Reviewed research article

Geothermal systems in Iceland

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Abstract — *Geothermal systems in Iceland have been classified as high and low-temperature. High-temperature systems are located within the belts of active volcanism and rifting whereas low-temperature systems are in Quaternary and Tertiary formations. The heat source for the high-temperature systems is high-level magma intrusions. Therefore these systems are volcanic by a commonly used classification of geothermal systems. From data on the natural heat output of several high-temperatures areas, it is concluded that the heat conduction layer between melt and the base of fluid convection is very thin, from several tens of meters to a few hundreds at the most. Most high-temperature fields lie astride active fissure swarms where these swarms intersect the lithosphere plate boundary. Central volcanic complexes have formed at some of these points of intersection and calderas have developed in several of them. Most low-temperature activity in Iceland is known to be associated with recent sub-vertical fracturing and faulting of older crust of the North-American plate and the Hreppar microplate. These systems may therefore be classified as tectonic. Their heat sources are hot rocks at depth. Convection in high-temperature areas is density driven. It is also density driven in some low-temperature areas but in others by hydraulic head. Recorded maximum temperature in drilled high-temperature fields is >380°C at ∼2 km depth. The highest temperature recorded in a low-temperature field is 175°C at 2 km depth. Many fossil high and low-temperature systems in Quaternary and Tertiary formations have been exhumed by erosion providing important information on the geological structure of such systems. The energy current from below Iceland has been estimated as ∼30 GW (1 GW = 10^9 W), corresponding to 5-fold the world average heat flux per unit area. At the surface, this energy current is split as follows; 7 GW comes from rising magma, 8 from fluid flow in geothermal areas and 15 GW from heat conduction. The estimated amount of thermal energy stored in the crust down to 10 km depth is ∼1.2 EJ (1 EJ = 10^{24} J). Above 3 km depth, the energy stored in high-temperature fields is estimated as ∼0.1 EJ. Geothermal fluids in Iceland are meteoric, seawater or mixtures thereof in origin. Low deuterium content in some of these fluids is due to the presence of a Pre-Holocene water component. Primary geothermal fluids that do not contain a seawater component are low in Cl and other dissolved solids. The cause is the low Cl content of the host basalt. Geothermal energy constitutes a very important energy resource in Iceland involving both direct uses of geothermal water and power generation by geothermal steam. In 2005 the annual direct use of geothermal heat was ∼6800 GWh (9% of the worldt's total) and the installed capacity 1844 MWt. At present the installed capacity of power plants using geothermal steam is 484 MWe (5% of the world's total). Four countries use more geothermal heat directly than Iceland and six have higher installed capacity of geothermal power plants.*

INTRODUCTION

Geothermal energy is economically a very important energy resource for some countries, in particular those located in active volcanic regions, although it is not so on a global scale. Today geothermal energy accounts for ∼0.3% of the annual global energy consumption. Deep drillings for geothermal fluids have provided important information on the nature of geothermal systems down to depths of a few km, including their size, temperature conditions, lithology, geological structure, permeability, as well as fluid origin, composition and migration, its boiling and reaction with rock-forming minerals.

Pioneering studies of geothermal processes and the development of geothermal resources were carried out in Italy, California and New Zealand around the middle of the 20th century and earlier. They were linked to the developments of the Larderello, Geysers and Wairakei geothermal fields for power generation. During the Second World War use of geothermal heat on a large scale started in Iceland with space heating in the capital of Reykjavík and few years later in several other towns. A decade earlier use of geothermal water for greenhouse farming started in Iceland.

In 1960 the United Nations organized a symposium in Rome on new sources of energy. This initiative led to much enhancement of geothermal energy development in many countries. The present state of knowledge on studies of geothermal systems and their exploitation is presented in the proceedings of the World Geothermal Congress held in 1995, 2000 and 2005. Further important published information on geothermal resources can be obtained from the proceedings of the Geothermal Resources Council and the Stanford and New Zealand Geothermal Workshops.

The present contribution summarizes the present-day understanding of the nature of geothermal systems in Iceland with emphasis on the geological features that characterize them and lead to their formation. A summary of the current use of geothermal resources in Iceland is also given.

TYPES OF GEOTHERMAL SYSTEMS AND THEIR LOCATION

Bödvarsson (1961) classified geothermal systems in Iceland into high and low-temperature types. According to Fridleifsson (1979) temperatures are below $\sim$150°C in the uppermost 1000 m of low-temperature systems but >200°C above 1000 m depth in high-temperature systems. The classification of Bödvarsson (1961) is a practical one in the sense that potential use of geothermal resources mostly depends on reservoir temperatures. Yet, this classification is related to geology, most of the high-temperature systems are volcanic, i.e. the heat source is high-level magma bodies intruded into brittle crust and permeable rock. Most of the low-temperature ones are associated with young tectonic fractures.

Several classification schemes have been proposed for geothermal systems in the world. Thus, White *et al*. (1971) divided geothermal systems into hot-water (also termed liquid-dominated) and vapour-dominated systems. Goff and Janik (2000) divided geothermal systems into five categories as (1) young igneous systems, (2) tectonic systems, (3) geopressurized systems, (4) hot dry-rock systems and (5) magma tap systems. All of the nine high-temperature systems that have been drilled so far in Iceland are of the hot-water type. It is considered very unlikely that any vapour-dominated systems exist in Iceland. In his early classification of geothermal systems, Bödvarsson (1961) considered four systems, which are by the margins of the active belts of volcanism and rifting, as borderline cases between high and low-temperatures types. Arnórsson (1985) estimated temperatures in two of these areas (Geysir Area and Hveravellir in central Iceland, see Figure 1C) by chemical geothermometry as >250°C so they may truly be regarded as high-temperature. This has, however, not been verified by drilling. Drillings in the other two areas, Hveravellir in Reykjahverfi in NE Iceland and Hveragerdi in SW Iceland, have verified that the first is of the low-temperature type but the latter high-temperature.

The regional geology of Iceland is basically the product of the relative movement of the Iceland mantle plume and the Mid-Atlantic divergent plate boundary. The mantle plume is migrating eastward relative to the plate boundary (Saemundsson, 1974; Jóhannesson, 1980). This relative movement has caused the volcanic belts in Iceland to shift towards east with time because crustal dilation preferentially occurs above the mantle plume (Figure 1A). This eastward shift of the volcanic belts with time is clearly reflected in their present position relative to the crest of the Mid-Atlantic Ridge to the north and south of Iceland (Figure 1B). The direction of the tensional forces responsible for the separation of the North-American and Eurasian plates is generally not at right angles to the axes of the volcanic belts, i.e. the plate boundary as defined by the central volcanic complexes within them. As a consequence, zones of fissuring and normal faulting developed by the crustal tension cut obliquely across the plate boundary. Recent volcanic activity is characteristically most intense where these zones cut across the plate boundary. In the eastern active volcanic belt, central volcanic complexes have formed at the intersection point but not in the western belt, except possibly for Hengill (see Figure 1C).

The high-temperature geothermal systems are located in the central parts of the belts of active volcanism and rifting except for three that are located close to their margin (Figure 1C). In the eastern volcanic belt, these systems are typically found within volcanic complexes at the intersection of fissure swarms and the plate boundary. The northern part of the western volcanic belt between Lake Thingvallavatn and the Langjökull glacier is devoid of high-temperature activity but in the southern part of this belt, south of Lake Thingvallavatn and on the Reykjanes peninsula, several high-temperature systems occur. As in the eastern belt, they are located at points of intersection between fissure swarms and the plate boundary. No volcanic complexes have, however, developed here except for Hengill. Dilation is diminishing in the western volcanic belt north of Lake Thingvallavatn (LaFemina *et al.*, 2005).

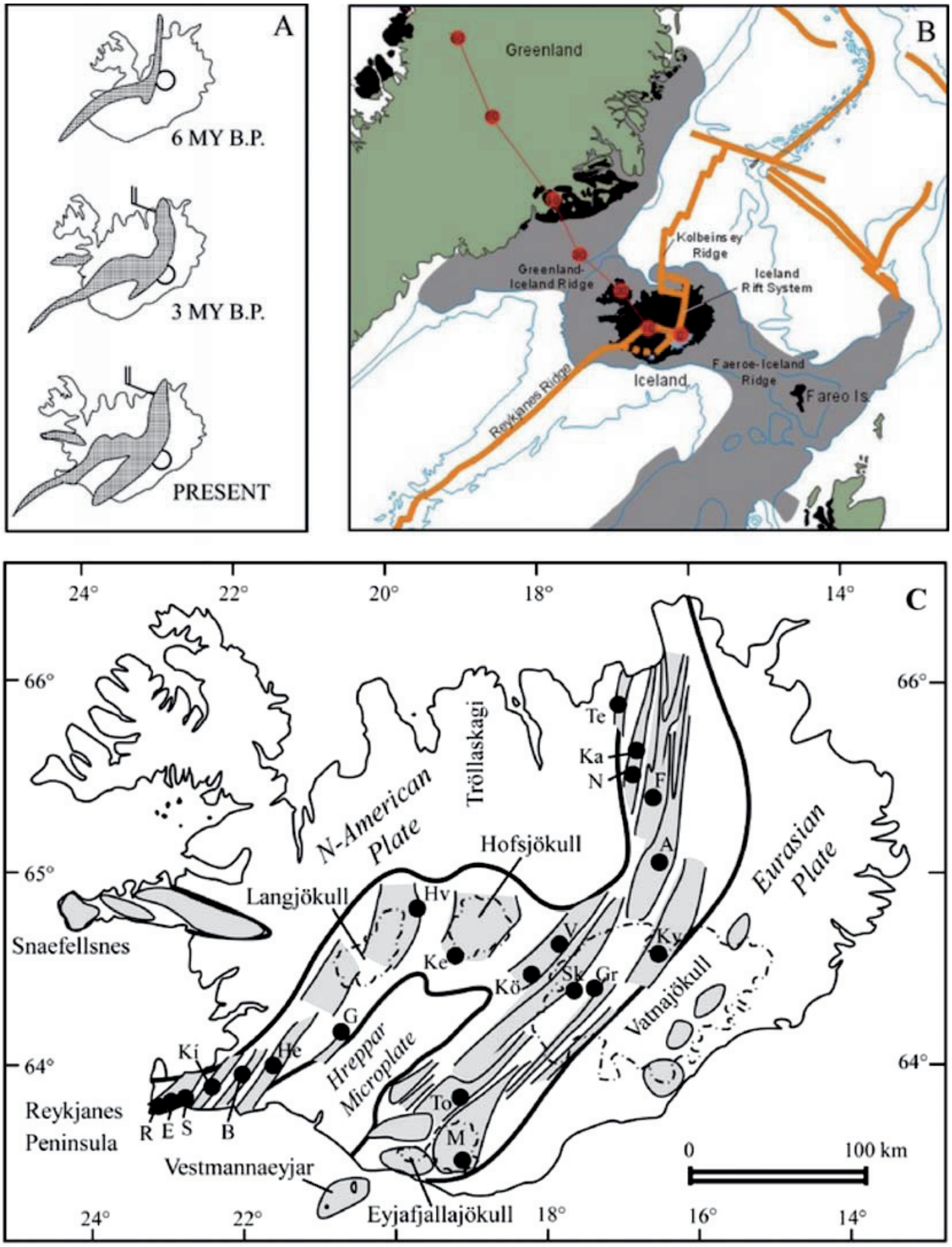
Low-temperature activity is widespread in the Quaternary and Tertiary formations of Iceland (Figure 2) but its intensity, that is the combined magnitude of temperature and flow rate shows a distinct pattern. The activity is largely confined to the North-American plate and the Hreppar micro-plate in south and southwest Iceland (Figure 3). The low-temperature fields with the highest heat output are located on both sides of the volcanic belt in southwest Iceland. These fields also have the highest temperatures, $\sim$150°C. Pálmason (1973) considers that the relatively intense low-temperature activity in southwest Iceland is the consequence of high temperature gradient in this area (Figure 4). It has also been proposed that the intense activity in the Tertiary formations west of the belt is due to intrusion of magma from the volcanic belt along stratigraphic horizons (Arnórsson and Ólafsson, 1986). A fault system extending west from Langjökull, however, rather favours dyke intrusion being responsible (Saemundsson, 1967).

The low-temperature activity is frequently seen to be associated with active fractures and faults. Some systems are located within active fissure zones that run obliquely across the active volcanic belt and into older formations. Other systems form where build-up of stress in the crust by the plate movement leads to deformation by fracturing. Björnsson *et al.* (1990) and Arnórsson and Gíslason (1990) conclude that regional tectonics and the resulting local stress field constitute the main control of the distribution of low-temperature activity. However, low-temperature geothermal reservoirs have also been discovered by deep drillings in several places in Tertiary formations where there are no surface manifestations (Axelsson *et al.*, 2005).

ENERGY CURRENTS AND STORAGE OF HEAT IN THE CRUST

The geothermal systems of Iceland are manifestations of heat stored in the crust as well as heat flowing through it. The energy of these two heat components was estimated separately in the 1980's. Bödvarsson (1982a) estimated the size of the total heat flow through the crust while Pálmason *et al.* (1985) estimated the amount of thermal energy stored in the crust. Both studies were based on available data on thermal gradients in holes drilled into impermeable formations where temperatures were considered to be unaffected by flow of ground water. A recent map of thermal gradient in Iceland is shown in Figure 4. It is largely based on Flóvenz and Saemundsson (1993).

By the methods of Bödvarsson (1982a) and Pálmason *et al.* (1985), the geothermal potential of Iceland is estimated quite differently, which in fact reflects the dual nature of geothermal activity. Stefánsson (2000) combined the results of the two stud-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ies in a unified presentation (Figure 5). He concluded that the high-temperature geothermal systems likely represent a renewable energy resource. This is considered here to apply to those systems that withdraw heat from magma through a thin conductive layer (see section on heat transfer below). However, some high-temperature systems may have developed a large body of hot rock and hot fluid during their lifetime and utilization of such systems will involve, at least partly, extraction of heat from this hot body of rock through enhanced cold water recharge brought about by reservoir drawdown in response to exploitation.

