and the chloride content in selected well waters

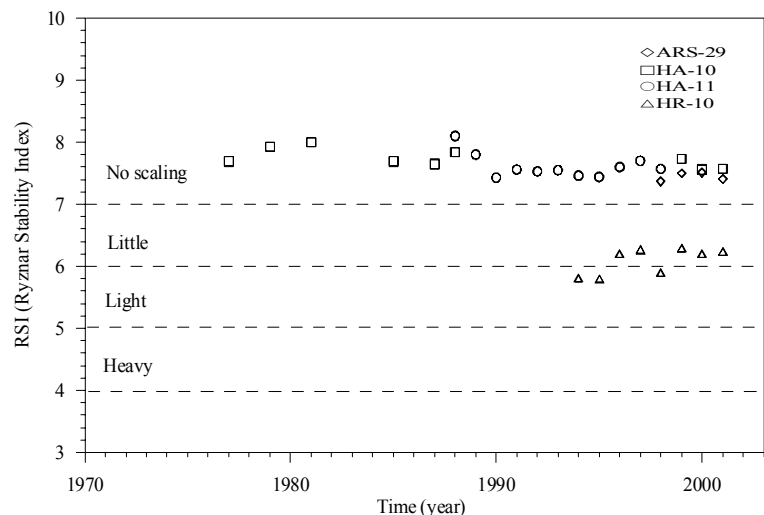


FIGURE 14: The Ryznar stability index correlated with time (years) in selected well waters

computer programs WATCH and SOLVEQ and plotting the log (Q/K) diagrams, possible equilibrium and non-equilibrium situations between studied fluid components and minerals can be identified. The chemical geothermometers which have been used to estimate the subsurface temperatures of the geothermal fluids, give an overall picture of the fields. The chalcedony geothermometer temperatures are in the range 49-113°C, the quartz geothermometer temperatures in the range 81-141°C, the Na-K geothermometer temperatures 24-123°C, and the Na-K-Ca geothermometer temperatures 80-101°C. The geothermal fluids from wells ÁRS-29, HA-10 and HA-11 plot in the peripheral waters region, suggesting a bicarbonate-sodium water type, but from well HR-10 they plot near the chloride corner, suggesting a sodium-chloride water type. The solute geothermometers and the log (Q/K) method give reasonable estimates for the subsurface temperatures of the well discharges, but their use for geothermal fluids are limited by mixing processes in the upflow, that have been detected, by the use of two mixing models. The Na-K-Mg ternary diagram reservoir temperatures (T_r) are in the range 110-125°C for well ÁRS-29; 110-120°C for well HA-10; 80-115°C for well HA-11; and 115-125°C for well HR-10.

Studies of the calcite saturation index (*SI*), and the Ryznar stability index (*RSI*) were carried out to check on scaling tendency. The results indicate that scaling might have occurred in a few samples from Hrísey but scaling is not expected in Dalvík nor at Árskógsströnd.

6. THE KHANGAI AREA, MONGOLIA

6.1 General background

Mongolia has an average altitude of 1580 m above sea level, with the highest point at Huiten peak in the Tavan Bogd mountain range in the north. The lowest point is the Hoh Nuur (Blue lake), lying at an altitude of 532 m above sea level. The latest active tectonic period in Mongolia started at the end of the Mesozoic and the beginning of the Oligocene, due to the simultaneous development of the south Siberian plate (mountain part) and the Baikal lake region. At that time, intense tectonic development gave the Mongolian mountains their present appearance. A geophysical survey on the crustal structure has established that accumulative thermal sources (magma lumps) are located near the surface under the Khangai Khentii mountain region. Geothermal resources in Mongolia are mainly distributed in Khangai, Khentii, around the Khubsugul, Mongol Altai plate forms, Dornod-Dariganga and Orkhon-Selenge regions, due to developments during the second geodynamic Cenozoic age. The Khangai geothermal system has attracted the interest of researchers, and its location is favourable with regard to social and economic conditions. The Khangai geothermal system, which is a block structure system limited by the Tarvagatai, Baidrag, Tamir, and Orkhon rift, is located in the central fold mountains (Ministry of Agriculture and Industry of Mongolia, 1999). Figure 15 shows hot spring locations and a tectonic rift map of the Khangai area (Geodesy and Cartographical Institute, 1980).

Mongolia consists of 21 *aimags* (provinces), and each aimag consists of 12-22 *soums* (a territorial administrative unit subordinate to an aimag). Khangai geothermal field extends into 8 aimags. The Khangai area in Central Mongolia has wide open valleys with huge rounded mountains, the highest reaching an altitude of 3905 m. The Khangai mountain range is over 750 km long, stretching from west to south-east in the central portion of Mongolia, with peaks at 3200-3500 m. The main mountain range has several big branch ranges. In the north, high ranges such as Tarvagatai and Bulnai run east to west. The average altitude of the ecoregion is 2500-3000 m, with the highest peak, Zurhiin Hunh Mt. at 3227 m. The ecoregion is bordered by the Selenge-Orkhon forest steppe in the north and meets the Khangai mountain alpine meadows in the south. The elevation of the Khangai area's hot springs is in the range 1335-2500 m a.s.l. (Popov, 1963). From 36 hot springs of the Khangai area, seven were selected (Tsenkher, Tsagaan sum, Shivert, Chuluut, Khuremt, Khujirt and Noyonkhangai. This area of hot springs has relatively highly developed infrastructure with access of central electric network and a highway located nearby. Furthermore, the area is rich with historical places, making it more attractive for tourism development. The locations and general features of the hot springs considered in this study is summarized below:

