CONCEPTUAL MODEL OF THE HVERAGERDI GEOTHERMAL RESERVOIR BASED ON GEOCHEMICAL DATA

Kristján Geirsson and Stefan Arnórsson

Science Institute, University of Iceland, Dunhagi 3, IS 107 Reykjavík Iceland

ABSTRACT

A combined study on the chemistry of geothermal water and steam along with temperature data froin wells results in a conceptual model of the Hveragerdi high-temperature geothcrrnal field in southern Iceland. The reservoir water enters the system from the north and flows towards the south. On the way the water cools due to mixing with cold groundwater. Boiling is confined to the uppermost 200 meters in upflow zones. Water entering wells at depth is generally in equilibrium with secondary minerals. The Cl/B ratio is higher in water from springs than in water from wells due to mixing between geothermal water and old groundwater of a marine origin in upflow zones. The complicated effects of boiling and mixing diminish the value of solute geothermometers in estimating subsurface temperatures. Three mixing models were applied to evaluate the temperature of the reservoir water. They all indicate an initial temperature of 230-250 °C. Gas geothrrnnoineters reflect decreasing temperatures from north and they indicate that the temperature of the reservoir water that enters the Hveragerdi iield is 240-250 °C.

1. INTRODUCTION

Gcothcrmal activity in Iceland is divided into high- and low- temperature gcothermal fields (Bödvarsson, 1961). Low- temperature fields (t < 150°C in the uppermost 1000 m) occur outside the active volcanic zones, in places where recent active tectonics has enhanced the bedrock permeability (Arnórsson and Gislason, 1990). The high-tempcrature geothennal fields (t > 200°C at < 1 km depth) are located in the volcanic zones, generally associated with ccntral volcanoes.

The Hveragerdi high-temperature geothermal field is on the eastern margin of the western rift zone, close to the southern Iceland seismic zone. The geothermal activity consists of fumaroles dominating in the northern part of the system, and hot springs that are more common below 100 m altitude in the southern part. Numerous drillholes have been sunk into the Hveragerdi field (Fig. 1). Temperature profiles in all deep drillholes are reversed where the temperature reaches a maximum and subsequently decreases with increasing depth (Fig. 2). The highest tempcrature is observed in the northernmost drillhole. H1. The temperatures of the maxima decreases southward, and the depth to the temperature maximum diminishes. From the drillhole tcmpcrature data it is expected that boiling of the rising hot water is confined to the uppermost 200 m.

2. WATER CHEMISTRY

2.1 Mobile Elements

Chloride and boron are considered to behave as mobile elements in Icelandic geothermal systems (Arnhrsson and Andrésdóttir, submitted). Their sources of supply *are* seawater spray and aerosols in the precipitation, the rock being dissolved and. in some cases, marine groundwater (Arnórsson, 1987; Arnórsson *et al.*, 1989, 1993). Experimental work on water-rock interaction at high temperatures reveal that Cl and B are easily removed from the rock, limited only by their concentration in the rock and the water/rock mass ratio (Ellis and Mahon, 1964, 1967).
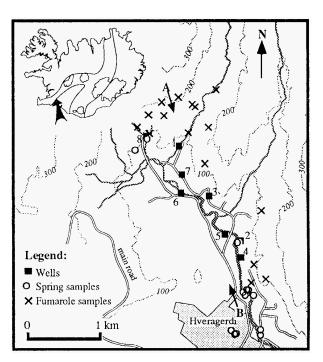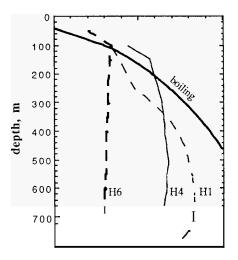

Fig. 1. The Hveragerdi geothermal field in southern Iceland. Location of wells, fumaroles and springs included in this study. The section A—B is drawn in Fig. 10.

