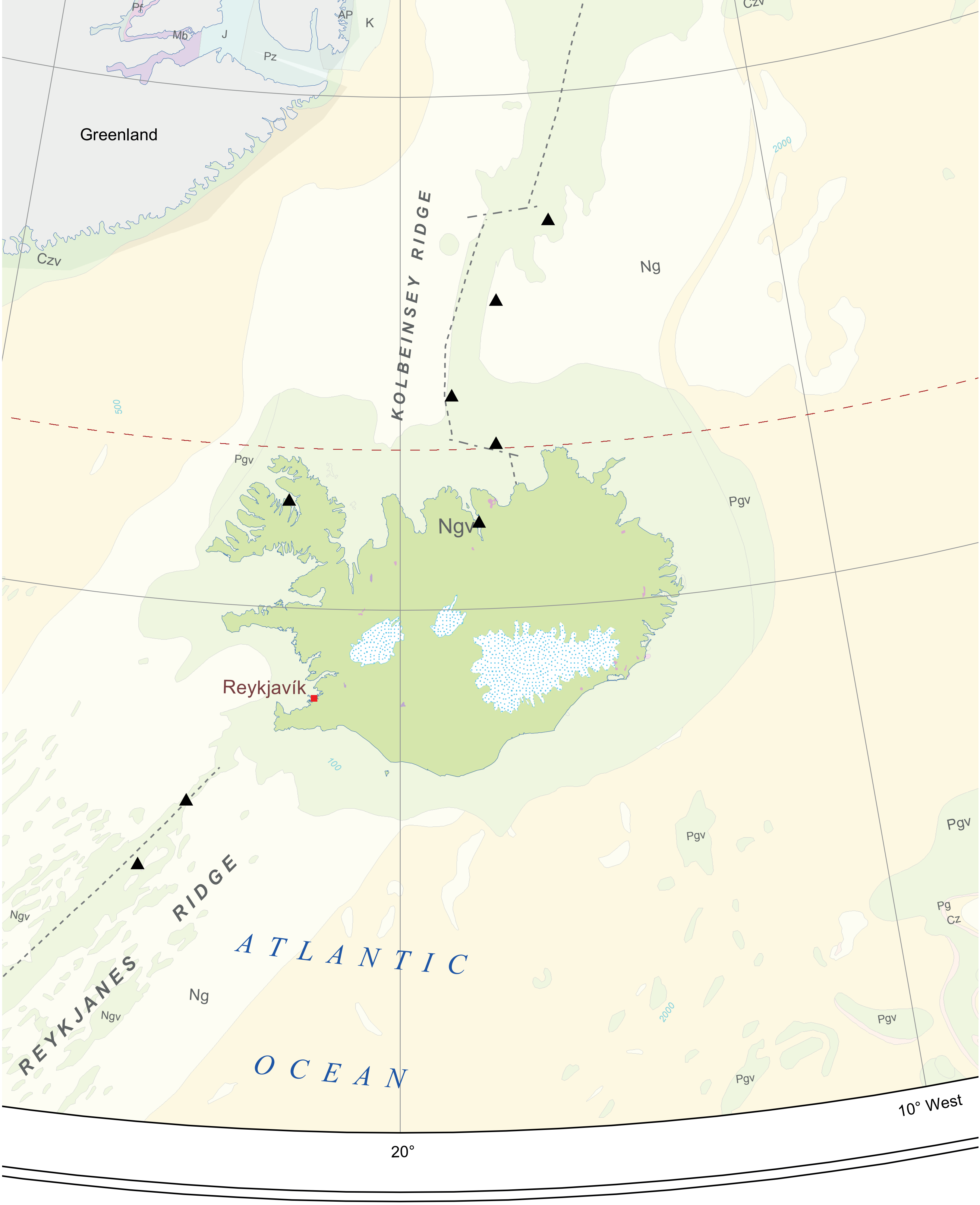

Deposits

▲ Hydrothermal fields

North Atlantic Ridge

- - - - - Ridge

— · — Transform fault

Scale 1:5 000 000

Stereographic North Pole Projection
Standard Parallel 70°N Coordinate System WGS 1984
Prime Meridian: Greenwich (0.0), Central Meridian 20°W

CHAPTER 4
ICELAND

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GEOLOGICAL OUTLINE

Iceland lies within the North-Atlantic Igneous Province (NAIP) about 63-67°N, and owes its existence to an underlying mantle plume and to the Mid-Atlantic Ridge that runs through the country (Figure 1). The trace of the mantle plume, which at least dates back from the initial opening of the North-Atlantic Ocean (63 my), runs from the eastern coast of Greenland in the west across to Faeroes and Scotland in the southeast. The crustal structure of Iceland differs somewhat from the oceanic crust. It is considerably thicker due to anomalous volcanic accumulation. The rift zones of Iceland consist of fissure swarms commonly with a volcanic complex residing in the central part (Figure 2). These complexes are the locus of anomalous volcanic and intrusive activity and frequently form caldera subsidence structures. In these volcanic complexes the rock types range from basaltic to more evolved rocks such as andesite and rhyolite, either formed through fractional crystallization or partial melting of the lower crust. There the relatively shallow magmatic heat-sources and volatiles, along with highly permeable strata lead to the formation of vigorous high-temperature systems. The lifetime of a volcanic complex along with its multiphase geothermal system ranges from 0.3 to over 1 million years, where its extinction is related to the drifting out of the volcanic zone. The average rifting velocity in Iceland is about 1 cm/y in each direction and the age of the crust increases gradually away from the rift zones to a maximum of about 15 my on either side (Saemundsson 1979, Hardarson et al. 2008). The geological map in Figure 2 depicts several formations. The volcanic zones with their postglacial lava fields are surrounded by late Pleistocene areas (< 0.8 my). Farther out are the early Pleistocene areas (0.8-2.6 my) (Hjartarson & Saemundsson 2014). Evidence of 20-30 individual glaciations have been found interbedded in the Pleistocene stratapile along with inter-glacial stages (Eiriksson 2008). The extent of the ice sheet has been traced to the outer limit of the insular shelf as indicated in Figure 1. Besides putting weight on the crust, the ice sheet changes the hydrological condition of the underlying crust, causing a variable exchange between fluid availability, permeability and heat-sources and the high-temperature systems, which may in some instances induce favorable conditions for metal deposition. No major glaciations are evidenced in the volcanic sequence in the Tertiary regions (2.6-15 my) as indicated by the dominance of sub-aerial lavas. The main erosion in Iceland is due to the repeated glaciations, and this increases away from the volcanic zones, from no erosion to 1-2 km deep erosion valleys in the older regions.

Fluid geochemistry plays an important role when considering metalliferous epithermal environment. Sulfur is common in the high-temperature systems and is considered to be the main carrier of metals. Salinity is important transport agent of metalliferous complexes. This is not the case in the Icelandic environment where meteoric water dominates in the geothermal systems with chlorine content less than 200 ppm. The geothermal systems on the outer Reykjanes Peninsula in SW-Iceland, however, have higher salinities, approaching that of seawater (Hardardottir et al. 2009).

Primary gold concentrations in Icelandic volcanic formations range from 0.5 to about 20 ppb with an average value of 3.6 ppb (Zentilli et al. 1985, Nesbitt et al. 1986, Geirsson 1993). These are higher values than the 2 ppb quoted for regular mid-oceanic ridge basalt (MORB) (Peach et al. 1990).

Iceland presents a very different geological environment in comparison with Scandinavia, Greenland, Canada, Alaska and Russia. It can, however, be viewed as a "living" example of the geological environment and processes which, in earlier geological periods, created some of the metallogenic provinces now hosting deposits which are being mined elsewhere. It also illustrates processes that are to be found within the oceanic crustal environment in several parts of the Arctic Ocean.