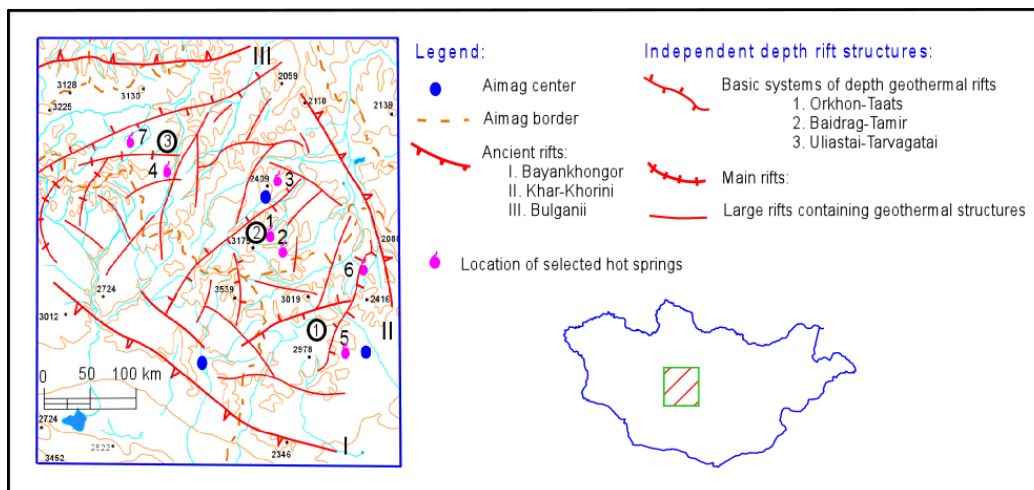


FIGURE 15: Khangai area, Mongolian hot spring locations and tectonic rift

- The Tsenkher hot spring is the biggest of the Mongolian hot springs and located at $47^{\circ}20'N$ and $101^{\circ}39'E$ in the Tsenkher soum of the Arkhangai province. It is 25 km southeast of the Tsetserleg province center, with an elevation of 1860 m a.s.l.. The average surface temperature is $85^{\circ}C$, and the flow rate is 10 l/s.
- The Tsagaan sum hot spring is located at $47^{\circ}4'N$ and $102^{\circ}6'E$ in the Khotont soum of the Arkhangai province. The average surface temperature is $69^{\circ}C$, and the flow rate is 8 l/s.
- The Shivert hot springs are located at $47^{\circ}38'N$ and $101^{\circ}31'E$ and 21 km northeast from the Tsetserleg province center of the Arkhangai, at an elevation of 1710 m a.s.l.(Dorj et al., 2003).
- The Chuluut hot spring is located at $47^{\circ}5'N$ and $100^{\circ}15'E$, and is 25-30 km north of the Chuluut soum in Arkhangai province at an elevation of 2190 m a.s.l. The average surface temperature is $44^{\circ}C$, and the flow rate is 1.2 l/s. The hot spring is surrounded by mountains like Bayanzurkh mount, Angarkhain Buluu, Baatar Khaan, and Samlin, which are parts of the big Khangai forest mountain systems.
- The Khuremt hot spring is located at $46^{\circ}16'N$ and $102^{\circ}47'E$ in the Taragt soum of Uvurkhangai province, at an elevation of 1900 m a.s.l., 5 km south from the Taragt soum center and northwest of Arvaikheer province center. The average surface temperature is $55^{\circ}C$, and the flow rate is 5 l/s. The water has been tested for H_2S , and no trace was observed from more than 10 different sources producing small, warm water streams.
- The Khujirt hot spring is located at $46^{\circ}54'N$ and $102^{\circ}46'E$ at an elevation of 1660 m a.s.l., an average surface temperature of $53^{\circ}C$, and flow rate of 16 l/s. As a result of drilling in 1945-1946, hot water of $48-52^{\circ}C$ was extracted at a depth of 37.4 m. A big tourist center is operated nowadays at the base of this source.
- The Noyonkhangai hot spring is located at $47^{\circ}45'N$ and $99^{\circ}25'E$ in the Khangai soum of the Arkhangai province at an elevation of 2370 m a.s.l. The average surface temperature is $37^{\circ}C$, and the flow rate is 6 l/s. The hot spring is located 10 km southwest of the Khangai soum center, and 240 km of northwest of the Tsetserleg province center. This spring branches from 10 small springs, and flows into the river of Noyonkhangai (Dolgorjav, 2002).

6.2 Properties of the geothermal fluids

6.2.1 Chemical composition

From twenty-one hot springs in the Khangai area, seven were selected with the intention to be evenly distributed over the study area. The samples were collected from hot springs with artesian flow rate in the range 1.2-16 l/s. The temperatures measured during the collection of the sample ranged from 37 to $86.5^{\circ}C$. The pH is generally slightly alkaline (8.5-9.45). The TDS of the samples is in the range 200- 342 ppm. All analytical results are presented in Table 12. Ionic balance calculated by the WATCH program (Arnórsson et al., 1983a; Bjarnason, 1994) gives information regarding the quality of the analysis. For

the selected samples the values were ranging from -11.03 to 32.87. These are relatively high but the samples nevertheless have been used for interpretation.

