Application of soil measurements and remote sensing for monitoring changes in geothermal surface activity in the Reykjanes field, Iceland

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Faculty of Earth Sciences
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Application of soil measurements and remote sensing for monitoring changes in geothermal surface activity in the Reykjanes field, Iceland

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90 ECTS thesis submitted in partial fulfillment of a Magister Scientiarum degree in Geology

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I hereby declare that this thesis is written by me and is based on my own research. It has not before been submitted in part or in whole for the purpose of obtaining a higher degree.

___________________________________

Auður Agla Óladóttir
Abstract

The objective of this study was to summarize the available data on the surface activity in the Reykjanes geothermal area for the last eight years and reveal the changes which have taken place there during this period and also to evaluate the methods which have been used. Annually since 2004, soil measurements were carried out on soil temperature at 15 cm depth and CO$_2$ emission through soil on a measurement grid with a 25 x 25 m grid spacing except in 2011 when two such datasets were made. Total heat flow, total CO$_2$ flux and uncertainty was calculated for each year’s measurements and the distribution shown on maps. A thermal infrared image, obtained in May 2011 from Reykjanes, snowmelt tracks, done in March 2011 and results from temperature loggers which obtained data during four periods from May 2011 to April 2012 were also used. The soil measurements show that heat flow has increased from $17 \pm 1.4$ MW to $36.1 \pm 2.5$ MW and the CO$_2$ flux has increased from $13.5 \pm 1.7$ tons per year to $36.6 \pm 3.9$ tons per year and the area where surface activity is present has grown, especially to the south and southeast. These changes can mostly be traced to the geothermal power plant in the area even though changes of this order of magnitude are known to be able to take place in Reykjanes without any utilization. The production involves withdrawal of large volumes of geothermal fluid which causes pressure lowering in the system. One of the consequences of the pressure lowering is the formation or increase of a boiling zone in the upper part of the system which can result in more pathways for steam towards the surface, increased heat flow and CO$_2$ emission.

The TIR image from 2011 was calibrated from few soil measurements which were carried out at the time of the flight. The image was compared to soil temperature measurements, snowmelt tracks and a similar TIR image obtained in 2004 from Reykjanes. The comparison revealed a weak relationship with the results of the soil temperature measurements but these two methods showed a similar temperature distribution when visually compared. Soil temperature distribution obtained by snowmelt tracks showed a very similar distribution to the TIR2011 image. The comparison of the TIR2011 image and TIR2004 image shows that soil temperature has risen in most parts of the area and also shows an area north of the Gráa lónið lagoon which has not been included in the soil measurements in recent years where the soil temperature has risen greatly. The average soil temperature from a comparison area has risen from 7.4°C in 2004 to 10.1°C in 2011. This
correlation shows that repeated TIR imagery from the same area can give reliable data to monitor changes in surface temperature.

To evaluate the effects of the grid spacing of the measurements from Reykjanes the datasets from 2009 and 2010 were split up to get two different datasets for each year. For the magnitude of temperature and CO$_2$ flux anomalies observed in the Reykjanes geothermal area, a grid spacing of 30-50 m is not small enough to map the distribution without possible interference from randomly located measurements. Measurements carried out with a 25 m grid spacing seem to give reliable results when visually compared to the TIR image. Measurements carried out with a smaller grid spacing (17-20 m) gives a comparable distribution even though the image is more detailed. To evaluate the total CO$_2$ flux and heat flow the measurements carried out on a 30-50 m grid spacing seem to be able to give statistically good results but that might be doubted. Measurements on a 25 m grid spacing give statistically similar results and a dataset with a smaller grid spacing does not give statistically better results or lower uncertainty.

The results from the continuous temperature measurements show a correlation between soil temperature and precipitation and the correlation is at a maximum for 96 hour precipitation indicating that precipitation lowers the soil temperature for a longer time than previously thought for Reykjanes. Temperature loggers which were located in warm soil (> 40°C) showed sudden temperature drops of tens of degrees and then rises to the previous values in a very short time (hours). These events could not be correlated to any known weather parameter but seem to depend on random micro-scale changes in heat flow.
Ágrip


hækkað úr 7,4°C árið 2004 í 10,1°C árið 2011. Þessi samanburður sýnir að endurteknar hitainnrauðar myndir sem safnað er af sama svæði nýtast vel til að fylgjast með breytingum á jarðvegshitastigi.

Fyrir útbreiðslumynstur jarðvarma og koltvisýringsflæðis á jarðhitasvæðinu á Reykjanesi dugir mælinet með möskvasterð um 30-50 m ekki til að kortleggja útbreiðslu þeirra án þess að tilviljanakennd dreifing mælinga geti haft áhrif. Mælingar sem gerðar eru á mælineti með möskvasterð um 25 m gefa góða mynd af dreifingu jarðhita sem með sjónrænu mat fellur vel saman við hitainnrauða mynd. Með mælingum sem eru gerðar á þéttara mælineti þar sem möskvasterðin er um 17-20 m fæst sambærileg mynd eins og þegar mælingar eru gerðar í 25 m mælineti en þó aðeins nákvæmari mynd af útbreiðslu anomálía. Til þess að meta heildarflæði á koltvisýringi og eða varmaflæði virðast mælingar sem gerðar eru á mælineti með 30-50 m möskvasterð geta gefið tölfraðilega sambærilegar niðurstöður en það er ekki öruggt. Mælingar sem gerðar eru mælineti með 25 m möskvasterð gefa tölfraðilega sambærilegar niðurstöður og þó að mælinet sé þétt niður í 17-20 m möskvasterð gefa þær niðurstöður ekki tölfraðilegra betra mat á heildarflæðinu né lægra óvissumat.

Niðurstöður síritanna sýna greinileg tengsl jarðvegshitastigs og úrkomu og tengslin hámarkast við fjögurra daga úrkomu sem bendir til þess að úrkoma hafi áhrif á jarðvegshita lengur en talið var. Þetta sýnir mikilvægi þess að mælingar af þessu tagi séu gerðar við þurrar og eins stöðugar veðuraðstæður og mögulegt er til að lámarka áhrif þeirra á mælingarnar. Áhrif dægursveiflu sem kemur fram í síritum hesur lítil áhrif á heildarvarmamát. Síritar sem staðsettir voru í heitum jarðvegi (> 40°C) sýndu hegðun þar sem hitastig fell skyndilega um tugi gráða og rauk svo upp aftur á mjög skómmum tíma (< 24 klst.). Þessir atburðir tengjast ekki þekktum veðurþáttum heldur virðast þeir tengjast tilviljanakenndum og mjög staðbundnum breytingum á varmaflæði.
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1 Introduction and geological settings

Geothermal systems are known to form in connection with volcanoes and calderas and it is considered that under most Icelandic high temperature geothermal areas there is a magmatic body that has intruded into brittle crust and permeable rock, and is the heat source for these systems (Pálmason, 2005; Arnórsson et al. 2008). 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 (Arnórsson et al. 2008). In their nature, geothermal systems are transient phenomena and they can persist of thousands of years to hundreds of thousands of years, new systems form and others become extinct when the heat source does not supply energy to the system anymore. During their lifespan the activity of the geothermal systems is variable, often related to intrusive events and the surface activity is known to be even more variable, e.g. changes in the activity of geysers and other surface features are often related to earthquakes.

For about the last thirty years, there has been a growing interest in studying the CO₂ degassing of the Earth. The release of anthropogenic CO₂, which is among the main greenhouse gases, is considered among the most serious environmental problems in the world and for better understanding of the present-day atmospheric C-budget, the role of Earth degassing (volcanic and non-volcanic) in the atmospheric CO₂ budget has to be quantified. It is known that CO₂ can escape from depth through a different pathway and the best known way is through volcanoes. Non-volcanic degassing can occur with the escape of gases from the upper mantle, from carbonate bearing rocks in the crust, from hydrocarbon reservoirs in the sedimentary beds and from surface deposits and surface processes (Mörner and Etiöpe, 2002; Chiodini et al. 2010).

It is known that changes in the behaviour of geothermal systems and its surface activity can occur in relation with utilization of a geothermal system (Pálmason, 2005; Hunt 2001; Giroud and Arnórsson, 2005; Jones, 2006). A study in New Zealand has shown that the exploitation of the Wairakei system significantly increased heat flow through surface (Allis, 1981), which if heat flow is considered as a proxy for CO₂ emissions it could lead to
the conclusion that the exploitation has increased the natural CO\textsubscript{2} emissions (Sheppard and Mroczek, 2004). Other studies have demonstrated that vast quantities of CO\textsubscript{2} are released naturally and in many cases, the natural emissions by far exceed the emissions from geothermal power production (e.g., Seaward and Kerrick, 1996; Delgado et al., 1998; Bertani and Thain, 2002).

To be able to evaluate and quantify changes in geothermal areas due to utilization it is essential to understand natural changes in geothermal systems and then follow closely any changes in its behaviour when the system is utilized. This is possible by monitoring geothermal areas, both natural and utilized and building up knowledge of geothermal systems and how to develop them in a responsible way. Information on gas emissions from natural geothermal and volcanic areas is not only important for estimation of the contribution of CO\textsubscript{2} from volcanic and hydrothermal sources to the global carbon cycle, but also for volcano monitoring, geothermal exploration, delineation of fault and fracture zones. In active and quiescent volcanoes, gas is released not only from craters and fumaroles but also from well-defined areas on the flanks and at the base of volcanoes where CO\textsubscript{2} is the main component of geothermal and volcanic gas (Chiodini et al. 2001; Frondini et al. 2004). Monitoring CO\textsubscript{2} emissions from geothermal areas is one of the fundamental ways of understanding changes in geothermal systems. It has been demonstrated that CO\textsubscript{2} degassing on flanks of volcanoes is sensitive to changes in magmatic activity of the volcano itself, and therefore providing one method for monitoring volcanoes. For geothermal systems, soil diffuse CO\textsubscript{2} degassing has been shown to be a good indicator of the energetic state of the system (Brombach et al. 2001; Chiodini et al., 2001) and monitoring changes of soil CO\textsubscript{2} degassing can therefore lead to better understanding of the behaviour of undisturbed geothermal systems and the response of such systems to geothermal power production.

It is important to evaluate and measure the total natural CO\textsubscript{2} emissions from volcanic and geothermal systems to be able to estimate possible effects of utilizations or changes in the system’s heat source. The scale of the total CO\textsubscript{2} flux value of Icelandic volcanic and geothermal areas is also relevant in the environmental context and all discussions about the global CO\textsubscript{2} budget, renewable and green energy sources and climate change. Two independent estimations have concluded that the total natural CO\textsubscript{2} emissions from Icelandic volcanic and geothermal systems amount to 1 – 2 x 10\textsuperscript{9} kg year\textsuperscript{-1} (Arnórsson and
Gíslason 1994; Óskarsson 1996). Direct measurements of CO₂ emissions have been carried out on 5 of about 40 geothermal areas / volcanic systems in Iceland using different methods. Two subglacial geothermal/volcanic systems, two systems have been measured directly and one has been partly measured. The results of these studies are detailed in section 2.2. To obtain information on gas emission from other geothermal and volcanic systems in Iceland, measurements in each volcanic system are needed, owing to the diversity of the geothermal and volcanic areas.

Heat flow through the Earth’s crust has an average rate of nearly 60 mW m⁻² due to upward convection and conduction of heat from the mantle and the core of the earth (Pollach and Chapman, 1976). Magmatic intrusions increase the normal heat flow locally and are associated with areas of geologically recent volcanic events or geothermal areas (Tester et al. 2005). It has been argued that gas discharge and heat flow from geothermal areas could be correlated because gas species are transported to the surface by steam and advective steam flow is a very efficient heat transport mechanism (Brombach et al. 2001; Chiodini et al. 2001). The studies have demonstrated that CO₂ flux estimated from measured heat flow from geothermal systems agrees reasonably well with measured CO₂ discharge. Accordingly, Arnórsson (1991) argued that measured or estimated heat loss could be used to estimate total steam discharge from particular areas and thereby gas discharge if gas concentration in the steam were known. However, according to Eygerður Margrétardóttir (2005) there is a very weak correlation between gas discharge and heat flow from the Reykjanes geothermal system.

Besides CO₂ emissions, changes in temperature and the quantification of the heat flow are important factors when monitoring geothermal areas, and following changes in their activity. Several attempts have been made to estimate and measure heat flow from geothermal systems in Iceland using volume estimation (Bödvarsson, 1961; Pálmason et al. 1985) or model simulations (e.g. Bödvarsson et al, 1991) but values from these estimations do not agree. Then thermal infrared imaging (TIR) has been used to determine heat flow from a few geothermal systems including Reykjanes (Margrétardóttir, 2005), Torfajökull (Pálmason et al. 1970) and Kverkfjöll (Friedman et. al. 1972). Values obtained by investigations using a combination of TIR imagery and ground measurements have been shown to agree well with values obtained by direct measurement techniques (Harris and Stevensson 1996; Sorey and Colvard, 1994). One of the advantages of using the TIR
method is that it can cover the observation area completely, including inaccessible features, and the survey does not take a long time. It gives an instant overview and shows the temperature variability clearly.

1.1 Objectives

The main objective of this study is to improve methods to monitor environmental changes in geothermal areas, with emphasis on temperature and natural CO₂ emission by comparing available data on surface activity. In this study geostatistical methods that have been used are evaluated and improved by analysing the results of 8 years of annual soil measurements, from 2004 to 2011 on temperature and CO₂ flux in the Reykjanes geothermal area. Furthermore, a thermal infrared image from the same area obtained in 2011 is compared to the soil measurements, data of snowmelt cover and an older TIR image from 2004. The relationship between CO₂ emissions through soil, measured temperature and observed surface temperatures based on TIR imaging is estimated. The other objective of this study is to use these available data to reveal and quantify the changes in CO₂ flux and heat flow in the Reykjanes geothermal area from 2004 to 2011 and put into context with an extensive geothermal production in the area which began in 2006.

1.2 The study area and geological settings

Iceland is located on the Mid Atlantic Ridge, which marks the present divergent boundary of the North American and Eurasian plates. The volcanic rift zone stretches from southwest to northeast of the country and most of the volcanic activity takes place within these areas although volcanoes are also found in off-rift flank zones where little or no spreading occurs (Sigmundsson and Sæmundsson, 2008). The volcanic activity in Iceland is rather unique when compared to other areas above sea level because of its location on the spreading ridge and the effects of the Icelandic mantle plume, as it is more akin to activity below sea level on mid-oceanic ridges.
The Reykjanes peninsula is located on the southwest tip of Iceland. It consists of four volcanic systems which are from east to west; the Hengill, Brennisteinsfjöll, Trölladyngja and Reykjanes systems. The systems have a northeasterly strike and extend across the peninsula. In the Reykjanes Peninsula there have been several basaltic eruptions, the most recent volcanic episodes in the late 12th and the early 13th century, although a volcano, sensu stricto, has not been formed (Sigurgeirsson 1995, 2004).

The study area in this study is the Reykjanes geothermal area in the Reykjanes volcanic system, located on the south western tip of the Reykjanes Peninsula (Figure 1). The extent of the geothermal manifestations in the Reykjanes geothermal system has been estimated to be around 2 km² (Pálmason et al. 1985) and they are closely associated with tectonic fractures, (Björnsson et al. 1971) but most of the current surface activity is concentrated in an area of approximately 0.3 km², sometimes called the Gunnuhver area, which is characterised by extensive normal faulting and high-temperature geothermal activity (Björnsson et al. 1971).

![Figure 1](image)

**Figure 1** A map of Reykjanes Peninsula showing the location of the Reykjanes volcanic system and Reykjanes geothermal area.

The Reykjanes geothermal area has been known for a long time for its variable and abruptly changing surface activity. Investigations and descriptions go back to the middle of
the 19\textsuperscript{th} century and this is now among the most studied geothermal areas in Iceland (Fridriksson et al, 2010). Systematic investigations began there between 1960 and 1970, 30 geothermal wells have been drilled in Reykjanes, the deepest one being 3,082 m (www.hsorka.is) and now there is a 100 MW power plant that started electrical energy production in May 2006 using 12 production wells.

The study area and its surroundings are mostly covered with Holocene lavas and tuff formations. The landscape is dominated by recent lava flows and volcanic crater rows but a few Pleistocene hyaloclastite ridges stand through the younger Holocene lavas. The lavas are between 10,000 and 2,000 year old except for the youngest volcanic formation at Reykjaness, the 4.5 km long Yngri-Stampar cone row, along with the lavas and the tuff it produced. These were formed in the so-called Reykjanes Fires, a major volcanotectonic episode within the Reykjanes system between 1210 and 1240 (Franzson, 2004). Because of the high permeability of the bedrock, proximity to the ocean and low topographic relief the geothermal system is recharged by seawater.

