The Thickness of the Seismogenic Crust in Iceland and its Implications for Geothermal Systems

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ABSTRACT

Iceland is one of the most favorable places in the world for geothermal energy production, thanks to favorable geological conditions. The location of Iceland on an intersection between the Mid-Atlantic Ridge and the Iceland hot spot create the unique geological condition for geothermal activity in the country. It is known that the hydrothermal activity extends at least down to 3 km depth but the lower limits for utilization is still not known. The background temperature gradient in Iceland in the uppermost 1,5 km is ranging from 40 to 150°C and shows no sign of a general decrease with depth. Linear extrapolation of these gradients leads to partial melt at 8 – 30 km depth.

In order to get information on the temperature depth relationship below the depth of the presently utilized geothermal systems, we have analyzed the relationship between the maximum focal depth of earthquakes and the near surface temperature gradient. This is done by analysis of the SIL earthquake catalogue, but the Icelandic Meteorological Office runs the SIL seismic network. The seismic data cover the period from 1991 to June 2003. Of these around 108000 events met our requirement regarding maximum error estimate in the depth determination. By comparison of the temperature gradient, the maximum focal depth of earthquakes and available estimates of the temperature at the bottom of the seismogenic crust worldwide, we conclude that the temperature gradient in Iceland must decrease considerably with depth below the depth penetrated by boreholes. We suggest two possible explanations, increased thermal conductivity with depth or heat production by repeated intrusions into the lower part of the brittle crust. The proposed increase in thermal conductivity could partly be due to reduction in porosity with depth in the upper crust and partly due to increased share of mantle material in the lower crust. For possible deep geothermal systems in Iceland this means that temperature at 3-5 km depth are lower than previously was expected but still very high and exploitable by deep holes.

1. INTRODUCTION

The crust in Iceland is considerably thicker than on the ocean ridges, 19-40 km compared to less than 10 km on the ocean ridges (Kaban et al. 2002) . The background temperature gradient in Iceland decreases with increasing age of the crust from 100-150°C/km close to the ridge axis to 40-50°C/km in the 10-15 my. old crust (Flóvenz and Sæmundsson, 1993). These values are mostly based on shallow boreholes, usually 50-100m deep outside areas of known geothermal activity but occasionally supported by up to 1,5 km deep holes. However practically no direct measurements of heat flow are available from the neovolcanic zone since the uppermost 1-1,5 km of it consists of flow that removes all the conducted heat from below. Boreholes and geothermal utilization is still limited to the uppermost 3 km of the crust and the temperature conditions below that are poorly known and highly disputed. The heat flow and geothermal processes in Iceland have been described in many articles e.g. Pálmason (1973), Bödvarsson (1982) and Flóvenz and Sæmundsson (1993). The general model of the geology in Iceland includes an interaction between the North Icelandic hot spot with mantle plume center under the eastern part of Central Iceland and the spreading axis of the Mid Atlantic Ridge (MAR), that crosses the country from SW to NE, forming the active zone of volcanism and rifting, the neovolcanic zone (e.g. Sæmundsson, 1979, Pálmason 1973, Kaban et al 2002, Wolfe et al 1997). Due to proposed eastward movement of the mantle plume relative to the plate boundary the neovolcanic zone in Iceland has repeatedly jumped to the east, away from the natural prolongation of the MAR axis, creating transform faults in the south Iceland lowlands, known as the South Iceland Seismic Zone (SISZ) and along northern coast, known as the Tjörnes fracture zone (TFZ). The seismically active areas are mainly restricted to the plate boundaries and the greatest activity is in the transform zones where the largest earthquakes occur. In the neovolcanic zone the seismic activity is usually associated with central volcanoes. Considerable seismic activity is also in the western volcanic zone (WVZ). That is a zone of rifting and recent volcanism trending NE and joins the SISZ and the active zone along the Reykjanes peninsula in a triple junction in an area called Hengill.

There are two main types of geothermal fields in Iceland, the high temperature (HT) fields, that are related to volcanic centers in the neovolcanic zone and the low temperature (LT) fields which are almost exclusively found to the west of the present active rift zone, i.e. on the American plate. Both types are characterized by fracture permeability in rocks of otherwise very low permeability as well as active fluid convection in the fracture system. The heat sources of the HT fields are believed to be magmatic intrusions at shallow levels in the volcanic system while the heat source of the LT fields seems to be the general high background heat flow.

The nature and the temperature of the earth below Iceland have been subject to fruitful discussions among earth scientists. Temperature of 1000-1200°C would be reached at 10-25 km depth if the surface temperature gradients are linearly extrapolated (Pálmason 1973, Beblo and Björnsson 1980, Flóvenz and Sæmundsson 1993, Flóvenz, 1992). This implies that a layer of partial melt should exist at these depths. This interpretation is strongly supported by the presence of low resistivity layer at these depths revealed by MT measurements (Beblo and Björnsson 1980, Eysteinsson and Hermance, 1985). In contrast to this, it has been proposed on the basis of seismic wave propagation and especially the high Q values observed, that the crust in

Iceland is much cooler, probably just around 450°C and the hypothesis of a widespread partial molten layer is abandoned (Menke and Sparks, 1995, Menke et al., 1997). In that case the relatively high background heat flow must then be due to active hydrothermal convection. A model with more moderate temperatures at deep crustal level has been derived by joint interpretation of topography, seismic and gravity data (Kaban et al, 2002).