TABLE 12: Chemical composition of hot springs in the Khangai area, Mongolia (in ppm)

No.	Sample no.	T (°C)	pH	SiO ₂	Na	K	Ca	Mg	Fe	NH ₃	CO ₂	SO ₄	Cl	F	H ₂ S	I.b (%)	TDS
<i>Tsenkher hot spring</i>																	
1	1977-0102	86.5	8.8	133.1	107	4.2	2.12	0.5	-	0.9	61.6	49.0	17.0	24.5	15.89	1.69	338
2	2002-0103	84.3	8.9	113.8	82.4	2.9	2.4	0.4	0.2	0.2	49.3	38.7	17	25.5	10.19	-11	283
<i>Tsagaan sum hot spring</i>																	
3	1977-0202	69	8.9	106.3	97.2	3.2	2	0.3	-	1	58.9	47.6	7.5	19	16.21	6.24	284
4	2002-0203	69.1	9.3	83.13	72.2	1.9	3.21	0.24	0.2	0.8	49.8	28.8	2.48	23.5	11.65	-10.8	216
<i>Shivert hot spring</i>																	
5	1977-0302	48	8.8	106.3	123	6	2.4	0	-	0.2	51.9	72.7	15.8	15.5	8.75	25.35	342
6	2000-0303	57.3	9.5	86.88	96.3	4.1	3.2	1.2	0.2	0.8	48.4	77.4	19.9	17.5	9.1	-9.41	307
7	2002-0304	57.3	9.1	74.56	105	3.9	4	1.2	1	0.6	43.1	78.3	21.88	-	-	24.2	290
<i>Chuluut hot spring</i>																	
8	1977-0402	45	8.7	91.87	105	3.9	1.9	0	-	0.8	68.2	51.1	12.1	17.0	10.68	11.7	284
9	2002-0404	44	9.3	64.94	89.4	3.3	4	1.2	-	1.4	55.61	55.73	18.7	-	-	26.45	239
<i>Khuremt hot spring</i>																	
10	1977-0503	56	8.7	98.13	92.5	2.2	1.3	0.8	-	0.5	80.9	43.4	8.48	8.5	12.97	7.18	256
11	2002-0504	54	9.5	83.12	80.6	1.5	1	0.6	0.8	0	65.6	32.9	5.5	6.2	10.1	3.67	212
<i>Khujirt hot spring</i>																	
12	1973-0603	54.5	8.5	79.38	117	5	2.0	0.1	-	0.8	64.9	31.1	14.7	17.0	12.47	32.87	267
13	1977-0604	55	8.7	103.8	97.9	3.9	1.6	0.1	-	0.3	72.13	32.1	7.17	13.5	11.55	18.83	260
14	2002-0605	48.5	9.35	104.38	82.4	2.8	2	0.2	0.2	0.08	54.17	33.3	15.2	16.1	7.86	-4.77	257
<i>Novonkhangai hot spring</i>																	
15	2002-0702	37	8.8	56.56	60.8	2.1	5.1	2.3	-	0.2	48.1	50.99	11.19	10.5	3	-2.13	200

6.2.2 Ternary diagrams

Figure 16 shows the Cl-SO₄-HCO₃ ternary diagram for the samples from the Khangai area. Most samples plot in the peripheral waters region. Samples number 5, 6 and 7 (1977-0302, 2000-0303 and 2002-0304) from the Shivert hot spring are though in steam heated waters region. The Na-K-Mg ternary diagram is shown in Figure 17. As seen from the diagram, all plots are located between the boundaries between immature and fully equilibrated waters. The Na-K-Mg ternary diagram yields a reservoir temperature range of 120-180°C for all the hot springs.

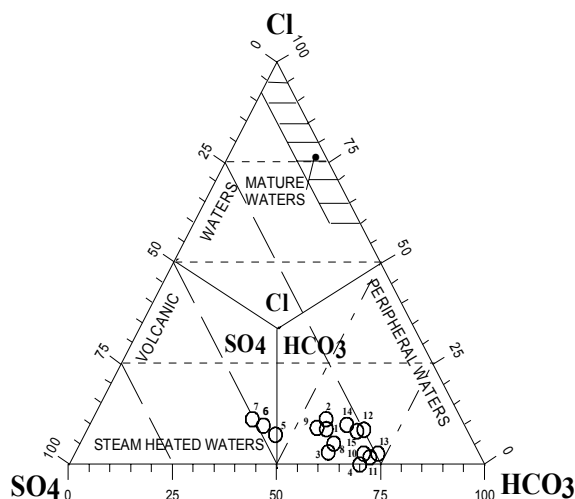


FIGURE 16: Cl-SO₄-HCO₃ ternary diagram for the hot springs in the Khangai area

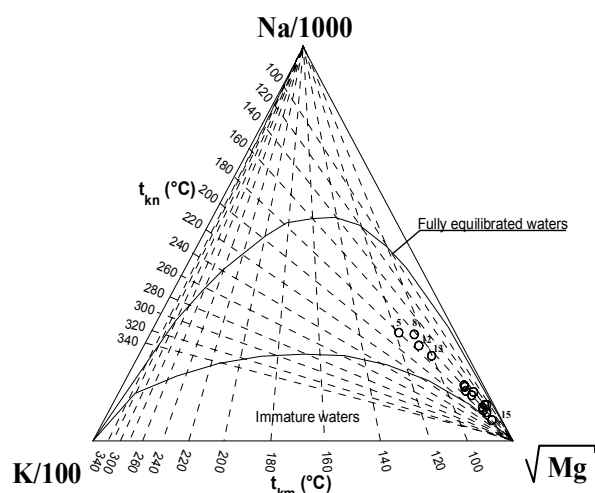


FIGURE 17: Na-K-Mg ternary diagram for the hot springs in the Khangai area

6.2.3 Geothermometry

Subsurface temperatures in the study areas have been estimated by chemical geothermometry. As has been discussed, for various reasons different geothermometers may give different results. Table 13 and Figure 18 show the results for different solute geothermometers.