**Figure 2** Overview of the study area at Reykjanes. Roads and tectonic features are shown. A green line marks the area where surface measurements were carried out in 2011. Red symbols show wells, well RN-13 marked due to weather station.
In the Gunnuhver area, where most of the current surface activity takes place, surface manifestations include steam heated mud pools, steam vents, fractures and warm ground (Fridriksson et al. 2006). Seawater geysers and boiling springs were active on and off from 1906 to 1980 (Friðriksson et al. 2010) but there are no boiling springs currently active in the area. Steam vents and steam heated mud pools are the dominating features in the two most active parts of the Gunnuhver area, in the southeastern and northwestern ends (Figure 2). The most intensive surface activity is in the southeastern end, characterized by intense steam vent activity and steam heated mud pools while the northern end, that lies on the flats close to the brine pond (Gráa lónið lagoon), is characterised by mud pools but much fewer steam vents. Between these two areas and also to the north and to the south, the ground is warm in large patches (Fridriksson et al., 2006). Where the geothermal activity is intense the areas are mostly unvegetated and the soil consists of wet geothermally altered clay but where the ground is not very much affected by the geothermal activity, the characteristic vegetation consists of green moss (Hypnum jutlandicum) and creeping thyme (Thymus praecox arcticus) (Elmarsdóttir et al., 2003). Geothermal signs are also present outside the study area, mostly warm and moist air rising through small fissures e.g. south of the study area towards Skálafell but on the northeastern side of the Gráa lónið lagoon, around Rauðhólar, the soil is geothermally altered and also on the north side of the Gráa lónið lagoon.

*Figure 3* Overview of the Reykjanes geothermal area (picture from June 2011). The bright and red soil is geothermal clay and mud. The Gráa lónið lagoon is seen to the left. Buildings belonging to the Reykjanes power plant are seen and most of the steam (in this photograph) comes from the separator towers.
2 Literature Review

2.1 Studies of natural degassing from geothermal areas

Increased emphasis on the global C budget and climate effects has led to studies of volcanic and non-volcanic CO₂ emissions. The lower limit of global CO₂ discharge from subaerial volcanism has been estimated at ∼300 Mt yr⁻¹ (Mörner and Etiope, 2002). However, this estimate of CO₂ emissions did not include emissions from volcanic lakes and in order to improve this information, Péres et al. (2011) performed a study on CO₂ emissions from volcanic lakes worldwide and concluded that the total CO₂ emissions were 117±19 Mt yr⁻¹. It has become clear, since the work of Irwin and Barnes (1980) that there is a close relationship between active tectonic areas and anomalous crustal emissions of carbon dioxide. The high crustal permeability faults act as pathways for the upward migration and eventual release of deep gases into the atmosphere.

In general, there is a constant flux of microbially and plant-respired CO₂ through soil to the atmosphere but here it is considered as a background flux (Bond-Lamberty and Thomson, 2009). The soil respiration rate varies greatly according to soil vegetation type but the values go from <1 g/m²/day in tundras to 1-10 g/m²/day in moorland and up to 12-20 g/m²/day in tropical moist forests (Raich and Schlesinger, 1992; Sotta et al. 2004; Margrétardóttir, 2005). Elevated levels of CO₂ through soil, higher than previously mentioned values have been used in geothermal exploration as an indicator of geothermal activity and can be useful for the purpose of delineating fractures or other structures that direct flow of fluids in the geothermal reservoir.

Numerous studies have been focused on the CO₂ soil diffuse degassing from quiescent volcanic/geothermal areas (e.g. Brombach et al., 2001; Chiodini et al., 1998, 2001a; Hernández et al., 1998; Gerlach et al., 2001; Salazar et al., 2001) and most of them show that gas is not released uniformly from the whole volcanic area, but rather from relatively restricted regions. Such areas have been called diffuse degassing structures by Chiodini et al. (2001b). Studies also suggest that significant amounts of CO₂ are released to the
atmosphere by quiescent degassing of volcanoes and soil diffuse degassing from geothermal systems compared to the CO$_2$ released from fumaroles (e.g. Baubron et al., 1990; Kerrick, 2001; Salazar et al., 2001; Inguaggiato et al. 2011) and some studies have stated that diffuse emissions can be of similar magnitude as from fumaroles and craters e.g. in Etna (Allard et al., 1991) and in Fossa cone (Chiodini et al., 1996) on the magnitude of tens to hundreds of tons per day (e.g. Allard et al., 1991, Chiodini et al., 1996). It has also been shown that geochemical signals of volcanic unrest have been clearly identified before, during and after the effusive activity, e.g. at Stromboli (Inguaggiato et al., 2011).

A recent Icelandic study, of soil diffuse degassing, gas discharge through steam vents, and steam heated pools showed that most of the CO$_2$, by far (97.4%), is emitted through soil diffuse degassing (Fridriksson et al., 2006). Although volcanic CO$_2$ degassing can also take place from cold areas (Rogie et al., 2001; Sorey et al., 1998; Hernández et al., 2003), many areas of anomalously high diffuse CO$_2$ flux are characterised by elevated soil temperature (Notsu et al. 2005; Frondini et al. 2004; Chiodini et al. 2001, 1996; Brombach et al. 2001; Cardellini et al. 2003; Granieri et al. 2010) indicating that the CO$_2$ flux correlates to the condensation of the water vapour in the soil. When hot hydrothermal fluids rise towards the surface, the steam condenses mostly near the surface and releases thermal energy and CO$_2$ is diffusively degassed through soil, suggesting that elevated soil temperature is a result of convective transport of water vapour rather than thermal conduction (Frondini et al. 2004). The energy associated with the diffuse degassing process supplies an important contribution to the energy budget of geothermal and volcanic system in a quiescent state (Brombach et al. 2001; Favara et al., 2001; Chiodini et al., 2005) and therefore, monitoring of CO$_2$ diffuse degassing may play a significant role in the observation strategy.

2.2 Direct observations on CO$_2$ emissions from geothermal systems in Iceland

In general, volcanic and geothermal systems can be considered as geochemical reservoirs of CO$_2$. In many places, metamorphic decarbonation of marine limestone and
decomposition of organic sediment can be an important source of CO$_2$ in geothermal systems but carbon isotope ratios of CO$_2$ in Icelandic geothermal fluids indicate that degassing of mantle-derived basaltic magma is the dominant source of CO$_2$ in these systems (Ármannsson 1998; Ármannsson et al., 2005). About 35-40 volcanic / geothermal systems are defined in Iceland but direct measurements of total CO$_2$ discharge are available from five of these systems. Ágústsdóttir and Brantley (1994) studied gases including CO$_2$ from the Grímsvötn subglacial geothermal system in Vatnajökull ice cap, one of the most active volcanic systems in Iceland with at least six eruptions in the 20$^{th}$ century. Later Gíslason (2000) estimated the amount of CO$_2$ emissions from the subglacial caldera of Eyjafjallajökull during the period from 1993-2000. In both cases, the CO$_2$ release occurs in subglacial calderas where the CO$_2$ is dissolved in glacial meltwater and the release can be determined by analysing the flow rate and the total carbonate content of the meltwater.

Ágústsdóttir and Brantley (1994) concluded the average flux of CO$_2$ from the Grímsvötn volcanic system was $1.9\times10^8$ kg year$^{-1}$ between 1954 and 1991. The observed CO$_2$ flux from Eyjafjallajökull, a much less active system with three known eruptions in historical time, is between $2.6\times10^6$ and $2.6\times10^7$ kg year$^{-1}$ during the quiescent period from 1993 to 2000 (Gíslason, 2000). In 2004 CO$_2$ emissions and heat flow through soil, steam vents and fractures, and steam heated mud pools were determined in the Reykjanes geothermal system. CO$_2$ through soil was measured by soil flux equipment, heat flow from steam vents and fractures was determined by quantifying the amount of steam emitted from the vents by direct measurements of steam flow rate and the heat loss from the steam heated mud pools was determined by quantifying the rate of heat loss from the pools by evaporation, convection and radiation (Fridriksson et al., 2006). They stated that $5.1\times10^6$ kg year$^{-1}$ of CO$_2$ were emitted from the Reykjanes system, with more than 97% released through soil. Ármannsson et al. (2007) determined the natural CO$_2$ emissions from the Krafla geothermal system of the order of $8.4\times10^7$ kg year$^{-1}$. In addition to these studies, measurements of partial CO$_2$ discharge the Hekla volcanic system have been carried out. About $7\times10^7$ kg/year is estimated to be released from the Hekla magma chamber into the overlying groundwater system (Gíslason et al., 1992).
2.3 Estimates of total CO\textsubscript{2} emissions from Icelandic geothermal systems

Estimations of total CO\textsubscript{2} emissions from natural geothermal and volcanic systems in Iceland have been done but they have differed by an order of magnitude. Ármannsson (1991) estimated that total CO\textsubscript{2} via steam vents in Icelandic geothermal systems, assuming that all CO\textsubscript{2} is emitted from steam vents, was $1.5 \times 10^8$ kg/year and this value was obtained by using measurements from active geothermal manifestations in Krafla to extrapolate for other geothermal areas in Iceland and using observed CO\textsubscript{2} concentrations in steam from individual systems. Arnórsson (1991), and later Arnórsson and Gíslason (1994) used estimated heat flux from Pálmason et al. (1985) to reach an estimate between $1.0 \times 10^9$ and $2.1 \times 10^9$ kg/year of emitted CO\textsubscript{2} assuming that natural heat loss in those areas is mainly due to convective flow of the steam. By combining the results of analyses of tectonic modelling and fluid inclusion, Óskarsson (1996) estimated a total CO\textsubscript{2} flux of $2.2 \times 10^9$ kg/year for Iceland. The disagreement between the estimate of Ármannsson (1991) and then Arnórsson and Gíslason (1994) and Óskarsson (1996) can be explained by the fact Ármannsson (1991) considered emissions from steam vents only in his estimate while the others include diffuse emissions through soil and by bubbling from surface water bodies.

It is worth noting that the order of magnitude varies between the five measured geothermal / volcanic system in Iceland which are very different in size and activity so it is not appropriate to infer the total CO\textsubscript{2} on the basis of these five studies only. Nevertheless, the sum of the measured and/or estimated natural CO\textsubscript{2} flux from the Grímsvötn, Eyjafjallajökull, Reykjanes, Krafla and Hekla systems is about $3.5 \times 10^8$ kg year\textsuperscript{-1} implying that the estimate of total natural CO\textsubscript{2} emissions from Icelandic geothermal systems is of the right order of magnitude. These five systems are about 12.5% to 15% of the total 35-40 volcanic/geothermal systems in Iceland and their sum gives about 15-35% of the estimated total CO\textsubscript{2} in Iceland.
2.4 CO$_2$ flux measurements and geostatistical methods

During the past two decades, many studies have been focussed on measuring CO$_2$ flux through soil, map its areal distribution and estimate the total CO$_2$ emission from volcanic and geothermal areas and in these studies a wide range of measurement technologies and statistical methodologies have been applied to accomplish these goals (Farrar et al. 1995; Giammanco et al. 1997; Chiodini et al. 1998; Gerlach et al. 2001, Salazar et al. 2001; Chiodini et al. 2001, 2010; Cardellini et al. 2003; Frondini et al. 2004; Granieri et al. 2010). The choice of methodology is an important factor and may affect the CO$_2$ flux measurements and characterization of their natural spatial variability, and the total CO$_2$ emission rate estimated for a given area (Lewicki et al. 2005). The reasons for the variability in the results of CO$_2$ flux measurements are different methodology and the natural variability of surface and subsurface parameters that can influence the gas flow. These surface and subsurface parameters are biological respiration, meteorological parameters such as atmospheric pressure, temperature and wind speed, physical properties of the medium (e.g., permeability, porosity) and the deep CO$_2$ source.

In order to measure CO$_2$ flux, an accumulation chamber method has become a routine studying and monitoring tool at many volcanic and geothermal sites for the last 15 years (e.g. Chiodini et al., 1998; Evans et al., 2001; Lewicki et al., 2005; Giammanco et al., 2010; Mazot et al., 2011). The use of the accumulation chamber with a gas analyser provides a simple and rapid measurement which is based on the rate of CO$_2$ increase inside the chamber. This is an absolute method that does not require corrections linked to the physical characteristics of the soil (Giammanco et al. 2007). The limitations of this method include the measurement’s small spatial scale and lack of ability to monitor continuously over longer period. CO$_2$ flux through soil has a large variability, even on a small spatial scale, variation which is commonly associated to the permeability of the soil and the micro-fracturing system of the soil that occurs even in apparently homogeneous ground (Chiodini et al., 1998; Granieri et al., 2010). The eddy covariance / EC method, a micrometeorological technique traditionally used to measure CO$_2$ fluxes across the interface between the atmosphere and a plant canopy (e.g. Baldocchi, 2003) has been proposed as feasible technique to monitor volcanic CO$_2$ and heat fluxes in conjunction
with the chamber method (Anderson and Farrar, 2001; Werner et al., 2006). The benefits of this method are that it does not interfere with the ground surface and is averaged over both time and space, with a much larger spatial scale (m^2-km^2). The underlying theory assumes the homogeneity of surface fluxes, flat terrain and temporal stationary, conditions which are normally not characteristic of geothermal and volcanic environments. However, in Lewicki et al. (2008) described an observation in which the accumulation chamber method and EC (Eddy Covariance) were compared and concluded that the results indicated that despite complexities at their study area, EC can be reliably used to monitor background variations in volcanic CO\textsubscript{2} fluxes associated with meteorological forcing. Other comparative measurements of CO\textsubscript{2} fluxes in volcanic areas have not been presented in the literature, although laboratory tests on controlled CO\textsubscript{2} fluxes using multiple measurement techniques have been described (Evans et al. 2001).

CO\textsubscript{2} flux measurements are often used to map its areal distribution and such maps may be produced using a variety of methods. Due to circumstances in volcanoes and geothermal areas, CO\textsubscript{2} flux measurements are often made at widely and/or unevenly spaced intervals within the measurement area. In these cases, geostatistical methods must be used to interpolate for CO\textsubscript{2} flux at unmeasured locations. This has commonly been accomplished by a kriging algorithm which is focused on providing the best fit in the minimized least square sense, hence unique, without considering the resulting spatial statistics of all the estimates taken together and producing a set of estimated values whose variogram, a tool that quantifies spatial correlation, does not match with the original dataset (Salazar et al., 2001; Gerlach et al., 2001; Cardellini et al. 2003). Other limitations of the kriging algorithm is that it is incapable of detecting spatial uncertainty (Delbari et al. 2009) and it smooths out the extreme values of the dataset, large values are underestimated and small values are overestimated and might therefore hide some important spots (Cardellini, 2003).

More recently a stochastic simulation algorithm has been used to process gas flux measurements and other measurements in soil science, in which the spatial variability of the measured attributes has to be preserved (Goovaerts, 2001). The simulations are usually performed by using the sequential Gaussian simulation algorithm (sGs) (Cardellini et al., 2003; Fridriksson et al., 2006; Mazot et al., 2011). The sequential Gaussian simulation is a method used to interpolate or fill in the areas between measuring nodes and is a suitable tool to model soil diffuse degassing (Frondini et al. 2004). The sGs method uses the dataset
to generate a great number (chosen by the user) of equiprobable representations or realizations of the spatial distribution of the CO₂ flux. It operates using a sampled attribute (e.g. CO₂ flux) and the variable is simulated at each unsampled location by random sampling of a Gaussian conditional cumulative distribution defined on the basis of the original data. The sGs process needs a multigaussian distribution and therefore CO₂ fluxes have to be transformed into a normal distribution, which then is used in the simulation process. Simple kriging estimate and variance, computed according to the variogram model of normal scores are used to define a Gaussian conditional cumulative distribution function at each location. After the simulation itself, the simulated normal scores are transformed back into values expressed in the original dataset unit applying the inverse of the normal score transform. Using the same sample set again, a requested number of equiprobable representations or realizations is generated and then used to draw a map representing the average of the requested number of simulations. The advantage of using this method is that it results in more realistic values for uncertainties in the total flux and preserves certain values of the dataset, including averages (Cardellini et al., 2003). It also allows one to evaluate the spatial uncertainty through generation of several equally probable stochastic realizations (Delbari et al. 2009).