The Cl concentration of surface waters in Iceland is variable according to location, ranging from 15 ppm by the coast to less than 2 ppm in the central highlands (Airo, 1982). The Cl/B is similar to that of seawater (4350), or somewhat lower. Chloride concentrations in geothermal waters lie mostly in the range 10-100 ppm if associated with basaltic rocks but up to 500 ppm when associated with silicic rocks (Arnórsson *et al.*, 1978, 1989). The Cl/B in low-temperature water is similar to the ratio in the bedrock (80-120). High temperature geothermal waters sometimes have much lower values of Cl/B, as low as 3. The reason is contribution of B from the magma heat source (Amdrsson and Andrésdóttir, submitted).

In contrast to most geothermal systems, the Cl/B in water from the Hveragerdi field is not uniform. In the wet-steam wells the Cl/B ≈ 180, but 230 in hot and warm springs (Fig. 3). The ratio is higher than in basalts and much higher than in most other high-temperature fields in Iceland. In the Nesjavellir field, north of Hveragerdi, the Cl/B is about 38 (Arnórsson and Andrésdóttir, submitted).
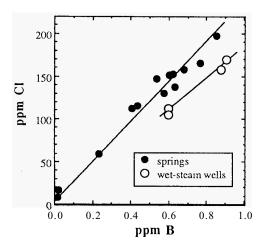

Fig. 3. Chloride vs. boron. Empty and filled circles represent wells and springs respectively.

Both measured temperature and chloride concentrations in wells decrease southward. This indicates that cooling of the geothermal water occurs by mixing with cold groundwater, low in chloride. In upflow zones feeding springs, boiling and/or further mixing with cold groundwater affects the chemical characteristics of the geothermal water. Such processes are not expected to affect the Cl/B of the water, so an explanation of the higher Cl/B in springs, relative to wells, must be sought. The only processes known that can increase an initial Cl/B water ratio of 180 is mixing with seawater, or leaching of marine sediments. It is considered that the deep geothermal water contains a component of old groundwater of a marine origin, like what is observed in the low-tempemture fields in southern Iceland (Arnórsson *et al.*, 1993; Arnórsson and Andrésdóttir, in press, submitted).

2.2 Reactive Elements

All chemical characteristics of water from wet-steam well discharges described below are of calculated deep-water composition, obtained with the WATCH chemical speciation program (Arndrsson *et al.*, 1982). The calculated deep water generally appears to be close to equilibrium with secondary minerals as exemplified for calcite in Fig. 4. On the other hand water from hot springs in the area are relatively far removed from equilibrium with common hydrothermal minerals at the discharge temperature. This is attributed to reactions with the wallrock in the upflow, cooling and boiling, and mixing with shallow groundwater or surface water. Two of the sampled hot springs represent steam heated surface water, and their water is far removed from solution — mineral equilibrium.
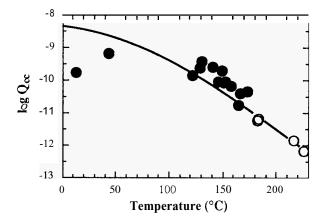

Fig. 4. Calcite solubility data. Legend as in Fig. 3.

Silica concentrations vary approximately linearly with Cl and B concentrations. A better relation is observed between SiO_2 and B than with Cl (Fig. **5).** the reason being dilution of the geothermal water with cold water containing variable chloride but negligible boron concentration as previously discussed. The SiO_2-B relationship indicates that the dilution process has generally not involved significant precipitation of SiO_2 nor dissolution. However, one sample is low in SiO_2 relative to B. The low SiO_2 is best accounted for by precipitation from solution subsequent to mixing. Mixing of equilibrated geothermal water with cold water leads to supersaturation (Arnórsson, 1975) and tends, therefore, to cause silica to precipitate from solution.

2.3 Solute Geothermometry Results

The silica and the Na-K geothermometers were used to estimate subsurface temperatures from hot spring data applying the respective equations from Arnórsson *et al.* (1983b). The equations use Na and K concentrations in the case of the Na-K geothermometer, and in the case of the silica geothermometer unionized silica Concentration derived from the analyzed total silica and measured pH at room temperature, assuming that the pH is only buffered by **aqueous** silica at all temperatures. Silica and Na-K temperatures have also been calculated for well discharges.