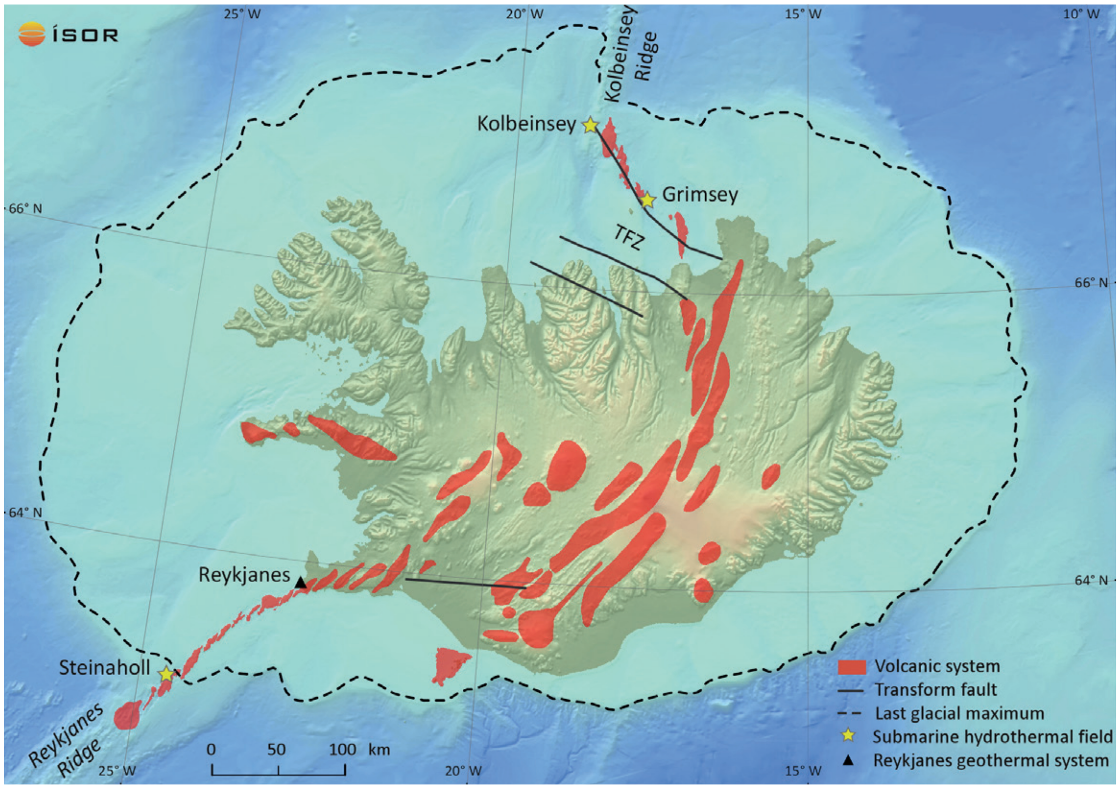
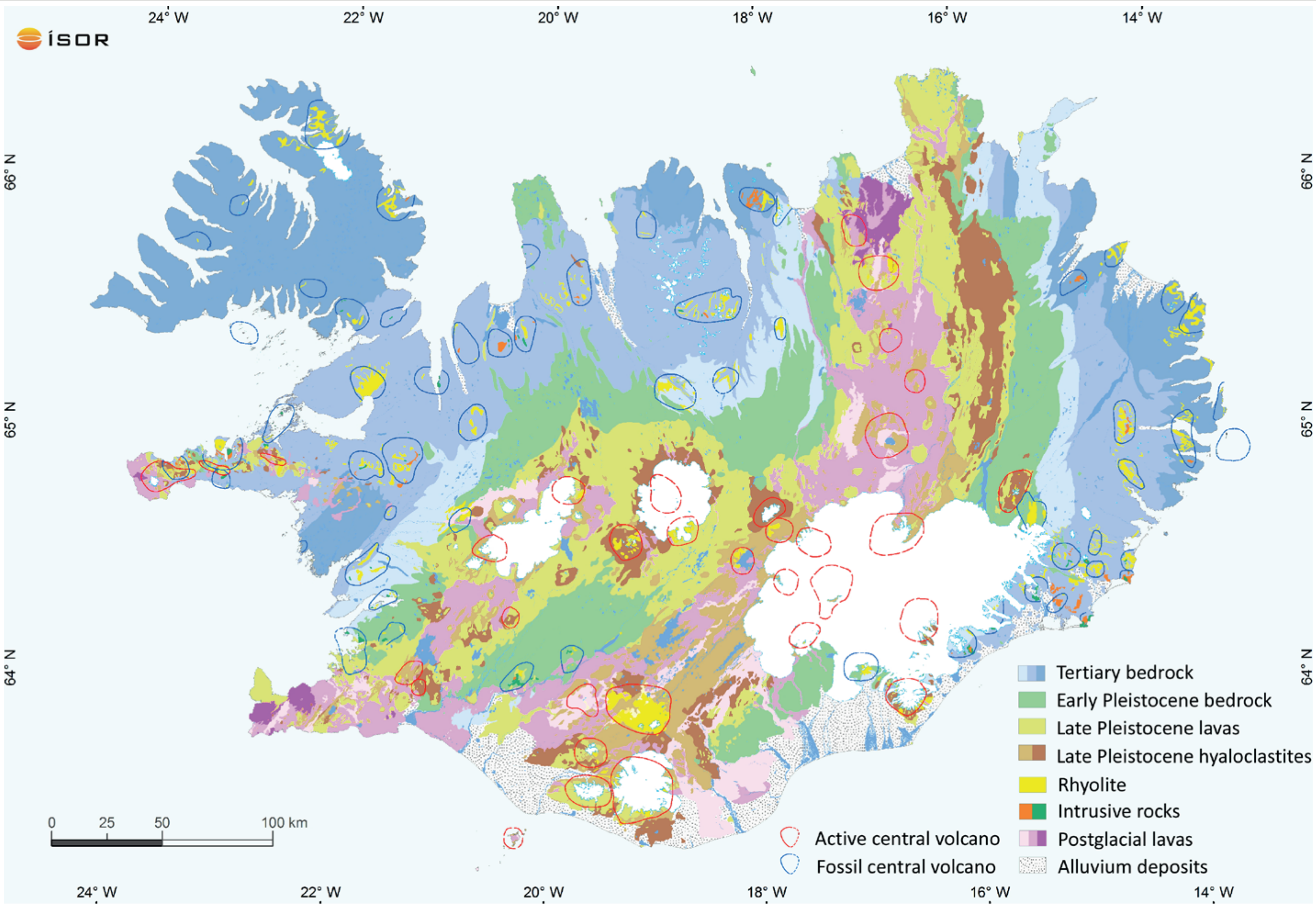


Figure 1. Map of the Mid-Atlantic Ridge on- and offshore Iceland (map modified from Hjartarson & Erlendsson 2016).

Figure 2. Geological map of Iceland showing Quaternary, including the volcanic zones, and Tertiary formations. Also shown are active central volcanoes and eroded, fossil central volcanoes (map modified from Hjartarson & Saemundsson 2014).



Exploration for gold in Iceland dates back to the early 20th century (Kristjansson 1929, Fridleifsson et al. 1997). The early prospectors were e.g. Björn Kristjansson, a politician and a bank manager, and Einar Benediktsson, a famous poet and entrepreneur who found a number of gold-rich localities. The most notable of these are in Mogilsa and Thormodsdalur, northeast and east of the capital Reykjavik (Figure 3). The latter, a multiple quartz-vein system, was explored by tunneling and surface excavations. The exploration stopped, apparently due to the economic recession in Europe during and following World War I. Gold exploration within the city of Reykjavik led, however, to the import of a drilling rig which (though mineable gold was not found) became very useful in drilling for geothermal water in Reykjavik (Jonasson 2006).

Renewed interest for gold exploration in Iceland developed about 1989 when the close connection between thermal activity and gold deposition, even in low salinity environments, became apparent. This led to limited reconnaissance surveys financed by private and governmental sources, later followed up by the exploration companies Malmis/Melmi and Sudurvik which made extensive, low-density reconnaissance surveys in eroded sections of the country (Franzson et al. 1992, Franzson & Fridleifsson 1993, Oliver 1993, Franzson et al. 1995, 1997).