According to Bödvarsson (1982a) the energy current from below Iceland is about 30 GW (1 GW = 10^9 W). This includes 24 GW (210 TWh/a[1]) by flowing magma and 6 GW (53 TWh/a) by heat conduction (Figure 5). He only included the parts of the country above sea level, while considerable additional energy also flows up through the ocean floor around the island. At the surface, 7 GW (61 TWh/a) of the total heat flow is due to volcanic activity, 8 GW (70 TWh/a) is transported by water and steam convecting in geothermal areas, and 15 GW (131 TWh/a) by heat conduction. According to these numbers the energy flux through Iceland per unit area is about five times the world average. Bödvarsson (1982a) also pointed out that the 8 GW transported by water and steam in Icelandic geothermal areas is about 1/10 of the corresponding energy flow through all land-based geothermal areas worldwide.

The principal result of Pálmason *et al*. (1985) is that the total energy stored in the crust of Iceland, from surface down to 10 km depth, amounts to about 1.2 EJ (1 EJ = 10^{24} J). Above 3 km depth the energy stored is only about 0.1 EJ ($27*10^6$ TWh) of which $6*10^6$ TWh are assumed to be accessable and $1*10^6$ TWh harnessable (Figure 5). Again these results only apply to the crust below Iceland. The energy density (concentration) is greatest within the volcanic zone, in particular in the high-temperature systems. The thermal energy stored in 5 of the largest high-temperature systems is estimated to account for 70% of the total

[1] Terawatthours per annum

Figure 1. (A) Evolution of the volcanic belts (shaded) in Iceland in relation to the Iceland mantle plume (circle). Based on Óskarsson and Sigvaldason (1985). (B) Position of the active belts of volcanism and rifting in Iceland in relation to the crest of the Mid-Atlantic Ridge to the north and south of the country (yellow lines). Also shown (by red line) is the relative eastward movement of the Iceland mantle plume. The numbers inside the red dots indicate millions of years before present. Based on Saunders *et al*. (1997). (C) Active volcanic belts (delineated by thick lines), volcanic systems (shaded) and known *high-temperature* fields (dots) in Iceland. R: Reykjanes, E: Eldvörp. S: Svartsengi, Kí: Krísuvík and Trölladyngja, B: Brennisteinsfjöll, He: Hengill with three wellfields (Hellisheidi, Hveragerdi and Nesjavellir), G: Geysir Area, Ke: Kerlingarfjöll, Hv: Hveravellir, M: Mýrdalsjökull, To: Torfajökull, Gr: Grímsvötn, Sk: Skaftárkatlar, Kö: Köldukvíslarbotnar, V: Vonarskard, Kv: Kverkfjöll, A: Askja, F: Fremrinámur, N: Námafjall, Ka: Krafla with three wellfields (Leirbotnar, Sudurhlídar and Hvíthólar), Te: Theistareykir. Local names mentioned in the text are shown. Based on Arnórsson (1995a). – *(A) Vensl þróunar gosbelta (skyggð svæði) á Íslandi við íslenska möttulstrókinn (hringur). (B) Lega virku gosbeltanna miðað við Mið-Atlantshafshrygginn fyrir norðan og sunnan landið (gular línur). Einnig er sýnd (með rauðri línu) austlæg hreyfing möttulstróksins miðað við flekaskilin. Tölurnar inni í rauðu hringjunum sýna legu stróksins í milljónum ára fyrir nútíma. (C) Virk gosbelti (afmörkuð með sverum línum), eldstöðvakerfi (skyggð svæði) og þekkt háhitasvæði (punktar) á Íslandi. R: Reykjanes, E: Eldvörp, S: Svartsengi, Ki: Krísuvík og Trölladyngja, B: Brennisteinsfjöll, He: Hengill með þremur vinnslusvæðum (Hellisheiði, Hveragerði og Nesjavellir), G: Geysissvæði, Ke: Kerlingarfjöll, Hv: Hveravellir, M: Mýrdalsjökull, To: Torfajökull, Gr: Grímsvötn, Sk: Skaftárkatlar, Kö: Köldukvíslarbotnar, V: Vonarskarð, Kv: Kverkfjöll, A: Askja, F: Freminámur, N: Námafjall, Ka: Krafla með þremur vinnslusvæðum (Leirbotnar, Suðurhlíðar og Hvíthólar), Te: Þeistareykir.*

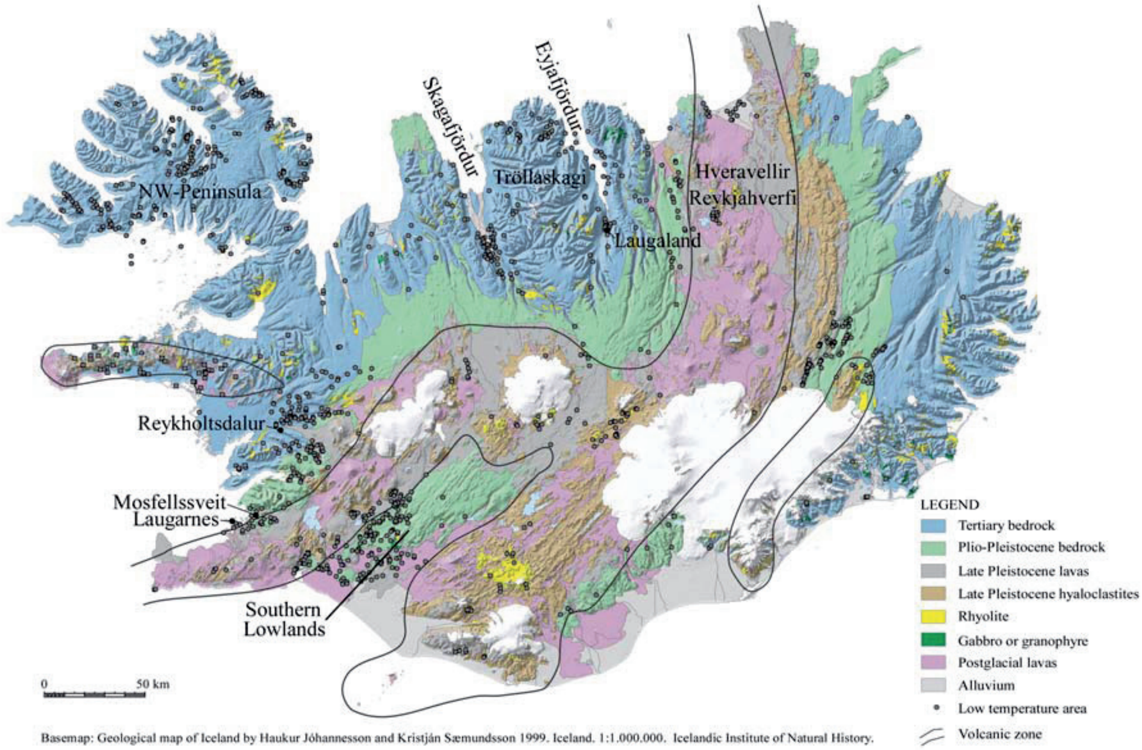

Figure 2. Distribution of low-temperature activity in Iceland. – *Dreifing lághitasvæða á Íslandi.*

energy stored within all high-temperature systems in Iceland.

By combining the estimated thermal energy stored above 10 km and the heat flow, it is possible to estimate roughly how long it has taken the thermal energy to accumulate in the crust, or the time-scale on which this stored energy is renewed. In this way an estimate of the order of 1.3 million years is obtained. But it is clear that energy replenishment takes place at drastically variable time-scales, depending on the mode of heat transportation involved. Thus energy replenishment through the flow of magma, water or steam is several orders of magnitude faster than replenishment by heat conduction alone.

GEOLOGICAL STRUCTURE OF HIGH-TEMPERATURE GEOTHERMAL SYSTEMS

Active systems

Most of the high-temperature areas in Iceland are located at the plate boundary where it is intersected by a fissure swarm (see Figure 1C). Volcanic activity is most intense at these points of intersection. In the eastern volcanic belt, central volcanic complexes that rise above the lava plateau have formed at these points of intersection. In some of them calderas have formed (Figure 6). In the western active belt of volcanism and rifting, volcanic complexes have not formed except possibly for Hengill (Figure 1C). Few high-temperature systems are located close to the margin of the active volcanic belts. These systems are con-

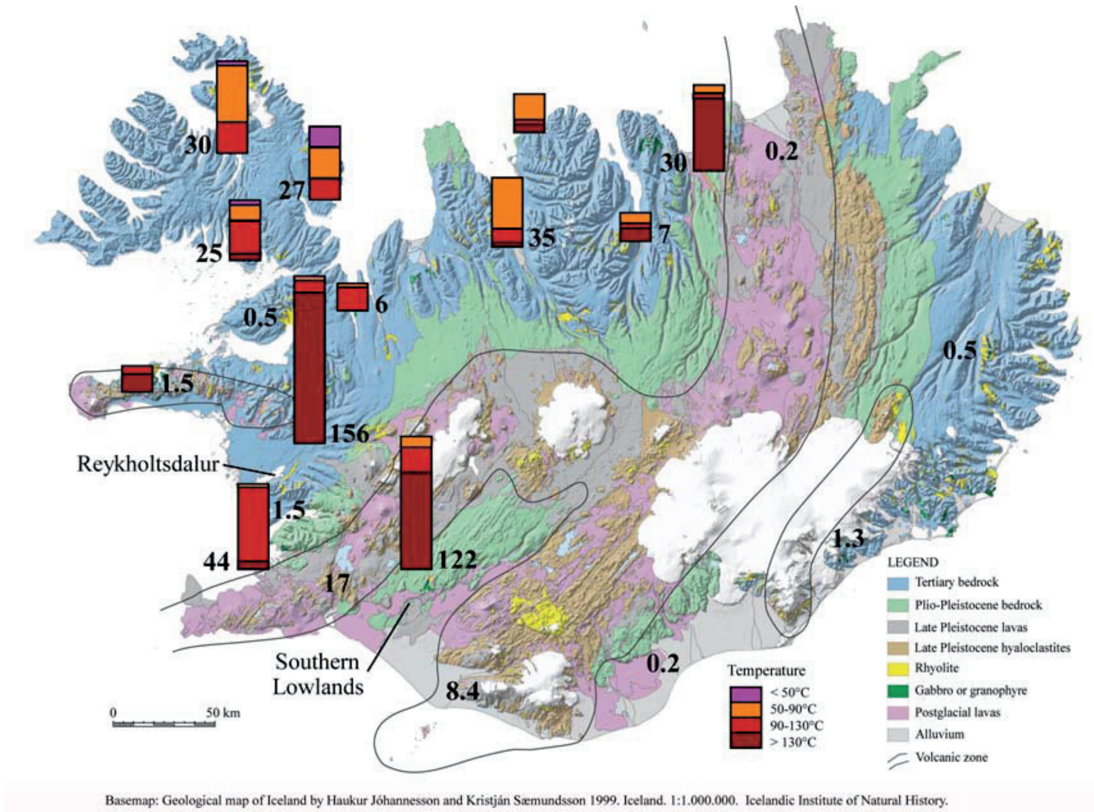

Figure 3. Heat output (in MW above 15°C; numbers by columns) of thermal springs in the low-temperature areas of Iceland by county as they were before drilling. Far the largest low-temperature fields are located in SW-Iceland on both sides of the western volcanic belt, in Reykholtsdalur and the Southern Lowlands. The columns represent the temperature of the geothermal waters as estimated by the chalcedony chemical geothermometer. The height of the columns is proportional to flow rate. – *Varmaútstreymi (í megawöttum ofan við 15°C; tölur við súlur) frá heitum uppsprettum á lághitasvæðum á Íslandi eftir sýslum eins og það var áður en boranir hófust. Langstærstu lághitasvæðin eru á Suðvesturlandi beggja vegna við vestara gosbeltið, í Reykholtsdal og á Suðurlandsundirlendi. Súlurnar sýna hita vatnsins samkvæmt kalsedón-efnahitamælinum. Hæð súlanna er í réttu hlutfalli við rennsli.*

sidered to be mature. Likely they are drifting out of the volcanic belts and in the process of cooling down because they have been cut off from their magmatic heat source.

In some of the larger high-temperature areas, silicic volcanics are abundant including Torfajökull and Kerlingarfjöll. At Torfajökull a prominent gravity low is associated with the silicic volcanics but within this low there is a gravity high. Walker (1974a) considers that this high reflects mafic intrusives below the silicic volcanics that constitute the heat source for the geothermal system. The silicic volcanics acted as a density trap for rising mafic magma, thus explaining association of intense geothermal activity with silicic rocks. Mafic intrusions underlying the silicic volcanics may also constitute the heat source to

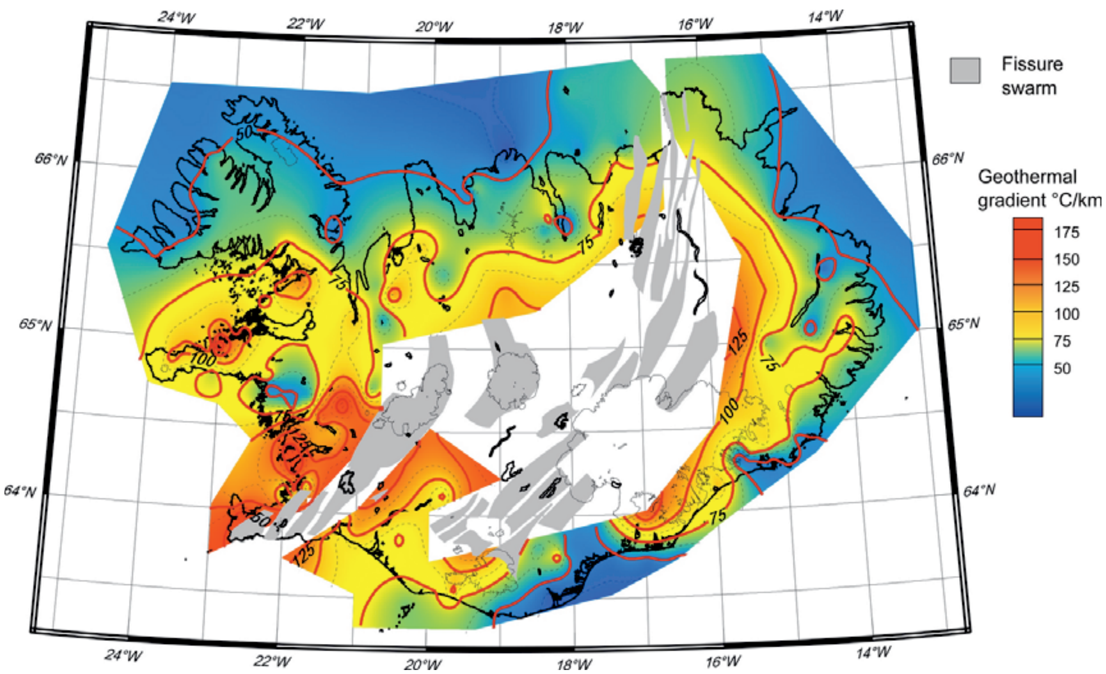

Figure 4. Map of regional geothermal gradient in Iceland. Within the young formations of the active volcanic belts and Upper-Pleistocene rocks (blank area) permeability is high so the geothermal gradient is much disturbed by ground water flow. Based on Flóvenz and Saemundsson (1993). – *Kort af svæðisbundnum hitastigli á Íslandi. Í yngri jarðmyndunum innan virku gosbeltanna og síðkvarteru bergi (autt svæði á mynd) er lekt há þannig að hitastigullinn er truflaður af grunnvatnsrennsli.*

the geothermal system at Kerlingarfjöll. Sediments within the caldera of Torfajökull dip away from its center by as much as 30° suggesting up-doming by rising magma (Saemundsson, 2007).