Aquifer temperatures in wells and Na-K temperatures of their discharges compare well. On average, the Na-K geothermometer yields 1.5°C lower values. For the coolest well (157°C), chalcedony equilibrium temperature is vely similar to the aquifer temperature but for other wells (aquifer temperatures 181-225°C) the quartz geo-
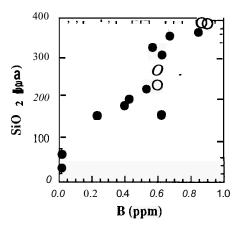

Fig. **5.** Silica and boron concentrations in wet-steam wells and springs. Legend as in Fig. 3.

thermometry temperature compares better being on average 6.9°C higher whereas the chalcedony temperatures are 15.2°C lower.

Geothermometry temperatures for hot springs differ considerably more than those of drillhole discharges (Fig. 6). Silica concentrations in hot spring waters point to subsurface temperatures from 134°C to over 200°C. Many hot springs give values in the range 180 to 200°C so it is uncertain whether chalcedony, rather than quartz-equilibrium temperature is applicable. Na-K geothermometry temperatures for hot springs range from 121 to 167°C. A warm and a slightly tepid spring yield, however, higher temperatures, 199 and 177°C respectively. These high values, which correspond to low Na/K, are considered to result from approximately stoichiometric dissolution of Na and K from the rock rather than equilibration with albite and K-feldspar, the minerals of the Na-K geothermometer.

Many of the hot spring waters are boiling and their content of CO_2 is considerably lower than that of the well discharges indicating that boiling and, therefore, steam loss in the upflow is significant. For this reason chalcedony/quartz geothermometry formulas which assume adiabatic boiling in the upflow are considered to reflect better subsurface temperatures than those which assume conductive cooling (Arndrsson *et al.*, 1983b).

Correlation between silica and Na-K temperatures is rather poor. The silica geothermometer tends to indicate higher temperatures (Fig. 6). There is no observable geographic pattern of the geothcrmometry temperatures across the field. Variations in the silica temperatures are much affected by mixing with cold water in the upflow. Large variations in chloride and boron concentrations of the spring waters constitute the evidence for such mixing. Mineral precipitation/dissolution reactions and boiling also affect the silica geothermometry results but to a lesser extent than mixing. The Na-K geothermometer is not as sensitive to mixing processes as the silica geothermometer. However, waters containing a large cold water component tend to yield high Na-K temperatures (Arnórsson, 1985). The Na/K ratio in cold waters tends to be controlled by stoichiometric dissolution of the respective elements from the primary rock constituents, giving Na-K temperatures in the range of 150-250°C for cold waters in basaltic rocks.
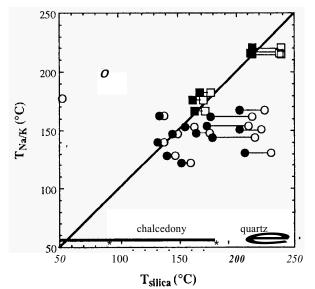

Fig. 6. Comparison between the silica and the Na/K geothermo-meters calculated using formulas from Arndrsson *et al.* (1983b). Silica values higher than 180°C are obtained with the quartz geothermometer. others with chalcedony. Squares and circles represent wells and springs respectively. Filled symbols represent values obtained when boiling is accounted for, open symbols represent non-boiled water.

Due to the mixing that evidently occurs in the upflow in the Hveragerdi system, spring waters highest in silica should present minimum values for the hot water component in the mixture and, therefore, subsurface temperatures, or about 205°C. It has been observed that Na-K temperatures are often lower than silica temperatures of hot springs in high-temperature fields in Iceland (Amdrsson, 1985), the reason most likely being precipitation of K-feldspar in the upflow. This explanation does not hold for the Hveragerdi geothermal field. Sodium and potassium show a relatively good linear correlation with chloride concentrations, and are probably just diluted by the mixing, but precipitation is limited. This explanation also holds for silica which, with a few exceptions, correlates linearly with the chloride concentrations.