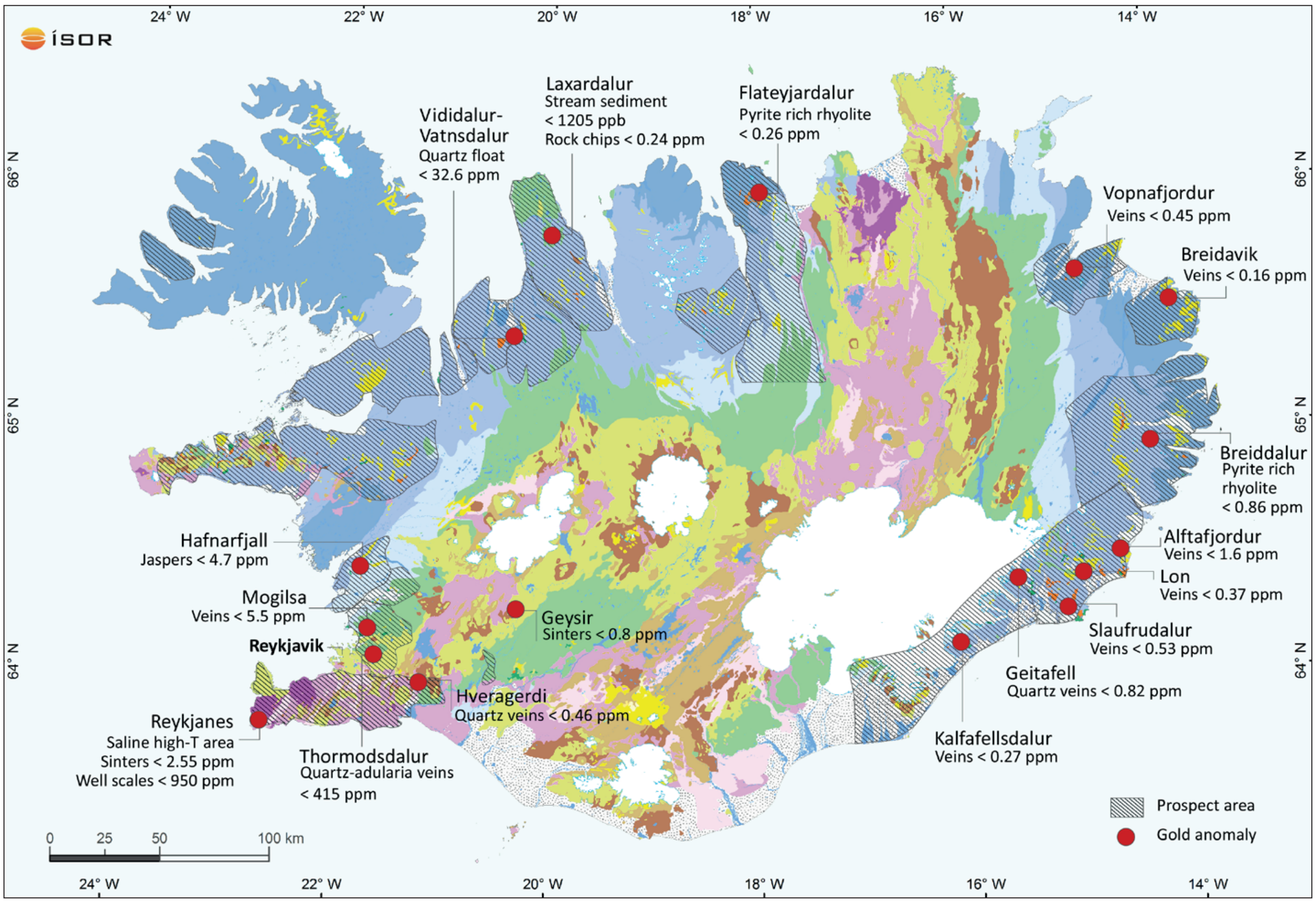


Figure 3. Map of gold anomalies discovered in Iceland as of 2015 and the prospect areas which have been explored to variable degree (Franzson et al. 1992, Karajas 1998, Fleming et al. 2006, Hardardottir 2011; map modified from Hjartarson & Saemundsson 2014).

GOLD PROSPECTS

Conditions for exploration in Iceland are good as the country is relatively barren, with good bedrock exposures, and limited sedimentary and vegetation cover. The target areas are fossil high-temperature systems within exhumed -central volcanoes. The basic theme for exploration has been the epithermal origin of gold which would imply targets limited to relatively shallow parts of the geothermal systems, i.e. above about 1000 m depth (Franzson et al. 1992, Franzson & Fridleifsson 1993). Geothermal systems in Iceland have been viewed as multiphase, extending up to 1 my in age, with thermal conditions changing according to the appearance of heat sources and/or renewed pathways for the in-/outflow of geothermal fluids. Areas around larger intrusive bodies at the base of the central volcanic complexes are also of interest for exploration, mainly for possible evidence of metalliferous magmatic volatiles. It should though be taken into consideration that magmas from oceanic crustal environment are expected to have lower volatile contents than those derived from subduction or continental environments. Base metals in high concentrations have not been found in Iceland, except at two locations in the southeast, both associated with volatiles from a rhyolitic magma source (Jankovic 1970, Franzson & Fridleifsson 1993).

The main exploration methods used are stream-sediment, rock and grab sampling. A summary of the exploration results shown in Figure 3, is mainly derived from the data collection of Melmi (e.g. Karajas 1998, Franzson et al. 2013). These show the maximum gold concentrations in individual rock or sediment samples from some of the licence areas. Many of these range from 0.2 to a little over 1 ppm Au. Four areas show more anomalous values, i.e. Vididalur-Vatnsdalur in N-Iceland with up to 32.6 ppm, Hafnarfjall in W-Iceland with samples reaching up to 4.7 ppm, Mogilsa in SW-Iceland up to 5.5 ppm and last, but not least, Thormodsdalur in SW-Iceland with values reaching up to 415 ppm. Exploration in the two last-named areas has been studied further, as described below.

Mogilsa prospect

The Mogilsa prospect is located on Esja Mt. some 20 km northeast of the capital Reykjavik (Figure 3). The interest in the area started in 1875 when mining of a thick calcite vein system for production of lime began (Figure 4). Later studies, around 1917, indicated the presence of gold in the veins, but efforts to establish a gold mine were aborted due to lack of belief in the analytical data.

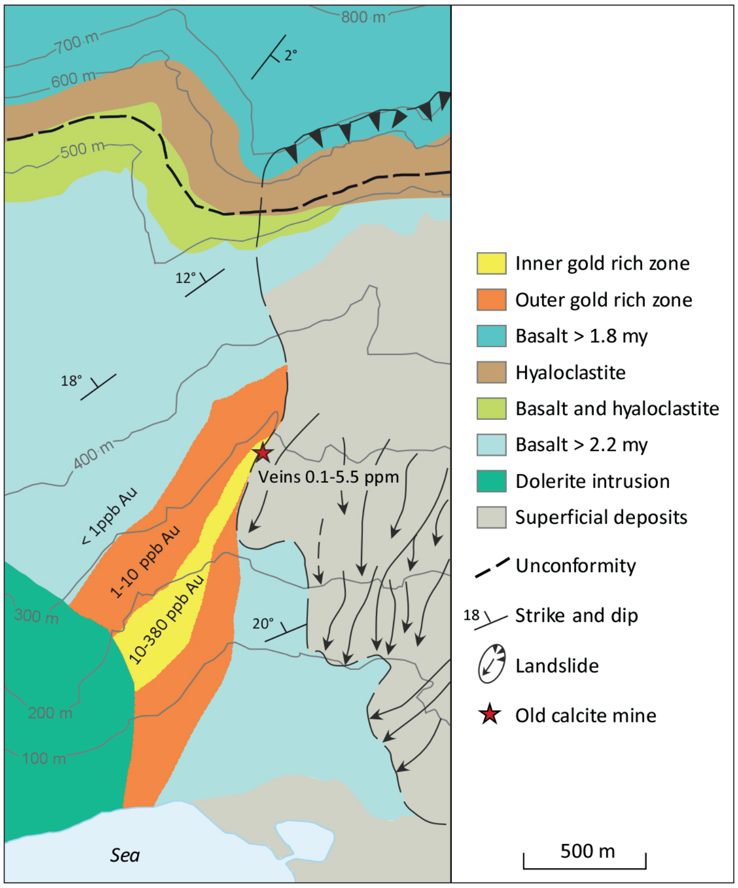


Figure 4. Map of the Mogilsa gold anomaly showing the BLEG gold values in the inner and outer zones (Franzson et al. 1992; map modified from Torfason et al. 2000).

Renewed gold exploration in the late 1980s confirmed the presence of gold in the area (Franzson et al. 1992, 2008; Franzson & Fridleifsson 1993). The host rocks are Quaternary (~2 my) and apparently part of the Kollafjordur Intrusive Complex to the south (Fridleifsson 1973). The area shows high-temperature alteration with chlorite-epidote alteration around sea level and chlorite reaching up to 400 m elevation, mainly related to a NE-SW-trending, heavily sulphidized zone. This zone contains gold enrichment which was defined by profiles using BLEG (Bulk Leach Extractable Gold) analytical methods and shown in Figure 4. The anomaly is concealed towards the northeast, where it disappears under a recent landslide. The BLEG gold values in the outer zone range from 1-10 ppb and in the inner zone from 10-380 ppb. This is concomitant with increasing sulphidization, which is most intense in the inner zone, where sampling of veins shows gold contents from 0.1 to 5.5 ppm. Breccia in the core of the vein system suggests the presence of a hydrothermal explosion breccia. Hydrothermal alteration of the rocks indicates intense chloritization at the time of gold enrichment. A fluid inclusion study shows a Th-temperature range of 200-270°C which conforms to a boiling-point curve depth of 300-500 m. An unconformity occurs some 300 m above the main anomaly, seen both as a change in strata inclination and alteration. The unconformity probably represents the surface of the geothermal system at the time of ore formation, during a state of intense boiling, a very favourable condition for gold precipitation. Figure 5 shows the conceptual model of the geothermal system and the location of the anomaly deduced from the field study. It predicts the presence of a broader, underlying geothermal reservoir narrowing upwards along a tectonic lineament. The intense sulphide zone indicates the presence of an underlying, degassing magma intrusion. The Mogilsa area lies within a very popular mountain hiking route, which may lead to public reservations regarding permits for exploration drilling and exploitation.