During Quaternary times, hyaloclastite edifices were formed by sub-glacial eruptions. These hyaloclastite formations have lower bulk density than basalt lava flows. One may speculate that these sub-glacial formations act as density traps for rising mafic magma causing intrusive activity and therefore high-temperature geothermal activity to be more intense in Iceland in Quaternary and Recent times than during earlier geological periods.

Evidence of phreatic explosions is abundant in some high-temperature fields, such as Krísuvík (see Figure 1). They may be purely hydrothermal but it is also common that magma is brought to the surface by these eruptions. In the latter case, these eruptions bear witness of intrusion of magma into ground water bearing formations. Poor permeability of these formations may have led to accumulation of vapour above a boiling hot-water reservoir. This causes vapour pressure at deep levels to be transferred to higher levels and if the roof yields, a hydrothermal explosion results. Hydrothermal explosions triggered by another mechanism may have been particularly common around the end of the Pleistocene when the ice-sheet was melting. The ice melting reduced hydrostatic head in the geothermal systems and brought rock with temperature above 100°C to the surface. Contact of water with this rock could have produced super-heated steam and in this way led to hydrothermal explosions until this rock had cooled down by the circulating water to a temperature corresponding with the boil-

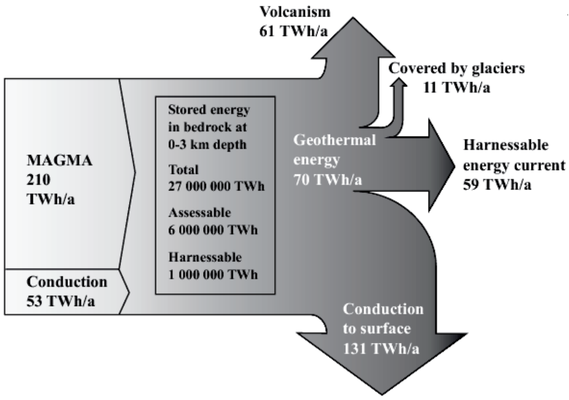

Figure 5. Simplified sketch of the terrestrial energy current in Iceland and heat stored in the crust. From Stefánsson, 2000. – *Einfölduð mynd af varmastraumum gegnum jarðskorpu Íslands og uppsafnaður varmi í skorpunni*.

Deep drillings into active high-temperature systems show without exception that fine to medium grained intrusive bodies become abundant at depths as shallow as 1 km. Below some 1.5 km, such intrusions as well as gabbro, and occasionally granophyre, dominate the succession (e.g. Arnórsson, 1995a; Franzson, 1995, 1998; Franzson, 2004). These high-level intrusions must have contributed heat to the geothermal system over a period of time as they represent many intrusive events. It is difficult to infer the shape of these intrusions from the study of drill cores and drill cuttings. By comparison with fossil high-temperature systems in eroded central volcanic complexes, these intrusions are considered to consist of dykes, sills, stocks, cone sheets and irregularly shaped bodies when the high-temperature systems hosted in such complexes. In contrast, the intrusions are likely to be predominantly sheeted dyke complexes in high-temperature areas where such complexes have not formed. In the high-temperature systems the flow paths for the convecting fluid are predominantly provided by tectonic fractures but also by contraction fractures, permeable sedimentary layers between lava flows, scoriaceous tops of lava flows and brecciated rocks around intrusions. The fluid stored in the systems predominantly occurs in vesicles, and micro-fractures of extrusives and intrusives. Fluid convection is density driven, i.e. by the difference in the density of a hot fluid column within the system and a cooler and therefore denser liquid water column outside the system.

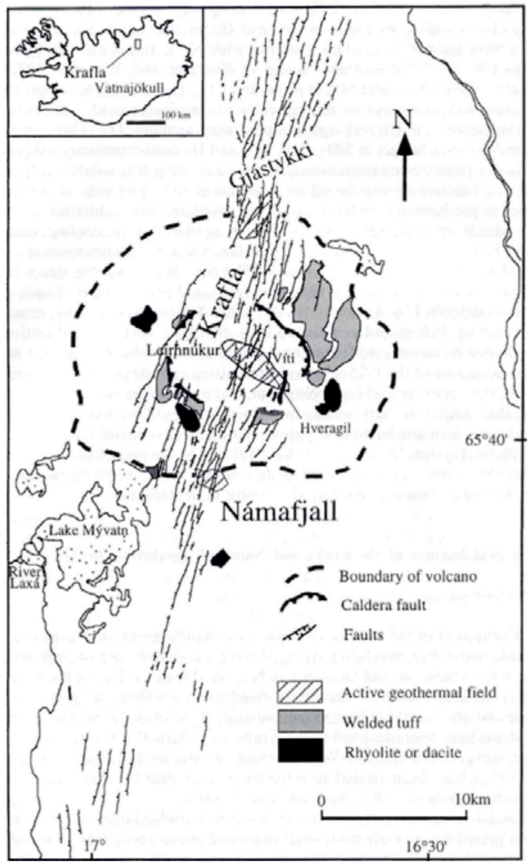

Figure 6. Structural map of the Krafla and Námafjall geothermal fields showing the caldera structure of the Krafla central volcanic complex and associated faults and fissure swarm. Mapped by K. Saemundsson. From Björnsson *et al.* (1979). – *Tektónískt kort af jarðhitasvæðunum við Kröflu og í Námafjalli sem sýnir Kröfluöskjuna og tengdan sprungusveim. Kortlagt af Kristjáni Sæmundssyni.*

Fossil systems

Many fossil high-temperature geothermal systems in Tertiary and Lower-Quaternary formations have been exhumed by erosion (Figure 7). The classic example is that hosted in the Breiddalur central volcano in eastern Iceland (Walker, 1963). Examples of other central volcanic complexes that hosted high-temperature geothermal activity include Setberg in western Iceland (Sigurdsson, 1966), Vídidalur-Vatnsdalur in northern Iceland (Annells, 1969), Thingmúli in eastern Iceland (Carmichael, 1964) and Geitafell in southeastern Iceland (Friðleifsson, 1983). The volcanic complexes are typically embedded within flood basalt sequences formed by fissure eruptions. These complexes are distinguished from the regional flood basalts by difference in dip and abundance of silicic and sometimes intermediate volcanics. The silicic rocks are considered to have formed by partial melting of basaltic rocks overlying major magma reservoirs at the base of the crust (Óskarsson *et al.*, 1982, 1985, Steinthórsson *et al.*, 1986; Sigmarsson *et al.*, 1992; Geirsson, 1993). Some of the intermediate rocks may have formed by magma mixing but others by crystallization differentiation of the primary basalt magma (Sigmarsson *et al.*, 1992).

The fossil high-temperature systems are represented by an aureole of alteration minerals enveloping and overlying a complex of minor intrusions within the central volcanic complex. Major gabbroic bodies, like those found in the most deeply eroded formations in the southeast part of the country, probably correspond to relatively high-level magma chambers that fed higher-level minor intrusions. These bodies in turn were likely fed by larger magma reservoirs at the base of the crust (Gudmundsson, 1987). The formation of the high-level magma chambers is facilitated by density barriers, which are reflected in the p-wave velocity structure. It has also been suggested that the occurrence of stress barriers may lead to the formation of thick sills. Such sills, if not solidified will absorb magma rising from deeper levels, thus evolving into magma chambers (Gudmundsson, 1987).

The altered lavas and intrusions of fossil high-temperature systems are characteristically green in colour due to abundance of chlorite, and sometimes epidote. Other common hydrothermal minerals include calcite, quartz and sulphides, mainly pyrite, but many other hydrothermal minerals have been identified including actinolite, adularia, albite and garnet. This mineral assemblage belongs to the lower-greenschist facies and is indicative of temperatures in excess of 250°C. The alteration is sometimes pervasive, but it is more common that the rock has been partially altered in which case intrusions may be quite fresh as well as the massive central part of lava flows whereas the amygdaloidal upper parts of the lavas, which have the highest porosity and permeability, are typically intensely altered (Neuhoff *et al.*, 1999).

The depth of intrusion may be inferred from estimates of the original top of the lava pile. Thus at Breiddalur, it is deduced from the work of Walker (1963) that minor intrusions formed at depths as shallow as $\sim$1 km. At Setberg, extrapolation of cone sheet swarms, indicate that the depth to the top of the feeding magma chamber is $\sim$2.6 km. The top of the major gabbroic intrusion at Geitafell is little over 1 km (Fridleifsson, 1983) and the top of the Vesturhorn and Austurhorn gabbros are as little as $\sim$2 km from the top of the original lava plateau surface. In this area the laumontite zone, which belongs to the zeolite metamorphic facies, is close to the top of the mountains. According to Walker (1974b), the top of this zone is at about 1700 m below the original basalt plateau surface in eastern Iceland. All these observations are consistent with results obtained from presently active high-temperature geothermal systems within the active volcanic belts. They indicate multiple intrusion of magma into brittle rock to form variously shaped bodies that constitute the heat source to the high-temperature geothermal systems. Intrusion of magma into brittle rock favours effective heat extraction from this magma by convecting ground water. Seismic studies by Pálmason (1973) indicate updoming of seismic "layer 3" (seismic velocity $\sim$6.5 km/s) under fossil high-temperature geothermal systems in Tertiary formations. Pálmason (1973) considers that "layer 3" represents basaltic crust composed essentially of intrusive bodies. Accordingly, central volcanic complexes are characterized by anomalous abundance of intrusions at shallow crustal depths.

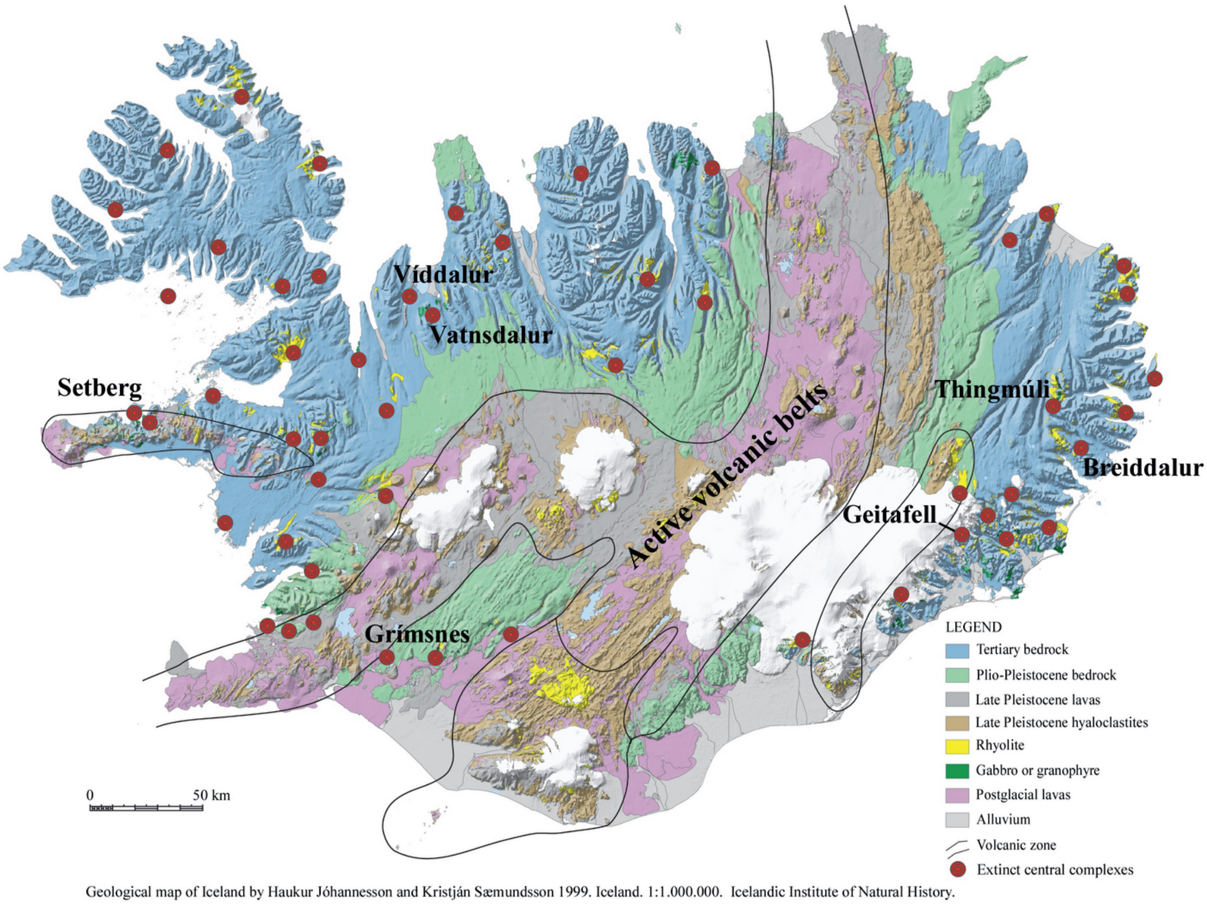

Figure 7. Fossil high-temperature geothermal systems in Quaternary and Tertiary formations exhumed by erosion. From Haukur Jóhannesson, unpublished data. – *Forn, rofin háhitasvæði í kvarterum og tertíerum jarðmyndunum. (Frá Hauki Jóhannessyni, óbirtar upplýsingar).*

THE GRÍMSNES BORDERLINE CASE

An example of a borderline geothermal field by the definition of Bödvarsson (1961) is provided by the Grímsnes area in the Southern Lowlands. This field is located on the east margin of the western volcanic belt as are the Geysir and Hveragerdi fields which Bödvarsson (1961) classified as borderline cases. Yet, today they are defined as high-temperature on the basis of available data. All three fields represent high-temperature fields, which are declining.