When compared to other computational applications in geosciences, sequential simulation is not extremely computationally intensive (Nunes and Almeida, 2010). Lewicki et al. (2005) compared six geostatistical methods for processing soil degassing data (arithmetic and minimum variance unbiased estimator means of uninterpolated data, arithmetic means of data interpolated by the multiquadric radial basis function, ordinary kriging, multi-Gaussian kriging and sGs) and they concluded that sGs yields the most realistic representation of the spatial distribution of CO₂ flux. They also suggested that if similar measurement instrumentation and protocol are used, grid measurements of diffuse CO₂ degassing can be used as a tool to monitor volcanic emissions with relatively low uncertainty (Lewicki et al., 2005). Teixeira et al. (2011) reached the same conclusion and stated that their data presented a best representation when estimated by sGs. According to Goovaerts (2001), the difference among all simulated maps can be used to calculate the uncertainty of the flux estimation.
2.5 Thermal infrared imaging and natural heat flow from the soil

All materials at temperatures over absolute zero (0 K) emit energy. This energy is in the form of electromagnetic waves and is usually classified by wavelength intervals where infrared wavelength from the edge of visible red light at 0.74 µm extends conventionally to 300 µm. Energy of electromagnetic waves is, according to Planck’s law, inversely proportional to its wavelength, i.e. the longer the wavelength, the lower the energy (Lillesand and Kiefer, 2000). In thermal remote sensing, radiations emitted by ground objects are measured for temperature estimation. The temperature observed by thermal remote sensing then reproduces only the surface temperature of the objects. The measurements give the radiant temperature of a body which depends on two factors; kinetic temperature and emissivity. The emissivity is the emitting ability of a real material compared to that of a black body, a theoretical object that absorbs and then emits all incident energy of all wavelengths. The emissivity is a spectral property that varies with composition of material and geometric configuration of the surface. The atmosphere between the surface and the TIR scanner do have some effects on the imagery. Atmospheric absorption and scattering tend to increase the energy that reaches the TIR scanner, making the ground appear colder than it is. But atmospheric emission may add to the radiation sensed, making the objects appear warmer than they are. These temperature effects depend on atmospheric conditions during imaging but they can also cancel out each other to some degree (Lillesand and Kiefer, 2000).

Theoretical or empirical atmospheric models can be applied to thermal scanner calibration in order to minimize atmospheric effects but due to the complexity of such models, these effects are generally eliminated by correlating the scanner data with surface measurements. These reference measurements have to be done at the same time as the TIR flight in points where temperature is assumed to be constant over larger areas, making water surfaces preferable (Lillesand and Kiefer, 2000) and it is also important to be able to identify the reference point on the images (Árnason, 1997). A calibration curve is constructed relating the scanned output value to the corresponding ground surface temperature and this calibration relationship is used to estimate the temperature at all points in the TIR image (Lillesand and Kiefer, 2000).
According to Dawson (1964) the natural heat discharge from geothermal areas appears as a heat flow through soil, heat loss from water surfaces, heat loss through fumaroles, through overflow from geysers and springs and seepage to lakes and rivers. This happens through three main heat transfer mechanisms; advection, conduction and radiation. According to Sorey and Colvard (1994) the dominant mode of heat loss differs between geothermal areas due to different surface characteristics, such as manifestations, alteration and ground cover. Thermal infrared data only show the heat flow through radiation. This is therefore not a complete method to map the total heat flow from an area, but maps the extent of surface thermal anomalies and gives an overview of the distribution of the heat flow. Thermal infrared images (TIR) images also represent a suitable tool to monitor changes in surface activity of geothermal systems with time (e.g. Chiodini et al. 2007).

As far back as the 1960’s, remote sensing has been used for geothermal mapping. With the technology to acquire TIR images the scope of application of remote sensing expanded by allowing thermal conditions of ground surfaces to be sensed remotely (Sabins 1997). The first geothermal area which was studied with TIR technology was the Wairakei geothermal area in New Zealand where heat flow from the thermal area was studied (Dawson and Dickinson, 1970). TIR data was used to map the extent of thermal features, and field measurements were used to obtain temperature data. The first TIR data from Iceland were obtained in six geothermal areas (including Reykjanes geothermal field) in 1966 and again in 1968 and 1973. This was an experiment carried out by the US Geological Survey, the United States Army, University of Michigan and National Energy Authority of Iceland (Pálsson et al., 1970; Friedman et al. 1972). The purpose of these three surveys was to evaluate the infrared technique as an exploration tool for mapping thermal anomalies. At this time the images were all collected on film sensitive to the 4.0-5.5 μm wavelength bands and the data collected in these surveys still exist but they are difficult to use quantitatively. This is because these images are difficult to georeference, since information about flight lines are missing and surface features (such as roads and buildings) cannot be identified. For TIR images it is necessary to quantify them with field temperature measurements and for these surveys in Reykjanes, ground measurements were carried out at the same time as these TIR images were collected. Pálsson et al. (1970) state that when weaker anomalies in the Reykjanes imagery have been filtered out away, a very clear
picture of the stronger anomalies emerges showing clearly the distribution of heat anomalies.

In the early 1990’s the department of engineering in the University of Iceland experimented with an airborne TIR sensor. Árnason et al. (1994) reported on the development of the instrument and its application over populated areas, over sea and over geothermal areas. From 1990, many experiments were carried out using TIR imaging for mapping some geothermal areas, among them the Reykjanes geothermal field (Árnason, 1994, 1995, 1996, 1997). In 2004, TIR images were obtained from the Reykjanes geothermal field and compared with ground measurements on 25x25 m grid in the MS thesis of Margrétardóttir (2005). She estimated heat flux and surface temperature based on soil temperature measurements and the relationship between surface temperature and CO₂ emissions was used to estimate the total CO₂ emissions from her study area. This resulted in emissions only 3.3% lower than value determined from direct measurements of CO₂ flux through soil. In her work, she states that the CO₂ flux as a function of surface temperature appears to have bimodal distribution; most of the area having surface temperatures lower than 5°C, according to the image at the time of the TIR survey, with low CO₂ flux values, i.e. with flux less than 10 g m⁻² d⁻¹. At higher surface temperatures, ranging from 5° to 79°C the CO₂ fluxes generally range from 10 to 1000 g m⁻² d⁻¹ without much apparent correlation between CO₂ flux and surface temperature. Margrétardóttir (2005) concludes that TIR images are useful in gas and heat flow estimations but says it is clear that geological information and ground measurements for each area will always be needed. Owing to interaction between cold groundwater and steam removing a large proportion of the geothermal heat before it reaches the surface at Reykjanes, it would be likely that the Reykjanes geothermal area is not representative for the correlation between surface temperature and CO₂ emissions in other geothermal areas in Iceland or worldwide (Margrétardóttir, 2005).
2.6 Methods for estimating heat flow through soil

Dawson (1964) developed an empirical method to determine the total heat flux through the ground surface, in which water vapour convection is the main transport mechanism. Dawson (1964) assumed that only a small percentage of the heat discharged to the atmosphere from soil was in the form of radiant heat. Furthermore, he assumed that a large proportion of the radiant heat was absorbed by the vapour and thus detected by the calorimeter if it were placed firmly on the soil surface. The method which was calibrated by direct measurements at the Wairakei thermal field, New Zealand, is based on the correlation between measured soil temperature at 15 cm depth in °C, and heat flow measurements using a portable calorimeter ($Q_s$ in W/m$^2$). The relationship was resolved, ending with the equation

$$Q_s = 5.19 \times 10^{-6} T_{15}^4,$$

Equation 1

which applies when $T_{15}$ (temperature at 15 cm depth in °C) is lower than 97°C. For higher values than 97°C at 15 cm depth, the depth to the point where the soil temperature reaches 97°C (in cm) allows the estimation of the heat flow through soil using:

$$Q_D = 10^{(\log d_{97} - 3.557) - 0.894)},$$

Equation 2

where $d_{97}$ is the depth in cm where the soil reaches 97°C temperature. Guðmundsdóttir (1988) measured soil temperature and heat flow with a calorimeter at Nesjavellir and compared her results to those of Dawson (1964). She showed that the relationship between $d_{97}$ and heat flow at Wairakei, determined by Dawson applied reasonably well to her measurements at Nesjavellir although the heat flux at the same depth of 97°C soil temperature was slightly higher at Nesjavellir.
3 Data and methods

In this study, several large datasets were used to explore the surface activity changes that have taken place in the Reykjanes geothermal area over the last 8 years. Annual measurements on soil diffuse CO$_2$ flux and soil temperature have been made in this area since 2004 and the objective of these measurements is to evaluate and quantify changes in the surface activity. The soil diffuse CO$_2$ flux has been measured using a close-chamber method and a digital thermometer has been used to measure the soil temperature. These measurements have been processed using statistical methods (sequential Gaussian simulations) and the results displayed on maps. Then a thermal infrared (TIR) image was obtained in 2011 for evaluating surface temperature but another TIR image from the area obtained in 2004 was also used here for comparison. These methods and data are described thoroughly in sections 3.1 and 3.2. In 2011-2012, continuous soil temperature measurements were done, using loggers that measure temperature and can be placed in soil and measure the temperatures for longer periods (weeks). Then in March 2011, mapping of the edges of the snow cover in the geothermal area was done using handheld gps unit and digital thermometer and these data were used to compare with the TIR image.

3.1 Thermal infrared image of the tip of the Reykjanes peninsula

A TIR image of the tip of the Reykjanes peninsula was collected May 25$^{th}$ 2011 in good weather conditions to obtain an image of the precise temperature distribution. Four soil temperature data loggers were placed in and around the Reykjanes geothermal field as groundwork for the TIR image collection on May 5$^{th}$ (and placed there until June 10$^{th}$, period I). These data loggers are HOBO® U20 Loggers from Onset Computer Corporation with operation range from -20°C to 80°C, accuracy of 0.37°C at 20°C and response time of 3.5 minutes (Hoboware, 2009). The purpose was to collect a time series of soil temperature measurements near the soil surface for imagery calibration purposes. The loggers were located in cold water, warm water, cold ground, and warm ground. The loggers were placed as close to the surface as possible, which resulted in 5 to 15 cm below the surface and attempting to use locations with as homogeneous temperature as possible, like water
bodies. Figure 6 shows the locations of the data loggers during this period. The loggers were set up to log both temperature and pressure at 5 minutes intervals.

On May 25th 2011, just after sunset, the TIR image collection was carried out by Kolbeinn Árnason at the University of Iceland from an aircraft specially customised for aerial photography. The images were collected using the TIR scanner owned by the Engineering Research Institute at the University of Iceland (described in Margrétardóttir, 2005, section A.3.1). The TIR imagery was collected in 10 overlapping strips that covered the 4 km wide tip of Reykjanes peninsula. The time of day was chosen in order to minimize the effects of reflected sunlight in the TIR data. Figure 5 shows the area of the data collection. The objective was to collect data with around 1 m resolution, which required the flight altitude to be around 1000 feet but due to overlapping of the scanned lines the effective resolution of the image is higher, each pixel covering 0.40 m². The area was covered in 10 flight lines from the north-east to the south-west. During the survey the sky was almost clear, wind was calm and the air temperature was about 6°C at sea level.

Ground temperature measurements were carried out in the evening of May 25th while the TIR data were collected. The purpose was to collect more ground data to support the calibration of the TIR data. About 100 measurements were carried out with handheld digital thermometers, scattered widely around the geothermal area located both in water and homogeneous soil spots. The temperature was measured as close to surface as possible, the lowest observed temperature was 5.4°C and the highest 99.7°C. Figure 5 shows the locations of temperature measurements.

3.2 Ground measurements and methods

3.2.1 Soil temperature and CO₂ flux ground measurements

Since 2004, annual measurements of soil temperature and CO₂ flux through soil in the Gunnuhver area have been carried out. The soil temperature measurements were performed with a handheld digital thermometer which was put 15 cm into the soil and held there until it gave a constant temperature value. Surface CO₂ flux was measured using a WEST Systems fluxmeter (West Systems, 2003) based on a close chamber method with
repeatability of ± 10% (Chiodini et al., 2003). The CO₂ flux was measured directly using a closed-chamber CO₂ flux meter from West Systems equipped with a LICOR LI-820 single-path, dual wavelength, non-dispersive infrared gas analyser. The flux measurement is based on the rate of CO₂ increase in ppm sec⁻¹ inside a 3.06 x 10⁻³ m³ chamber. The increase in ppm sec⁻¹ was calculated to moles m⁻² day⁻¹ and the measurements were converted to g m⁻² day⁻¹ by multiplying the value by the mole weight of CO₂ (44 g mole⁻¹).

These measurements have been carried out systematically every summer on a grid that covers the part of the area that shows the most significant surface activity. The size of the measured area has changed from year to year due to changes in the distribution of the surface activity but has covered on average about 0.3 km² and annual number of measurement nodes has been around 400-500 with grid spacing about 25 x 25 m. The data has been used to map the soil temperature distribution using the Surfer software and the CO₂ flux using geostatistical methods (sequential Gaussian simulations). The whole dataset has been used to evaluate changes from year to year (Fridriksson et al., 2010, Óladóttir et al., 2010, Óladóttir and Snæbjörnsdóttir, 2011).

### 3.2.2 Sequential Gaussian simulations and geostatistics

To process the CO₂ measurement data and evaluate the total CO₂ emission from the measured area, the software WinGslib (from Statios) has been used. For each year (each dataset), 100 realizations have been calculated for the CO₂ flux on a model grid with a 2 by 2 m grid spacing using the sGs algorithm of the sgsim code by Deutsch and Journel (1998) and the average value for each cell presented on maps.

Similar procedures have been used to study the datasets and same software has been used since 2004. First, a normal score transformation of the dataset is made and thereafter, the experimental values are used to make a semi-variogram. The semi-variogram and its parameters are being described here. It shows how CO₂ flux changes with distance between two measurement nodes, concluding that two measurements lying closely together are more similar than two measurements with great distance. Where the distance is so great that the correlation of two measuring values is completely random, the variogram has reached a certain constant value, referred to as sill (c). The h value (on x-axis of the variogram) at which the semi-variogram reaches the sill is called range (a). Theoretically, all variables separated by distances larger than the range are therefore uncorrelated.
The vertical jump from the value of zero at the origin of the variogram small separation distance is the nugget parameter \( (c_o) \). The \( h \) tends towards zero but the semi-variogram does not tend towards zero and this discontinuity is known in geostatistics as the nugget effect (Figure 4).

\[ \gamma(h) \]

**Figure 4** Classical shape of a semivariogram showing the model parameters, the nugget, the sill and the range.

The nugget effect can be attributed to spatial sources of variations at distances smaller than the sampling interval. Natural phenomena can vary spatially over a range of scales and will appear as part of the nugget effect. A variogram model of the normal scores that fits the experimental data is made, using the nugget, sill and range value that best fits each dataset. The four most commonly used semi-variogram models are the spherical, the linear, the exponential and the Gaussian semi-variogram models and the one that fits the experimental data best is chosen (Armstrong, 1998). Here, spherical and exponential semi-variogram models are both used, the best fit chosen for each dataset.

For a given dataset the most common method of estimating the semi-variogram is called the classical or method-of-moment estimator of the semi-variogram and according to Morgan (2006) it is defined as follows, given \( n \) measurements of a spatial variable:
Equation 3

\[ \gamma (h) = \frac{1}{2N(h)} \sum_{(x',x) \in N(h)} (z(x') - z(x))^2 \]

where

\[ N(h) = \{(x',x) : x' - x = h; i, j = 1, 2, \ldots, n\} \]

is the number of distinct pairs located a vector distance \( h \) apart. The variogram (\( \gamma \)) shows the correlation between measured values and the distance between measuring nodes. The parameter \( \gamma \) is calculated as the average squared difference of values separated approximately by distance \( h \). This is called the experimental variogram and is later used to define the variogram model which is used for the interpolation process.

The variogram can be used to evaluate how well the dataset covers the variance in the CO\(_2\) flux in the measured area and shows graphically how similar measurements at certain distance are. A low value of the nugget parameter and an evenly shaped curve towards the sill parameter describes a dataset in which there is a strong correlation between neighbouring values and the correlation decreases with increased distance between measured nodes, indicating that the measurements cover well the variability or the anomalies in the area. A high value of the nugget describes a dataset where there is weak correlation even between neighbouring values at the selected grid spacing, indicating that values of CO\(_2\) are distributed more randomly in the area. For CO\(_2\) flux, it would not be realistic to expect two values, measured at a small distance (< 25 m) to give very similar results in all cases due to dissimilarity in the soil and possible underlying fissures and therefore it is not realistic to expect the nugget effect very low (\( \gamma < 0.1 \)). CO\(_2\) flux through soil is highly dependent on the type and nature of the soil, especially permeability of the soil. In nature, soil permeability is often very heterogeneous, at least on the scale of the CO\(_2\) flux measurements (the basal area of the fluxmeter is 0.03 m\(^2\)). Hence there can be great difference in measurements carried out within small distances and therefore it cannot be expected that two CO\(_2\) flux surveys on different grids would give exactly the same picture of the distribution of gas flux through soil even though all conditions were the same. On the other hand it is possible to conclude that more nodes measured on a tighter grid will give results closer to reality.
3.2.3 Measurements in 2011

In 2011, the annual soil measurements were carried out in the Reykjanes geothermal area but more data was collected during that year to obtain fuller information on the methods and the activity in the area. In March 2011, the extent of snowmelt in the geothermal area was tracked, by walking along the snow edges with a handheld GPS unit for the purpose of mapping the distribution of the soil temperature. Soil temperature at 15 cm depth was measured at 50 spots on the edges of the snowmelt to record the temperature at the margin. The data was collected in good weather conditions with no precipitation and a slow easterly wind and the snow cover was sufficient to show clearly the outlines of the warm soil.

![Figure 5](image_url) Overview of the location of the measured nodes. Red crosses represent nodes from dataset 2011 A and blue triangles show nodes from dataset 2011 B.