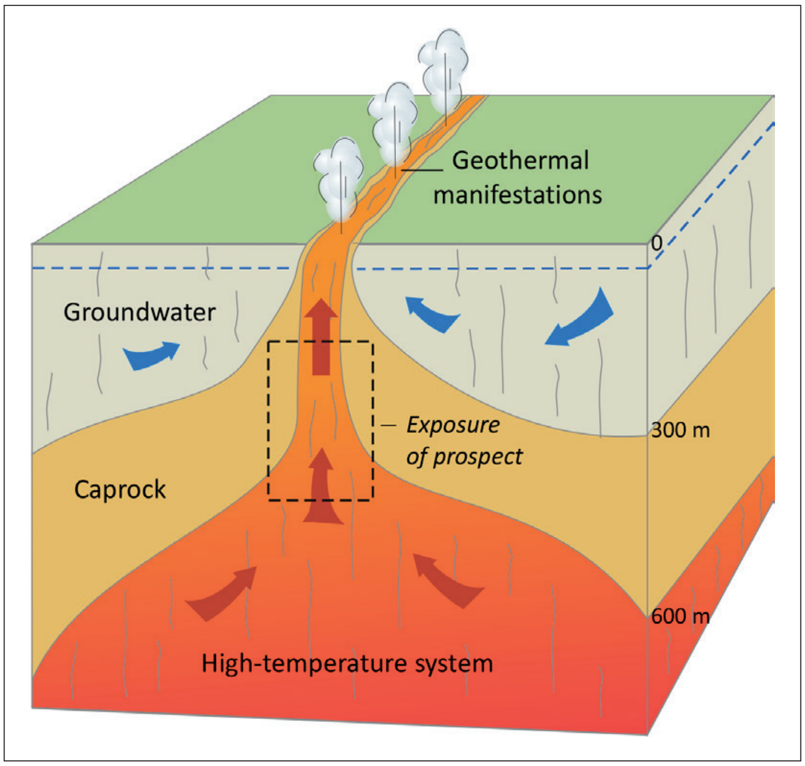


Figure 5. Conceptual model of the Mogilsa geothermal system (modified from Franzson et al. 1992).

Thormodsdalur prospect

The Thormodsdalur prospect is, as yet, anomalous in Iceland in relation to gold enrichment. Its location is about 10 km from the outskirts of the capital Reykjavik (Figure 3). Initial exploration of the locality was made by a local farmer and his family and then further aided by the poet and entrepreneur Einar Benediktsson: Over 300 m of excavations and tunneling were achieved during 1908-1925 by three consecutive exploration companies. The rocks were at one stage exported to Germany, but reports of their Au content remain speculative (Fridleifsson et al. 1997).

The country rock is dominantly pillow-rich hyaloclastites with subordinate sub-aerial lavas. The strata dips about 12° SE. The area may belong to the Stardalur Central Volcano (1.5-2 my) located about 6 km to the northeast (Fridleifsson 1973). The prospect area is within the chabazite-thomsonite alteration zone, indicating a low temperature environment (30-50°C) and a burial depth of 300-500 m. However, data from nearby wells to the north show an underlying propylitic alteration due to a fossil high-temperature reservoir, belonging to the Stardalur Central Volcano.

The area is densely faulted, mostly by NE-SW-trending faults parallel to the rift fractures. More northerly normal faults with a dextral strike-slip component and fracture trends are also evident. Their occurrence may be related to the structural change from a normal rift to the hybrid rift-transform environment of the Reykjanes

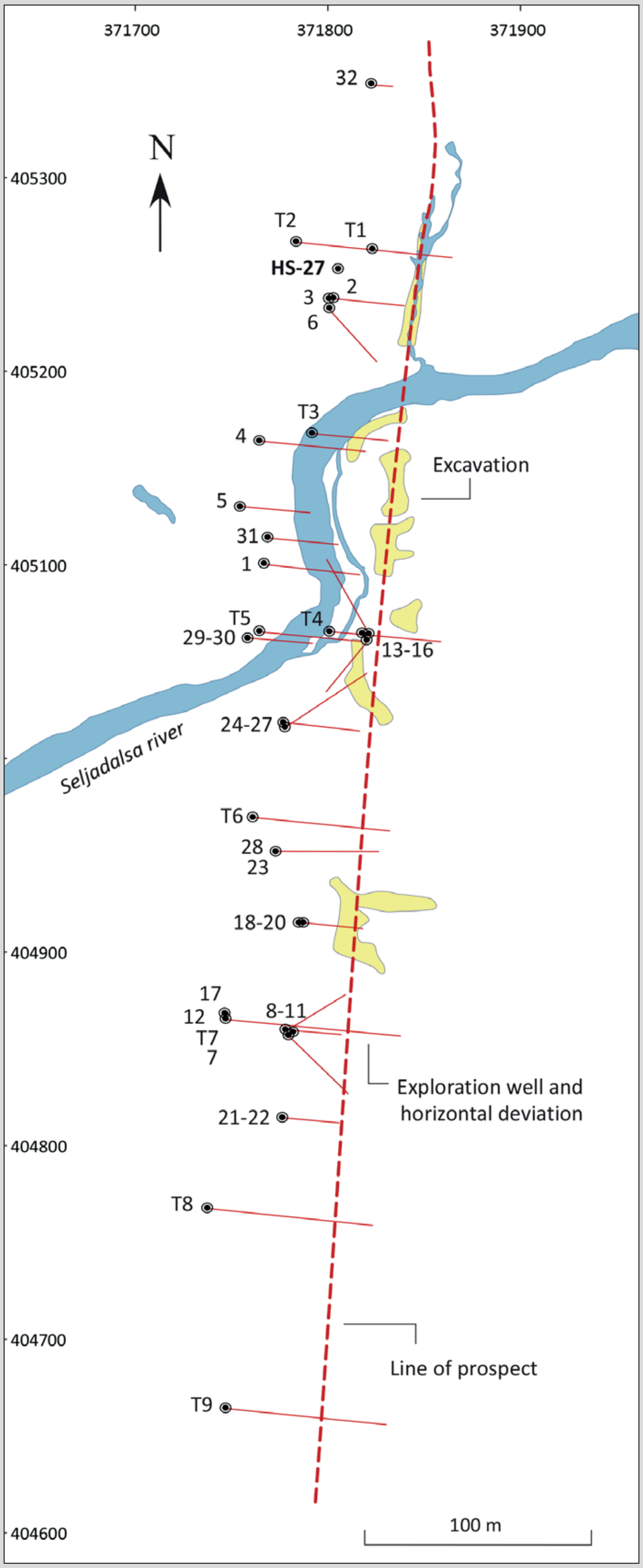


Figure 6. Map showing the alignment of the quartz-adularia vein and the location and horizontal projection of the 41 drillholes at Thormodsdalur. Excavation north of the river shown in Figure 8 (modified from Fleming et al. 2006).

Figure 7. The Thormodsdalur prospect area looking north along the trace of the quartz-adularia vein system (photo: Hjalti Franzson ca. 2000).