The Grímsnes field measures about 10 km at its widest from NW to SE. A NE-SW trending fissure swarm is associated with it. Several wells have been sunk in the area. The maximum recorded downhole temperature is 185°C which is significantly higher than that observed in the hottest low-temperature systems in Iceland (∼150°C). The lithology penetrated by wells (maximum drilled depth is 1100 m) includes a significant proportion of silicic rocks and shows high-temperature alteration mineralogy. Post-glacial lavas have been erupted in the area of which Seydishólar are among the youngest, ∼6200 years old (Jakobsson, 1977). A high CO_2 concentration in the fluid of this system is unique for Iceland, the maximum being over 1%, yielding a CO_2 partial pressure of about 35 bars (unpublished data of the first author). The nearby young volcanic edifices, the silicic rocks, the high-temperature alteration and the high CO_2 content of the fluid all point towards a volcanic origin for this system. The combined data on temperatures at depth, hydrothermal alteration and the location of the field on the margin of the active western volcanic belt suggests that the geothermal system is in the process of cooling down.

LOW-TEMPERATURE GEOTHERMAL ACTIVITY

Complicated stress fields exist in Quaternary and Tertiary formations of the North-American plate and the Hreppar micro-plate that cause deformation by fracturing. The stress field is produced by the combined effects of the shape of the active belts of volcanism and rifting, the direction of the tensional forces across the plate boundary and the displacement of this boundary by transform faults. Much of the hot springs in the low-temperature areas are known to be associated with young fractures that formed under the existing stress fields (Arnórsson and Gíslason, 1990; Björnsson *et al*., 1990). Low-temperature geothermal systems have, however, been discovered in Tertiary formations in many places where there are no surface manifestations (Axelsson *et al*., 2005) and away from well-defined recent fracturing. Small thermal springs are known throughout much of the country.

As already pointed out, temperatures at depth in the low-temperature areas, as deduced from drillhole data and chemical geothermometers, decrease in general with distance from the volcanic belts (see Figures 2 and 3). Likely this is the consequence of decreasing temperature gradient away from these belts rather than depth of water circulation. Flow rates also tend to be highest where temperatures are highest. The reason for this is considered to be twofold; enhancement of density driven convection with increasing temperature and decreased kinematic viscosity of liquid water with increasing temperature. The latter effect leads to increased flow at the same permeability and the same pressure gradient.

According to Arnórsson and Gíslason (1990), low-temperature geothermal activity in Iceland is the consequence of one or more of the four following processes:

(1) Deep circulation of ground water from higher to lower elevation along fractures or other permeable structures driven by hydraulic head.
(2) Convection in young fractures formed by deformation of older crust.
(3) Drift of high-temperature geothermal systems out of the active volcanic belts accompanied by cooling due to displacement from their magmatic heat source.
(4) Intrusion of magma into fractures or other permeable formations by the margins or outside the volcanic belts.

The development of some low-temperature systems may be the consequence of more than one of the processes listed above. Process (2) is certain to play a role in both of processes (3) and (4). Bödvarsson (1982b) concluded that process (1) cannot explain the existence of many of the more powerful low-temperature systems and that they must in essence represent transient phenomena. Furthermore, Bödvarsson (1983) reasoned that data on the temperature/flow statistics of low-temperature activity in Iceland was inconsistent with process (1). Björnsson *et al*. (1990) concluded that the heat source for the low-temperature activity is simply the abnormally hot crust of Iceland, but that the local stress field and the existence of open fissures and faults determines whether a low-temperature system develops in a particular location. In active systems, fractures that are kept open by continuously ongoing tectonic activity play an essential role by providing the channels for the water circulating through the systems. Many of the low-temperature systems are considered to be local density driven convection systems, wherein heat is transported from depth to shallower formations. In other systems water circulation is driven by hydraulic head, or by a combination of both processes.

Below a brief outline is given of the structural control of the most important low-temperature systems. These include: (1) the Central North Iceland fields, which are controlled by fault systems extending into the area both from north (Kolbeinsey domain) and south (fissure swarms of Central Iceland volcanoes, (2) the Reykjavík geothermal fields, which occur in the distal parts of fissure swarms on the Reykjanes Peninsula, (3) the Reykholtsdalur geothermal field, which owes its existence to trans-tensional forces acting on the area west of Langjökull glacier towards Snaefellsnes and (4) the Southern Lowlands geothermal fields, which are largely controlled by normal and transform faulting acting between the east and west volcanic belts.

Central North Iceland

Active N-S trending normal faults are prominent in the outer parts of the valleys of Skagafjördur and Eyjafjördur in northern Iceland and in the mountain area of Tröllaskagi between them (Figure 2). They represent branches of the Kolbeinsey spreading axis (the Mid-Atlantic Ridge north of Iceland). Hot spring activity in the mentioned valleys is associated with these fault zones (Arnórsson and Gíslason, 1990; Arnórsson *et al.* 2002; Björnsson *et al.*, 1990). In the case of the inner Skagafjördur Valley, the hot springs lie on young fractures, which form an extension of the Hofsjökull volcanic system in Central Iceland (see Figure 1C).

Circulation of water in the young fractures is in some, perhaps most instances, maintained by hydraulic head but in other instances it is density driven, or by a combination of the two processes. Thus, ground water flow feeding fresh water hot springs that emerge under the sea in Eyjafjördur must be mostly maintained by hydraulic head. By contrast, at Laugaland in Eyjafjördur, circulation is thought to be largely density driven as deduced from temperature distribution in deep wells (Figure 8C).

The Reykjavík areas

The low-temperature systems of Laugarnes and Mosfellssveit within and in the vicinity of Reykjavík in southwest Iceland are hosted by fossil high-temperature systems, as deduced from alteration mineralogy. Temperatures in deep wells indicate that the water circulation is essentially density driven (Figure 8A and B). The systems lie astride the active fissure swarms of Reykjanes and Krísuvík, respectively, which run obliquely across the active volcanic belt on the Reykjanes peninsula and into Quaternary and Tertiary crust to the north of this belt (Arnórsson *et al.*, 1992; Björnsson *et al.*, 2000). In Mosfellssveit, recharge to the geothermal systems is concentrated along the fissure swarm from the southwest and the northeast (Figure 9). Arnórsson *et al.* (1992) and Björnsson and Steingrímsson (1995) have described in some detail a conceptual model for this field. In the Laugarnes field reservoir drawdown caused by exploitation has led to an increase in fluid salinity in some wells due to recharge of a seawater component (Gunnlaugsson, 1988). This recharge is from the southwest, presumably along the Reykjanes fissure swarm.

Reykholtsdalur

This is the largest low-temperature field in Iceland in terms of heat output ($\sim$400 L/s of 100°C water, Figure 3). Hot springs are characteristically lined along young N-S trending fractures. The area is tectonically active. The last swarm of earthquakes occurred just north of the thermal field in 1974 (Einarsson *et al.*, 1977). The curvature of the plate boundary in the east leads to tensional stress in the hot spring area (see Figure 2). As already pointed out, rifting is insignificant in the northeastern part of the neighbouring volcanic belt as compared to its southwestward continuation. This is expected to cause a trans-tensional movement in the area of deformation. The N-S trending active faults in the Reykholtsdalur geothermal field may be the consequence of fault plane resolution at the surface of the slip-strike movement at depth.

Southern Lowlands

The Southern Lowlands region (see Figures 1 and 2) constitutes a part of the Hreppar microplate, which occupies the area between the two active volcanic belts of southern Iceland. The western active volcanic belt lies to the north and west of the Southern Lowlands.

Active fractures of this volcanic belt, trending NNE-SSE, extend into the Southern Lowlands. Farther south, active fractures and faults form a zone of N-S trending arrays across the Hreppar microplate from east to west, known as the South Icelandic Seismic Zone (Stefánsson *et al.*, 1993). It is the expression of a left lateral fracture zone that extends from Hekla in the eastern volcanic belt to the Reykjanes Peninsula. Earthquakes of magnitude 6-7 occur in this seismic zone about once a century. The numerous hot springs in the region are known to be located at the active faults in the area and most of over 70 large and small geothermal heating supplies are based on holes drilled into them (Axelsson *et al.*, 2005). Earthquakes in the area tend to maintain fracture permeability and may renew it as was observed in the last large earthquake of the South Iceland Seismic Zone in the year 2000.

HEAT TRANSFER TO FLUID CONVECTING IN GEOTHERMAL SYSTEMS

The general model for high-temperature (volcanic) geothermal systems envisages deep circulation of liquid water above and to the sides of a magmatic heat source (Fournier, 1989; Lister, 1983; Fournier and Pitt, 1985) and a layer between magma and the base of water circulation through which heat is transferred conductively. On the basis of estimates of the natural heat output of the Hengill geothermal system in SW-Iceland Bödvarsson (1951) concluded that this conductive layer had to be thin.

Data obtained from drillings into the molten lava from the 1973 eruption on the island of Heimaey off the south coast of Iceland provide important information on the mechanism of cooling of magma through its interaction with water. Seawater was pumped onto the new lava with the purpose of diverting its flow away from the town on Heimaey. Drilling into the lava revealed a temperature of 100°C down to a certain depth followed by a sharp rise to as much as ∼1050 °C in a layer that was as thin as 1-2 m (Jónsson and Matthíasson, 1974; Figure 10). The interpretation of the temperature profile is as follows: the temperature of 100°C was controlled by rising steam at atmospheric pressure.

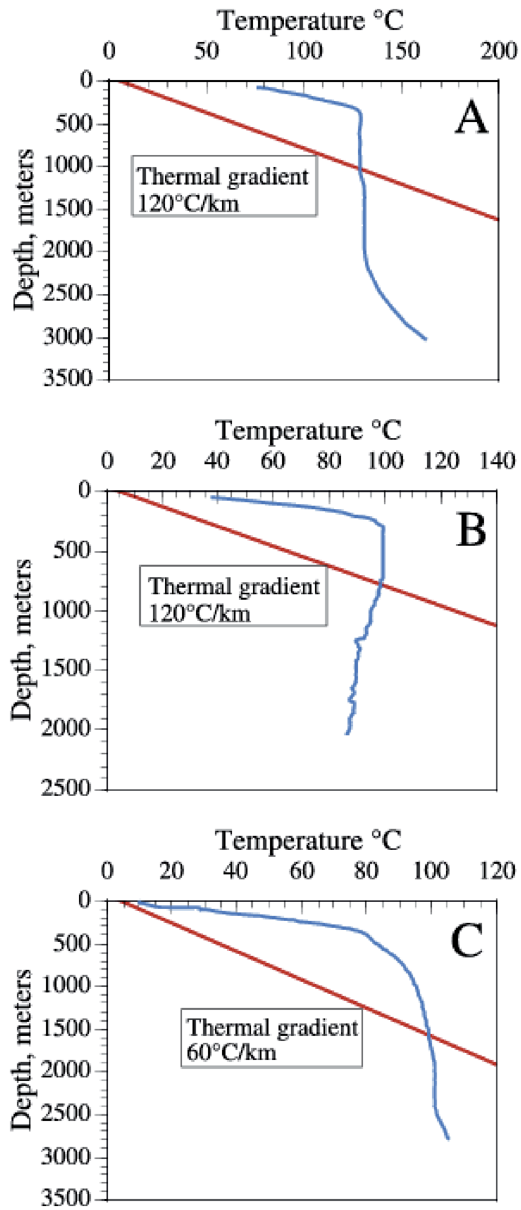

Figure 8. Temperatures in deep holes drilled into low-temperature systems where water convection is density driven. A: Laugarnes in Reykjavík, B: Reykir in Mosfellssveit, C: Laugaland in Eyjafjördur. – *Mældur hiti í djúpum borholum á lághitasvæðum þar sem hræring vatnsins er drifin með eðlisþyngdarmun á kaldri súlu grunnvatns utan svæðanna og heitri innan þeirra.*

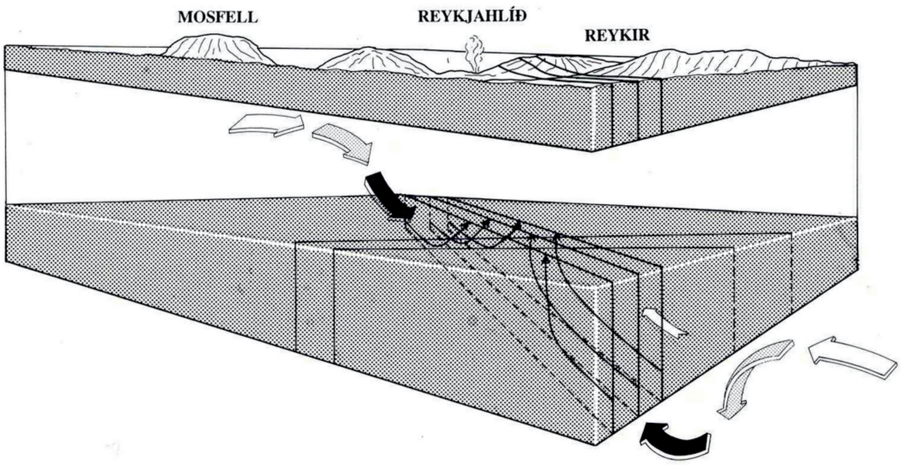

Figure 9. Simplified conceptual model for the low-temperature systems in Mosfellssveit (Reykir and Reykjahlíd). Large white, grey and black arrows represent cold, warm and hot water recharging the geothermal system along active fractures shown by three parallel straight and broken lines. Slim arrows represent ascending hot water. The two parallel sets of lines, which are at right angles to the lines representing the active fractures, represent an inferred caldera boundary. From Arnórsson *et al*. (1992). – *Einfaldað mynsturlíkan af lághitasvæðunum í Mosfellssveit (Reykir og Reykjahlíð). Stórar hvítar, gráar og svartar örvar sýna kalt, volgt og heitt vatn sem streymir inn í jarðhitakerfið eftir virkum sprungum sem sýndar eru með þremur beinum heilum og slitnum línum. Grannar örvar sýna uppstreymi á heitu vatni. Samsíða línurnar tvær sem eru hornréttar á virku sprungurnar tákna öskjumisgengi.*

The water pumped onto the lava infiltrated it until it had absorbed sufficient heat to be converted into steam. The layer where the temperature increased from 100° to 1050°C represents a layer through which heat was transferred conductively from the molten lava to the base of the circulating H_2O. A balance of heat transfer was established between the conductive and convective layers, respectively, higher fluid through-flow leading to thinner conductive layer. With time the water circulation caused the conductive layer to migrate downward.