As seen in Table 13, the chalcedony geothermometer temperatures vary from 69 to 123°C (Fournier, 1977); and from 86 to 126°C (Arnórsson et al., 1983b). Quartz geothermometer temperatures are in the range 100-149°C (Fournier and Potter, 1982); 108-154°C (Fournier, 1977); and 94-142°C (Arnórsson et al., 1983b). The Na-K geothermometer temperatures are in the range 69-132°C (Arnórsson et al., 1983b); 58-122°C (Truesdell, 1976); 111-170°C (Fournier, 1979); and 126-181°C (Giggenbach, 1988). The Na-K-Ca geothermometer temperature varies from 114 to 156°C (Fournier and Truesdell, 1973).

TABLE 13: Results of different geothermometers for samples from hot springs in the Khangai area, Mongolia

Hot spring	Number	T _{meas}	T _{qtz} ¹	T _{qtz} ²	T _{qtz} ³	T _{chal} ¹	T _{chal} ²	T _{NaK} ¹	T _{NaK} ⁴	T _{NaK} ⁵	T _{NaK} ⁶	T _{NaKCa} ⁷
Tsenkher	1977-0102	86.5	149	154	142	123	126	115	105	154	167	145
	2002-0103	84.3	139	145	132	112	116	108	98	148	161	135
Tsagaan sum	1977-0202	69	136	141	128	108	112	102	92	143	156	135
	2002-0203	69.1	116	127	114	87	99	88	77	130	144	118
Shivert	1977-0302	48.0	136	141	128	109	112	132	122	170	181	156
	2000-0303	57.3	114	129	117	85	101	121	111	160	172	144
	2002-0304	57.3	100	121	108	69	93	111	101	151	164	137
Chuluut	1977-0402	45	128	133	120	101	104	111	101	151	164	143
	2002-0404	44	104	114	101	73	86	111	101	151	164	135
Khuremt	1977-0503	56	132	136	124	104	108	83	73	125	140	125
	2002-0504	54	111	127	114	81	99	69	58	111	126	114
Khujirt	1973-0603	54.5	122	125	112	93	96	122	111	160	172	151
	1977-0604	55.0	135	139	127	108	111	116	106	156	168	146
	2002-0605	48.5	126	140	127	98	111	105	95	146	159	135
Noyonkhangai	2002-0702	37	104	108	94	74	79	106	96	147	160	125

- 1) WATCH program;
- 2) Fournier (1977);
- 3) Arnórsson et al. (1983b);
- 4) Truesdell (1976);
- 5) Fournier (1979);
- 6) Giggenbach (1988);
- 7) Fournier and Truesdell (1973).

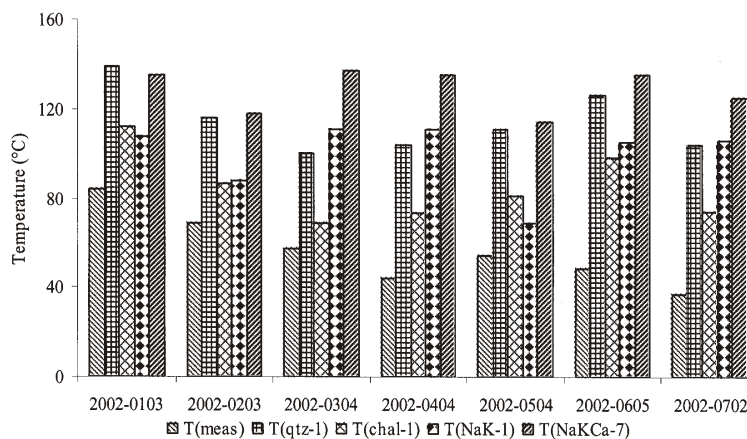


FIGURE 18: Comparison of various geothermomete temperatures with measured temperatures

Compared to the measured values, on average, the quartz geothermometer yields 66°C higher values, the chalcedony geothermometer gives 37°C higher values, the Na-K geothermometer gives 49°C higher values, and the Na-K-Ca geothermometer gives 79°C higher values (see Figure 18). The fact that the

chemical geothermometers persistently predict temperatures significantly higher than observed temperatures in the springs suggests that the geothermal waters may have cooled on their way from the reservoir. The calcedony geothermometer is probably more appropriate than the quartz geothermometer for these hot springs because of the low temperatures of the solutions. Comparison of the resulting temperatures of the chalcedony and Na-K geothermometers shows that they generally agree within some 10 to 20°C. This is encouraging considering the relatively poor charge balance of the chemical analyses (see Table 13).

6.2.4 Log (Q/K) diagrams

The WATCH and SOLVEQ programs were used to calculate the calcite saturation index. The SOLVEQ program was used to calculate saturation indices of 47 minerals. Five main minerals were selected, calcite, cristobalite, diopside, magnesioferrite and quartz. The results obtained are shown in Figure 19 with the broken line indicating measured temperature. For the multi-component chemical equilibria in the Khangai area hot spring waters, the results can be summarized as follows:

- *Tsenkher (2002-0103)*: Convergence of quartz, cristobalite, magnesioferrite and diopside with intersection at 80°C; convergence of quartz, cristobalite and calcite at 100°C;
- *Tsagaan sum (2002-0203)*: Convergence of quartz, cristobalite, magnesioferrite, diopside and quartz at 40°C, and intersection at 60°C;
- *Shivert (2002-0304)*: Convergence of quartz and diopside at 40°C; convergence and intersection of cristobalite and calcite at 60°C;
- *Chuluut (2002-0404)*: Convergence of diopside and quartz at 20°C;