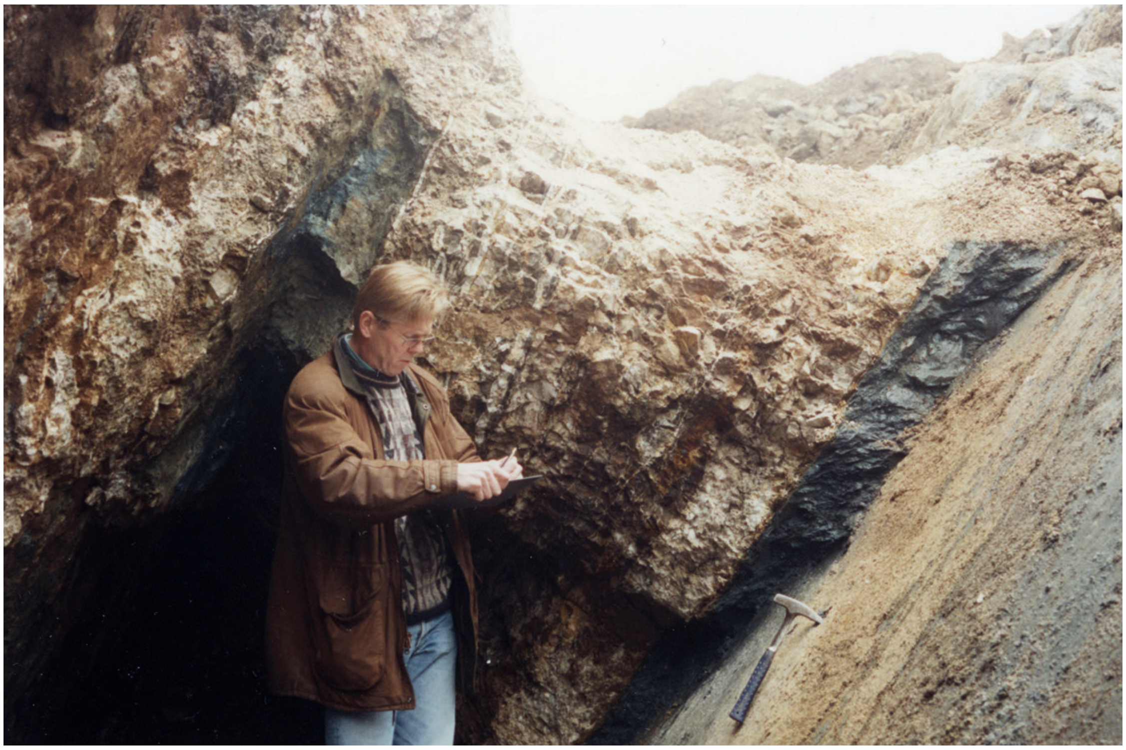


Figure 8. Excavation of the multiple quartz-adularia vein system north of the river at Thormodsdalur (photo: Hjalti Franzson ca. 2000).

Peninsula to the south. The Thormodsdalur structure belongs to the latter northerly trend and has thus a transform character. This fault has been traced for about 700 m (Figures 6 and 7).

Petrographic and XRD studies show the evolution of the vein system from a zeolite assemblage to quartz-adularia indicating progressive heating of the system and lastly to a minor calcite. The Au-enriched zone belongs to the quartz-adularia assemblage. A preliminary SEM study shows Au-grains up to 20 µm across. Temperature estimates based on mineral zonation and a limited fluid inclusion study suggest a range of 180-230°C which concurs with boiling conditions in the geothermal system at approx. 300 m depth (Franzson & Fridleifsson 1993). A review of the data suggests that the deposit is categorized as a low-sulphidization, adularia-sericite, epithermal Au-Ag type (Corbett 2004). The limited wall rock alteration suggests that this part of the geothermal system may have been relatively short-lived, but intense.

Forty-one cored holes have been drilled into the vein system, totalling nearly 3000 m. These drill holes generally extend to <100 m, many of them inclined. They indicate significant grades and thicknesses confined to two shoots along the vein structure (Fleming et al. 2006). A 450 m deep, temperature-gradient hole (HS-27, see Figure 6) was, in addition, drilled slightly to the west of the vein system, where it intersected Au-enriched veins at depths down to 450 m. The Au grades of the veins in the holes are variable, which is not surprising considering the mineral evolution discussed above: they range from <0.5 ppm to a maximum of 415 ppm (40 cm core sample in one hole).

Reykjanes geothermal system

The Reykjanes sub-aerial, high-temperature field is located at the westernmost tip of the Reykjanes Peninsula, where the Mid-Atlantic Ridge emerges on land (Figure 1). It is an extensively drilled seawater dominated geothermal field, and has a geological succession made up of submarine strata of volcanic origin which could be an analogue to oceanic crust, and has furthermore been proved to precipitate well scales that are almost identical to those found in typical black smokers (Hardardottir et al. 2009, 2010; Hardardottir 2011). It is thus an ideal location to study internal structures of a high-temperature system and reservoir characteristics for comparison with equivalent submarine hydrothermal systems.

The main geological features of the field are shown in Figure 9. The area is largely covered by sub-aerial basalt lavas erupted in postglacial time along with low level hyaloclastite ridges from the last glacial stage. Tectonic structures are related to both dilation rift and transform structures. Surface geothermal manifestations include mud pools and steam vents. Heavily altered ground is found within an area extending over about 1 km² as shown by the broken line in Figure 9.

The geothermal field is harnessed for geothermal steam to produce electricity, and for that 35 wells have been drilled into the reservoir extending from a few hundred to a maximum 3200 m depth. A study of the stratigraphy shows a dominance of pillow basalt formations in the lower part but gradually changing to a succession of tuffaceous volcanic formations of Surtseyan type above about 1000 m depth with intervening shallow water fossiliferous tuffaceous sediments. These predominate up to about 100 m depth where sub-areal lavas top the sequence. The stratigraphic cross section (Figure 10) of the shallower wells shows this character which depicts an accumulation of volcanic products from deep (pillow basalt) to shallow depths (tuffaceous sediments) which finally reaches above sea-level to produce sub-aerial lavas (Franzson et al. 2002).

Intrusions are commonly found in the succession below about 800 m depth. These are mostly fine to medium grained basalt dykes. An abundance assessment suggest they may reach up to 60% of the succession at deeper levels. These act both partly as heat source and permeability structures (Franzson et al. 2002). The increasing abundance of intrusions with depth is in line with what is expected within an oceanic crust.

Hydrothermal alteration shows a progressive intensity with depth from fresh rocks through zones characterized by smectite-zeolite, mixed layer clay, chlorite-epidote, epidote-amphibole to amphibole zone at deepest level (Tomasson & Kristmannsdottir 1972, Franzson et al. 2002, Marks et al. 2011). The temperature in the geothermal system mostly follows the boiling point