The Heimaey lava heat transfer model has been applied to estimate the thickness of the conductive layer in a few high-temperature geothermal systems in Iceland for which natural heat output has been reasonably accurately estimated. Björnsson *et al*. (1982) concluded from the estimated heat output of 5000 MWt for the Grímsvötn sub-glacial geothermal system that the thickness of the conductive layer between magma and geothermal system over the Grímsvötn caldera (10 km^2) was only 13 m. Careful measurements of the natural heat flow from the relatively small (1 km^2) geothermal field at Reykjanes, Iceland, of 130±16 MWt (Fridriksson *et al*., 2006) yields a thickness for the conductive layer as about 50 m. The heat output of the Námafjall system is estimated as ∼280 MWt. Taking the areal extent of this field to be 4 km^2 the conductive heat transfer layer is ∼100 m. For all the above calculations the top of the magma body was taken to form a horizontal surface over the estimated lateral extent of the geothermal system. In reality the top surface of the magma is expected to be irregular and consist of dykes, sills and stocks, even cone-sheets forming a larger contact area between magma and conductive layer than the lateral extent of the system thus leading to a correspondingly thicker conductive layer.

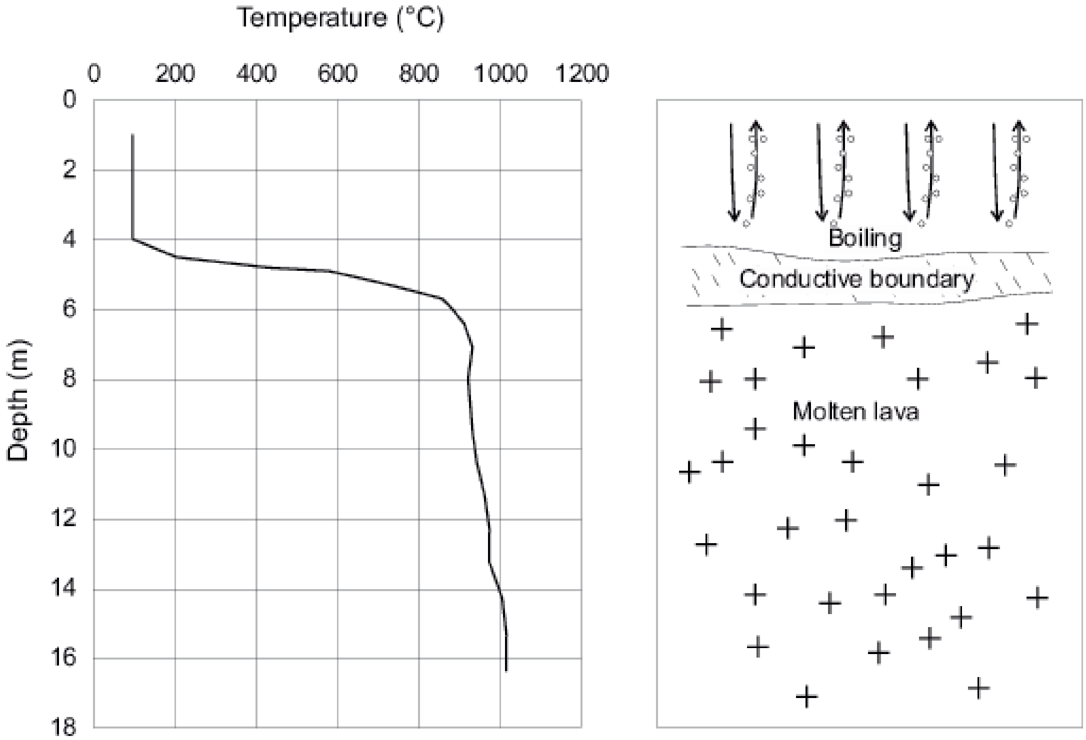

Figure 10. Temperature in a hole drilled into the molten lava that formed on the island of Heimaey (Vestmannaeyjar) in 1973. From Arnórsson *et al*. (2007). – *Mældur hiti í holu sem boruð var í bráðið hraun í Vestmannaeyjagosinu 1973*.

Temperatures at deep levels in wells drilled to depths as great as 3000 m in several low-temperature (tectonic) systems in Iceland are much lower than estimates made by extrapolating temperature gradients measured in shallow holes in the vicinity of these areas (Figure 8). Thus in the Laugarnes field, which is located within the city of Reykjavík, temperature at 3000 m depth is close to 160°C but the regional gradient in the area is ~120°C/km (Pálmason, 1973; Björnsson *et al*., 1990). Comparable results have been obtained for deep wells in the Mosfellssveit area just east of Reykjavík and at Laugaland in Eyjafjördur, North-Iceland (Figure 8). These profiles indicate that convecting water in fractures cools the rock at deep levels, and as the water gains heat by cooling the rock, it rises by buoyancy and heats up the rock at shallow levels, thus leveling off temperatures over a long depth range in the geothermal systems.

These results show that the heat source to these systems is hot rock towards the base of the water convection. The rock in the roots of these low-temperature systems has been cooled by the circulating water producing an anomaly of low temperature relative to the surrounding rock at the same depth. The source water must therefore plunge down within the system or very close to it. Likely the recharging water is local, but if derived from a distant source, it must flow at shallow level to the geothermal system. Tectonic systems of the type just described will have a given lifetime determined by the rate of circulation of the ground water and the nature of the permeability that determines contact area between circulating water and rock. With continued cooling, density driven convection will gradually die out and so does the geothermal system.

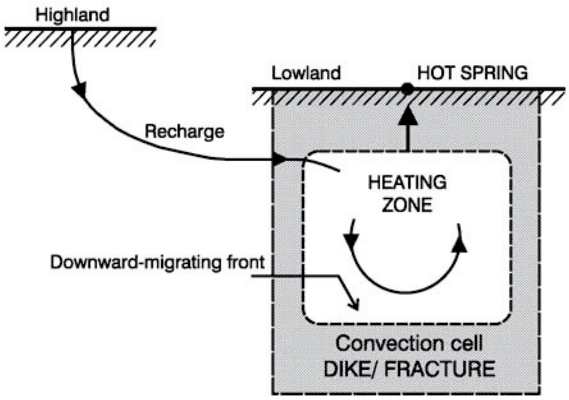

Figure 11. Conceptual model for water convection in a low-temperature geothermal system according to Bödvarsson (1983). – *Mynsturlíkan af hræringu grunnvatns í lághitakerfum samkvæmt Gunnari Böðvarssyni (1983).*

Bödvarsson (1982b, 1983) proposed a model for the heat-source mechanism of the low-temperature activity, which can explain the high heat output of many systems. This model appears to be consistent with the data available at present on most of the low-temperature systems (Björnsson *et al.*, 1990). According to his model, presented in Figure 11, taking into account deuterium data (see below), the recharge to a low-temperature system is shallow ground water flow from the highlands to the lowlands. Inside the geothermal area, the water sinks through an open fracture, or along a dike, to a depth of a few km where it becomes heated and ascends. The fracture is closed at depth, but opens up and continuously migrates downward during the heat-mining process by cooling and contraction of the adjacent rock. As discussed in the following section, the low deuterium content of water in many low-temperatures systems need not reflect its origin on high ground inland but could be the consequence of a deuterium depleted ice-age component in the water. Thus, low-deuterium water circulating in fractures of low-temperature systems could essentially be local precipitation in origin.

Theoretical calculations based on Bödvarsson's model by Axelsson (1985) and Björnsson *et al.* (1990) indicate that the existence and heat output of low-temperature systems is controlled by the temperature and stress conditions in the crust, in particular the local stress field, which controls whether open fractures are available for the heat-mining process and how fast these fractures can migrate downward. Given the abnormal thermal conditions in the crust of Iceland, it appears therefore that the regional tectonics and the resulting local stress field are the main factors controlling the low-temperature activity.

ORIGIN OF THE GEOTHERMAL FLUID

The water of geothermal fluids in Iceland is meteoric water or seawater by origin or mixtures thereof. In some high-temperature fields, the geothermal fluid may contain a small magmatic water component.

The δ^2H content of most low-temperature waters is lower (more negative) than that of local precipitation (Árnason, 1976, 1977, Sveinbjörnsdóttir *et al.*, 2001; Arnórsson and Gíslason, 1990). Árnason (1976, 1977) considered this to indicate that the water originates as precipitation inland from the low-temperature fields and at higher elevation but due to the so-called inland and altitude effects, the δ^2H content of precipitation in Iceland decreases inland and with elevation (see e.g. Árnason, 1976). Arnórsson and Andrésdóttir (1995) observed that many low-temperature waters emerging in low-lying areas, stretching relatively far inland, had very negative δ^2H-values, sometimes more negative than that of any precipitation in Iceland today, and that there was a negative correlation between the δ^2H-values and the Cl content of the water. They considered that such waters were mixtures of three components; present-day meteoric water, seawater that infiltrated the bedrock when these low-lying areas were transgressed by the ocean in early Holocene times and Pre-Holocene glacial melt water. As deduced from hydraulic gradient in these low-lying areas and global permeability derived from drillhole data, it takes ground water considerably more than 10,000 years to flow underground from inland areas to the coast (Björnsson *et al.*, 1990). Therefore ground water hydrology data support the

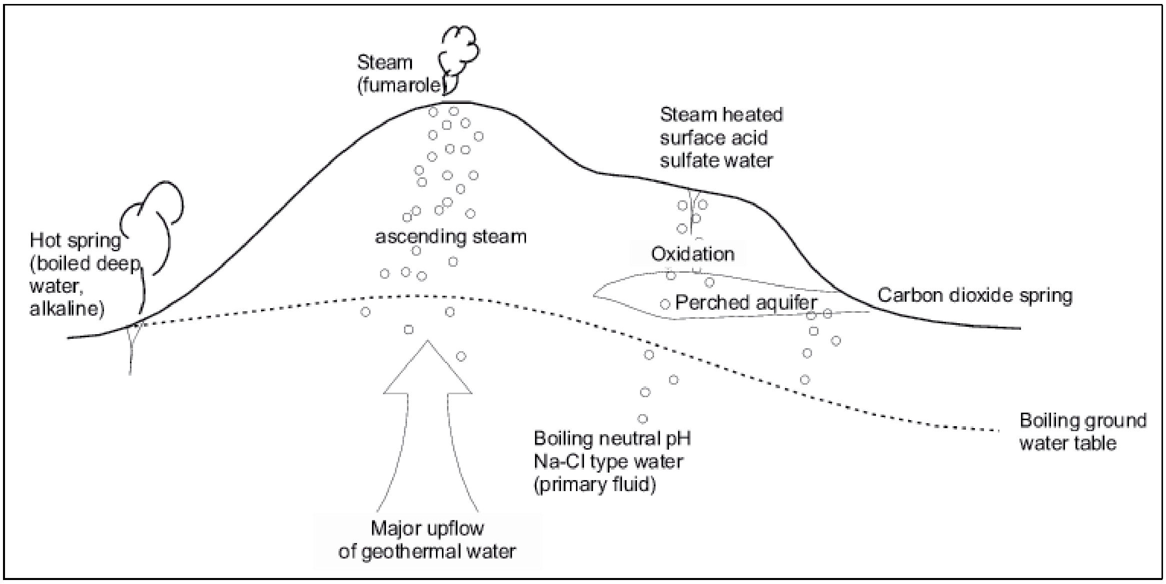

Figure 12. Schematic cross section through the uppermost part of a high-temperature geothermal system depicting the formation of various types of secondary geothermal fluids. From Arnórsson *et al*. (2007). – *Einfaldað snið gegnum efsta hluta háhitasvæðis sem sínir myndun ýmissa gerða af síðmynduðum jarðhitavökva.*

geochemical data that low δ^2H waters with elevated marine Cl component contain a Pre-Holocene fresh water component. In areas with steeper topography or at higher elevation low-temperature waters are solely meteoric in origin.

The fluid in the high-temperature field of Reykjanes on the southwest tip of the Reykjanes peninsula is seawater that has been modified chemically by interaction with the basalt host rock (Björnsson *et al.*, 1972) although low δ^2H values of the geothermal fluids suggest it was a meteoric (glacially) dominated system in Pre-Holocene times. The waters of other high-temperature systems on the Reykjanes peninsula are mixtures of seawater and local meteoric. In other drilled high-temperature fields, the source water is essentially local meteoric water (Arnórsson, 1995a). At Nesjavellir, however, the recharging water is from the nearby Lake Thingvallavatn and the mountain of Hengill. This lake is essentially fed by ground water from the permeable formations of the active volcanic belt to the northeast of the lake and far from the geothermal field.

GEOTHERMAL FLUID CHEMISTRY

Geothermal fluids that reach the deepest level of penetration were termed primary geothermal fluids by Arnórsson *et al*. (2007). The maximum depth penetrated by the fluid was termed base-depth by Bödvarsson (1961) and the temperature at this depth the base-temperature. Primary geothermal fluids may be a mixture of as much as three fluid components: meteoric water, seawater and magmatic volatiles. When primary fluids rise towards the surface they can undergo boiling and mixing forming secondary geothermal fluids. The most important processes leading to the formation of secondary geothermal fluids include:

1) Depressurization boiling to yield a gaseous vapour phase and boiled liquid.
2) Vapour condensation in shallow ground water or surface water to produce steam-heated acid-sulphate, carbon-dioxide or sodium bicarbonate waters.
3) Mixing of CO_2 gas from a deep source with thermal ground water.