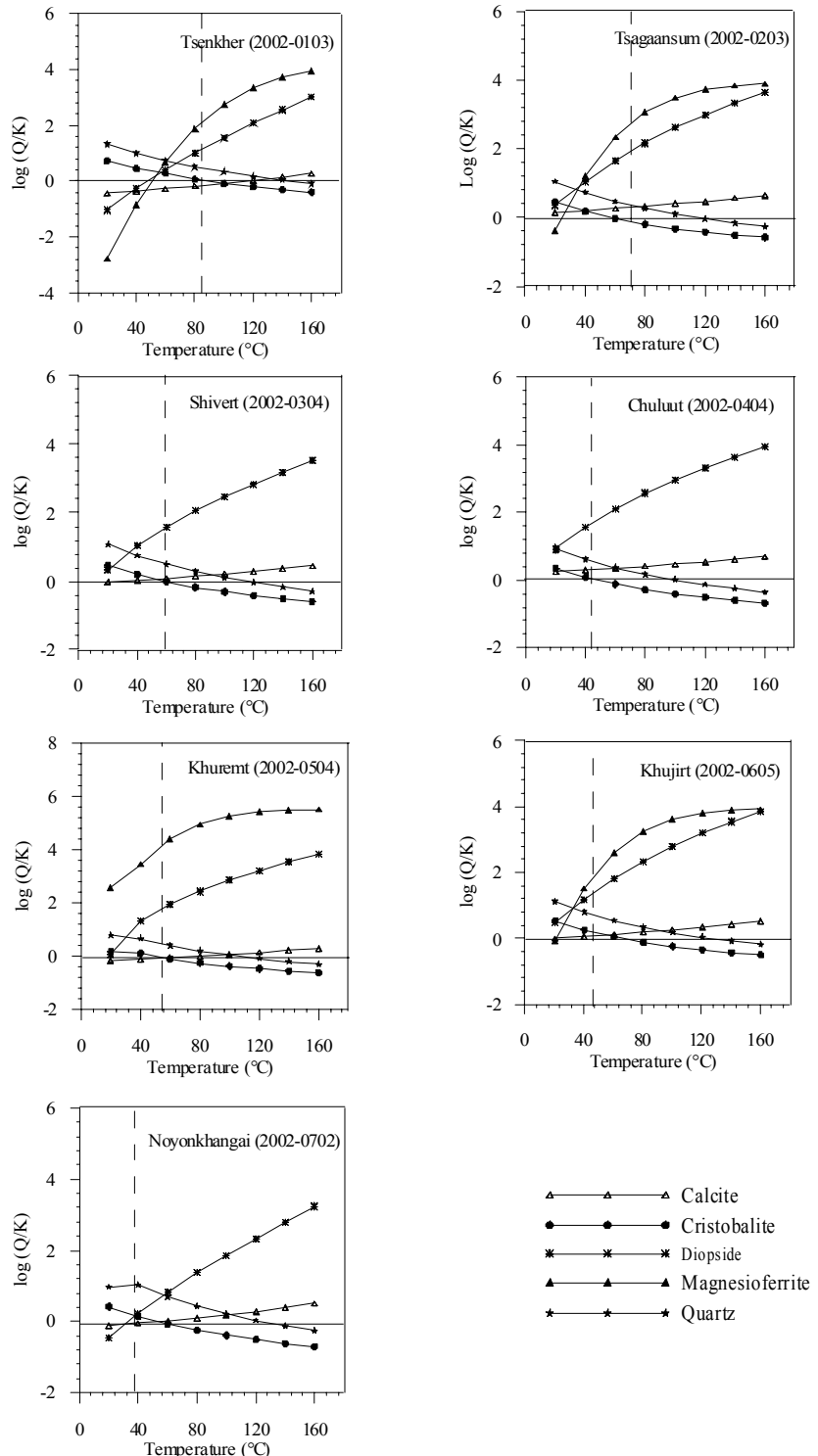


FIGURE 19: Mineral equilibrium diagrams for hot springs in the Khangai area

- convergence of calcite and cristobalite at 40°C, and intersection at 45°C;
- *Khuremt (2002-0504)*: Convergence of diopside and quartz at 30°C, and intersection at 55°C; convergence of calcite and cristobalite at 60°C;
- *Khujirt (2002-0605)*: Convergence of magnesioferrite, diopside and quartz at 35°C; convergence and intersection of cristobalite and calcite at 55°C;
- *Noyonkhangai (2002-0702)*: Convergence of cristobalite, calcite and diopside at 40°C and intersection at 50°C; convergence and intersection of quartz and calcite at 100°C.

6.2.5 Mixing models

Two mixing models were applied to the water in this study; the silica-enthalpy and the silica-carbonate mixing models:

Silica-enthalpy mixing model: In order to estimate the reservoir temperature, the silica-enthalpy mixing model (Truesdell and Fournier, 1977) was used for comparison and further classification. Figure 20 depicts the silica-enthalpy mixing model. The cold water point (A) is assumed to represent the cold ground water ($T=6^{\circ}\text{C}$, and $\text{SiO}_2=12$ ppm) in the study area (Pisarskii, 1986). The intersection point with the solubility curve for chalcedony gives the silica content and enthalpy of the deep hot water component, and its temperature is obtained from a steam table. For the hot springs of Khangai area, two mixing lines, a and b, are possible. Line a intersects the quartz solubility curve at point P_1 , with an enthalpy of 1142 kJ/kg corresponding to a temperature of about 261°C . Line b intersects the quartz solubility curve at P_2 , with an enthalpy of 908 kJ/kg corresponding to a temperature of about 212°C . Evidence of mixing is not obvious in the log (Q/K) diagrams, but the Na-K-Mg diagram suggests that there might be mixing. Again a different choice of silica activity buffer, i.e. chalcedony instead of quartz would have resulted in less predicted mixing and lower predicted reservoir temperatures.