2.4 Mixing Models

The different processes responsible for the chemical characteristics of the Hveragerdi waters diminish the reliability of solute geothermometers and their indication of the subsurface temperature. The chloride and boron variations in the geothermal waters are indicative of effective mixing processes Arndrsson (1985) applied three different mixing models to estimate subsurface temperatures in selected high-temperature fields in Iceland. These models have also been applied to the Hveragerdi data to evaluate subsurface temperatures of the geothermal system.

The silica-enthalpy warm spring mixing model developed by Truesdell and Fournier (1977), is based on the silica-enthalpy relationship of hot spring waters. It assumes that conductive cooling and changes in aqueous silica concentrations do not occur in the upflow subsequent to mixing. According to this model the hot water component of the mixed hot spring waters is 215-225°C. The lower value is based on the sample which is highest in silica and assumes adiabatic boiling to 100°C prior to mixing whereas the higher value assumes conductive cooling for hot spring samples lowest in silica.

The carbonate-silica mixing model of Arnórsson (1985) is based on the relationship between silica and total carbonate concentrations in hot spring discharges to estimate subsurface temperatures. The model is also useful to distinguish boiled water from conductively cooled water. The basic assumption for this model is that both aqueous silica and CO_2 concentrations are fixed by temperature dependent equilibria with hydrothermal minerals and that at elevated temperatures (above 200°C), CO_2 is the dominating carbon bearing aqueous species (Arnórsson *et al.*, 1983a). Boiling leads to instant degassing of CO_2 but increasing concentration of silica and points fall above the curve in Fig. 7, whereas mixed waters fall below the curve as both silica and carbonate concentrations decrease. If sufficient samples of mixed, unboiled water are obtained, the aquifer temperature can be estimated by linear extrapolation of the data points to the equilibrium curve. The carbonate-silica mixing model indicates a aquifer temperature of 235°C. The model can also be used to divide the samples into three groups; water from hot springs that has boiled before sampling, predominantly mixed water, and steam-heated, carbonate-rich water low in silica, and with low pH.

The chloride-enthalpy mixing model assumes that the concentrations of the mobile element chloride in geothermal waters are fixed by external variables acting on the system, including boiling, mixing, and its sources of supply. Leaching is not important in the reservoir of an evolved geothermal system as the Hveragerdi field, *so* the variation of the chloride concentration is largely expected to be due to the effects of boiling and/or mixing. Difficulties arise from the influences of the marine groundwater on the chloride concentration in some samples. This addition is not considered to be important in well discharges and the possible increase in chloride concentrations in spring samples has no major effects on the results obtained from the model.

This model indicates an original temperature of 257 °C and that the unmixed geothermal water contained about 170 ppm Cl. The water samples inside the triangle on Fig. 8 represent boiled and mixed waters. A few data points plot above the mixing line between the proposed aquifer water and cold water. There is ample evidence that those samples (263.272) are steam-heated and therefore with higher enthalpy than can be explained by mixing.
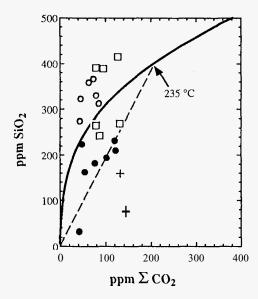

Fig. 7. Carbonate-silica mixing model. The curve is calculated the same way as described in Árnórsson (1985) except that the temperature dependence of silica was assumed to be controlled by quartz solubility according to the data of Foumier and Potter (1982). The broken line indicates silica-CO_2 relationship in mixed, undegassed water and the intersection with the equilibrium curve is a measure of the temperature of the initial hot component. This model also divides the spring samples into 3 groups. Degassed water above the curve (empty circles), mixed water (some partly degassed) plotting between the equilibrium curve and the mixing line (filled circles), and steamheated, CO_2- rich samples (crosses). Empty squares represent wet-steam wells.
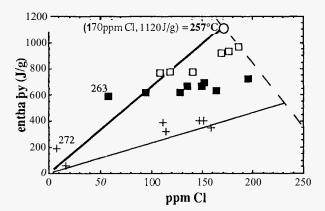