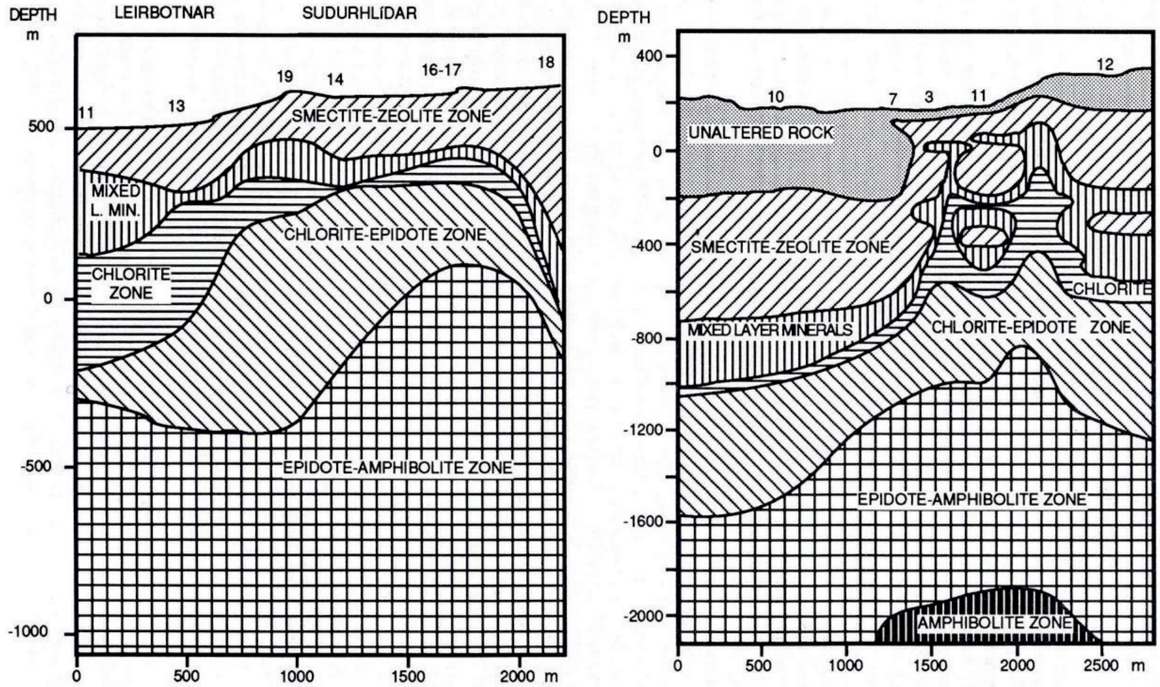

Figure 13. Hydrothermal mineral zonation in the Krafla (left) and Nesjavellir (right) geothermal fields. From Ármannsson *et al*. (1987) and Franzson (1988), respectively. – *Lagskipt jarðhitaummyndun í jarðhitasvæðunum við Kröflu (til vinstri) og á Nesjavöllum (til hægri).*

4) Mixing of geothermal water with shallower and cooler water.

In some geological environments, phase separation of saline fluids into concentrated brine and more dilute vapour solution may also produce secondary fluids (Heinrich *et al*., 2004) but it seems unlikely that such a process is operative in Iceland because basaltic magma is low in HCl and other volatiles that can produce brine upon interaction with common silicates.

Geothermal waters, both high and low-temperature, containing a seawater component in excess of that derived from precipitation, are all located in low-lying coastal areas. On the Reykjanes peninsula seawater infiltration occurs under present-day hydraulic conditions although meteoric water (ice-melt) circulated through the system during the last glaciation (Pope *et al*., 2008). In most other areas, which are low lying, it is considered that seawater infiltration occurred around the end of the last glaciation when these areas were transgressed by the ocean (Arnórsson and Andrésdóttir, 1995).

Primary geothermal fluids in Iceland that have reacted only with basaltic rock are relatively low in chloride (Cl) compared to geothermal fluids worldwide (Tables 1 and 2 and data presented by Ellis and Mahon, 1977 and Arnórsson *et al*., 2007). Waters in low-temperature fields typically contain 10-30 mg/L Cl if not affected by seawater infiltration. In high-temperature waters, Cl concentrations are typically higher, up to 200 mg/L. The reason for the low Cl content of the Icelandic geothermal waters hosted by basalt is the low content of this element in this rock, but Cl is the only common anion in any ground water system forming soluble salts with the major cations of common rock types. Waters that have reacted with silicic rocks, such as those in the Geysir Area and

Landmannalaugar are higher in Cl than water that has reacted with basalt only, the cause being higher Cl content of the silicic rock. In coastal areas up to 10 mg/L of the Cl in geothermal waters, even more, may be derived from seawater spray, that is from the precipitation, but <2 mg/L in areas located farthest inland.

Geochemical studies of many geothermal fields in Iceland, both high and low-temperature have shown that local chemical equilibrium is closely approached between solution and hydrothermal minerals at temperatures as low as 50°C for all major components except Cl (Arnórsson *et al.*, 1983; Arnórsson, 1995a, 1995b; Gudmundsson and Arnórsson, 2005). Aqueous Cl concentrations are determined by their supply to the water. Close approach to mineral-solution equilibria involving the other major components causes activities of uncharged species and activity ratios of ionic species that do not involve Cl to be fixed by temperature and pressure alone. Pressure in the range occurring from surface to the base of geothermal systems (several km) has small effect relative to temperature. Effectively, therefore, temperature and supply of Cl to geothermal waters represent the independent chemical thermodynamic variables, which control the composition of primary geothermal fluids.

The gas content of fumaroles and well discharges is highly variable. In low-temperature systems, the main gases are N_2 and Ar (Arnórsson, 1986). They are considered to be of atmospheric origin dissolved from the atmosphere into the parent water to the geothermal solution. Yet, a small fraction of the N_2 may come from decaying organic matter. Carbon dioxide, H_2S and H_2 are present in measurable concentrations in vapour formed by boiling of the hottest (∼150°C) low-temperature fluids. In high-temperature systems, the gas concentrations are higher than in low-temperature fluids and they tend to increase with increasing aquifer temperature. Carbon dioxide is the most abundant gas followed by H_2S and H_2. Methane is always detectable, its concentration typically being <1 mmole/kg in vapour samples collected from wet-steam wells. Argon concentrations are similar to those of air-saturated water whereas N_2 concentrations may be slightly higher, either due to supply of N_2 from decaying organic matter or degassing magma (Arnórsson, 1995a). The aqueous concentrations of CO_2, H_2S and H_2 in deep aquifers most often are controlled by temperature dependent equilibria with hydrothermal mineral assemblages (Arnórsson and Gunnlaugsson, 1985; Stefánsson and Arnórsson, 2002; Gudmundsson and Arnórsson, 2002, 2005; Arnórsson *et al.*, 2007). On the other hand, departure from gas-gas equilibrium involving CO_2 and CH_4 ($CO_2 + 4H_2 = CH_4 + 2H_2O$) is typically large (Stefánsson and Arnórsson, 2002), the cause little doubt being the slow rate of this redox reaction. Sometimes gas-mineral equilibria are not closely approached due to high flux of gases from the magmatic heat source (Ármannsson *et al.*, 1982; Arnórsson *et al.*, 2007). A classic example is provided by the Krafla high-temperature system in northern Iceland. During a volcanic episode in 1975-84, fresh magma was intruded into the roots of this system (Björnsson *et al.*, 1977). This led to an increase in the gas content (mostly CO_2 and H_2) of fumarole steam and well discharges (Ármannsson *et al.*, 1982, 1989). Thirty years later this gas pulse is still observable but it has diminished much.

Secondary fluids are common, in particular in some high-temperature geothermal fields (Figure 12). The most prominent type is steam-heated surface water. Such waters are typically low in Cl but high in sulphate (SO_4). Their pH is low, in extreme cases as low as 1. Oxidation of hydrogen sulphide (H_2S) in the steam to SO_4 produces the acidity. The acid water reacts effectively with the rock-forming minerals, transporting most major elements away from the site of alteration leaving a residue enriched in silica, titanium (in the case of basalts) and aluminium (Sigvaldason, 1959; Arnórsson 1993). This alteration product constitutes the most prominent thermal manifestations in the high-temperature areas together with high-discharge fumaroles. Warm springs with CO_2 water are common in some high-temperature fields, particularly by their margins. They may represent shallow ground water or surface water heated by H_2S deficient steam or a mixture of un-boiled deep water and shallow ground water (Arnórsson, 1985).

Table 1. Chemical composition of selected samples from hot-water wells and hot springs in high-temperature areas. Concentrations are in mg/kg for major components (SiO_2 to F) and in μg/kg for trace elements (As to Zn). – *Efnasamsetning valinna sýna úr heitavatnsholum og hverum af lághitasvæðum. Styrkur aðalefna (SiO_2 til F) er í mg/kg en í μg/kg fyrir snefilefni (As til Zn).*

Sample no.	96-005	97-001	02-041	03-005	01-221	01-201	01-241
Temp.°C	40	89	69	127	96	87	96
pH/°C	10.19/21	9.47/24	9.49/26	9.58/18	9.00/26	9.41/25	9.63/26
SiO_2	70.2	123.1	66.7	139	435	534	255.2
B	0.480	0.468	0.337	0.0479	0.493	1.047	5.082
Na	64.5	75.1	356.1	52.5	149.4	238.8	423.9
K	0.27	1.90	3.98	1.80	7.17	22.62	13.45
Ca	2.40	1.72	35.76	2.36	3.88	0.77	1.03
Mg	0.0030	0.0020	0.0045	0.0020	0.0136	0.0038	0.0026
Al	0.0461	0.0904	0.0482	0.193	0.072	0.695	0.174
Fe	0.0044	0.0049	0.0034	0.0015	0.0045	0.0072	0.0115
SCO_2	14.5	31.7	4.1	18.5	49.7	143.3	51.8
SH_2S	0.086	0.788	0.049	0.425	2.238	2.98	17.5
SO_4	35.3	50.7	116.0	15.6	153.3	128.4	24.8
Cl	30.7	27.4	527.0	28.5	59.5	124.3	498.3
F	1.78	2.20	1.32	0.69	2.50	8.85	23.39
As	8.80	9.50	9.21	1.55	46.6	84.5	252.0
Ba	0.316	0.205	0.96	0.51	0.31	0.13	1.75
Br			2090.0	101.0	115	349.0	2010.0
Cd	0.009	0.022	0.037	<0.002	<0.004	<0.002	<0.002
Co	0.06	<0.01	0.007	<0.005	<0.005	<0.005	0.079
Cr	0.14	0.08	<0.01	<0.01	<0.01	0.03	0.09
Cs			0.080	0.372	0.26	12.60	13.40
Cu	0.67	0.23	0.16	0.12	<0.10	0.26	0.34
Ga			2.22	5.34	4.26	9.10	30.30
Ge			2.20	1.73	13.2	21.4	54.7
Hg			0.050	0.005	0.008	0.008	0.008
I			392.0	5.0	4.8	11.6	93.8
Li			9.77	8.96	141.0	356.0	199.0
Mn	4.87	0.39	0.597	0.073	0.10	0.30	6.30
Mo	13.99	34.35	51.00	3.25	29.8	41.00	5.82
Ni			0.03	0.05	<0.05	0.05	0.12
P	<1.0	5.1		<1.0	2.2	4.6	7.4
Pb	0.036	0.053	0.022	0.064	0.020	0.540	0.39
Rb	0.328	8.96	6.83	7.81	18.8	176.0	150.0
Rb	0.328	8.96	6.83	7.81	18.8	176.0	150.0
Sb			0.018	0.026	1.40	3.92	30.60
Se			0.024			0.01	
Sn			0.047	0.079	0.009	0.033	0.281
Sr	2.37	7.23	42.3	22.1	7.80	9.20	9.9
Ti	0.12	0.18	<0.02	0.04	0.23	0.53	19.00
Tl			0.015	<0.030	<0.01	0.124	<0.03
V	3.33	0.48	0.85	1.05	6.19	8.95	2.88
W	0.064	9.50	3.21	1.04	15.8	27.70	88.00
Zn	0.66	14.92	1.96	0.65	<0.03	0.034	1.04

96-005: Daufá, Skagafjördur. 97-001: Varmahlíd, Skagafjördur. 02-041: Blesastadir, Southern Lowlands. 03-005: Laugarnessvaedi, well 19, Reykjavík. 01-201: Geysir, Geysir Area, Southern Lowlands. 01-221: Eyvindarhola, Hveravellir. 01-241: Eyrarhver, Landmannalaugar. The first four samples are from wells in low-temperature areas in basaltic terrain. The last three samples are from hot springs in high-temperature fields, the first in basaltic terrain, the second is associated with mafic and silicic volcanics and the last with silicic volcanics only.