In June 2011, soil measurements were carried out from 7\textsuperscript{th} to 28\textsuperscript{th} June 2011 in relatively homogeneous and stable weather conditions with an average air temperature of 11.4°C, a mild wind from the north and no precipitation (Reiknistofa í Veðurfræði, 2011). The same
equipment was used as in previous years, a digital thermometer and a closed-chamber CO₂ flux meter from West Systems. In order to obtain information about the effects of the extent of the grid spacing on the results the area was measured twice, first normally on a 25 x 25 m grid covering about 0.42 km² and then the same area was measured again, locating the 25 x 25 m grids in between the first ones as shown in Figure 5. This allows geostatistical comparison of the two datasets and also to put them together in a tight dataset with approximately 17 m between points for more details. Both soil temperature at 15 cm depth and CO₂ were measured in each node of the grid. The total measurements were carried out in 908 measurement nodes, about 450 in each dataset, covering around 0.42 km² area.

According to Granieri et al. (2003), various external factors can influence the CO₂ soil flux, e.g. rainfall barometric pressure, air and soil temperature, air and soil humidity and wind speed. To minimize these effects and avoid potential reduction of the observed flux due to water saturation of the soil, the measurements were only carried out when at least 24 hours had passed without any rain. The ground covered by the chamber was chosen to be as flat as possible to avoid changes in volume inside the chamber, and to prevent contamination from the atmosphere the chamber was pressed firmly against the ground during measurements.

However, to obtain more information about weather related effects on the soil temperature in the Reykjanes geothermal area and to try to understand normal variations in the geothermal soil temperature at 15 cm depth on a short term scale (hours-days), the four HOBO data loggers, which were also used during the TIR image collection, were used to gather information about soil temperature over longer periods than previously. The four loggers obtained data from June 20th to August 16th 2011 (period II), randomly located around the Gunnuhver area, again from October 24th to December 20th (period III), this time located in a warmer ground than before, and then from January 24th to April 2nd 2012 (period IV), during which time all four were located in soil where the temperature was over 40°C. The loggers were programmed to measure temperature (in °C) and pressure (in kPa) regularly at 5 minute intervals during period II and III and 10 minute intervals during period IV. They were always located at approximately 15 cm depth in clayish soil, with no overlying vegetation or water. A weather station measuring air temperature, wind direction, wind speed and precipitation every 10 minutes (Verkfræðistofan Vista, 2011) is
located at well RN-13. Information from this weather station was used to correlate weather parameters with data from the loggers. The locations of the data loggers and the weather station at well RN-13 are shown in Figure 7.

Figure 6 Overview of the HOBO loggers’ locations, categorized by measuring periods.
4 Results

4.1 Results of temperature studies

4.1.1 Surface temperature - the 2011 thermal infrared imagery

The data from the TIR imagery, obtained in May 2011, were corrected to get an orthographical TIR image of the surface temperature using the ArcGIS software. Geometric projection from a cylinder to a plane was carried out using the tangent technique (Árnason 2011, pers. comm.). The image was geo-referenced in steps, first it was referenced roughly using an “affine projection”. This method uses a linear function to correlate between unreferenced image coordinates to spatial coordinates and requires only 3 points to make the correlation. Then more detailed projections were applied with up to 40 links used to generate a 3 order polynomial projection. In order to finalise the geo-referencing process a “rubber stamp” method was used to correlate the TIR image strips. The 10 image strips were conflated into a single file. In this processing step the data was reworked in the overlapping area and a new value is generated in the overlapped pixels. The resulting value is determined from an algorithm that is weight based and dependent on the distance from the pixel to the edge within the overlapping area. This creates a raster image of uniform resolution of 0.63 by 0.63 meters per pixel. The data covers the entire Reykjanes geothermal area, around the coast and is around 24 km² in total. Due to vibration caused by the aircraft’s motors, some features in the data, e.g. roads had a sinus curve shape. These effects where filtered out as well as noise and pixels at the edges of the flight lines.

The TIR-scanner encodes data as 16 bit numbers which means that the raw data has values which range from 0 to 65536. The data was then processed to 8 bit range, i.e. 256 grey levels with 0 for black (coldest value) and 255 for white. The image’s greyscale values (ranging from 0-255) can be converted into temperature by looking at the relationship between the 8 bit values at the location where the soil temperature was measured as described in section 2.5. (Lillesand and Kiefer, 2000). The in situ soil temperature, measured at the same time as the TIR-data was collected, did not reveal any clear
relationship. The soil temperature measurements can be found in Appendix A. Figure 7 shows the measured soil temperature plotted against the pixel value of the TIR image.

![Pixel value vs measured soil temperature (°C)](image)

**Figure 7** Results of the 94 in situ soil temperature measurements correlated with pixel value from the TIR image in the same coordinates in order to find a relationship to convert pixel values to temperature values. This dataset could not be used to define correlation between soil temperature and grey scale pixel value.

The correlations derived from this dataset did reveal a fit with a very low correlation coefficient ($R^2$), and in the end it was not used for calibrating the TIR image. This poor outcome can be explained mostly by resolution difference, the soil on the measured spots was not homogeneous enough, the pixel size of the TIR image is 0.63 x 0.63 meters and that can include a wide temperature range and finally the thermometers were inserted into soils at some (however shallow) depth while the image displays temperature from surface. The emitting ability of different surfaces is variable even though the difference for the surface type involved here are considered to be similar and it is also known that water bodies are better suitable for scaling purposes than heterogeneous soil (Árason, 2011, pers. comm.). Therefore it was decided to use the data from the HOBO loggers (two of them located in water bodies and other two located in rather homogeneous areas) as well as few other measured locations. Table 1 shows the coordinates of the locations where surface temperature was measured and the corresponding image value. Still, the temperature from
the logger located in warm soil was excluded because the data logger was located at 10-15 cm depth and therefore likely showed a higher temperature value than the one observed in the TIR data at the surface. By involving both temperatures from water bodies and from surfaces with other emitting ability the effects of different emissivity is lowered.

Table 1: Coordinates of the measured locations and the corresponding image value used for correlating the TIR image

<table>
<thead>
<tr>
<th>Description</th>
<th>Logger</th>
<th>x</th>
<th>y</th>
<th>Temperature (°C)</th>
<th>Image value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cold water</td>
<td>Logger 1</td>
<td>317793</td>
<td>373631</td>
<td>7.5</td>
<td>23</td>
</tr>
<tr>
<td>Gray lagoon</td>
<td>Logger 2</td>
<td>318605</td>
<td>374039</td>
<td>46.7</td>
<td>96</td>
</tr>
<tr>
<td>Cold soil</td>
<td>Logger 4</td>
<td>319274</td>
<td>373916</td>
<td>5.96</td>
<td>12</td>
</tr>
<tr>
<td>Warm soil</td>
<td>Logger 2</td>
<td>318721</td>
<td>373724</td>
<td>35.97</td>
<td>14</td>
</tr>
<tr>
<td>Gunnuhver</td>
<td>Measured</td>
<td>318633</td>
<td>373705</td>
<td>100</td>
<td>242</td>
</tr>
<tr>
<td>Outlet water temperature</td>
<td>Measured</td>
<td>317276</td>
<td>375446</td>
<td>50.8</td>
<td>92</td>
</tr>
<tr>
<td>Sea temperature</td>
<td>Measured</td>
<td>317165</td>
<td>375872</td>
<td>8</td>
<td>13</td>
</tr>
</tbody>
</table>

The pixel values were plotted as a function of measured temperature and a best fit function through the data was made (presented in Figure 8).

Figure 8 Pixel values plotted as a function of measured temperature and a best fit function through the data. The grey dot shows the measured value and corresponding pixel value for warm soil. Since it was obtained in soil at approximately 10 cm depth, it was excluded.
This resulted in the temperature calibration equation describing the relationship between the measured ground temperatures and the corresponding image pixel values in the TIR image:

\[
\text{Equation 4} \quad -0.0007755l^2 + 0.60997l - 2.13714 = t
\]

where \( l \) is the pixel value from the image and \( t \) is the temperature in °C. The surface temperature based on the orthographically corrected TIR image was obtained for all pixel values in the TIR image by using Equation 4 and the software ArcMap. The resulting surface temperature ranged from 0 to 102°C and the temperature map is shown in Figure 9. The image clearly shows the temperature anomalies in the Reykjanes geothermal area. The highest temperature captured is in a big cratered fumarole in the Gunnuhver area with a temperature of 102°C. The main area showing high temperature is around the Gunnuhver area, north direction from wells RN-1 to RN-3. The area with high temperatures extends towards northwest and well RN-2. Then close to the Gráa lónið lagoon is a warm area where surface activity is dominated by mud pits. North of the Gráa lónið lagoon, around RN-28 there is an area with a significantly elevated temperature. High temperature is also observed in the Gráa lónið lagoon, where effluent from the separation plant flows into the lagoon.
Figure 9 Calibrated TIR image of the geothermal area in Reykjanes from May 2011.

4.1.2 Comparison with snowmelt track

A map of the snowmelt tracks from March 2011 is shown in Figure 10. It shows the areas, where the soil temperature at 15 cm depth on the snow – bare earth border reached at least 20°C (the lowest measured value) but the average value was 36.3°C on the edges of the snowmelt track.
Figure 10 Snowmelt areas in the geothermal area in Reykjanes in March 2011. Green dots are locations of measured nodes and numbers the corresponding temperature values at 15 cm depth.

The snow melt track is overlain on the TIR image in Figure 11. The figure shows that the snow melt track is in excellent agreement with the outline of the thermal anomaly as detected by the TIR image. This suggests that snowmelt tracking of the Reykjanes geothermal field is a good method to map the extent of the surface thermal anomaly. The greatest discrepancy appears in the eastern part of the area where the snowmelt track extends more towards east than is seen in the TIR image. This might be caused by a westerly wind blowing the geothermal steam towards east and melting the snow, signifying that good weather and snow conditions are important to get good results.
Figure 11  Snowmelt tracks overlying the 2011 TIR image. It shows that mapping the distribution of areas with elevated surface temperatures by snowmelt tracking fits well with the distribution from the TIR image.

4.1.3 Comparison with an older TIR image

A TIR image from the Reykjanes geothermal area, obtained in April 2004 and thoroughly described by Margrétardóttir (2005) was used here for comparison. This TIR image does not cover the whole peninsula’s tip as the TIR image from 2011, but shows the greatest part of the geothermal area as can be seen in Figure 12. This TIR image from 2004 was obtained using the same infrared scanner as in 2011 and it was projected like the TIR image from 2011 (Margrétardóttir 2005), and the temperature scale displayed here is the same. Visual comparison reveals that in 2004 the area of elevated surface temperature is smaller than in 2011. The area with the highest temperature around the Gunnuhver area appears strongly in both images and so does the area furthest northwest, closest to the Gráa lónið lagoon. The area there between, close to well RN-2 shows much higher temperatures in 2011 than in 2004. The area north of the lagoon does not appear warm at all in 2004 but
in 2011 it clearly shows elevated temperature values. In 2005, this part of the Reykjaness geothermal area was included in the soil measurements but since then this area has not been included in the soil temperature and CO$_2$ flux measurements so this is the only data from after the commissioning of the power plant that covers this part of the area.

![Image](image_url)

**Figure 12** TIR image from April 2004 covering the area with the most significant surface activity in the Reykjaness geothermal area.

Figure 13 shows an image in which the 2004 TIR image has been subtracted from the 2011 TIR image, showing clearly where the temperature has increased (yellow and red areas) and where it has decreased (green and blue areas). The area north of the lagoon appears strongly yellow and red indicating a strong temperature increase. The strongly red coloured phenomenon which appears by the northeast tip of the Gráa lónið lagoon is the separator station which is a part of the power station, and built after the 2004 TIR imagery but before the commissioning of the power plant in spring 2006. This image also shows that the middle part of the area, close to well RN-2 also has clearly warmed up during this period. The only area, where the temperature has decreased significantly during this 8 years period is a small area close to well RN-3.
Figure 13 The 2004 image has been subtracted from the 2011 image showing the difference between surface temperatures. Red and yellow indicated areas where the surface temperature is higher in 2011 and green and blue colours show areas where the surface temperature was higher in 2004. The pink line defines the “south-part” and brown lines mark the “north-part” (see text below).

To be able to evaluate better the changes between 2004 and 2011 recorded by the TIR images, two areas were selected for a closer look. Firstly the main activity area was defined. This area is marked with a pink line in Figure 13 and is hereafter referred to as the “south-part”. This area was also covered by the soil measurements in 2011. Secondly, the areas north and east of the Gráa lónið lagoon were defined but ground measurements have not been carried out in these parts in recent years. These two parts are shown inside a brown line in Figure 13 and are from now on referred to as one part, the “north-part”. Straight through the north-part there appears to be a warm pipeline that extends from the boreholes and into the separator station that was non-isolated during the time of the TIR imagery because of maintenance (see Figure 13). The pipeline was excluded from the average temperature evaluation due to its unnatural origin. Average temperatures,
according to the TIR images were calculated for both parts. For the south-part, the average temperature has increased from 7.4°C in 2004 to 10.1°C in 2011. For the north-part, the difference is much greater, the average temperature has increased from 5.7°C in 2004 to 14.0°C in 2011.

### 4.1.4 Comparison with soil measurements

The results of all the soil temperature measurements, carried out in June 2011 at 15 cm depth, were used to map the distribution of soil temperature. There were 908 measurements and the temperature ranged from 7.1 to 101.1°C with the average value of 49°C. The software Surfer (from Golden Software) was used to create maps and the kriging method for interpolation. The resulting temperature map is shown in Figure 14.

![Figure 14](image-url)  
*Figure 14* Thermal map compiled from 2011 soil temperature measurements using kriging interpolation.
When comparing the TIR data to the soil measurement result it is obvious that the temperature observed in the TIR data is lower than what is observed from the soil measurements. By taking the temperature from the TIR image in the same coordinates as that from the soil measurements a rather weak correlation was detected as shown in Figure 15.

![Figure 15](image)

**Figure 15** Correlation between surface temperature from the TIR image and results of soil temperature measurements at 15 cm depth. Red line shows a linear fit through the dataset.

The average measured soil temperature is 37.3°C higher than the temperature calculated from the TIR image of the 908 measured nodes. Margrétardóttir (2005) showed very similar results with the average remotely obtained surface temperature 30.3°C lower than the average soil temperature measured at 15 cm depth. As expected, the difference was greater in places where the measured temperature at 15 cm depth was high but much smaller where the measured temperature was low. This can be explained by the different methods and reasons. The TIR imagery took place on 25th of May 2011, around midnight with air temperature around 6°C but the soil measurements were done in June, during daytime with average air temperature around 11.4°C. The soil temperature measurements are carried out in-situ at 15 cm depth but not at the surface while the TIR scanner is collecting data that is radiated from the surface. The data collected by the TIR scanner is also affected by atmospheric parameters and steam. The warm geothermal soil can have a
colder mud crust overlying a much warmer ground underneath and therefore the TIR image loses some information on warm soil. The soil temperature measurements also miss some details due to the extent of the measurement grid resulting in much worse resolution and the interpolation methods used (here kriging), do have their limitations, described in section 2.4. Still, the same anomalies appear in both cases, the areal extent fits very well and some details can be seen, for example a cold area between wells RN-1 and RN-2.

### 4.1.5 Older soil temperature measurements in the Reykjanes geothermal area

The soil temperature data, collected annually since 2004, have been used to map the distribution of thermal anomalies and evaluate changes from year to year. The mapping has been effected using the software Surfer (from Golden Software) using the kriging interpolation. From 2004, some changes have appeared from year to year but there are still some main anomalies that have been more or less constant. The most prominent soil temperature anomaly is in all cases in the Gunnuhver area, in the south-eastern part of the studied area. This is the area with the most intense surface activity, including steam vents. In 2008, though, this anomaly was not as obvious as in other years. Furthermore, there is an anomaly at the western end of the study area, by the west end of the Gráa lónið lagoon where mud pits dominate the surface activity. This anomaly has become more obvious, especially in 2009, 2010 and 2011. Even though these anomalies are generally similar from year to year, significant differences are noticeable. Some of the variations are related to the commissioning of the Reykjanesvirkjun Power Plant but others cannot be explained easily. The distribution maps are presented in Appendix A.

Between the first two years, 2004 and 2005, the results show little changes in soil temperature and the distribution was nearly the same apart from a slight decrease within the hottest part of the area. The measurements from 2006 were carried out right after the commissioning of the Reykjanes geothermal power plant and by then, the surface activity and steam flow were significantly more intense than before. The area with warm soil extended further to the south and southeast and the temperature readings were generally higher than in previous years. Between 2006 and 2007, the area stretched even more to the south and southeast as well as to the northeast and the whole area seemed to be warmer.
Measured soil temperatures in summer 2008 were significantly lower than in 2007 although an elongated anomaly, which was not noticed in 2007, stretched from the warm area south of Kísilhóll towards Skálafell. Despite this general cooling and reduction in the warm area from 2007 to 2008, the soil temperature anomaly was obviously of a larger areal extent in 2008 than in 2006. It is not clear what could have caused this reduction in the soil temperature between 2007 and 2008. Comparison of soil temperature measurements from 2006 and 2009 shows that the size of the soil temperature anomaly is greater in 2009 than in 2006 but the temperature is somewhat lower around Gunnuhver, the warmest area. Between 2009 and 2010 some changes appear in soil temperature. The area around Gunnuhver was warmer in 2010 than in 2008 and 2009 and there was a possible increase in temperature to the southwest, in the area between wells RN-2 and RN-4.