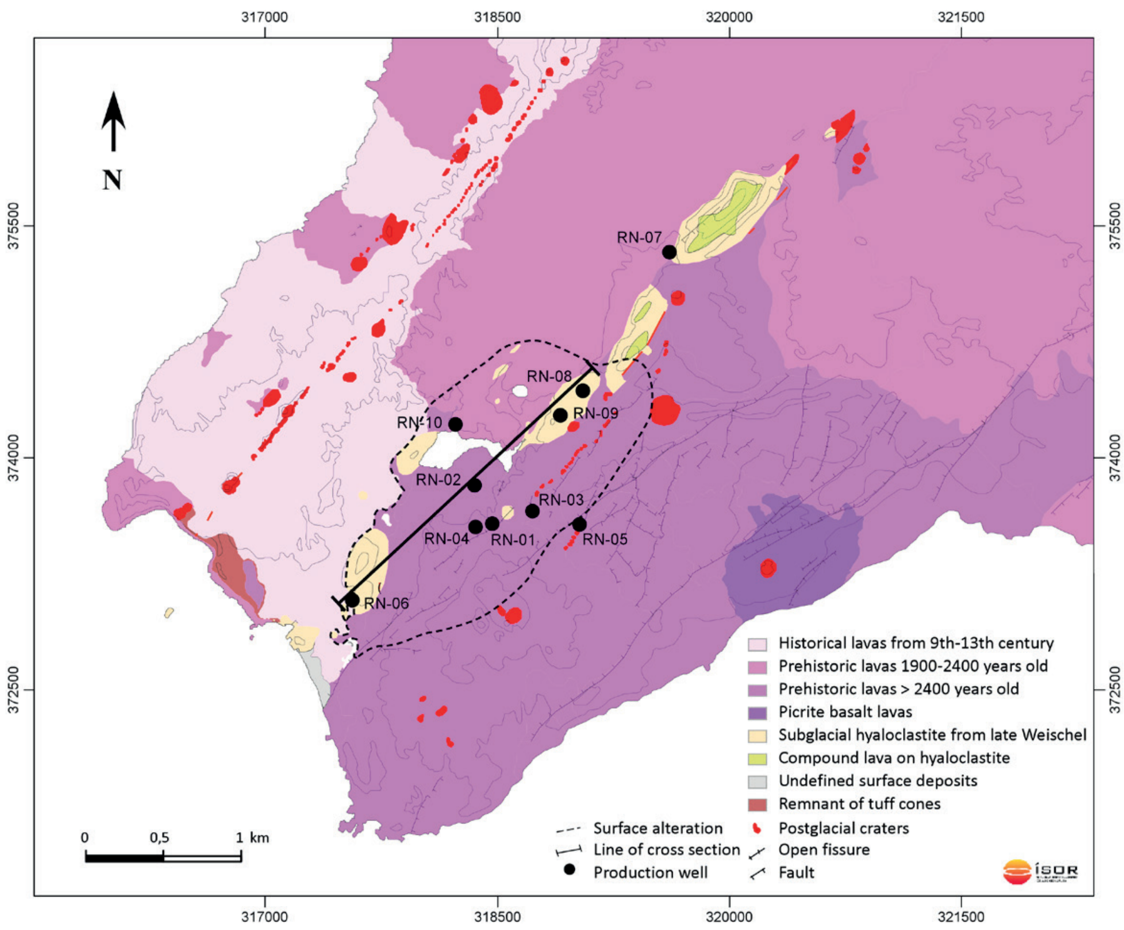
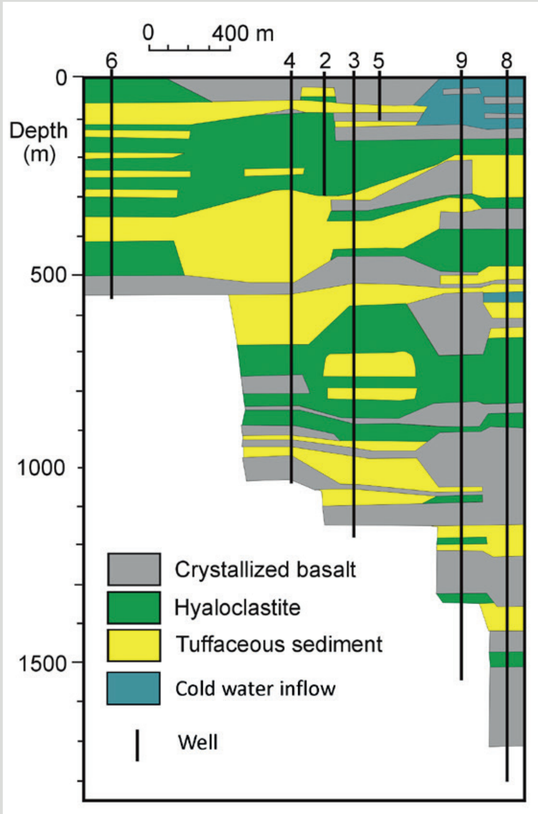


Figure 9. Map of the main geological surface features of the Reykjanes high-temperature field, including ten wells (Franzson et al. 2002; map modified from Saemundsson et al. 2010).

curve down to about 1200 m depth below which it becomes more water dominated. Highest temperatures found are around 340°C at about 3 km depth. The seawater dominated geothermal fluid is mined from deep aquifers which enter the wells at high pressures and ascends to the surface. At the wellhead the fluid is decompressed resulting in steam flashing which results in

Figure 10. NE-SW cross section of the Reykjanes high-temperature field based on the early wells. The location of the cross section is shown in Figure 9 (Tomasson & Kristmannsdottir 1972, Lonker et al. 1993, Hardardottir 2002).



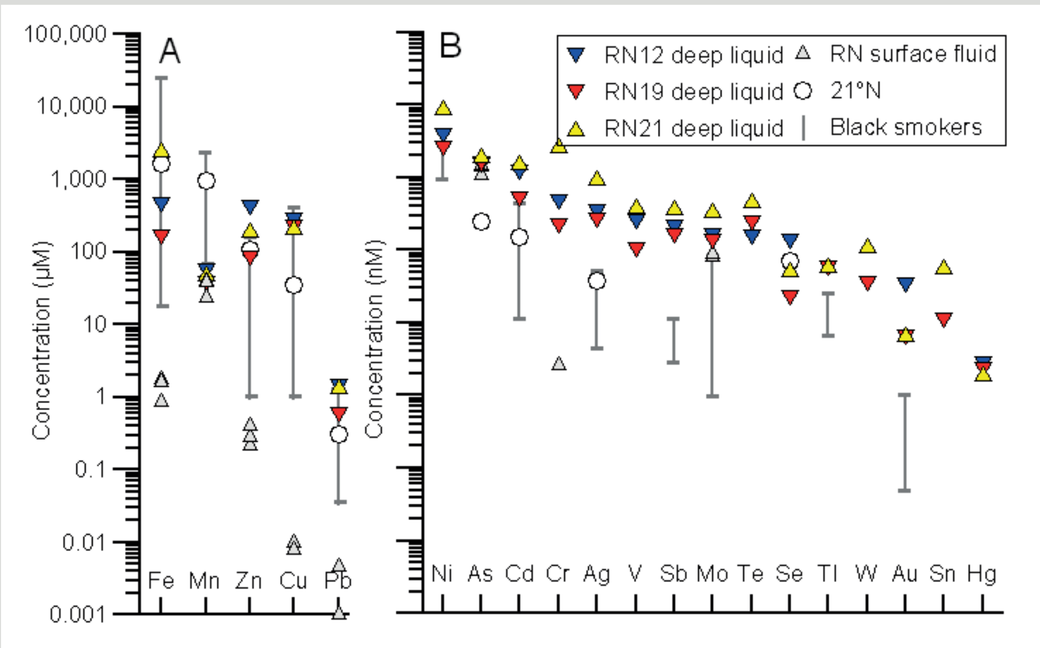


Figure 11. Metal and trace-element concentrations (in µMoles and nMoles) from three wells in the Reykjanes geothermal system compared to black smokers (21°N EPR, TAG and 5°S at MAR) (Hardardottir et al. 2009).

heavy pipe scaling. The analysis of these scales show high-percentage of Zn, Fe, Cu, Pb, and S, and Au concentrations up to 950 ppm and Ag up to 2.5 %. The fluids at the wellhead are modified as they have boiled from 1200 m depth resulting in the precipitation of sulphides (mainly ZnS) within the wells. Sampling of deep unboiled fluid in wells has been done (Hardardottir et al. 2009, 2013) to assess the unmodified reservoir fluid composition (Figure 11). A comparison between black smoker fluids compositions and those of Reykjanes show that they are very similar in most elements. It has even been argued that due to sampling problems of true black smoker fluids, the Reykjanes samples may be better representatives of such fluids (Hardardottir et al. 2009).

The Reykjanes geothermal system has been harnessed for steam for a few years allowing a monitoring of its behavior. One of the results has shown an effective permeability barrier between the system and the surrounding groundwater systems. This is explained firstly by the rapid mineral deposition in the upflow channels forming a caprock. Secondly, at the side of the geothermal system where inflow is expected into the system, studies of wells show clear evidence of very massive anhydrite deposition in all tectonic veins, which clearly indicates a near instantaneous clogging of fracture permeability. Replenishment of fluids into the geo<   thermal system through tectonic fractures, at least in the upper “2 km” should therefore be considered to consist of small short term dosages rather than long term massive inflows.

SUBMARINE HYDROTHERMAL SYSTEMS

Three submarine hydrothermal systems found on the Mid-Atlantic Ridge close to Iceland are shown in Figure 1. They are located at relatively shallow waters and if at boiling condition, the temperatures would lie along the hydrologically controlled boiling point curve. Reliable data is available from these three systems; Steinaholl hydrothermal field on the Reykjanes Ridge, Grimsey hydrothermal field north of Iceland, and Kolbeinsey hydrothermal field near the southern end of the Kolbeinsey Ridge. Another feature shown on Figure 1 is the outline of the maximum extent of the Icelandic ice sheet during the last glacial period. The presence of these is evidenced by glacial debris, end-moraines and erosional features (Hjartarson & Erlendsson 2015). The latter implies that hydrothermal precipitation may have been eroded and those observed today may be limited to postglacial times. This would apply to the Kolbeinsey and Grimsey hydrothermal fields, while the Steinaholl field on the Reykjanes Ridge is found at the outer margin of the ice sheet.