Fig. 8. Chloride-cnthalpy mixing model. Enthalpy values are derived from measured temperatures in wells (empty squares) and cold and warm springs $T < 90°C$ (crosses). For boiling hot springs (filled squares) the enthalpy is derived from the Na/K geothermometer calculated by WATCH. Solid and dotted lines indicate mixing and adiabatic boiling respectively. The large circle represents the estimated chloride (170 ppm) and enthalpy (1120 J/g) of the initial deep-water that flows into the geothermal system.

3. GAS CHEMISTRY

Several gaseous species are dissolved in the aquifer water of geothermal systems. During boiling of the water, these gas components are transferred from the water phase into the steam. When the two phases are separated, the concentration of the gaseous components in the steam reflect the equilibrium conditions in the undegassed geothermal water. During upflow of the steam, however, some components react with the surrounding rock, soil or groundwater, leading to a modification of the original composition of the steam Condensation, oxidation, and reactions due to cooling and pressure drop, are the major processes affecting the chemical characteristics of the steam during ascent.

The concentrations of CO_2, H_2S, and H_2 are controlled by temperature dependent fluid-mineral equilibrium. At temperatures exceeding 200°C and in water of low salinity, the CO_2 concentration is controlled by the mineral assemblage calcite t quartz t epidote + prehnite, but H_2S and H_2 by pyrite + epidote + prehnite +pyrrhotite (Arnórsson and Gunnlaugsson, 1985). Methane concentration is generally low in Icelandic waters and experience has shown that the Fischer-Tropsch equilibrium $CO_2 + 4H_2 = CH_4 + 2H_2O$ is not attained. The reason is probably limited supply of CH_4, slow rate of the reduction process of CO_2, and the mineral control on the concentration of CO_2 and H_2 (Arnórsson, 1986; Arnórsson and Gunnlaugsson, 1985). The concentration of N_2 is not controlled by fluid-mineral equilibria, and is considered to be largely air-derived. In a few cases, magma degassing is shown to have led to increasing concentrations. This is the case with the Krafla geothermal system in northern Iceland, where a rifting episode and magma ascent in the Krafla-Námafjall volcanic system led to an increase in N_2 concentrations in the years between 1975 and 1983 (Gislason et al., 1978; Ármannsson et al., 1982). Argon, like N_2, is introduced into the geothermal system from air and from magmatic degassing. It is generally assumed that the N_2/Ar in undegassed geothermal waters in Iceland should be around 37, the ratio in 5°C water in equilibrium with the atmosphere (Arnórsson, 1986, 1987). Oxygen concentrations are negligible. Detectable O_2 in gas samples suggest air contamination during sampling. Elevated N_2 and Ar concentrations and N_2/Ar ratios are also indicative of air contamination, although not decisive.

3.1 General Characteristics of Steam.

The average concentration of CO_2 in fumarole steam from the Hveragerdi field is about 90% of the volume of total gas, ranging from 60 to 160 mmoles/kg steam. H_2S, H_2 and N_2 range from 0.1 to 9.4 mmoles/kg steam, averaging 3.7, 2.3, and 4.1 % of the total gas volume. CH_4, Ar and O_2 are generally less than 1% of the total gas. Gaseous compounds make only about 0.02% of the total volume of the fumarole steam, considerably less than observed in most other high-temperature geothermal fields in Iceland. Another characteristic of the gas chemistry of the Hveragerdi field is the low concentration of H_2S and H_2 compared to the neighbouring fields at Nesjavellir and Hengladalir (Arnórsson, 1986, Armannsson et al., 1986).