In June 2011, the soil temperature was measured twice and these two datasets were compared. In Figure 16 and Figure 17 the two different datasets are used to create temperature anomaly maps. Visual comparison of the datasets does not reveal any great differences. The soil temperature distribution is very similar, but with some minor differences e.g. elevated temperature stretching little bit further south in Figure 17. The temperature anomaly outlines in the map from dataset B are slightly stronger and the areal extent of the thermal anomalies a little bit greater than in the map from dataset A. That corresponds to the dataset parameters, in which the average temperature measured in dataset A is 46.8°C and the median 41.4°C while the average temperature is 51.3°C and the median 45.8°C in dataset B.
Figure 16  Dataset 2011A used to map the soil temperature distribution at 15 cm depth in Reykjanes using the kriging interpolation.

Figure 17  Dataset 2011B used to map the soil temperature distribution at 15 cm depth in Reykjanes using the kriging interpolation.
Visual comparison between 2011total (Figure 14) and previous years (see Appendix A) indicates a warmer ground in 2011 than before. The anomalies around well RN-2 and both north and south of it, are stronger in 2011 than in 2010. More detailed anomalies can be seen in the figure from 2011total, due to the tighter grid but the areal extent is very similar in 2010 and 2011.

4.1.6 The heat flow through soil

The heat flow through soil ($Q_{T15}/d_{97}$) was calculated from the soil temperature measurements in 2004 and again in 2007-2011, assuming that convection is the dominant heat transport mechanism using Equation 1 of Dawson (1964). In 2005 and 2006 $d_{97}$ was not measured making it impossible to calculate the heat flow using this method. For 2004 and 2007-2011, Equation 2 was used for the nodes for which the soil temperature exceeded 97°C at 15 cm depth. The results are shown in Table 2. Since two datasets were collected in 2011, as described in section 3.2.3, heat flux calculations were also carried out for 2011A and 2011B separately. By using the Dawson method it is assumed that each value is representative for the same size of area and for that to be reasonable, the grid spacing has to be regular. In 2011, the measurements were done on grid with approximately 25 m x 25 m spacing but due to circumstances in the area, regular grid spacing was impossible in some parts of the area. Therefore, to take the irregular spacing into account and also to be able to estimate the uncertainty of the heat flow, 100 simulations or realizations using the sGs algorithm of the sgsim code by Deutsch and Journel (1998) were performed on each annual dataset for which the heat flux had been calculated using the Dawson method (except 2005 and 2006) in the Reykjanes geothermal area.

The total heat flux for each year was derived from the soil temperature measurements and 100 sGs realizations and the uncertainty with 95% confidence level was evaluated by using the realizations. The heat flux was determined for each year’s measurements coverage but for comparison, two different parts were chosen. Firstly, an area defined tightly around the part of the whole area which has been measured every year (Tight Comparison Area, TCA). Since the measured area has not been the same though the years due to changes that have appeared in areas stretching in one direction or another, these are excluded in such a comparison, and another larger comparison area was also chosen (Large Comparison Area,
LCA). These areas are shown in Figure 18. To be able to compare areas where measurements were not performed during some years, a background value of Q=0.02 which corresponds to 5°C which marks the background temperature threshold in Reykjanes (Margrétardóttir, 2005) was used for the heat flow determination and used to complement for the areas where no measurements took place.

Table 2  The heat flux from the Reykjanes geothermal field derived from soil temperature measurements, 100 sequential Gaussian simulations and the uncertainty evaluated using the simulations. Note that no data is available for 2005 and 2006.

<table>
<thead>
<tr>
<th>Year</th>
<th>Heat flow in MW according to 100 Gaussian simulations from TCA</th>
<th>Heat flow in MW according to 100 Gaussian simulations from LCA</th>
</tr>
</thead>
<tbody>
<tr>
<td>2004</td>
<td>16.9$^a$ ± 1.4</td>
<td></td>
</tr>
<tr>
<td>2005</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2006</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2007</td>
<td>39.0 ± 10.7</td>
<td>40.1 ± 10.8</td>
</tr>
<tr>
<td>2008</td>
<td>19.4 ± 2.6</td>
<td>20.6 ± 2.7</td>
</tr>
<tr>
<td>2009</td>
<td>31.0 ± 6.0</td>
<td>34.8 ± 6.8</td>
</tr>
<tr>
<td>2010</td>
<td>28.0 ± 4.3</td>
<td>29.0 ± 4.2</td>
</tr>
<tr>
<td>2011 total</td>
<td>34.3 ± 2.6</td>
<td>36.1 ± 2.5</td>
</tr>
<tr>
<td>2011A</td>
<td>33.0 ± 4.3</td>
<td>35.1 ± 4.7</td>
</tr>
<tr>
<td>2011B</td>
<td>39.3 ± 4.4</td>
<td>41.5 ± 4.5</td>
</tr>
</tbody>
</table>

$^a$ data from Fridriksson et al. 2006 with 300 realizations.

Figure 18  Heat flux derived from the 2011 total dataset. The TCA is shown inside the pink line and the LCA is shown inside the brown line.
As can be seen in Table 2 the heat flow computed from measured soil temperatures has more than tripled between 2004 and 2011. It has not increased evenly from year to year; on the contrary the results have been rather irregular and a noticeable reduction is seen in the heat flux in 2008. When the Gaussian simulations are used as an interpolation method to process the heat flux data to avoid assuming that every value is representative for the same size of area, it shows that a peak in heat flow (until 2011) may have been reached in 2007, a year after the commissioning of the Reykjanes power plant but the uncertainty value for this year is very high so it allows for heat flux values from 29.3 MW to 50.9 MW. In Figure 18 the heat flux according to 100 sGs realizations for the 2011 total dataset is shown.

![Graph showing the total heat flow from the large comparison area with time.](image)

**Figure 19** Graph showing the total heat flow from the large comparison area with time.

### 4.1.7 Results of continuous soil temperature measurements

The data from the four data loggers, which were used to obtain information about soil temperature at 15 cm depth from June 20th to August 16th 2011 (period II), October 24th to December 20th (period III) and January 24th to April 2nd 2012 (period IV) were analysed and compared to weather data (air temperature, air pressure and precipitation) obtained from a weather station on well RN-13. The data on wind speed from the weather station proved to be incorrect due to a technical problem in the weather station and is therefore unusable.
During period II, the data loggers all show diurnal variations, especially the three loggers with the lowest temperature, as can be seen in Figure 23. This is during the summer, July and August, hence diurnal variations are expected. The diurnal variations are on the scale of 1.5-6°C. Correlations between the loggers are very high, with a correlation coefficient, $R^2$ up to 0.84 between logger II-A and logger II-B but the correlation is somewhat lower when compared with logger II-D, which is in the warmest soil. Selected statistical parameters for the logger data of period II are shown in Table 3 and the correlations between each of them are shown in Figure 20.

**Table 3 Some statistical parameters for loggers I-IV from period II.**

<table>
<thead>
<tr>
<th>Period II logger</th>
<th>Average value (°C)</th>
<th>Min value (°C)</th>
<th>Max value (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Log II-A</td>
<td>19.4</td>
<td>15.1</td>
<td>23.0</td>
</tr>
<tr>
<td>Log II-B</td>
<td>34.5</td>
<td>29.1</td>
<td>39.6</td>
</tr>
<tr>
<td>Log II-C</td>
<td>35.1</td>
<td>24.1</td>
<td>44.7</td>
</tr>
<tr>
<td>Log II-D</td>
<td>51.1</td>
<td>21.8</td>
<td>58.9</td>
</tr>
</tbody>
</table>
For period III, some similar effects are noted. Loggers A, B and C are located in rather cold geothermally altered soil with an average temperature from 10 to 30°C while the warmest one has an average temperature of 61°C. This data is obtained during the fall, from late October to late December and the diurnal variation is not seen as clearly as during period II, yet it is still present and the variations are on the scale of 1-2°C. As for period II, the relative correlation between the loggers is rather high or up to $R^2 = 0.521$ but when the logger III-D, which obtained the highest temperature values, is compared to each of the others, the correlation is inverse and also very low.
Table 4 Some statistical parameters for loggers I-IV from period 3.

<table>
<thead>
<tr>
<th>Period III logger</th>
<th>Average value (°C)</th>
<th>Min value (°C)</th>
<th>Max value (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Log III-A</td>
<td>8.0</td>
<td>1.4</td>
<td>13.1</td>
</tr>
<tr>
<td>Log III-B</td>
<td>26.0</td>
<td>20.5</td>
<td>30.1</td>
</tr>
<tr>
<td>Log III-C</td>
<td>22.1</td>
<td>13</td>
<td>26.4</td>
</tr>
<tr>
<td>Log III-D</td>
<td>61.6</td>
<td>37.8</td>
<td>74.6</td>
</tr>
</tbody>
</table>

Figure 21 Correlations between loggers I-IV from period 3. The $R^2$ (correlation coefficient) is shown for each plot.

In 2012, the HOBO data loggers were placed in warm soil, all four loggers were placed in soil warmer than 40°C, in the Reykjanes geothermal area during the period from January 31th to April 2nd. For this period, the diurnal variations are not obvious most likely because
these data were collected during winter. As for the other periods, the relative correlation between the loggers is rather high or up to $R^2 = 0.786$. As seen in Table 3, 4 and 5 it appears that the range of temperatures value increases with higher temperature and loggers in soil with high temperatures can show temperature values in the 40°C range.

**Table 5 Some statistical parameters for loggers I-IV from period 4.**

<table>
<thead>
<tr>
<th>Period IV logger</th>
<th>Average value (°C)</th>
<th>Min value (°C)</th>
<th>Max value (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Log IV-A</td>
<td>39.6</td>
<td>32.9</td>
<td>48.3</td>
</tr>
<tr>
<td>Log IV-B</td>
<td>54.6</td>
<td>33.1</td>
<td>63.6</td>
</tr>
<tr>
<td>Log IV-C</td>
<td>44.9</td>
<td>40.6</td>
<td>49.1</td>
</tr>
<tr>
<td>Log IV-D</td>
<td>74.2</td>
<td>39.4</td>
<td>80.2</td>
</tr>
</tbody>
</table>

**Figure 22** Correlations between loggers I-IV from period 4. The $R^2$ (correlation coefficient) is shown for each plot.
Most loggers from all periods (except especially logger III-A and also loggers III-B and III-C) did show some correlation with precipitation. Figure 23 to Figure 25 show the temperature from each logger plotted with time and the 24 hour precipitation (in mm) is also shown. Visually, it is noticed that precipitation seems to cause a sudden lowering in the soil temperature and the warm soil seems to be especially sensitive. The highest value of $R^2$ was reached when the precipitation was accumulated over 96 hours (4 days) for the data from every logger and every period, except from logger III-C. Data from loggers III-A, III-B, III-C show an inverse relationship with precipitation but these loggers are among the ones with the lowest temperature and therefore relatively sensitive to air temperature. When looking at the relationship between air temperature and precipitation it is clear that most precipitation took place during the warmest air temperature periods, thus creating this inverse relationship.

![Figure 23](image)

**Figure 23**  Results of soil temperature measurements from data loggers from June 20th to August 16th (period II) with time. The 24 hour precipitation is also shown.
During all the measurement periods, the loggers located in soil warmer than 40-50°C show some intense temperature drops that cannot be fully related to precipitation or any other known parameter. These events occur over a very short time (hours) and the temperature can drop about 10-35°C and then rise as quickly again.

The loggers were all corrected to available weather parameters, precipitation, air pressure and air temperature, and it appears that there is very weak and in some cases no relationship found between loggers temperature and either air temperature or air pressure. This is shown in Figure 26, to Figure 28 in which there is also shown the correlation with 24 hour precipitation and 96 hour precipitation which in almost all cases maximized the $R^2$ value.
Figure 26 Correlations between results from loggers A-D from period II with air pressure, air temperature, 24 hour precipitation and 96 hour precipitation. The $R^2$ (correlation coefficient) is shown for each plot.
Figure 27 Correlations between results from loggers A-D from period III with air pressure, air temperature, 24 hour precipitation and 96 hour precipitation. The $R^2$ (correlation coefficient) is shown for each plot.
Figure 28 Correlations between results from loggers A-D from period IV with air pressure, air temperature, 24 hour precipitation and 96 hour precipitation. The $R^2$ (correlation coefficient) is shown for each plot.
4.2 CO₂ measurements in the Reykjanes geothermal area

4.2.1 CO₂ measurements from 2004 to 2011

On each annual dataset on CO₂ flux in Reykjanes geothermal area, 100 sequential Gaussian simulations have been performed using the sGs algorithm of the sgsim code (Deutsch and Journel, 1998). After a normal score transformation of each dataset, a variogram model of the normal scores that fitted the experimental values of each dataset was used. The simulation domain was divided into square cells, each having a surface of 4 m².

<table>
<thead>
<tr>
<th>Year</th>
<th>Nugget</th>
<th>Sill</th>
<th>Range</th>
<th>Type of variogram model</th>
</tr>
</thead>
<tbody>
<tr>
<td>2004</td>
<td>0.4</td>
<td>1</td>
<td>150</td>
<td>spherical</td>
</tr>
<tr>
<td>2005</td>
<td>0.36</td>
<td>1</td>
<td>360</td>
<td>exponential</td>
</tr>
<tr>
<td>2006</td>
<td>0.34</td>
<td>1</td>
<td>160</td>
<td>spherical</td>
</tr>
<tr>
<td>2007</td>
<td>0.43</td>
<td>1</td>
<td>180</td>
<td>spherical</td>
</tr>
<tr>
<td>2008</td>
<td>0.5</td>
<td>1</td>
<td>300</td>
<td>spherical</td>
</tr>
<tr>
<td>2009</td>
<td>0.68</td>
<td>1</td>
<td>125</td>
<td>spherical</td>
</tr>
<tr>
<td>2010</td>
<td>0.6</td>
<td>1</td>
<td>165</td>
<td>spherical</td>
</tr>
<tr>
<td>2011</td>
<td>0.5</td>
<td>1</td>
<td>175</td>
<td>exponential</td>
</tr>
</tbody>
</table>

The results of each 100 simulations were depicted on maps that show the mean CO₂ flux of individual cells in the model. The maps are in Appendix B. From 2004, some changes in CO₂ flux have been noticed from year to year but there are still some main anomalies that have appeared every year. First, in all cases there is an obvious anomaly north of the road at the Gunnuhver area, to the west of well number 3, in the south-eastern part of the study area. This is the area with the most intense surface activity, including steam vents. Furthermore, there is an anomaly at the western end of the study area, by the west end of the Gráa lónið lagoon where mud pits dominate the surface activity. This anomaly did not appear strongly in 2007 and 2008. Finally there is an anomaly that stretches south or southwest from the east tip of the Grey lagoon. Even though these anomalies are generally
similar from year to year, significant differences are noticeable. Some of the variations are coherent with the changes seen in the temperatures and is related to the commissioning of the Reykjanes Power Plant.

Between 2004 and 2005 the distribution of CO$_2$ flux was rather similar even though the shapes of the anomalies were not exactly the same. The results of the measurements from the summer of 2006 show an obvious increase in CO$_2$ flux, compared to previous years, especially around Gunnuhver and the anomalies stretched further south. In 2006, the CO$_2$ flux seemed to increase a little southwest of the Grey lagoon. The anomaly around Gunnuhver is very clear in 2006 and again in 2007. The results from 2008 are completely different from the other years, both before and after as CO$_2$ flux appears to have dropped drastically in that year. The results of the measurements from 2009 indicate that the CO$_2$ flux had reached the same level then as in 2006 and 2007. In 2009, however, the anomalies did not appear as clearly as in 2006 and 2007. The results from 2010 indicate an increase in the CO$_2$ flux compared to 2009. The anomalies are not well defined and not as clear difference between areas with high and low flux rates. In 2010, the distribution is similar to the 2009 observations.