Scarcity of hydrothermal manifestations on the Reykjanes Ridge south of Iceland has aroused speculation. According to German & Parson (1998) transform zones are near absent for most of the Reykjanes Ridge reflecting low incidence of venting on the ridge. One explanation for the absence of black smoker deposits may be that chemical signals from thermal plumes may be disguised by the overlying very strong ocean currents. The evidence of the very effective inflow barrier created by the anhydrite, as clearly shown in the Reykjanes geothermal system data, may also point to limited inflow into black smoker systems. Indeed this may imply that a creation of a hydrothermal system on the ocean floor may just as much depend on the availability of recharge seawater as the availability of a heat source. The link between the occurrences of hydrothermal systems on the ocean floor and the transform-rift locations further south may simply be due to these locations being the only ones where fluid recharge into a heat source is sufficient to form a hydrothermal system able to penetrate to the ocean floor. This may be the underlying reason for the scarcity of seafloor hydrothermal systems on the Reykjanes Ridge.

Steinaholl hydrothermal field, Reykjanes Ridge

The Steinaholl hydrothermal field is located at the crest of the Reykjanes Ridge at 63°06.0N and 24° 31.98W (Figures 1 and 12). The area was initially discovered by fishermen who caught hot stones in trawling nets. Following an earthquake swarm in 1991 a research team, led by Icelandic research institutions, revealed the existence of a manganese deposit consisting of todorokite and birnessite (Thors et al. 1992, Olafsson et al. 1991). The Reykjanes Ridge consists at this location of en-echelon arranged fissure swarms, the Steinaholl hydrothermal field is situated in the northern sector of one of these (Figure 1). A more detailed bathymetry map of the area depicting a heavily faulted rift segment, elongated volcanic ridges and smaller circular volcanoes is shown in Figure 12. There are no obvious signs of oblique tectonics in the neighbourhood of the hydrothermal field. The seafloor is at about 250-300 m depth which is quite shallow compared to the deeper vent fields further south along the Mid-Atlantic Ridge. No rock sampling has been carried out within the area, aside from the early manganese dredging mentioned above. The main evidence of the hydrothermal activity is the formation of bubble rich plumes, which have been monitored by high-frequency echo-sounder (German et al. 1994, 1996, Palmer et al. 1995, Ernst et al. 2000). These have detected CH_4, Mn and H_2 anomalies. These bubble rich plumes are indicative of an underlying boiling hydrothermal field, which suggests temperature of 250-260°C (Ernst et al. 2000, Hannington et al. 2005).

Grimsey hydrothermal field

The Grimsey hydrothermal field is located at 66°36.36N and 17°39.24W, about 20 km ENE of the island of Grimsey (Figures 1 and 13). It is located near the northeastern edge of the northerly trending Skjalfandafloi Trough where northerly-trending faults with a westerly down throw dominate. An age estimate of the fault structures suggests that the eastern margin of the trough has been very active in Holocene times (Magnusdottir et al. 2015, Riedel et al. 2001). The hydrothermal vents are aligned in a northerly trend being ca. 1 km long and ca. 350 m wide: the shape of the field indicates its relationship to underlying northerly-trending fault structures. The extent of the bubble plume is more confined as seen in Figure 14. The boiling condition of the vent field seen by the bubble plume at a depth of about 400 m indicates

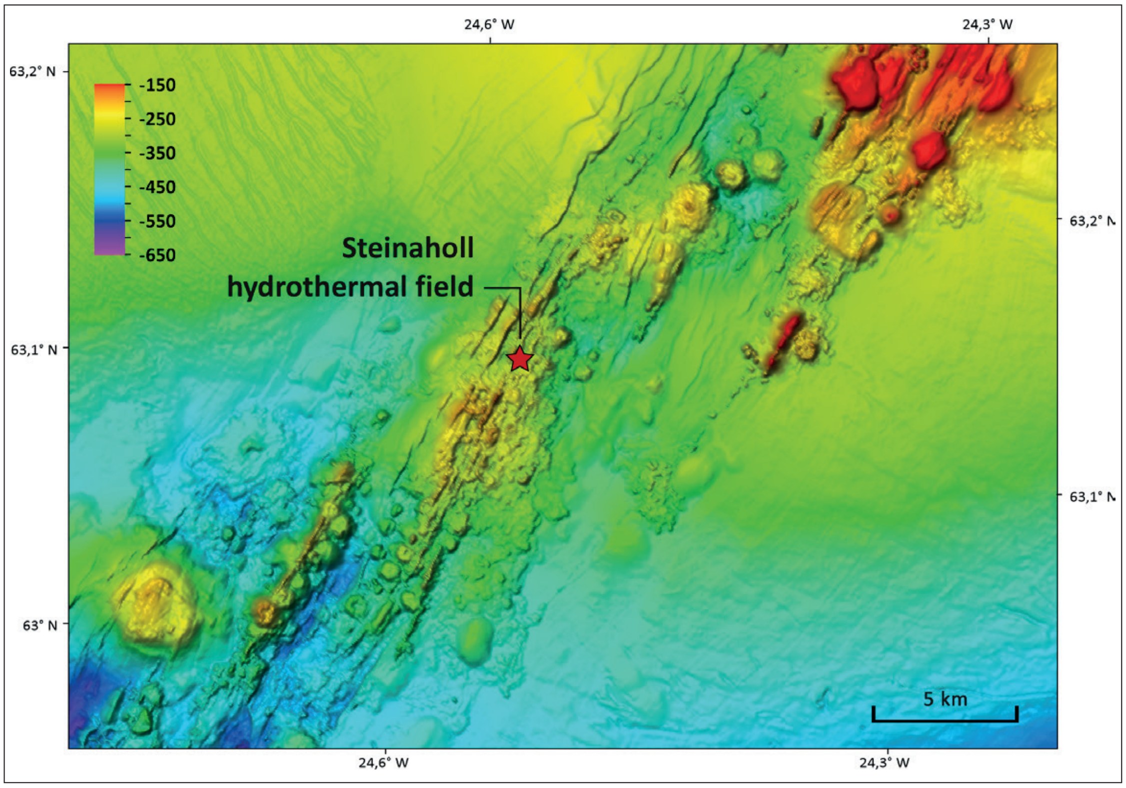


Figure 12. Bathymetry map of the area around the Steinaholl hydrothermal field, Reykjanes Ridge (multibeam data from Marine Research Institute).

temperatures of about 250°C in the uppermost part of the vent field (Hannington et al. 2001). Of the known submarine hydrothermal fields in Icelandic waters, Grimsey shows the largest amount of precipitates. The vent field is divided into three venting areas (Figure 14). The northernmost one shows isolated mounds and solitary anhydrite chimneys, the central one shows coalescing anhydrite mounds and the southern one shows apparently older, though still active mounds (Hannington et al. 2001). The topography of a typical mound is shown in Figure 15. The mineralogy of the precipitates is dominantly anhydrite and talc. Sulphides are relatively insignificant and are not seen to be precipitating in great quantity at the present time. It is, however, not possible to exclude the presence of a sulphide-rich deposit in the sub-seafloor of the hydrothermal field.

Strontium isotopic data along with water chemistry and fluid inclusion studies have shown that the anhydrite precipitation is due to shallow mixing of hydrothermal fluid and seawater (Kuhn et al. 2003).

Kolbeinsey hydrothermal field, Kolbeinsey Ridge

The Kolbeinsey hydrothermal field is located at 67°04.98N and 18°42.96W near the southern end of the Kolbeinsey Ridge as shown in Figures