Silica-carbonate mixing model: The silica-carbonate mixing model for this study is shown in Figure 21. In the Khangai area, a linear relationship between silica and carbonate is observed for cold groundwater (point A) and hot water from the Khujirt, Khuremt, Chuluut and Noyonkhangai hot springs, which indicates that the hot water component of the mixed water has not boiled. An extrapolation of a line through

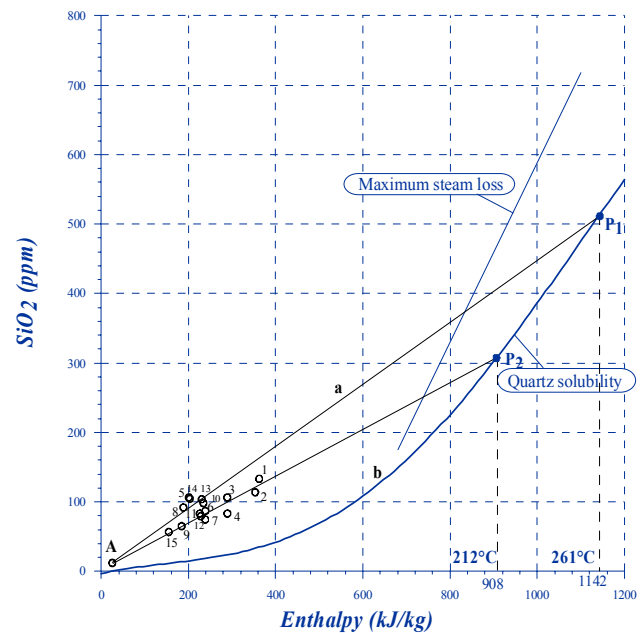


FIGURE 20: The silica-enthalpy mixing model used on samples from the Khangai area hot springs

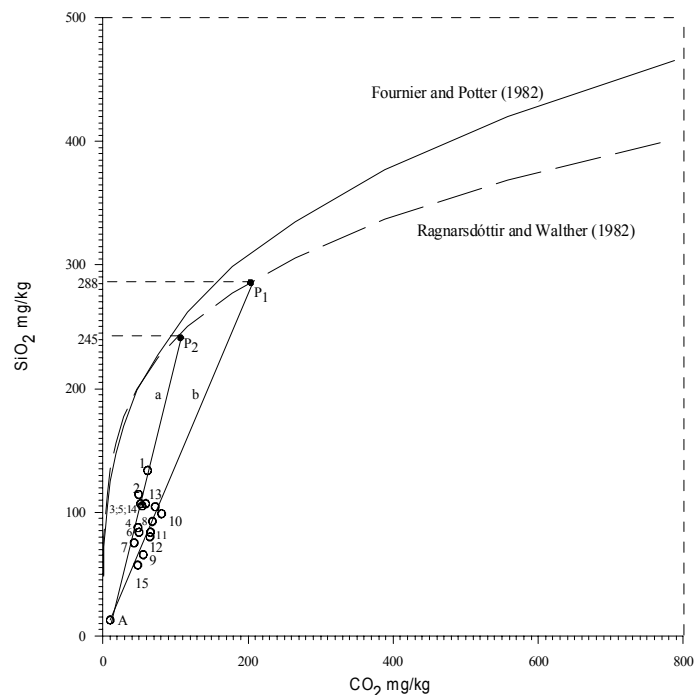


FIGURE 21: The carbonate-silica mixing model used on samples from the Khangai area hot springs

the data points (to point P₂) indicates that the temperature of the hot water component is about 188°C. These hot springs contain relatively high total carbonate (43.1-80.9 ppm). A linear relationship between silica and carbonate is also observed for groundwater (point A) and the hot water from the Thenkher, Tsagaan sum and Shivert hot springs. An extrapolation of the line through the data points (to point P₁) indicates that the temperature of the hot water component is about 201°C.

6.2.6 Scaling tendency

In order to study possible changes with time, available chemical analyses (Table 12) were entered into the WATCH program and the results were used to compute the saturation index. The WATCH program is a good way to study solution and calcite equilibria, especially chemical changes in water chemistry accompanying boiling, and variable degassing, and how these changes affect the calcite/solution equilibria. After that, the Ryznar stability index was calculated using Equation 15. The results are listed in Table 14. Figure 22