Table 2. Calculated chemical composition of aquifer fluid of selected wet-steam wells based on analyses of samples of liquid and vapour collected at the wellhead. Concentrations are in mg/kg for major components (SiO_2 to F) and in μg/kg for trace elements (As to Zn). – *Reiknuð efnasamsetning vökva í veitum gufuhola byggð á efnagreiningum sýna af vatni og gufu sem safnað varvið holutopp. Styrkur aðalefna (SiO_2 til F) er í mg/kg en í μg/kg fyrir snefilefni (As til Zn).*

Sample no.	**04-004**	**05-003**	**04-014**	**04-023**	**04-007**	**04-001**	**04-029**
Aquifer temp. °C	238	284	299	199	268	297	274
pH	5.22	4.39	6.62	7.40	6.83	6.83	7.04
SiO_2	426.6	567	648	337	536	638	596
B	7.36	5.67	1.21	0.49	0.306	1.63	1.50
Na	6813	8511	175.6	195.4	114.8	108.6	141.4
K	1009	1220	40.3	15.5	21.1	24.4	25.8
Ca	1023	1381	1.16	3.17	0.35	0.10	0.24
Mg	0.427	0.907	0.001	0.001	0.0040	0.001	0.001
Al	0.057	0.038	1.251	0.756	0.834	1.380	1.503
Fe	0.0082	0.247	0.0017	0.0024	0.0038	0.0118	0.0118
SO_4	25.0	20.5	125.2	213.3	3.2	2.3	8.8
Cl	12912	16241	110.4	40.3	93.1	109.7	151.1
F	0.143	0.116	1.031	0.729	0.451	0.691	0.939
$\Sigma CO_2{}^a$	318	2722	2485	74.5	667	461	318
ΣH_2S^b	7.30	57.8	644	26.4	193	582	763
H_2	0.02	0.27	34.4	0.03	32.0	26.7	0.54
CH_4	0.02	0.17	0.50	0.24	1.85	0.72	0.43
N_2	4.49	34.53	38.17	14.00	18.54	61.62	20.88
Ar	0.14	0.76	1.17	0.49	1.43	1.22	0.61
As	94.2	131.6	189.0	0.75	138.8	71.1	43.9
Ba	1920	5610	3.56	0.56	0.85	1.15	0.31
Br	45300	65800	213.0	109.2	328.1	450.6	562.0
Cd	0.134	0.779	<0.002	<0.002	0.003	<0.002	<0.002
Co	0.10	0.27	<0.005	<0.005	0.178	<0.005	0.004
Cr	1.84	0.23	0.042	0.018	0.122	0.076	0.053
Cs	57.6	52.0	6.06	3.84	2.36	6.53	5.69
Cu	0.91	1.47	0.35	0.10	0.59	0.70	0.34
Ga	0.06	1.93	5.87	3.24	3.01	4.37	4.18
Ge	12.5	19.4	29.5	28.6	22.2	38.0	26.6
Hg	0.009	0.058	<0.002	<0.002	<0.002	0.007	<0.002
I	120.5	253.0	6.09	3.04	9.56	10.3	10.0
Li	2700	3315	421.0	87.7	199.7	336.6	226.6
Mn	185	1265	1.47	0.32	2.10	0.36	0.17
Mo	9.15	78.0	1.94	1.28	0.85	3.29	6.18
Ni	2.27	1.12	<0.05	<0.05	0.67	0.26	<0.05
P	<10	13.0	1.74	<1.0	2.07	1.69	0.84
Pb	<0.1	0.844	0.015	0.022	0.039	0.111	0.013
Rb	2980	3543	217.4	160.0	87.8	142.2	143.1
Sb	1.67	1.16	2.69	0.23	1.10	1.55	0.56
Sn	<0.5	<0.5	<0.05	<0.05	<0.05	<0.05	<0.05
Sr	7083	7694	8.36	27.5	2.39	1.02	1.81
Ti	3.65	<0.10	0.22	0.06	0.26	<0.01	0.12
Tl	8.43	14.85	0.086	0.020	0.016	0.35	0.007
V	3.98	7.66	7.05	0.47	0.73	1.56	2.53
W	0.57	0.84	5.85	3.49	1.94	4.36	4.50
Zn	<2.0	42.9	0.48	1.53	0.40	0.39	0.57

04-004: Svartsengi well 19. 05-003: Reykjanes, well 15. 04-014: Krafla, well 32. 04-023: Krafla, well 24. 04-007: Námafjall, well 12. 04-001: Nesjavellir, well 13. 04-049: Nesjavellir, well 14. [a] ΣCO_2: Total carbonate carbon. [b] ΣH_2S: Total sulphide sulphur.

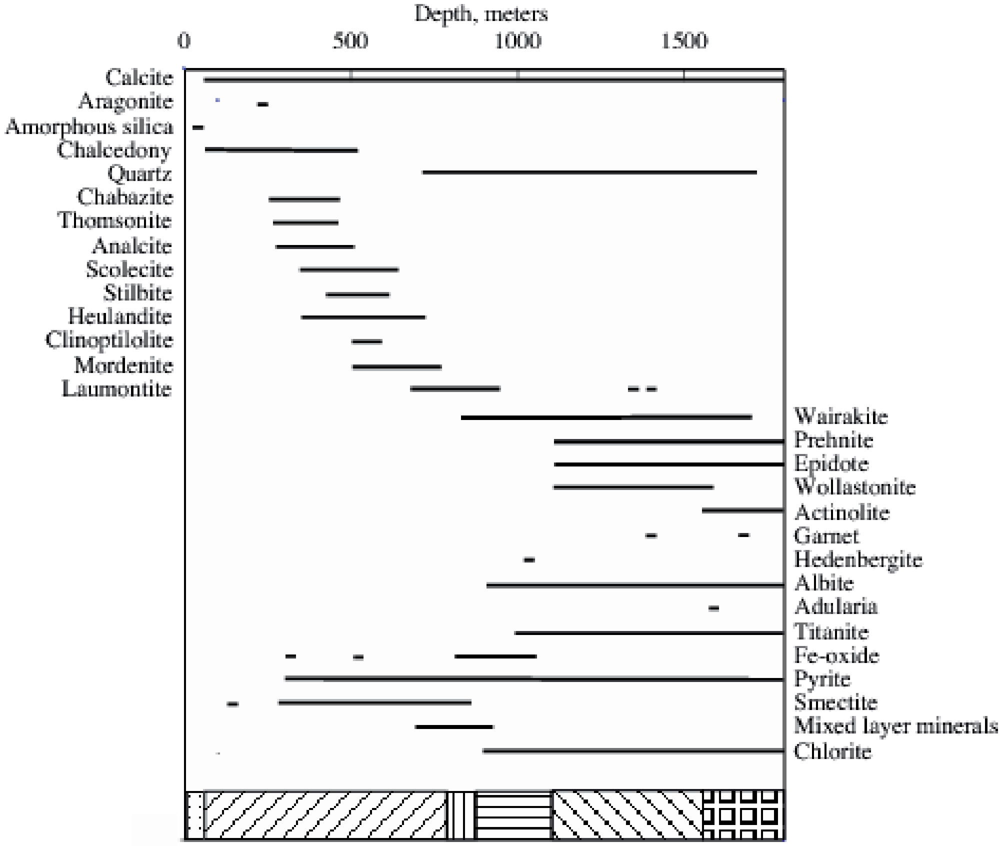

Figure 14. Depth distribution of hydrothermal minerals in basaltic rocks penetrated by well 15 in the Nesjavellir high-temperature field. Mineral zones are shown at the bottom. Their shading is the same as in Figure 13. From Steingrímsson *et al*. (1986a, 1986b). – *Dreifing ummyndunarsteinda sem fall af dýpi í basísku bergi sem hola 15 á Nesjavöllum var boruð gegnum. Ummyndunarbeltin eru sýnd neðst á myndinni með sömu táknum og á mynd 13*.

Of totally different origin are CO_2 waters found outside the active belts of rifting and volcanism. These waters may be thermal or non-thermal. They form by mixing of deep-seated CO_2 with ground or surface water. The ground water may be thermal, in which case it is deep-seated, or non-thermal, in which case it is shallow (Arnórsson and Barnes, 1983). Waters formed in this way are common in many parts of the world (Barnes *et al*., 1978; Marques *et al*., 2006). In Iceland, they are particularly widely distributed on the Snaefellsnes peninsula but also south of Eyjafjallajökull (see Figure 1C). Unlike primary geothermal fluids, secondary fluids may depart much from equilibrium with geothermal minerals. Yet, recent work indicates that even the highly reactive steam-heated acid surface water may be close to equilibrium with some amorphous phases (A. Stefánsson, pers. comm.).

HYDROTHERMAL ALTERATION

Hydrothermal alteration of igneous rocks involves their mineralogical transformation, that is dissolution of the primary rock-forming minerals and formation of secondary minerals. Some elements may be added to the rock being altered but others become depleted (Arnórsson, 1995a). Many researchers have studied alteration mineralogy at depth in active high-temperature systems in Iceland (e.g. Kristmannsdóttir, 1982, 1983; Björnsson *et al.*, 1972; Ragnarsdóttir *et al.*, 1984; Sveinbjörnsdóttir *et al.*, 1986; Franzson, 1988, 1990, 1995; Fridleifsson, 1990; Tómasson and Franzson, 1992; Lonker *et al.*, 1993). The hydrothermal minerals typically display a depth-zonal distribution as characterized by specific index minerals (Ármannsson *et al.*, 1987; Franzson, 1988). The zones may be irregular, such as at Svartsengi (Franzson, 1990) or oval in shape (Figure 13). Comparable mineral zonation has been observed in fossil high-temperature systems (e.g. Walker, 1963; Sigurdsson, 1966; Annels, 1967). The observed mineral zonation is well exemplified by data from the Nesjavellir field (Figure 14). Some minerals form over a large temperature range. Others, in particular Ca-Al-silicates, typically have limited temperature stability range. The number of hydrothermal minerals reported in the same sample, or in closely spaced samples from the same well, sometimes is greater than that permitted by the phase rule (Bird *et al.*, 1984; Bird and Spieler, 2004). This observation emphasizes the fact that the hydrothermal minerals found in the same specimen are the product of alteration formed over a period of time when temperature was undergoing changes. Thus the alteration mineralogy may provide information on the thermal history of a particular high-temperature system (e.g. Fridleifsson, 1983).

Regional alteration of the Tertiary formations in Iceland may be described as burial metamorphism. It is produced by low-temperature geothermal fluids and is characterized by a variety of zeolites, calcite, quartz and clay minerals including celadonite, smectite and mixed layer smectite-chlorite (e.g. Walker, 1960; Neuhoff *et al.*, 1999). These minerals occur as fracture and amygdale fillings but they also replace the primary minerals. Their formation reduces the porosity and permeability of the rock.

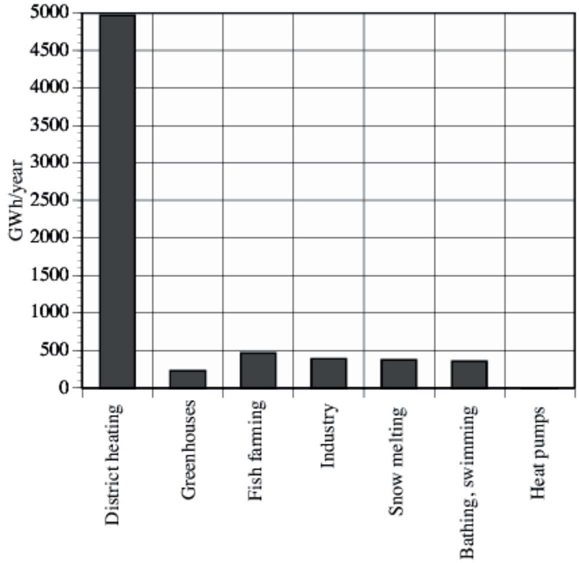

Figure 15. Direct use of geothermal heat in Iceland in 2005 by type. Based on Lund *et al.* (2005). – *Bein notkun jarðvarma á Íslandi árið 2005 miðað við tegund nýtingar.*

The zeolites have a well-marked zonal distribution in many areas of the Tertiary flood basalt sequences in Iceland (Walker, 1960; Weisenberger, 2005). The intensity of zeolitization and grain size increases with depth of burial. Deep drillings into Tertiary formations, away from central volcanic complexes, reveal that the zeolite zones are replaced by high-temperature alteration minerals typical of those found at shallower depths in high-temperature fields (Pálmason *et al.*, 1980). In the Tertiary basalts of eastern Iceland, zeolitization is more intense and crystal size larger than in other Tertiary basalt formations in the country. This is considered to be the consequence of the shifting of the active volcanic belts in Iceland. The basalts in eastern Iceland likely formed in the volcanic belt that lay between the basalts of the NW-Peninsula and Tröllaskagi in central North Iceland (see Figure 2). When this belt shifted eastwards 4–7 million years ago to its present position, it split the Tröllaskagi basalts from those in eastern Iceland causing re-heating of the latter, or maintaining it hot for a longer period thus prolonging the burial metamorphism of these basalts.

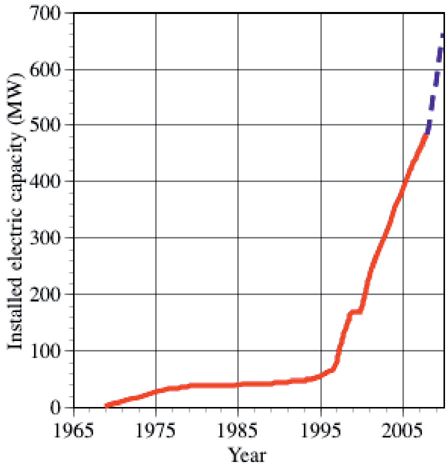

Figure 16. Production of electric power by geothermal steam in Iceland. The red line represents installed capacity by April 2008 and blue dashed line additional capacity of a power plant presently under construction. Based on Bertini (2005) and Ragnarsson (pers. comm.). – *Raforkuframleiðsla með jarðgufu á Íslandi. Rauða línan sýnir uppsett afl í apríl 2008 og bláa brotna línan viðbótar afl jarðgufuvirkjana sem eru í byggingu.*

SUMMARY OF GEOTHERMAL ENERGY UTILIZATION

Iceland is lacking subsurface Earth resources except for geothermal energy. This resource has proved to be very important economically for the country, especially since the oil crisis at the end of 1973. The use of hot-spring water in low-temperature areas has led to the development of small villages in many farming areas. Before the 1990's the main emphasis was on developing direct use of geothermal heat for house heating although other direct uses have also been important (Figure 15). The low-temperature reservoirs have been extensively utilized for this purpose but also three high-temperature fields. The last decade has seen enhanced use of geothermal steam for power production (Figure 16). The first geothermal power plant of 3MW was commissioned in 1969. By 1990 the installed capacity of geothermal power plants was 79 MW. During the last decade a 30 MW unit was added to the Krafla plant and 90 MW installed at Nesjavellir. In 2005 the total installed capacity was 202 MW and reached 484 MW in April of 2008. The large increase in power generation by geothermal steam during the last few years is by Icelandic power companies for aluminium smelters of international companies. Within the next ∼5 years further increase in power generation is expected, largely for the aluminium industry, perhaps by as much as 400 MW. In 2005 Iceland was in fifth place for direct annual use of geothermal heat (Table 3). The country was in eighth place for power generation by geothermal steam in 2005 but in seventh place in the latter half of 2007 (Tables 4 and 5). In 2007 the electric energy usage per capita in Iceland was ∼38,000 kWh which is the highest of any country in the world. The reason for this high number is production of power for various energy consuming industries, in particular aluminium smelters.