4.2.2 Geostatistical treatment of the 2009 and 2010 datasets

In order to evaluate geostatistically the accuracy and resolution of the methods used to process the data for the CO$_2$ measurements since 2004 and to evaluate the effects of grid spacing on the results of mapping the CO$_2$ flux anomalies, the existing datasets from 2009 and 2010 in Reykjanes were analysed. First, the datasets from 2009 and 2010 were resampled or divided into two identical parts with double grid spacing as compared to the original data set. The new data sets are referred to as dataset 2009A, dataset 2009B and dataset 2010A and 2010B. Exactly every other sampling node from the original grids was sampled into the two new data sets for each year. So, what had originally been a data set of N data points on a 25 by 25 m grid was now separated into two sets of N/2 data points on a grid with about 50 by 50 m grid spacing. The division was performed completely without any reference to the CO$_2$ flux value; it was just based on the location of the sampling. In this way, two datasets, covering the same area were prepared and these two datasets were then separately analysed and the results compared. The total datasets from 2009 and 2010 undivided, called 2009total and 2010total are here also for comparison.
When the datasets 2009A and 2009B had been prepared, statistical parameters for each dataset were computed and it appeared that the averages from the two datasets were quite different despite the arbitrary division into the two datasets. In Table 7 the average, median, highest and lowest values are shown for both datasets and the 2009 total in g/m²/day:

**Table 7 Some statistical parameters brought out from the divided datasets from 2009**

<table>
<thead>
<tr>
<th>Part</th>
<th>Average</th>
<th>Median</th>
<th>Highest value (g/m²/day)</th>
<th>Lowest value (g/m²/day)</th>
<th>Number of points</th>
</tr>
</thead>
<tbody>
<tr>
<td>2009 A</td>
<td>88.24</td>
<td>18.8</td>
<td>2418.11</td>
<td>1.592</td>
<td>184</td>
</tr>
<tr>
<td>2009 B</td>
<td>65.76</td>
<td>15.3</td>
<td>2550.08</td>
<td>0.796</td>
<td>182</td>
</tr>
<tr>
<td>2009 total</td>
<td>78.14</td>
<td>16.1</td>
<td>2550.08</td>
<td>0.796</td>
<td>367</td>
</tr>
</tbody>
</table>

The same was done for the 2010 datasets and the results are shown in Table 8 but here, the difference between the average and median of the two data sets is smaller than for the 2009 data.

**Table 8 Some statistical parameters brought out from the divided datasets from 2009**

<table>
<thead>
<tr>
<th>Part</th>
<th>Average</th>
<th>Median</th>
<th>Highest value (g/m²/day)</th>
<th>Lowest value (g/m²/day)</th>
<th>Number of points</th>
</tr>
</thead>
<tbody>
<tr>
<td>2010A</td>
<td>122.02</td>
<td>10.59</td>
<td>6755.48</td>
<td>0.0108</td>
<td>214</td>
</tr>
<tr>
<td>2010B</td>
<td>129.34</td>
<td>11.48</td>
<td>5804.47</td>
<td>0.0157</td>
<td>215</td>
</tr>
<tr>
<td>2010 total</td>
<td>125.69</td>
<td>10.89</td>
<td>6755.48</td>
<td>0.0108</td>
<td>429</td>
</tr>
</tbody>
</table>

The next step was preparing semi-variogram, variogram models and 100 sequential Gaussian simulations for each dataset separately. The nugget, sill and range are shown for each dataset.

**Table 9 Nugget, sill and range for each dataset**

<table>
<thead>
<tr>
<th>Year</th>
<th>Nugget</th>
<th>Sill</th>
<th>Range</th>
<th>Type of variogram model</th>
</tr>
</thead>
<tbody>
<tr>
<td>2009A</td>
<td>0.84</td>
<td>1</td>
<td>125</td>
<td>spherical</td>
</tr>
<tr>
<td>2009B</td>
<td>0.74</td>
<td>1</td>
<td>140</td>
<td>spherical</td>
</tr>
<tr>
<td>Year</td>
<td>Gamma</td>
<td>Alpha</td>
<td>Distance</td>
<td>Model</td>
</tr>
<tr>
<td>-----------</td>
<td>-------</td>
<td>-------</td>
<td>----------</td>
<td>--------</td>
</tr>
<tr>
<td>2009total</td>
<td>0.68</td>
<td>1</td>
<td>125</td>
<td>spherical</td>
</tr>
<tr>
<td>2010A</td>
<td>0.7</td>
<td>1</td>
<td>160</td>
<td>spherical</td>
</tr>
<tr>
<td>2010B</td>
<td>0.49</td>
<td>1</td>
<td>190</td>
<td>spherical</td>
</tr>
<tr>
<td>2010total</td>
<td>0.6</td>
<td>1</td>
<td>165</td>
<td>spherical</td>
</tr>
</tbody>
</table>

On Figure 29 semi-variograms (red connected dots) are shown and variogram models for each datasets (coloured curve).

![Experimental semi-variogram and variogram models for CO$_2$-flux for datasets 2009A, 2009B and 2009total are shown. On the x-axis there is the distance in meters and on the y-axis the $\gamma$. Red dots show an experimental semi-variogram derived from each dataset but the blue line (for 2009A), the green line (for 2009B) and the violet line (for 2009total) show the variogram model prepared for each dataset.](image-url)
As seen in Figure 29 there is a significant difference between the results from datasets 2009A and 2009B. The semi-variograms show that the first value or the nugget for both datasets is very high or more than 0.7 showing a very poor correlation between neighbouring values. It indicates that the data from either dataset 2009A or 2009B, with grid spacing around 30-50 meters, does not cover the distribution and local variance of the CO$_2$ flux not well enough, at least it does not show as good a correlation between neighbouring values as would be expected for the CO$_2$ flux anomalies in the Reykjanes geothermal area. The 2009total with a grid spacing of 25 x 25 meters gives a lower nugget indicating that this dataset covers better the CO$_2$ variation.

![Experimental semi-variogram and variogram models for CO$_2$-flux for datasets 2010A, 2010B and 2010total.](image)

**Figure 30:** Experimental semi-variogram and variogram models for CO$_2$-flux for datasets 2010A, 2010B and 2010total are shown. On the x-axis there is the distance in meters and on the y-axis the $\gamma$. Red dots show experimental semi-variograms derived from each dataset but the green line (for 2010A), the blue line (for 2010B) and the violet line (for 2010total) show the variogram models prepared for each dataset.
In Figure 30 it is shown that for dataset 2010A the value for $\gamma$ is 0.7 which is lower than that for the divided datasets from 2009. For 2010B it is even lower or $\gamma=0.49$ which indicates that this dataset covers the CO$_2$ flux anomalies in the area much better than 2010A.

When 100 sequential Gaussian simulations had been performed for each of these 6 datasets, the results were mapped and the distribution maps are shown in Figure 31 and Figure 32. A grey line was drawn indicating the anomalies appearing in the total dataset.
Figure 31 CO₂ flux from 2009 in the study area in Reykjanes, the first one from dataset 2009A, the second one from 2009B and the third one from 2009total. The scale bar is the same for all images and shows the flux in g m⁻² day⁻¹. The measured nodes are also displayed, sorted into five groups according to values. Grey lines show CO₂ flux anomalies, derived from the 2009total.

Figure 31 shows the CO₂ flux from the same area in Reykjanes, each using a different dataset from 2009. Visual comparison of the three images reveals that there is a significant difference in the outcome. The 2009A image was very scattered image with a low contrast between high flux areas and low flux areas, but still mostly following the grey line, anomalies derived from the 2009total, indicating similar anomalies as the 2009total. The image showing results from 2009B shows better defined anomalies with more contrast between high flux areas and low flux areas than in 2009A. The total CO₂ flux was calculated for each dataset.

By using the equally probable realizations the total CO₂ flux was calculated for each dataset but instead of 100 realizations calculated for mapping the distribution, 1000
realizations are used to estimate the total CO$_2$ flux to avoid possible random bias. The uncertainty is determined with 95% confidence level.

**Table 10** Total CO$_2$ flux in tons per day for the 2009 datasets, the uncertainty is evaluated with a 95% confidence level.

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Total CO$_2$ flux (tons/day)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2009A</td>
<td>26.0 ± 1.8</td>
</tr>
<tr>
<td>2009B</td>
<td>19.9 ± 2.2</td>
</tr>
<tr>
<td>2009total</td>
<td>21.6 ± 2.3</td>
</tr>
</tbody>
</table>
Figure 32 CO₂ flux from 2010 in the study area in Reykjanes, the first one from dataset 2010A, second one from 2010B and third one from 2010total. The scale bar is the same for all images and shows the flux in g m⁻² day⁻¹. The measured nodes are also displayed, sorted into five groups according to values. Grey lines show CO₂ flux anomalies, derived from the 2009total.
The graphical representations of the datasets from 2010 are interesting. Even though the semi-variogram for dataset 2010A indicates more poorly defined anomalies than 2010B and 2010total, a similar anomaly pattern appears in 2010A and in the 2010total. In 2010A they appear weaker and much more scattered with less difference between high flux and low flux areas. 2010B, whose semi-variogram indicated better defined anomalies, shows a different picture with a very well defined CO\textsubscript{2} flux anomaly which covers the middle part of the area heading NNV-SSA.

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Total CO\textsubscript{2} flux (tons/day)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2010A</td>
<td>35.6 ± 5.6</td>
</tr>
<tr>
<td>2010B</td>
<td>36.6 ± 8.3</td>
</tr>
<tr>
<td>2010total</td>
<td>34.4 ± 4.8</td>
</tr>
</tbody>
</table>

From these results it is clear that if the grid spacing is not small enough it is possible to get not only poorly defined anomalies as seen in 2009A but also misleading anomaly shapes as seen in 2010B. The tightness of the grid is therefore fundamental for reliable results in CO\textsubscript{2} flux measurements.

4.2.3 CO\textsubscript{2} and temperature measurements in 2011

In 2011 the CO\textsubscript{2} flux and soil temperature measurements were carried out on two different measurement grids, covering the same area (see section 3.2.3). The datasets were processed and interpreted separately and in combination. First, the average, median, highest and lowest values were brought out from each dataset as can be seen in Table 12. The difference between these two datasets is smaller than was seen in the divided datasets from 2009 and 2010 (above) but there is still a difference in all parameters.
Table 12 Statistical parameters brought out from the divided datasets from 2009

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Average CO₂ flux (g/m²/day)</th>
<th>Median</th>
<th>Highest value (g/m²/day)</th>
<th>Lowest value (g/m²/day)</th>
<th>Number of points</th>
</tr>
</thead>
<tbody>
<tr>
<td>2011A</td>
<td>165.9</td>
<td>9.38</td>
<td>5340.48</td>
<td>0.0134</td>
<td>459</td>
</tr>
<tr>
<td>2011B</td>
<td>159.31</td>
<td>10.96</td>
<td>7210.54</td>
<td>0.1474</td>
<td>449</td>
</tr>
<tr>
<td>2011total</td>
<td>162.64</td>
<td>10.32</td>
<td>7210.54</td>
<td>0.0134</td>
<td>908</td>
</tr>
</tbody>
</table>

The greatest difference is in the highest value. Such high values are rarely observed in the Reykjanes geothermal area and of the 908 nodes measured, only 17 exceeded 2000 g/m²/day. As before, semi-variograms, variogram models (see Figure 33) and 100 sequential Gaussian simulations were performed for both datasets 2011A and 2011B and for the 2011total. For 2011A the nugget is 0.4, the sill is 1 and the range is 190, for 2011B the nugget is 0.55, the sill is 1 and the range is 190 and for the 2011total, the nugget is 0.52, the sill is 1 and the range is 175.

![Variogram plots](image)

**Figure 33** Variograms for 2011A, 2011B and the 2011total dataset are shown. On the x-axis there is the distance in meters and on the y-axis the γ - value (nugget). The line with red dots shows results for measured points from the dataset but the yellow line is the result of the variogram model made from each dataset.
In the variograms in Figure 33 it can be seen that the nugget for 2011A is 0.4 but for 2011B it is a little higher or 0.55 which means that the anomalies in 2011B are not as well defined as in 2011A. 100 realizations were considered for each dataset and the results are shown on maps in Figure 34 to Figure 36. The map from 2011A has better defined anomalies than the one from 2011B which is consistent with a lower nugget value for the 2011A dataset. In 2011A, there are low-flux areas in edges of the study area, especially in the southern and south-western parts but also in the eastern part. The anomaly in the middle part of the area and stretches south-west from the Gráa lónið lagoon, close to well RN-2 is slightly larger in 2011A than in 2011B. The map based on the 2011 total data set shows more details of the anomalies with smaller nuances appearing like a weak anomaly with a NE-SW direction that lies next to the activity at the Gráa lónið lagoon in the north-western part of the area (Figure 36). Areas with strong CO₂ flux anomalies (red areas on maps) are larger in 2011 than in 2010, especially around well RN-2 and south of it. Further south, around well RN-4 and in the SW part of the study area, there is no increase in CO₂ flux visible; on the other hand it seems to be a little smaller than in 2010. The areal extent to the east and southeast is similar in 2010 and 2011.

![Figure 34 CO₂ flux based on 100 realizations from the 2011A dataset.](image-url)
Figure 35  $CO_2$ flux based on 100 realizations from the 2011B dataset.

Figure 36  $CO_2$ flux based on 100 realizations from the 2011 total dataset.
To reveal the difference between each dataset and the total dataset, 2011A and 2011B were subtracted from the 2011total. The resulting images show two different patterns. When 2011A was subtracted from 2011total it is clear that the 2011A overestimated CO₂ flux values in many places within the warmest areas, e.g. around Gunnuhver and also in the middle part of the area (blue areas) and underestimated the flux in the southwest area. For the 2011B dataset, this is in some contrast with the previous image. The outermost areas to the west and southwest show a blue colour, indicating that the 2011B values there were higher than the 2011total but in the area around Gunnuhver, it shows a red colour.

The total CO₂ flux from the measured area, shown in Figure 34 to Figure 36 was calculated and compared. The total CO₂ flux from 2011A is 35.3 ± 4.9 tons/day, about 7.6% less than the total CO₂ flux from 2011total which was 38.2 ± 4.2 tons/day. Despite this 7.6% difference the values are not statistically different, due to the uncertainty. For 2011B, the value is almost the same as the total dataset or 38.1 ± 5.3 tons/day (0.4% difference). It is worth noting that due to rather high uncertainty (at 95% confidence level) the values for these three datasets are all very similar and well within the error bars so they are not statistically different. It is also worth noticing that even with the 2011total dataset; it does not give results with much lower uncertainty. As for 2009 and 2010, 1000 realizations

Figure 37 To the left, the resulting map when 2011A had been subtracted from 2011total and to the right, the resulting map when 2011B had been subtracted from 2011total. A blue colour shows areas where 2011A values (left) or 2011B values (right) were higher than the 2011total values. A red colour shows areas where the 2011total values were higher than the other dataset values and white areas did not show a great variation between the two datasets subtracted.
were used to calculate the total CO$_2$ flux and the uncertainty, to minimize any random effects.

Table 13 Total CO$_2$ flux in tons per day for the 2011 datasets, the uncertainty is evaluated at 95% confidence level.

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Total CO$_2$ flux (tons/day)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2011A</td>
<td>35.3 ± 4.9</td>
</tr>
<tr>
<td>2011B</td>
<td>38.1 ± 5.3</td>
</tr>
<tr>
<td>2011total</td>
<td>38.2 ± 4.2</td>
</tr>
</tbody>
</table>

Figure 38 Histogram of the total mass flow of CO$_2$ from the 1000 simulations done for the 2011total.

Figure 38 shows a histogram displaying the results from the 1000 simulations in terms of the total mass flow of the CO$_2$ through soil resulting from each simulation. The uncertainty of the total CO$_2$ flux is calculated from the results of the 1000 simulations and it is 38.2 ± 4.2 tons day$^{-1}$ (at 95% confidence level). The 2011A and 2011B datasets both give the total
CO₂ flux well within the confidence level and the difference is therefore not statistically significant.

### 4.2.4 Total CO₂ flux from the Reykjanes geothermal area from 2004 to 2011

The total CO₂ flux has been calculated for each annual dataset since 2004. Numbers of equiprobable sGs realizations are used to evaluate the uncertainty, 100 for the datasets from 2004-2008 and 1000 for the datasets from 2009-2011. The same procedure was used for the total CO₂ flux determination as for the heat flow data (described in section 4.1.6). The area of each year’s measurement coverage was used to evaluate the total CO₂ flux for each year but for comparison, the same two areas were used, the tight comparison area (TCA) and the (LCA). For unmeasured areas the background value of 4.1 g/m²/day was used to represent the CO₂ flow and used to compensate for the areas not studied. This background value was estimated and calculated by Fridriksson et al. (2006) for the Reykjanes geothermal area. The results are presented in Table 14.