A geographical variation is observed in the chemical composition of fumarole gas within the Hveragerdi field. To the north, the total concentration of gaseous compounds is highest, as is the concentration of CO_2, H_2S, and H_2. The concentrations of these compounds decrease towards the south and reach a minimum in the lowlands around the village of Hveragerdi. The variation is most profound in the concentration of H_2, or by a factor of 90. The variation in concentrations of CH_4 and H_2S is also large, by a factor of 50. It is also apparent that samples collected at higher altitudes in the ridges between valleys contain less H_2S and H_2 than samples

collected at the valley bottoms. The most likely explanation for this is oxidation of these gaseous species above groundwater level. The concentrations of CO_2, N_2 and Ar show less variation, or by a factor of 3, 6, and 10, respectively. The concentration of N_2 and Ar is low, and no clear geographical variation is observed for their concentration or ratio over the field. The relatively stable concentration of CO_2, N_2, and Ar compared with the more reactive species (H_2, H_2S, CH_4) further indicates that the variation in the concentration of the latter species is largely controlled by chemical processes in the upflow zones (condensation, oxidation). The low concentration of CO_2, N_2, and Ar indicate that the extent of condensation of the ascending steam is low, and relatively constant throughout the field.

3.2 Evaluation **of** Condensation

Arnórsson (1987) used the concentrations of CO_2 and N_2 to evaluate the possible extent of steam condensation in upflows of geothermal systems, either by mixing with cold groundwater, or by conductive cooling. A major assumption in these calculations is that the concentration of N_2 in the geothermal non-boiled water at depth is fixed at the value for cold water in equilibrium with air. For Iceland, the mean temperature of the shallow groundwater is that of the annual mean temperature, or 5°C, corresponding to 0.71 mmoles/kg (Weiss, 1970). Such water would accordingly have N_2/Ar = 37. The formulas for the calculations are derived by Arnórsson (1987) and will not be described here. According to these calculations, condensation is not important, being on average 21% if caused by conduction, and 28% if it is assumed that the steam condensation has occurred in cold water. Some samples indicate "negative condensation", which may result from boiling of a previously degassed water.

3.4 Gas Geothermometers

The geothermometers from Arnórsson and Gunnlaugsson (1985), and Arnórsson (1987) were applied for the fumarole steam in the Hveragerdi field. The calculated temperatures divide the samples into two groups according to geographical location. Temperature values obtained from fumaroles in the northern part of the field are higher than from samples collected in the lowlands surrounding the village of Hveragerdi. This is well demonstrated in the temperatures obtained from the CO_2 and H_2 geothermometers (Fig. 9).

The CO_2-geothermometer gives highest temperature values, an average of 267°C for the northern part of the field and 247°C for the southern part. The H_2-geothermometer indicates aquifer temperatures of 230°C and 184°C for the northern and southern part, respectively. The temperatures obtained from the H_2-geothermometer are in an excellent agreement with measured temperatures in wet-steam wells, where for the southern part the temperature ranges from 181°C to 198°C, and in the northern part from 215°C to 230°C. The H_2S-geothermometer indicates aquifer temperature of 208°C with no significant difference between the northern and the southern parts of the field. The CO_2/H_2-geothermometer indicates lowest temperature values, or 205°C for the northern part of the field and 143°C for the southern part of the field. The CO_2/N_2 geothermometer gives temperature values slightly lower than those obtained from CO_2 concentrations. For the northern part of the field the obtained temperature is 251°C but 242°C for the southern part