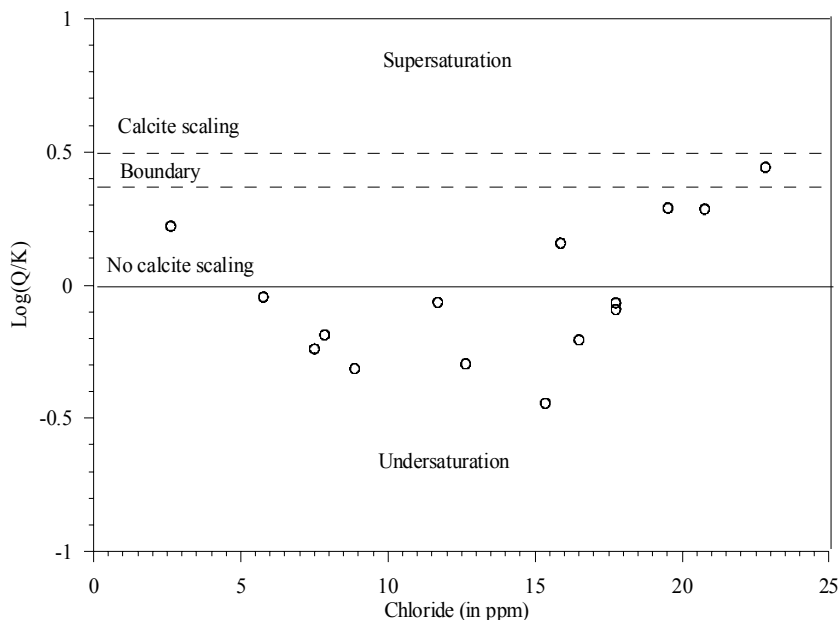


FIGURE 22: Calcite saturation index vs. chloride content of the Khangai hot springs

shows the correlation of calcite saturation index with chloride. According to the WATCH program the saturation index boundary between calcite scaling and no scaling in the selected hot springs is 0.36-0.5.

TABLE 14: Calcite saturation index calculation and Ryznar stability index for samples from selected hot springs, Mongolia

Hot spring	Sample ¹	Ionic balance (%)	LogK	LogQ	SI (Log(Q/K))	RSI	Cl (ppm)
Tsenkher	1977-0102	1.69	-	-	-0.067	7.19	17.0
		-11.03	9.858	9.925	-0.092	7.24	17.0
Tsagaan sum	1977-0202	6.24	-	-	-0.187	7.79	7.5
		-10.84	9.586	9.773	0.221	7.05	2.48
Shivert	1977-0302	25.35	-	-	-0.206	8.53	15.8
		-9.41	9.589	9.795	0.285	7.35	19.9
	2000-	24.2	-	-	0.442	7.6	21.88
Chuluut	1977-0402	11.7	-	-	-0.296	8.7	12.1
		26.45	9.449	9.745	0.289	7.64	18.7
Khuremt	1977-0503	7.18	-	-	-0.314	8.47	8.48
		3.67	9.509	9.823	-0.045	8.16	5.5
Khujirt	1973-0603	32.87	-	-	-0.444	8.54	14.7
		18.83	9.325	9.769	-0.24	8.42	7.17
	1977-	-4.77	-9.57	-9.81	0.157	8.06	15.2
Noyonkhan	2002-	-2.13	-	-	-0.065	8.29	11.19

If SI is higher than 0.5, calcite scaling problem occurs; if it is lower than 0.36, no scaling will occur. Figure 23 shows the relationship between saturation index and temperature for selected hot springs. Results from the SOLVEQ program were used. The Ryzner stability index was also used to analyse potential calcite scaling in low-temperature hot springs in Mongolia. The Ryznar index gives a qualitative estimate of the calcium carbonate scaling tendency of fluid and can be used in the temperatures range from 37 to 86.5°C. The results are shown in Figure 24. It shows that the Ryznar stability index increases with a decrease in the calcite saturation index. Calcite scaling occurs at low RSI values and high SI values. The results from the WATCH program confirm the results from the Ryznar stability index.

The general undersaturation with respect to calcite observed for many of the Mongolian hot spring samples is in qualitative agreement with the high temperatures predicted by the chemical geothermometers. This is because calcite solubility increases with decreasing temperature so that a geothermal solution in equilibrium with calcite in the geothermal reservoir will become undersaturated with respect to calcite upon isochemical cooling.

6.3 Summary

The reservoir temperatures indicated by the calculated chalcedony geothermometers (69-126°C) are closer to the measured temperatures of the water than the values given by other geothermometers. Mixing models were also used to estimate the reservoir temperature, and gave a temperature range of 181-261°C.

Most of the samples plotted in the peripheral waters region, except for the Shivert samples 5, 6 and 7 (1977-0302, 2000-0302 and 2002-0304) that plotted in the steam heated waters region. The Na-K-Mg ternary diagram yielded a reservoir temperature range of 120-180°C for the selected hot springs. The