<table>
<thead>
<tr>
<th>Year</th>
<th>Numbers of realizations</th>
<th>CO₂ in tons/day from TCA</th>
<th>CO₂ in tons/day from LCA</th>
</tr>
</thead>
<tbody>
<tr>
<td>2004</td>
<td>300</td>
<td>13.5 ± 1.7</td>
<td></td>
</tr>
<tr>
<td>2005</td>
<td>100</td>
<td>11.9 ± 2.9</td>
<td>14.3 ± 3.0</td>
</tr>
<tr>
<td>2006</td>
<td>100</td>
<td>15.3 ± 2.8</td>
<td>16.9 ± 2.8</td>
</tr>
<tr>
<td>2007</td>
<td>100</td>
<td>16.6 ± 2.2</td>
<td>18.7 ± 2.2</td>
</tr>
<tr>
<td>2008</td>
<td>100</td>
<td>6.6 ± 0.6</td>
<td>8.2 ± 0.6</td>
</tr>
<tr>
<td>2009</td>
<td>1000</td>
<td>17.2 ± 1.9</td>
<td>21.6 ± 2.3</td>
</tr>
<tr>
<td>2009A</td>
<td>1000</td>
<td>20.3 ± 1.6</td>
<td>26.0 ± 1.8</td>
</tr>
<tr>
<td>2009B</td>
<td>1000</td>
<td>15.5 ± 1.9</td>
<td>19.9 ± 2.2</td>
</tr>
<tr>
<td>2010</td>
<td>1000</td>
<td>27.6 ± 4.3</td>
<td>34.4 ± 4.8</td>
</tr>
<tr>
<td>2010A</td>
<td>1000</td>
<td>25.2 ± 5.2</td>
<td>33.5 ± 5.6</td>
</tr>
<tr>
<td>2010B</td>
<td>1000</td>
<td>28.6 ± 6.8</td>
<td>36.6 ± 8.2</td>
</tr>
<tr>
<td>2011total</td>
<td>1000</td>
<td>32.8 ± 3.8</td>
<td>36.6 ± 3.9</td>
</tr>
<tr>
<td>2011A</td>
<td>1000</td>
<td>31.1 ± 4.8</td>
<td>34.0 ± 5.1</td>
</tr>
<tr>
<td>2011B</td>
<td>1000</td>
<td>31.7 ± 4.6</td>
<td>37.1 ± 4.9</td>
</tr>
</tbody>
</table>

Table 14 The CO₂ flux from the Reykjanes geothermal field derived from soil temperature measurements, 100 sequential Gaussian simulations and the uncertainty evaluated using the simulations. Note that no data is available for 2005 and 2006.
4.3 Relationship between soil temperature / heat flow and CO$_2$ flux from the soil in Reykjanes

A number of studies (e.g. Brombach et al. 2001; Lewicki et al. 2003; Granieri et al. 2010) have shown that the CO$_2$ flux in many geothermal areas correlates strongly with soil temperature. The annual measurements in Reykjanes since 2004 have provided a large dataset of over 3800 measurements on soil temperature and CO$_2$ flux from the same geothermal area. This total dataset was used to explore the relationship between these two parameters in the Reykjanes geothermal area. The CO$_2$ flux as a function of soil temperature at 15 cm depth is shown in Figure 40. This did not reveal a strong relationship, and as seen on the figure. For instance, the observed CO$_2$ flux where the soil temperature is 100°C ranges from 0 to more than 7000 g/m$^2$/day.
Both parameters, the soil temperature and the CO$_2$ flux appear to be very scattered and this did not reveal any strong correlation. The reason for the poor correlation is probably interaction between cold groundwater and steam, where thermal energy in connection with ascending steam condensation is transported laterally out of the system by groundwater as reported by Fridriksson et al. (2006). Hence, direct measurements of the CO$_2$ emission from the area, without information on the amount of CO$_2$ dissolved in the groundwater, may be considered a minimum value for the total emission from the geothermal system.

Both parameters were filtered, using a 200 m x 200 m mean filter and then, these filtered datasets were plotted against each other. This figure, Figure 41, obviously shows a better relationship. It indicates that there is a strong correlation between soil temperature and CO$_2$ flux on regional scale, where looking at larger areas than the point measurements and indicates what is visually observed from mapping the distribution of these two parameters, i.e. that the anomalies or areas with intense surface activity are similar.
To summarize the calculation of heat flow and carbon dioxide flux, the total heat flow from 2004 and 2007-2011 was plotted vs. total CO$_2$ flux for the same period (see Figure 42). The linear fit added to the graph shows a rather scattered but rising trend with the values for the years 2004 but 2008 by far the lowest and the year 2007 stands out with a high value of heat flux (but very large uncertainty).
5 Discussion

5.1 Methods for monitoring soil temperature and CO₂ flux

5.1.1 The TIR image and its relationship with surface temperature, heat flow and CO₂ measurements

The TIR image from 2011 showed a very detailed picture of the surface temperature and provided an excellent record to compare with the TIR image from 2004. The comparison shows without any doubt that the surface temperature has increased in most parts of the area, especially in the middle part and then north of the Gráa lónið lagoon. The average temperatures derived from these images also indicate the difference. The distribution of surface temperature, obtained by snowmelt tracking in March 2011 fitted very well with the TIR image from May 2011 and the outlines of the elevated soil temperature are very similar in most cases. The soil temperature measurements (at 15 cm depth) done in June 2011 showed a very weak relationship with the TIR image when the temperature from the TIR image in the same coordinates are compared to the soil measurements (see Figure 15). However, when a map of soil temperature has been made from the soil measurements, the distribution is rather similar when visually compared with the TIR2011 image.

Margrétardóttir (2005) used heat flow derived with the Dawson method and d⁹⁷ from soil temperature measurements (QT₁₅/Qd⁹⁷) to assess the heat flow from the total coverage area of the 2004TIR image from Reykjanes. A linear regression fit was found by plotting the heat flow QT₁₅ / Qd⁹⁷ as a function of surface temperature from the TIR2004 image (T₀) and resulted in the following equation

\[ Q_w = 17.982T_0 - 44.411 \]

Equation 5

where \( Q_w \) is the estimated heat flux in W m⁻² based on fit through values of the whole measurement grid and \( T_0 \) is the surface temperature at any pixel in the image. The plot was rather scattered with \( R^2 = 0.29 \). This fit was applied to each pixel value of the TIR2004
image. It resulted in a total heat flow value of 17.3 MW which was only 1.5% higher than the heat flow determined by the soil temperature measurements from 2004 (Margrétardóttir, 2005). Even though she obtained some negative values which are of course unrealistic, no attempt was made to correct these values since they had little effect on the total heat flow estimate. Assuming that the correlation between surface temperature obtained by TIR technology and heat flow estimated from soil temperature measurements has not changed in the Reykjanes geothermal area since 2004, this equation was used to determine heat flow from the TIR2011 image, using the same methods.

For the TIR2011 image, two different areas were chosen. First, an area of the same size as the 2011 soil measurements were carried out was chosen to be able to compare these two different methods and secondly, a larger area was observed to include the area north of the Gráa lónið lagoon. The total heat flow from the 2011 study area (the south-part) was estimated by summarising the values of all the pixels within the area. As for the data from 2004 the heat flow derived from the TIR image gave higher values than the soil measurements but the difference between these two methods was larger in 2011 than in 2004. For TIR2011 the heat flow resulted in a value of 42.5 MW which is almost 18 % higher than the QT_{15}/Q_{d97} heat flux value which was 36.1 ± 2.5 MW. It is not possible to calculate the uncertainty for the TIR2011 image in the same way as for the QT_{15}/Q_{d97} heat flux evaluation but when considering the low R^2 value for the fit used for the TIR image it is certain that there is an uncertainty included in this value so this 18% difference between these two evaluations should be interpreted with caution.

The non-isolated pipeline in the north-part was excluded from the heat flow evaluation due to its unnatural origin. The heat flow from the north-part derived from the TIR2011 image using Equation 5 resulted in a value of 3.1 MW in 2004 but 11.4 MW in 2011. This indicates that heat flow in the north-part has more than tripled between 2004 and 2011. For 2011, the value corresponds to~ 25 % of the heat flow of the south-part obtained by soil measurements indicating that the increase during this period would obviously be greater if this part were included.
Figure 43 The heat flow from the study area based on the fit of surface temperature and the estimated heat flux using Equation 5 (Margrétardóttir, 2005).

To obtain an order of magnitude for the CO₂ flux that might be emitted from the north-part, a rough estimate was done. It was based on results from the master’s thesis of Margrétardóttir (2005). She found a fit by plotting CO₂ (g m⁻² day⁻¹) vs. of surface temperature from the TIR2004 image. Despite a wide scattering of the data and a very poor correlation (R²=0.04) she came up with an estimate of the total CO₂ emission which was only 3.3% lower than the value determined from the direct measurements of CO₂ flux from 2004. As for the heat flow it is assumed here that the correlation between CO₂ flux and surface temperature value has not changed in Reykjanes between 2004 and 2011 and the equation from Margrétardóttir (2005) is used:

Equation 6

\[ CO_{\text{TIR}} = 5.7422T_0 + 23.291 \]
where the $CO_{TIR}$ is the estimated CO$_2$ emission in g m$^{-2}$ d$^{-1}$ and $T_0$ is the surface temperature at any point in the calibrated TIR image (Margrétardóttir, 2005). This was used to recalculate the values of the TIR 2011 image and resulted in an estimate of 5.7 tons/day of CO$_2$ for the north-part that lies north of the Gráa lónið lagoon which is about 15% of the estimated total CO$_2$ flux derived from the soil measurements. The weak correlation between soil and surface temperature and gas flow is probably due to interaction between cold groundwater and steam, where thermal energy in connection with ascending steam condensation is transported laterally out of the system by groundwater as reported by Fridriksson et al. (2006).

The method of using TIR images to monitor changes in geothermal areas does have its limitations. Such images only represent the radiant temperatures, which is a function of the kinetic temperature and the emissivity. The emissivity differs between different objects and for the TIR images from Reykjanes, correcting for different emitting ability would be possible by dividing the area into small pixels and determine an emission factor representative for each pixel. Still due to the similarity of the emissivity of water and soil types in Reykjanes, this was not done here and considered to have little effects on the results. There is a weak correlation between soil and surface temperature because the natural heat discharge in Reykjanes is not only through radiation which makes quantification of heat flow from the Reykjanes geothermal area based on TIR images uncertain. It is also evident that other factors than soil temperature determine the flow of CO$_2$ from the system in the study area in Reykjanes. Using surface temperature obtained from TIR images to estimate natural gas flow from the area is therefore problematic. However, as has been shown here, repeated TIR images that are obtained from the same area offer a great possibility to compare possible changes in surface temperatures in geothermal or volcanic areas.

### 5.1.2 The effects of grid spacing on the results of soil measurements

The CO$_2$ flux and soil temperature measurements are done annually in Reykjanes to monitor changes in the surface activity in the area and the system’s reaction to the 100 MW$_e$ power production. The primary purpose is to estimate the volume of the total flow of CO$_2$ and heat flux from the Reykjanes geothermal area and secondly to map the
distribution of CO₂ flux and soil temperature in the area and monitor possible changes. When measuring parameters on a grid, the grid spacing can have critical effects on the measurements’ reliability. The grid spacing needed depends on the size of the anomalies and how reliable data is required. Visual comparison of CO₂ flux in the divided datasets from 2009 (2009A and 2009B) and especially 2010 (2010A and 2010B) shows that by using grid spacing with 30-50 meters, it can result in misleading gas flux anomalies depending on the arbitrary values and their locations. However, to be able to quantify the total CO₂ flux, the divided datasets seem to supply more similar results. For the datasets from 2010 the difference is not statistically significant but for the datasets from 2009 the total CO₂ flux estimate is not as consistent as the 2009A and 2009B datasets yield statistically different results.

The experiment in 2011 of making two datasets with a grid spacing of 25 x 25 meters each did reveal that for the CO₂ flux measurements the distribution from the two datasets are rather similar. They do not show exactly the same details but the distribution and extension is similar so it is concluded that 25 x 25 meters grid spacing is tight enough to obtain the CO₂ flux distribution accurately enough for the purpose of mapping the extension of the flux anomalies for the Reykjanes geothermal area. By using the TIR2011 as a proxy for the real temperature distribution and visually comparing the divided datasets from 2011 it is concluded that when the temperature anomalies are approximately 50 m in diameter they appear on the soil temperature distribution maps with a 25 x 25 m grid spacing but when the anomalies are less than 30 m in diameter they do not appear strongly on the soil temperature maps. This indicates that the grid spacing cannot be more than half of the size of measured anomalies. The total CO₂ flux estimation from the different datasets (2011A, 2011B and 2011total) did all give results that are statistically similar. It is concluded that the 2011total dataset did not supply a statistically better total CO₂ flux estimate nor significantly better results for mapping the distribution even though more details are seen in the 2011total than previously experienced. The two temperature datasets (2011A and 2011B) give a very similar distribution, both of which correspond well to the 2011total temperature map (see Figure 16 and Figure 17).

The heat flow from the Reykjanes geothermal area has been estimated using the Dawson method which uses the soil temperature to calculate heat flow with certain equations. Due
to irregularly spaced grids since 2005 due to access difficulties in the Reykjanes geothermal area, the sGs method was used to take the location on each node into account. When deriving heat flow from the two different datasets using the Dawson method, and 100 sGs, the difference between 2011A and 2011B is of the order of ~15% but the results are not statistically different and both correspond to the value derived from the 2011total. The Dawson method is sensitive to high values, especially values higher than 97°C at 15 cm depth and the depth in cm to 97°C and for dataset 2011A there are 49 measuring nodes with higher values than 97°C while 2011B has 61 such nodes. This difference of the two datasets indicates that the 25 x 25 meters grid spacing used in previous years is not tight enough to exclude variance of the order of at least 10% and that a tight grid and regularly spaced is needed when using the Dawson method to evaluate heat flow. When using sGs to calculate the uncertainty of the heat flow calculations it is clear that an uncertainty of the order of 7% to 25% cannot be excluded with measurements carried out on a 25 m x 25 m grid spacing. The tight dataset, 2011total gives a slightly lower uncertainty value than most of the other datasets indicating that such a tight dataset might result in a slightly better estimate of heat flow but as has been shown here the benefit of the tight 2011total dataset, which is twice as tight as the 25 x 25 m dataset, is small.

5.2 Changes in surface activity in the Reykjanes geothermal area

5.2.1 Temperature variations on a short term scale

The continuous temperature loggers show clear temperature changes on a short term scale and most of the variations can be correlated to precipitation and diurnal variations. The diurnal variations appear most prominently from period II, or during the summer when the annual soil measurements are done. The variations are of the order of 1.5 to 6°C but the smallest variations are obtained from loggers located in the warmest soil. The diurnal variations in soil temperature higher than 80°C at 15 cm depth in the Reykjanes geothermal area is unknown since the temperature loggers used here are not made for such high temperatures but it is assumed that the diurnal variations do not have a significant effect. To estimate the effects of the diurnal variations on the heat flow in Reykjanes two cases were considered. First, assuming that all temperature values in the 2011 dataset
would be 3°C too high because daytime measurements only would lower the total heat flow estimate of 2.5 MW but it is unrealistic to assume that the high temperature values, even nodes where boiling takes place, so that a difference of the order of 2.5 MW is an overestimate. Secondly, assuming that all temperature values lower than 50°C in the 2011 dataset would be 6°C too high (the maximal diurnal variations) would lower the total heat flow estimate to the order of 0.5 MW which is well within the uncertainty of the total heat flow from 2011 which is 36.1 ± 2.5 MW. In Equation 2, the estimated heat flow depends on the temperature value to the fourth power which means that the calculations are very sensitive to high temperature values. It would be possible to minimize these effects by measuring the temperature at a greater depth than 15 cm or by correcting for the diurnal variations for the time of day.

The correlation with precipitation maximizes with accumulated precipitation for four days, indicating that the effects of the precipitation can last for a few days. The reason for this long duration of precipitation effects might be due to the type of soil. As mentioned previously the loggers were all covered with dense clay and mud which gets even denser during precipitation and can take long to dry up again. It is known that CO$_2$ soil flux is strongly influenced by external factors, such as soil temperature and the amount of rain (Granieri 2003 et al.) and in order to try to minimize these effects on the CO$_2$ flux and soil temperature measurements the measurements were only carried out when at least 24 hours had gone without any rain. Considering the information from the loggers’ temperature, 96 hours from rain would provide the optimal conditions for measuring CO$_2$ flux and temperature but due to weather conditions on the Reykjanes tip it is unfortunately very unlikely to expect such circumstances to occur and remain long enough for 5-10 days of measurements every summer.

For the soil temperature data obtained from the loggers, the variations cannot be completely explained by precipitation. Their strong correlation to each other would indicate some parameter that can cause the temperature to fluctuate for up to 10°C. The loggers in the geothermally heated ground where the temperature is rather high (> 40°C) show some events that do not correlate with precipitation or any other weather related parameters. These events are characterized by significant temperature drops that occur on a very short time scale during which the temperature drops about 10-30°C in a few hours and
then warms up again as quickly as it drops. These events never appear at the same time in more than one logger. The heat flow from the heat source of the geothermal area to the surface depends largely on regional fissures and cracks but it also depends on micro scale cracks. These temperature events could be caused by microscopic changes in the pathways of the steam flow in the uppermost meters of the soil.

5.2.2 Soil temperature, heat flow and CO$_2$ flux changes and the utilization of the Reykjanes geothermal area

The natural CO$_2$ emissions from the Reykjanes geothermal system were quantified prior to the installation of the 100 MW$_e$ power plant in May 2006 to allow for the evaluation of possible changes in natural CO$_2$ emissions as a result of extensive production in the area. In 2004, the total CO$_2$ emission through soil and vents, fractures and pools was evaluated equal to $5.1 \times 10^6$ kg per year with most of the CO$_2$ emitted through soil or 97.4%. The natural emissions from Reykjanes were very small or of the order of magnitude lower when compared to other measured geothermal and or volcanic systems in Iceland. The natural emissions from Reykjanes have almost tripled during this 8 year period and in 2011 the emission through soil is estimated $1.4 \times 10^7$ kg per year. It is still a small portion of the total estimate of CO$_2$ emissions from Iceland which Ármannsson et al. (2005) and Gíslason (2000) estimated of the order of $2.1 \times 10^9$ tons per year.