The differences between the results of different gas geothermometers is in part due to condensation in the upflow but also loss of H_2 and H_2S from the ascending steam by oxidation and reaction with the surrounding rocks. Of the geothermometers used in this study, the CO_2- and CO_2/N_2 geothermometers are inert to chemical reactions in the upflow. Condensation leads to a higher concentration in the steam and thus higher calculated temperatures for the CO_2-geothermometer and it is expected to give too high temperature values. If condensation is taken into account, the CO_2 temperature decreases, averaging 256°C for the northern part and 240 for the southern part of the field. In this study, the temperature values obtained are considered to mark the maximal temperature of the Hveragerdi field, which then does not exceed 260°C. The apparently good correlation between measured temperatures in wells and values obtained from the H_2-geothermometer may be due to the fact that the effects of condensation and oxidation approximately cancel each other. H_2S is known to be lost from the steam due to oxidation and more so than H_2. The results from the H_2S-geothermometer are, therefore, probably too low, especially for samples collected on high ground. The values obtained from the CO_2/H_2-geothermometer are not reliable due to the complex effects of condensation and oxidation. The values are probably too low and the highest values obtained probably represent a minimum for the original temperature of the aquifer water entering the geothermal system, or 230°C.
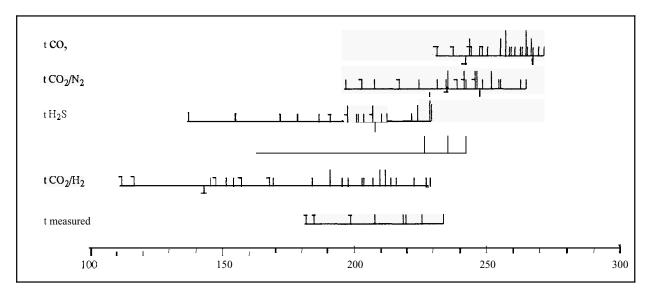

Fig. 9. Gas geothermometry results. Upward pointing bars represent results for individual samples, the shortest bars designate one sample. Bars with horizontal lines identify samples from the southern part of the Hveragerdi field. Downward pointing bars represent averages, for the northern and southern part of the field.

4. SUMMARY AND CONCLUSIONS

The Hveragerdi geothermal field lies on the margin of the active rift zone in western Iceland. Temperature profiles in drillholes show that the reservoir water enters the system from the north and cools as it flows towards south. The correlation between temperature and chloride concentration indicate that the water cools by mixing with cold groundwater. In upflow zones further mixing occurs, where old groundwater of a marine origin mixes with the reservoir water. The marine groundwater has high Cl/B and as a result this ratio is higher in springs than in well discharges.

The geothermal reservoir water is close to equilibrium with secondary minerals but boiling and mixing in upflow zones lead to departure from equilibrium for waters emerging in springs.

The characteristic feature of the geothermal steam of the Hveragerdi field is low gas concentrations, especially for H_2 and H_2S. Boiling is evidently not effective and lateral flow of the reservoir water results in degassing and departure of the more reactive gas components. Condensation appears to be limited in the upflow zones and is not expected to increase the gas content of steam in fumaroles much.

Gas geothermometry results reflect that the temperature of the reservoir water increases to the north. Due to the effects of boiling, condensation and chemical reactions, the CO_2/N_2 geothermometer is considered to reflect best the original temperature of the system, or 240 - 250 °C.

Mixing in upflow zones diminish the usefulness of solute geothermometers but mixing models are more reliable to estimate geothermal reservoir temperatures. The three mixing models applied all give similar results. or 230 - 250 "C.

A combined study of the chemistry of geothermal water and steam alongside with temperature data from wells result in the conceptual model presented in Fig. 10. A stream of geothermal water enters the Hveragerdi field from the north. The temperature of the water is 240 - 250 °C. The water flow is towards south and on the way, the water cools due to mixing with cold groundwater. Boiling occurs in the upflow zones above the main stream of the southerly flowing hot water. During upflow, the geothermal water mixes with old groundwater of marine origin and with high Cl/B, causing spring waters to attain higher Cl/B ratios than the reservoir water.
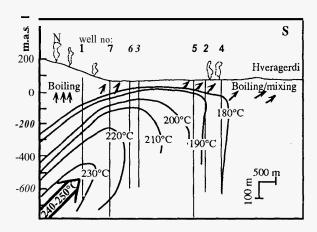

Fig. 10. A conceptual model of the Mveragerdi high-temperature geothermal field. Isograds are drawn according to well data. Shown is the appropriate location and depth of wells. The temperature of the initial reservoir water that flows into the system is derived from three mixing models and results from gas geothermometers. Designed from several sources, including Pálmason (1962). Ragnars et al. (1979) and Xi Xiang (1980).

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