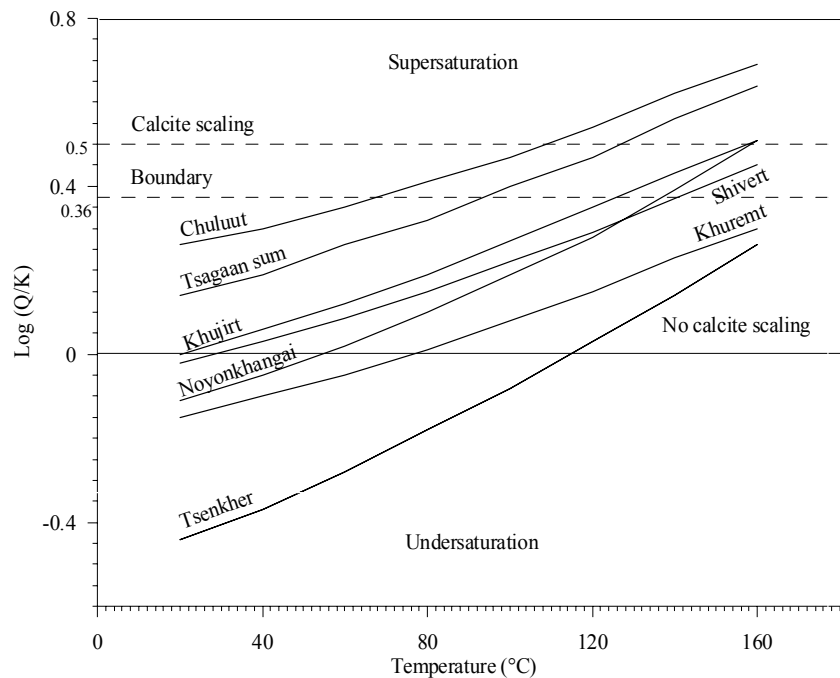


FIGURE 23: Calcite saturation index vs. temperature for the Khangai hot springs

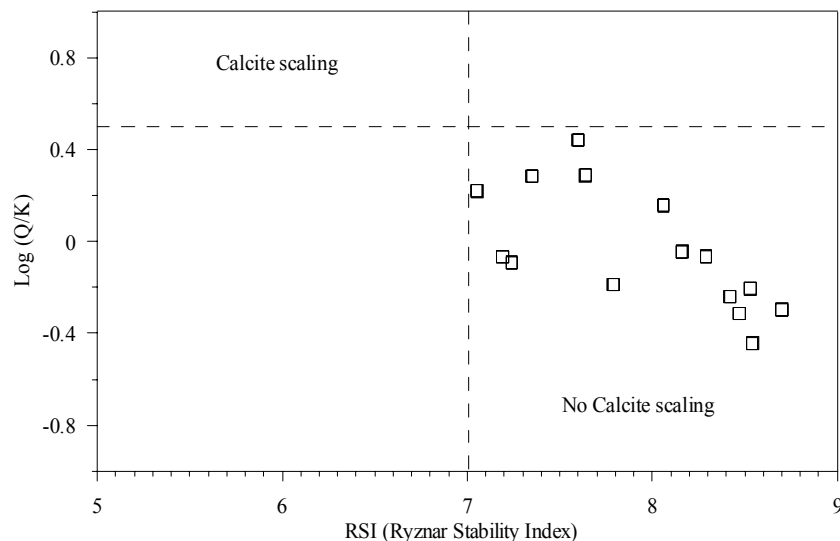


FIGURE 24: Ryznar stability index vs. calcite saturation index for the Khangai hot springs

Shivert hot spring is of sulfate-sodium type, while the other hot springs of the Khangai area are of bicarbonate-sodium type. For all the hot springs, the calcite saturation index values are low, between -0.444 and 0.289, and the RSI values high from 7.05 to 8.54, indicating that calcite scaling will clearly not become a significant problem in any of the geothermal solutions considered in this study.

7. CONCLUSIONS

The following summarizes the results of the geochemical studies of the Árskógsstönd (well ÁRS-29), Dalvík (wells HA-10, HA-11) and Hrísey (well HR-10) geothermal fields:

- The geothermal fluids from the wells in Árskógsströnd, and Dalvík can be classified as bicarbonate-sodium chloride waters, and those from the Hrísey geothermal field as sodium-chloride waters.
- Considering the major anions, the fluids from wells at Árskógsströnd and Dalvík are classified as peripheral waters, whereas those from well HR-10 Hrísey are close to the chloride corner.
- Taking into account the major cations, the geothermal fluids are classified as close to fully equilibrated, with geothermal fluids of the selected wells falling near Giggenbach's (1988) line for fully equilibrated waters.
- The chalcedony geothermometer gives the values closest to measured values compared to the quartz, Na-K, and Na-K-Ca geothermometers.
- The dissolved silica-enthalpy diagram for determining the temperatures of the hot water components mixed with cold water, indicates higher temperatures than the results given by the chalcedony geothermometer. This could possibly be due to a mixing of the hot water from the reservoir with infiltrated cold water in the upper layers. There are, however, arguments against high source temperatures, such as the geology of the area, which makes a powerful heat source unlikely, as well as the Na-K-Mg diagram. Thus, the results of mixing models should be treated with care.
- The results from the WATCH program agree with the results from the Ryznar stability index of scaling tendency. No scaling is predicted in the wells from Dalvík and Árskógsströnd, but scaling might occur in Hrísey.

For the Khangai hot springs in Mongolia, the results of the geochemical studies can be summarized as follows:

- The waters in the Shivert hot spring are of sulfate-sodium type, but in other hot springs waters of the bicarbonate-sodium type.
- Considering the major anions, the Shivert hot spring is classified as steam heated water, but water from the other hot springs as peripheral water.
- With reference to the major cations, the geothermal fluids are classified as partially equilibrated, with geothermal fluids of the selected hot springs lying between Giggenbach's (1988) line for fully equilibrated waters and Arnórsson's (1991) line.
- The chalcedony, quartz, Na-K, and Na-K-Ca geothermometer temperatures were calculated. The Na-K geothermometer gives very high and sometimes unrealistic temperatures for all hot springs, whereas results for the Na-K-Ca geothermometer indicate that this empirical geothermometer appears to be applicable to the low-temperature waters. Quartz geothermometer temperatures are realistic for hot springs in this area compared to estimated mineral-equilibrium temperatures. But the chalcedony geothermometer provides the most reliable temperatures for these fields, with predicted temperature values ranging from 69 to 123°C.
- The silica-enthalpy warm spring mixing model handles non-boiled and boiled mixed waters separately with mixing occurring after boiling, the boiling hot spring waters indicate enthalpies from 908 to 1142 kJ/kg for the hot water component (212-261°C).
- The calcite saturation indices and the Ryznar stability indices on the hot springs of the Khangai area show that there is no scaling tendency. The values for hot spring water vary, but fall in the 'no calcite scaling' range.

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