It is known that changes in the behaviour of geothermal systems and their surface activity can occur as a result of the utilization of a geothermal system and the development of geothermal energy has some impacts on the environment. The main environmental effects of geothermal development are related to surface disturbances, physical effects of fluid withdrawal, heat effects and discharge of chemicals (Kristmannsdóttir and Ármannsson, 2003). Exploitation of liquid-dominated high-temperature geothermal systems involves withdrawal of large volumes of geothermal fluid. A major consequence of the mass loss is a formation or rapid growth of a steam-water phase zone or boiling zone in the upper part of the reservoir and as the production continues this zone increases in size and the pressures in and below decreases causing increased boiling and degassing of the system (e.g. Goff and Goff, 1997; Hunt, 2001; Kristmannsdóttir and Ármannsson, 2003; Scott et
This pressure drop in the reservoir is an important cause of environmental changes at or near the surface (Glover et al., 2000; Hunt 2001; Kristmannsdóttir and Ármannsson, 2003). One of the consequences at the surface is that when the pressure declines, so does the amount of geothermal liquid reaching the surface resulting in a decline in the activity of geothermal manifestations such as geysers, hot springs and mud pools, e.g. in Wairakei and Ohaaki, New Zealand, Larderello, Italy and at The Geysers, USA (Hunt 2001). With increased size of the boiling zone the upward and lateral flows of steam into the overlying zones becomes stronger and the steam passes through fractures that had previously been filled with liquid resulting in an increase in the heat loss from areas of steaming ground (Allis, 1981; Hunt et al. 2009). Worldwide there is not much experience of how geothermal areas recover from the effects of geothermal exploitation but data from the Rotorua geothermal field in New Zealand suggest that some natural thermal features may recover from the effects of geothermal exploitation but other may not (Scott et al. 2005).

The effects of the production from the Reykjanes reservoir have been documented with results of pressure measurements in boreholes. Jónsson and Björnsson (2011), presented pressure changes in the Reykjanes system, measured at 1500 m depth below sea level in boreholes in Reykjanes since 2002. Prior to the production and until spring 2006, the pressure was around 122 bar-g but has dropped to 85 bar-g in 2011. This pressure drop is seen in the boreholes that are located in the most active part of the system. The pressure drop has clearly occurred fastest during the first month after the commissioning of the plant but the reduction has slowed down a lot since 2008 (see Figure 44).
Figure 44 Results of pressure measurements from wells in Reykjanes from 2002-2011 from Jónsson and Björnsson (2011).

While the pressure is dropping, the boiling level migrates downwards so the volume of rock where boiling takes place increases resulting in increased steam production in the system. This can result in more steam flow to the surface and a wider distribution of geothermally affected soil. A simplified image of possible circumstances is shown in Figure 45 and Figure 46.
Figure 45  *A simplified image of possible conditions in a geothermal system prior to geothermal development. The red zone marks the zone of boiling.*

Figure 46  *During geothermal utilization, lowering of water level leads to the formation or acceleration growth of a steam pillow or a boiling zone and subsequent degassing in the area.*

The continuing pressure drawdown, noticeable in the production well data from Reykjanes is likely to have effects on surface activity within the geothermal area and the relationship to the power plant could be seen already in the surface measurements from 2006, right
after the commissioning of the power plant. The increased surface activity in the area has been obvious to visitors, new mud pits split the road south of Gunnuhver and the changes have called for the ebuilding of tourist paths. The total CO₂ flux through soil and heat flow calculated from soil temperatures measured at 15 cm depth show that both parameters have at least doubled between 2004 and 2011. Still, this has to be considered as an underestimation since the area north of the Gráa lónið lagoon appears to have warmed up significantly during this period, as seen from the TIR images but this has not been included in the measurement grid.

A similar development has been seen in Wairakei geothermal system, New Zealand. During the first decade of operation there, a pressure drawdown of up to 20 bar developed. This pressure reduction resulted in widespread boiling and formation of segregate steam zones at the top of the reservoir. From 1975 to 1997 pressures in the deep liquid reservoir stabilized at 23–25 bars (2.3–2.5 MPa) below the original pressure. Areas with springs and geysers decreased and died but heat flow increased rapidly in thermal areas, during the first decade of production which resulted in an expansion of the area of thermal ground and the centres of the thermal activity appear to migrate randomly (Glover and Mroczek, 2009; Hunt, 2009). During the period from 1954 to 1964 the heat flow increased from 40 to 420 MW but declined in 1978 and stabilized at about 220 MW (Hunt et al. 2009) consistent with the pressure in the deep liquid reservoir. Springs and geysers in the Wairakei area have not increased in activity or appeared again.

A decline or disappearance of surface manifestations such as fumaroles and mud pools has not occurred in Reykjanes geothermal area after the commissioning of the power plant in 2006 even though soil temperature in some minor spots has dropped; according to the comparison of the TIR images from 2004 and 2011 (see Figure 13). However, the Reykjanes geothermal area has been known for its great variations in surface activity for the last 150 years and three periods of remarkable sea-water geysers are known from there. At least two of these geyser periods started after seismic events, in 1926 and again in 1967 (Friðriksson et al. 2010). Such seismic events occur at a few decades’ intervals on the Reykjanes tip but due to the production in the area and its effects it has to be considered unlikely that new geysers appear when the next seismic event takes place.
The total heat flow through soil in the geothermal area in Reykjanes has increased since the start of the production from 16.9 ± 1.4 MW in 2004 to 34.3 ± 2.6 MW in 2011 which is by far greater than the uncertainty of the measurements, but the increase has not been even during this period. The measurements from 2007 and 2009-2011 do not necessarily indicate an increase, on the contrary when the uncertainty is considered an increase might be doubtful. The increase in CO$_2$ flux is more explicit. Prior to the production, in 2004, the total CO$_2$ flux was estimated 13.5 ± 1.7 tons day$^{-1}$ and it has constantly increased until 36.6 ± 3.9 tons day$^{-1}$ in 2011 except for 2008 (as discussed in section 5.2.3). In Reykjanes, the increase in CO$_2$ flux and heat flow is expected to slow down from year to year similar to what was seen in the Wairakei geothermal system and also when considering the pressure measurements from wells in Reykjanes that show a great slow-drowning pressure reduction from 2009-2011. Even though the pressure reduction in the wells has already decelerated, the same is still not obvious from the soil measurements, especially not in the CO$_2$ flux measurements. Still, one might possibly experience some delay in the system, the effects from the pressure drop could appear later close to surface than in wells and that would explain the continuing increase in the soil CO$_2$ flux. For better understanding the development of the CO$_2$ flux and heat flow and the effects of the utilization in Reykjanes continued measurements are needed.

5.2.3 Measurements and events in 2008

The great reduction in 2008 in both heat flux from 40.1 ± 10.8 MW in 2007 to 20.6 ± 2.7 MW in 2008 and CO$_2$ flux from 18.7 ± 2.2 tons day$^{-1}$ to 8.2 ± 0.6 tons day$^{-1}$ is not detectable in temperature data from boreholes in the Reykjanes area, on the contrary, the borehole temperatures from 2008 are in good agreement with the borehole temperature data from 2007 and 2009 (Jónsson and Björnsson 2011).
These changes in soil diffuse degassing and soil temperature cannot be easily correlated to the production in the power plant, since no such changes occurred in the production during this period. On May 29th in 2008, researchers were carrying out the annual measurements in the area measuring CO₂ flux and temperature through soil. They observed no signs of significant changes in the surface activity. That same day, at 15:45, an earthquake of magnitude 6.1 took place in the south Iceland seismic zone, Southwest Iceland (The Icelandic Meteorological Office, www.vedur.is, June, 2011). The earthquake was clearly felt in the Reykjanes geothermal area (Daði Þorbjörnsson, pers. comm., 2012) and it might have caused short term changes in the heat flow and the CO₂ flux which could have affected the measurements. On June 10th a crater had formed in the Gunnuhver area, in a spot where the dominant steam vent in the area was found previously. The crater was 10-15 meters in diameter but the depth could not be estimated due to a very intense steam flow from the crater that completely blocked the view down to it, see Figure 48. The crater was obviously formed in an explosion, with mud splashes and pieces of rocks covering the area around the crater and up to considerable distances (tens of meters). The soil next to the crater, especially on the east size showed decreased activity indicating that the crater might be channelling steam from its surroundings and since the crater itself has never been
included in the soil measurements due to its size and the impossibility of measuring it might cause a reduction in the total estimate of heat flow and CO$_2$ flux.

![Figure 48](image)

*Figure 48* Photo of the crater in Reykjanes from 2008. The sightseeing paths and platforms seen here have plunged into the geothermal mud due to increased activity in this part of the area (photo: Ellert Grétarsson).

A phenomenon, often called hydrothermal eruptions (also called “hydrothermal” or “phreatic explosions”) are known from high-temperature liquid-dominated geothermal fields (e.g. Scott and Cody, 1982; Marshall, 1987; Bixley and Browne, 1988; Bromley & Mongillo, 1994; Scott et al. 2005). Although rare, hydrothermal eruptions occur during the natural evolution of high-temperature geothermal systems and small, shallow focussed events have been induced in some exploited reservoirs (Bixley and Browne, 1988; Bromley and Mongillo, 1994). In the Wairakei field, New Zealand, hydrothermal eruption activity began or significantly increased after development of the field began in 1958 and similar occurrences have been experienced in many geothermal areas after geothermal development (Hunt, 2001; Goof and Goff, 1997). The explosions occur when the steam pressure in the near-surface aquifers exceeds the overlying pressure and the overlying
material is then ejected, generally forming a crater 5-500 m in diameter and up to 500 m in depth but most are less than 10 m deep (Bromley and Mongillo, 1994).

The mechanism of hydrothermal eruptions is similar to the mechanism that drives geysers, indeed there seems to be a range of related styles of eruptions extending from geysers through to major deep-seated activity. The induced eruptions are of short duration (hours at the most) and shallow focus (a few meters); ejected material travels only up to 100 m from the vent (e.g. Scott and Cody, 1982; Bixley and Browne, 1988). Their genesis has been explained as a result of shallow hydrology changes with an increased steam flow from the deep reservoir due to deep pressure drawdown and the creation of a steam zone allowing increased flow of steam to the surface. This crater is still prominent in the Reykjanes area and is still emitting a large amount of steam but the steam activity from the crater seems to be variable.
6 Conclusions

A thermal infrared image which was obtained in May 2011 from the Reykjanes geothermal area shows a detailed picture of the surface temperature distribution. This image provides excellent data to compare with a TIR image from April 2004, also obtained from Reykjanes. The comparison of these two images shows without any doubt that surface temperature has increased in large parts of the Reykjanes geothermal area. A warm area lying north of the Gráa lónið lagoon has not been included in the soil measurements in previous years so that these images are the only concrete data showing the increase in this area. Snowmelt tracks were recorded in March 2011 to map the distribution of surface temperature and these tracks appear to fit very well with the TIR image from 2011. The results of the soil temperature measurements show a weak relationship with temperature derived from the TIR 2011 image; however, they show a similar distribution when visually compared. The heat flow derived from the TIR2011 image results in a slightly higher value than the heat flow value derived from the soil measurements and can be doubted because of a weak relationship between soil and surface temperature but it may give an idea of the order of magnitude of the heat flow from the area. Repeated TIR images that are obtained from the same geothermal area with some time interval offer a great possibility to visually compare possible changes in surface temperatures.

For the extent of the elevated temperature and CO$_2$ flux anomalies in the Reykjanes geothermal area, a grid spacing of 30-50 m is not small enough to give reliable results of the distribution of these parameters. With a grid spacing of 25 x 25 m the distribution seems to give results that are not dependent on a random factor but when the measured anomalies are less than 30 m in diameter they do not feature strongly in these soil temperature maps. A dataset with a grid spacing of 17-20 m does not give different results when visually compared even though more details are seen in maps from such a dataset. To quantify the total CO$_2$ flux or heat flow, measurements on a 30-50 m grid spacing can give statistically similar results but that may be doubtful. A dataset with grid spacing of 25 m, the total estimate are statistically similar and dataset with a grid spacing of 17-20 m does not give a significantly better estimate of total CO$_2$ flux or lower uncertainty than the one with a grid spacing of 25 m.
The temperature loggers showed that precipitation lowers the soil temperature and these effects last longer than previously thought for Reykjanes. It emphasizes the importance of choosing the steadiest and driest weather conditions possible, to try to minimize such effects. Loggers located at relatively cold locations (< 25°C) show a correlation with air temperature. The loggers in hot ground also show unexplained behaviour involving sudden pressure drops of tens of degrees for a very short time and this behaviour cannot be related to weather parameters. These events are most probably caused by random changes in the pathways of the steam flow on a micro scale. These micro scale features can also cause very different temperature variations in loggers located very close to each other (within 10 meters), indicating that different processes are affecting the temperature. On the contrary locations further apart (>100 m) can show strong correlation indications of common macro scale processes influencing the soil temperature at these locations. The most dominating external parameter that affects the soil temperature in Reykjanes appears to be precipitation but at high soil temperature (> 40°C) the variations could not be correlated to other meteorological parameters.

The eight years of annual measurements of soil temperature and CO$_2$ flux in the Reykjanes geothermal area have shown an increased activity both in soil temperatures and in CO$_2$ flux. The CO$_2$ flux has increased from 13.5 ± 1.7 tons per day to 36.6 ± 3.9 tons per day according to the results of the soil measurements and the heat flow has increased from 16.9 ± 1.4 MW to 36.1 ± 2.5 MW from 2004 to 2011 according to the results of the the soil measurements. The distribution of soil temperature and CO$_2$ flux anomalies has changed during that time, it stretches over a wider area, especially to the south and southeast, and an area which did not show elevated temperature values in 2004 and 2005, right north of the Gráa lónið lagoon, has appeared warm in 2011. The distribution variations seen from year to year in the area south of Gráa lónið lagoon seem to migrate randomly rather than being extending further in any certain direction.

These changes can be related to geothermal power plant production even though changes of this order of magnitude are known to have occurred in Reykjanes without any utilization (Fridriksson et al., 2010). Exploitation involves withdrawal of geothermal fluid and thus causes a rapid pressure drop and a major consequence of the mass loss is the formation of a
boiling zone in the upper part of the reservoir. With an increased steam-water phase zone steam flows more easily up through the surface due to its increased accessibility to pathways resulting in increased heat flow and CO₂ emissions. Pressure changes have been observed in wells in Reykjanes and have shown that pressure has not decreased much since 2009 but the CO₂ flux during that time has increased greatly and far exceeding the measurement’s uncertainty. The key to better understanding of the development of CO₂ flux and heat flow and the effects of utilization is to continue to obtain data from the area. Prominent surface features have not disappeared but the remarkable sea-water geysers that are known from the Reykjes geothermal area had already disappeared when utilization started. A hydrothermal eruption likely took place in the Reykjanes geothermal area at the end of May or beginning of June 2008, possibly related to an intense earthquake.

To be able to evaluate and quantify changes in geothermal areas in their natural state or due to utilization regular measurements are required. As has been seen here, annual soil measurements of temperature and CO₂ flux on a measuring grid, which has to be determined with respect to the extent of the anomalies, is essential and give quantitative results. However, these measurements have to be performed in dry and stable weather conditions and the resolution is of the order of meters or tens of meters. Thermal infrared images on the contrary can give very detailed information in high resolution on the temperature distribution and when combining these methods with snowmelt tracking, an overall state of the surface activity is well identified making it possible to evaluate changes in a geothermal area.
Appendix A

Soil temperature at 15 cm depth in 2004 from the Reykjanes geothermal area. The numbered marks show boreholes in the area.

Soil temperature at 15 cm depth in 2005 from the Reykjanes geothermal area.
Soil temperature at 15 cm depth in 2006 from the Reykjanes geothermal area

Soil temperature at 15 cm depth in 2007 from the Reykjanes geothermal area
Soil temperature at 15 cm depth in 2008 from the Reykjanes geothermal area

Soil temperature at 15 cm depth in 2009 from the Reykjanes geothermal area
Soil temperature at 15 cm depth in 2010 (Óladóttir et al., 2010)
Appendix B

$\text{CO}_2$ flux in 2004 from the Reykjanes geothermal area

$\text{CO}_2$ flux in 2005 from the Reykjanes geothermal area
$CO_2$ flux in 2005 from the Reykjanes geothermal area

$CO_2$ flux in 2006 from the Reykjanes geothermal area

$CO_2$ flux in 2007 from the Reykjanes geothermal area
CO₂ flux in 2008 from the Reykjanes geothermal area

CO₂ flux in 2009 from the Reykjanes geothermal area
$CO_2$ flux in 2010 from the Reykjanes geothermal area
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