Direct use of geothermal heat in Iceland is varied. However, by far the most important type of use is house heating (Figure 15, Table 6). Of the installed capacity for direct use 435 MWt is in three high-temperature fields but the rest (1409 MWt) is in many low-temperature fields, the most important of which are near the capital of Reykjavík and the town of Akureyri.

Far the greater part of the utilized low-temperature water is extracted from wells, most often using submerged pumps. Less than 5% is tapped from hot springs. With few exceptions, the water is low in dissolved solids and is neither corrosive nor forms scales. For this reason the water can be tapped directly from the wells to the radiator systems in buildings and greenhouses. On the other hand fluids from wells in high-temperature areas cannot be used directly in district heating systems. They contain unacceptably high concentrations of hydrogen sulphide, aluminium and some trace elements and amorphous silica frequently precipitates from the water to form troublesome scales. Heating of fresh water in heat exchangers is needed to utilize heat from high-temperature liquid and vapour.

The geothermal energy brought to the surface

Table 3. Summary of direct use of geothermal heat, May 2005. From Lund *et al*. (2005). – *Samantekt á beinni notkun jarðvarma í maí, 2005. Byggt á Lund o.fl. (2005).*

Country	**Capacity**	**Use**	**Capacity**	**Use**
	MWt	GWh/year	factor	% world total
China	3,687	12,605	0.39	16.6
Sweden	3,840	10,001	0.30	13.2
United States	7,817	8,678	0.13	11.4
Turkey	1,495	6,900	0.53	9.1
Iceland	1,844	6,806	0.42	9.0
Japan	822	2,862	0.40	3.8
Hungary	694	2,206	0.36	2.9
Italy	607	2,098	0.39	2.8
New Zealand	308	1,968	0.73	2.6
Brazil	360	1,840	0.58	2.4
World total	28,268	75,943	0.31	

Table 4. Summary of worldwide geothermal power generation in early 2005. From Bertani (2005). – *Samantekt raforkuframleiðslu með jarðgufu á veraldarvísu snemma árs 2005. Byggt á Bertani (2005).*

Country	**Capacity**	**Use**	**Load**	**Capacity**
	MWt	GWh/year	factor	% world total
United States	2,564	17,917	79.7	28.7
Philippines	1,930	9,253	54.7	21.6
Mexico	953	6,282	75.2	10.7
Indonesia	797	6,085	87.1	8.9
Italy	791	5,340	77.0	8.9
Japan	535	3,467	73.9	6.0
New Zealand	435	2,774	72.7	4.9
Iceland	202	1,483	83.8	2.3
Costa Rica	163	1,145	80.1	1.8
El Salvador	151	967	73.1	1.7
World total	8,933	56,786	72.5	

Table 5. Summary of geothermal power generation in Iceland, April 2008. Based on information from Árni Ragnarsson at Icelandic Energy and Utilities, Reykjavík. – *Samantekt á raforkuframleiðslu með jarðgufu á Íslandi, apríl 2008. Byggt á upplýsingum frá Árna Ragnarssyni hjá Samorku.*

Location	**Capacity**	**Use**	**Load**	**Commission**
	MWt	GWh/year	factor	date
Efri-Reykir	0.1			1998
Hellisheidi	123	17,917	79.7	2006, 2007[a]
Húsavík	2			2000
Krafla	60	9,253	54.7	1977, 1997[b]
Námafjall	3	6,282	75.2	1969
Nesjavellir	120	6,085	87.1	1998, 2001, 2005[c]
Reykjanes	100	5,340	77.0	2006
Svartsengi	76	3,467	73.9	1977, 1980, 1989, 1993, 1999, 2008[d]
Total	484	56,786	72.5	

[a]90 MW in 2006 and 33 MW in 2007. [b]30 MW in 1977 and 30 MW in 1997. [c]60 MW in 1998, 30 MW in 2001 and 30 MW in 2005. [d]2 MW in 1977, 6 MW in 1980, 3.6 MW in 1989, 4.8 MW in 1993, 30.0 MW in 1999 and 30.0 in April 2008. – *[a]90 MW árið 2006 og 33 MW árið 2007. [b]30 MW árið 1977 og 30 MW árið 1997. [c]60 MW árið 1998, 30 MW árið 2001 og 30 MW árið 2005. [d]2 MW árið 1977, 6 MW árið 1980, 3,6 MW árið 1989, 4,8 MW árið 1993, 30,0 MW árið 1999 og 30,0 MW á apríl, 2008.*

Table 6. Summary of direct use of geothermal heat in Iceland in 2005. Based on Lund *et al.* (2005). – *Samantekt á beinni notkun jarðvarma á Íslandi árið 2005.Byggt á Lund o.fl. (2005).*

	Capacity	**Use**	**Load**	**% of annual**
	MWt	GWh/year	factor	energy use
District heating	1,375	4,972	41.3	73.1
Greenhouses	50	233	53.2	3.4
Fish farming	65	467	81.9	6.9
Industry	60	389	73.9	5.7
Snow melting	215	378	20.0	5.6
Bathing, swimming	75	361	54.7	5.3
Heat pumps	4	6	15.8	0.1
Total	1,844	6,806		

through wells drilled in high-temperature areas, which is a mixture of water and steam, or steam only, is not used effectively in conventional geothermal power plants such as those erected in Iceland during the last few years. Only about 10-12% of the thermal energy is converted into electric power. The most effective utilization involves combined power generation and direct use of the heat, such as at Nesjavellir and Svartsengi. Basically, the power plant at Nesjavellir differs from that of conventional geothermal power plants through the use of heat exchangers instead of cooling towers. The heat exchangers have a dual role. They have the role of cooling towers, which is to condense steam coming from the turbine. At the same time they are used to heat up fresh water for space heating. By the design at Nesjavellir (Figure 17), emission of steam into the atmosphere is limited and together with injection of the spent fluid, which is partially exercised at present, surface installations are largely a closed loop through which the geothermal fluid passes. Only the gases in the steam escape into the atmosphere. If the gases were disposed of in an environmentally benign way or extracted to be used for some industries, the arrangement at Nesjavellir would offer a solution that is environmentally superior to that of conventional power plants (Figure 18). This way of using the geothermal resource also involves effective extraction of heat from the liquid and vapour discharged from drillholes. The economy of this scheme depends on the availability of ground water for the heat exchangers and a market for the heated water.

The steam discharged from wells in high-temperature fields forms essentially by depressurization boiling. A small fraction may be present in the initial aquifer fluid or form by conductive transfer of heat from the aquifer rock to flowing boiling water in the depressurization zones that forms as a consequence of reservoir pressure drawdown when wells are discharged. It is common that the discharge enthalpy of wells (the steam to water ratio of the discharge) drilled into high-temperature fields in Iceland is higher than that of the parent reservoir fluid. Sometimes they discharge dry steam only. Liquid water and vapour have different flowing properties which may influence the steam to water ratio of the fluid flowing into wells. Liquid water is preferentially adsorbed onto mineral grain surfaces due to the effect of capillary forces. This may cause liquid water to be partially or totally retained in the aquifer depending on the steam/water ratio of the flowing fluid and the surface area between fluid and rock.

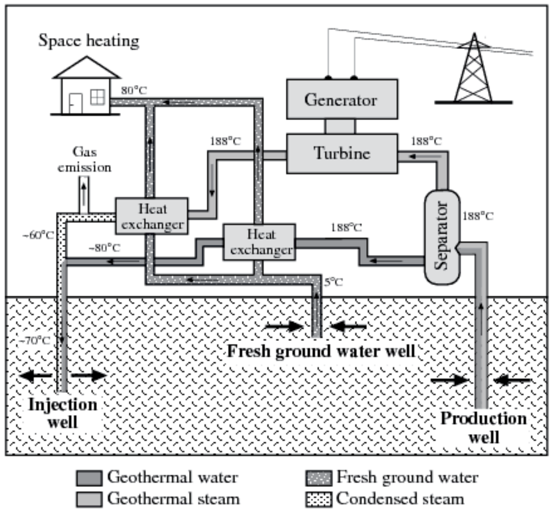

Figure 17. Simplified schematic layout of the Nesjavellir geothermal plant, Iceland. This layout offers reduced gas emission into the atmosphere and improves much efficiency of heat utilization compared to conventional geothermal power plants. From Arnórsson (2004). – *Einfölduð mynd af orkuverinu á Nesjavöllum. Fyrirkomulag virkjunarinnar býður upp á möguleika á að draga úr streymi gass í jarðgufu út í andrúmsloftið og bætir mjög nýtni varmans miðað við hefðbundnar jarðgufuvirkjanir.*

Utilization of low-temperature resources has insignificant environmental impact. This is, however, not the case with exploitation of high-temperature systems. Adverse environmental effects include scenery spoliation, soil disturbance, noise, visual and heat pollution. The kind of pollution of the greatest concern is, however, chemical, both water and airborne pollutants. To reduce surface spoliation and soil disturbance directional drilling has become the common practice that allows drilling of several wells from the same drill-pad. Disposal of spent fluids into special wells has been practised to reduce chemical pollution. Such disposal into the geothermal reservoir also has the advantage of maintaining reservoir pressures, thus counteracting decline in steam flow from production wells and increasing the lifetime of both individual wells and well fields as a whole.

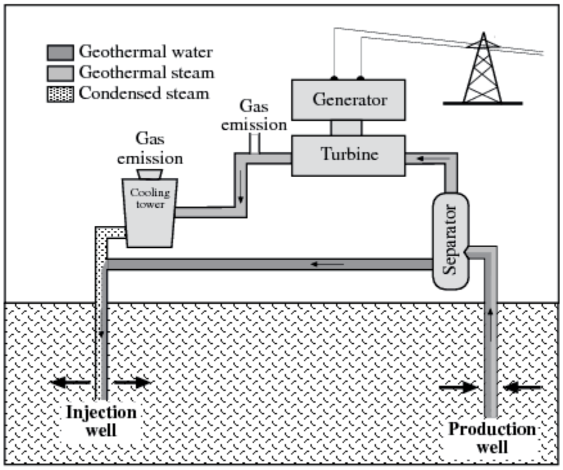

Figure 18. Simplified schematic layout of a classic geothermal power plant. From Arnórsson (2004). – *Einfölduð mynd af hefðbundinni jarðgufustöð.*

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ÁGRIP

Jarðhitasvæði á Íslandi hafa verið flokkuð í háhita- og lághitasvæði. Háhitasvæðin liggja innan hinna virku gos- og gliðnunarbelta en lághitasvæði er að finna í kverteru og tertíeru bergi. Hitagjafi háhitans er grunnstæð kvikuinnskot. Því teljast þessi svæði eldfjalla- eða kvikusvæði samkvæmt algengri alþjóðlegri flokkun jarðhitasvæða. Í ljósi gagna um náttúrulegt varmastreymi nokkurra háhitasvæða er talið að varmaleiðnilagið milli bergbráðs og vatns eða gufu í botni hræringar sé tiltölulega mjög þunnt, frá fáum tugum metra upp í nokkuð hundruð í mesta lagi. Flest háhitasvæðin liggja í virkum sprungusveimum þar sem þeir skera flekamót. Megineldstöðvar hafa myndast sumsstaðar á þessum skurðpunktum flekamóta og sprungusveima og í sumum þeirra hafa öskjur myndast. Vitað er að flest lághitasvæði á Íslandi tengjast ungum, nær lóðréttum sprungum í kverteru og tertíeru bergi á Ameríku-flekanum og Hreppasmáflekanum. Þessi jarðhitakerfi teljast því tektónísk samkvæmt algengri alþjóðlegri flokkun jarðhitasvæða. Varmagjafi þeirra er heitt berg í rótum þessara svæða. Í háhitasvæðum er hræring vatns knúin áfram af eðlisþyngri kaldrar vatnssúlu í berggrunni utan svæðanna og eðlisléttarri heitri vatnssúlu innan þeirra. Í mörgum lághitasvæðum er hringrás grunnvatns knúin með sama hætti en í öðrum með hæðarmun grunnvatnsborðs. Hæsti hiti sem mælst hefur í gufuholu á háhitasvæði er >380°C á $\sim$2 km dýpi (hámarkshiti mælanlegur með hitamæli var 380°C). Hæsti hiti sem mælst hefur á lághitasvæði er 175°C á 2 km dýpi. Mörg forn rofin há- og lághitasvæði finnast í kverterum og tertíerum jarðmyndunum. Athuganir á þssum svæðum veita mikilvægar upplýsingar um jarðfræðilega byggingu jarðhitasvæða. Varmaflæði gegnum jarðskorpuna undir Íslandi hefur verið áætlaður um 30 GW (1 GW = 109 wött). Samsvarar þetta flæði 5-földu meðalvarmaflæði jarðar á flatareiningu. Við yfirborð skiptist þetta orkuflæði þannig: 7 GW berast með bergbráð, 8 GW með jarðhitavatni og jarðgufu á jarðhitasvæðum og 15 GW berast með varmaleiðni. Sú varmaorka sem er í jarðskorpunni niður á 10 km dýpi undir Íslandi er $\sim$1,2 EJ (1 EJ = 10^{24} J) Ofan 3 km dýpis hefur þessi varmaorka verið metin sem 0,1 EJ. Jarðhitavatn á Íslandi er úrkoma að uppruna, eða

sjór eða blanda af slíku vatni. Lágt tvívetnisinnihald í sumu jarðhitavatni stafar af því að hluti þessa vatns er jökulbráð frá síðasta jökulskeiði. Frumjarðhitavatn sem ekki inniheldur sjó er tiltölulega snautt af klóríði (Cl) og öðrum uppleystum steinefnum. Orsökin er lágur styrkur Cl í basalti. Jarðvarmi er mjög mikilvæg orkuauðlind á Íslandi og felur bæði í sér beina nýtingu jarðvarmans og nýtingu jarðgufu til raforkuframleiðslu. Árið 2005 nam bein notkun jarðvarmaorku $\sim$ 6800 GWh (gígawattstundir; 9% af hnattrænni notkun) og uppsett afl var 1844 MWt. Í apríl 2008 var uppsett afl jarðgufuorkuvera 484 MWe (5% af heild á hnattræna vísu). Fjögur ríki nýta jarðvarma beint meira en Ísland og í sex ríkjum er uppsett afl jarðgufuorkuvera hærra en á Íslandi.

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