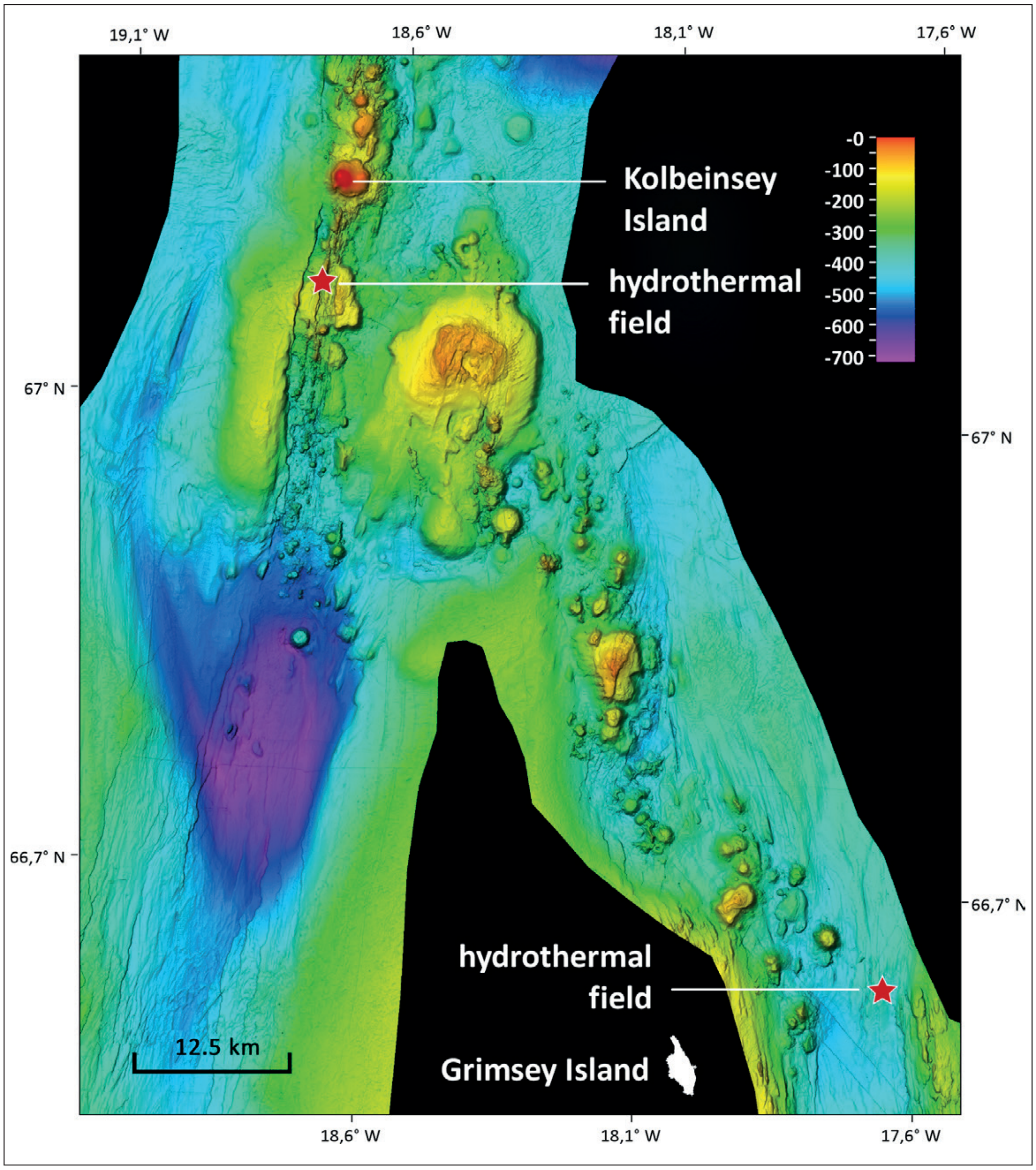
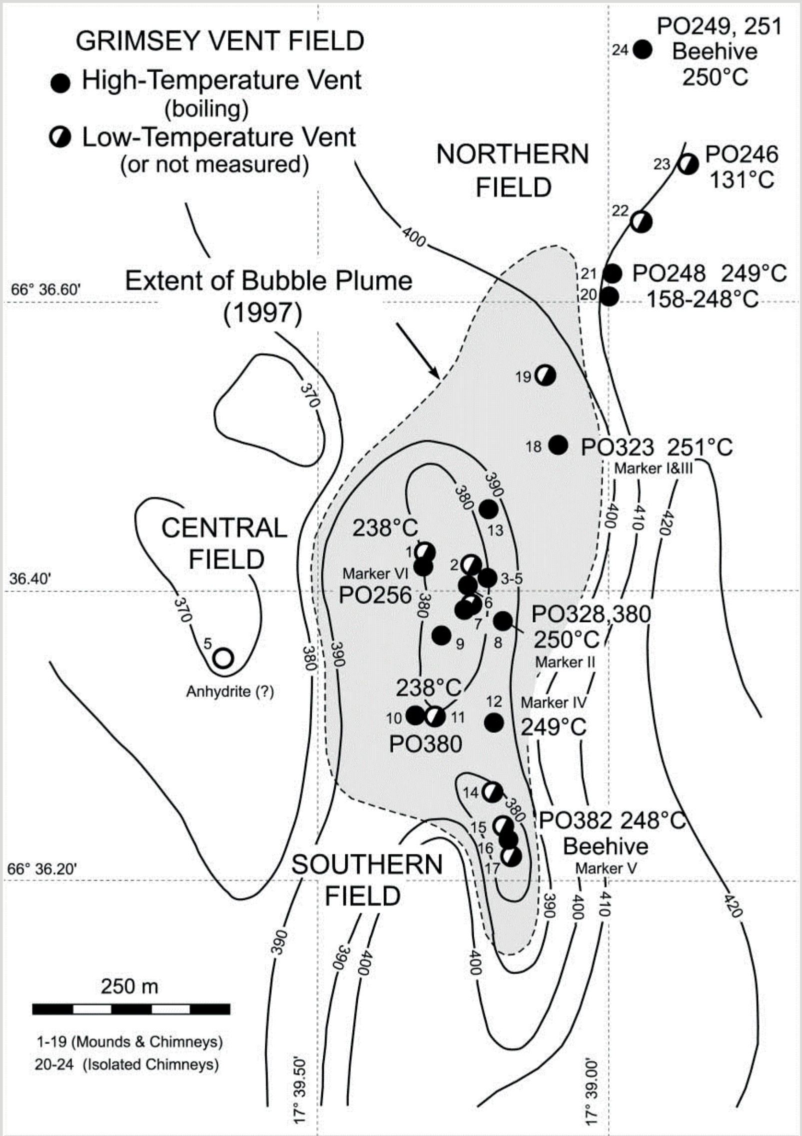


Figure 13. Bathymetry map north of Grimsey Island, with the location of the Grimsey and Kolbeinsey hydrothermal fields (multibeam data from Marine Research Institute, University of Iceland, National Energy Authority and Iceland GeoSurvey).

Figure 14. The Grimsey hydrothermal field showing depth contours, geothermal features and the area of the bubbling plume (Hannington et al. 2001).



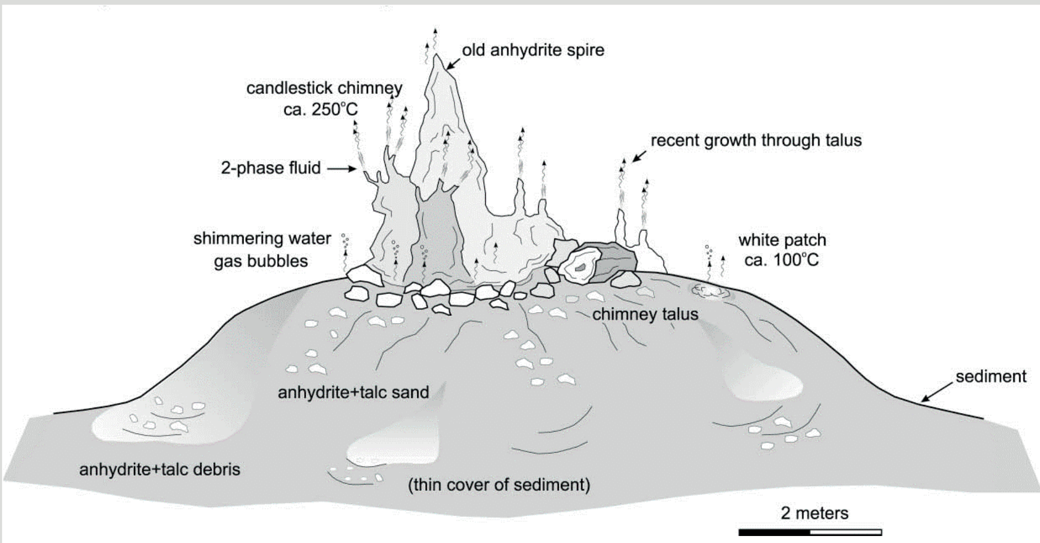


Figure 15. Sketch showing the main characteristics of an anhydrite hydrothermal mound found at Grimsey hydrothermal field (Hannington et al. 2001).

1 and 13. It lies along the western margin of a large, northerly-striking graben structure. The hydrothermal field appears to be confined to the southern part of a small volcanic ridge which rises from 200 m up to 90 m b.s.l. Samples from the field are mostly highly altered, fragmented lava flows, with precipitates of orange-reddish mud and yellow-reddish iron-hydroxides staining the altered basalts and minor chimneys (Lackschewitz et al. 2006). Mineral deposition is limited, which may indicate a young age of the hydrothermal field (or the volcanic formation). The surface temperature of the manifestations is assessed from its boiling condition at 90 m depth which is equivalent to ca. 180°C (Hannington et al. 2001). Higher temperatures can be expected in the underlying system if it follows the boiling point curve. The isotopic composition of the CO_2 and He gas bubbles emitted from the hydrothermal field show a definite mantle fingerprint (Botz et al. 1999).

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