For the estimate of the geothermal potential in Iceland, the knowledge of the temperature and transport properties of heat below the presently exploited 2-3 km of the crust is of high importance. This estimate will both affect the reservoir models of the geothermal fields and the possibilities of extracting high enthalpy fluid, even supercritical, at 3-5 km depth in the crust of Iceland.

The study presented here is motivated by the question of the temperature below 3km depth in the crust below Iceland. It has been pointed out (Stefánsson et al., 1993) that there exist a general relationship between the surface heat flow in the southern lowlands of Iceland and the maximum focal depths of earthquakes. Tryggvason et al (2002) analyzed the seismic data from the SIL network and concluded that by linear extrapolation of surface temperature gradients in the SISZ the temperature at the depth of maximum focal depth was close to 550°C.

It is generally accepted that it is the temperature that mainly defines the lower boundary of the seismogenic earth. In addition the strain rate is important as well as the pressure. It could be expected that at high strain rates and lower pressure the seismogenic boundary shifts to higher temperature. For continental crust the limiting temperature for the seismogenic zone is 250-450°C and for the oceanic lithosphere 600-800°C (Chen and Molnar, 1983). In the latter case the boundary is within the mantle where peridotites are the main constituents but data from basalts and gabbros, as expected in the lower crust of Iceland, are missing.

2. DATA

2.1 Seismic data

The recorded events for the period 1991-2003 are close to 220.000. The precision of the depth estimate is variable depending mainly on the distance from an earthquake to the nearest seismic station but also on the nature of the earthquakes, quality of the time pick and other factors. The velocity model used for the location is a model based on seismic profiling in south Iceland (Bjarnason et. al., 1993). It is known that it is not suitable for all seismic areas in Iceland, which may cause some systematic errors in the absolute depth estimates, particularly in some areas off the coast in north Iceland. Further, in areas where the velocity structure is complex, e.g. in the vicinity of central volcanoes, systematic errors in the depth estimates may exist. The size of the assumed error in the absolute depth estimate due to these factors is in general not known and in this work we use the depth estimate as it is in the catalogue.

In selecting usable events for this study the main criteria is the ratio of the error in the depth estimate and the depth. The ratio of 0.6 was used and it reduced the number of events by half. Further selections were based on the distance to the closest recording station for each event (events with greater distance than 60 km were not used), depth (events at 2 to 30 km depth were selected) and magnitude (events with local magnitude larger than –2 and moment magnitude larger than -1.6 were selected). Finally, known explosions were excluded. The final dataset consists of 108000 events (figs.1 and 2).

The analysis was carried out on subareas, which were selected in three different ways:

1. Regular cells with size of 0.16° NS and 0.4° EW (approximately 17 by 19 km).
2. Areas within certain radius from boreholes outside the temperature disturbances of the geothermal fields. The chosen radii were 5, 10, 15 and 20 km.
3. Selected areas of special interest.

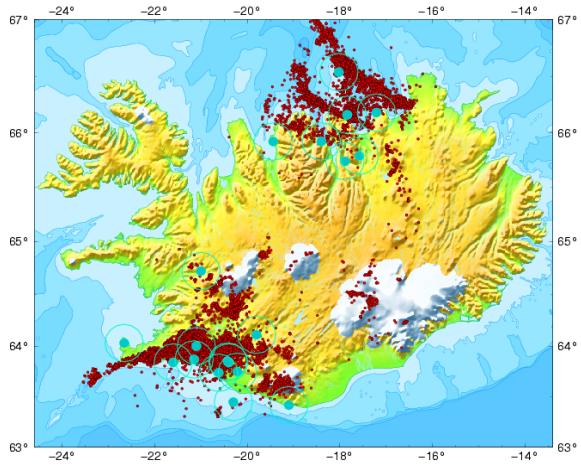

Figure 1. Earthquakes and boreholes used in this investigation. The earthquakes are shown with red dots and the boreholes with green filled circles. The filled circles have radius of 5 km. Further, a circle of radius of 20 km is shown.

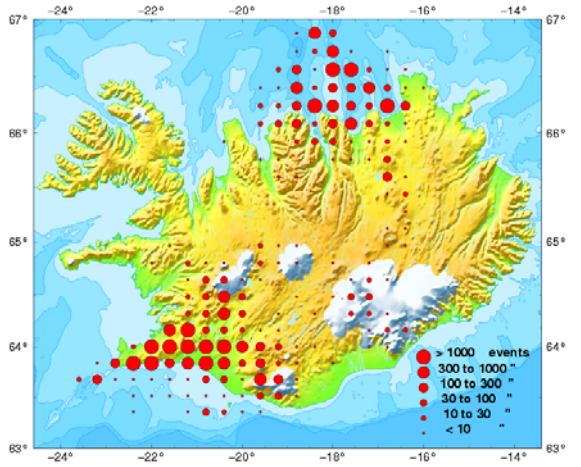

Figure 2. Number of earthquakes within cells of approximately 17 by 19 km.

In figure 3 examples of the depth distribution of earthquakes within 10 km radius of two boreholes are shown. It is clear from the cumulative curves that there is a sharp curvature or break close to the 0.95 fraction (95%). This is similar for most of the subareas selected and it was chosen that this level, i.e. the depth above which 95% earthquakes occur, defines the thickness of the seismogenic crust. In the analysis, subareas with less than 10 events were automatically excluded and some areas with up to 100 events were also excluded based on anomalous distribution.

More examples of depth distributions are shown in appendix 1.

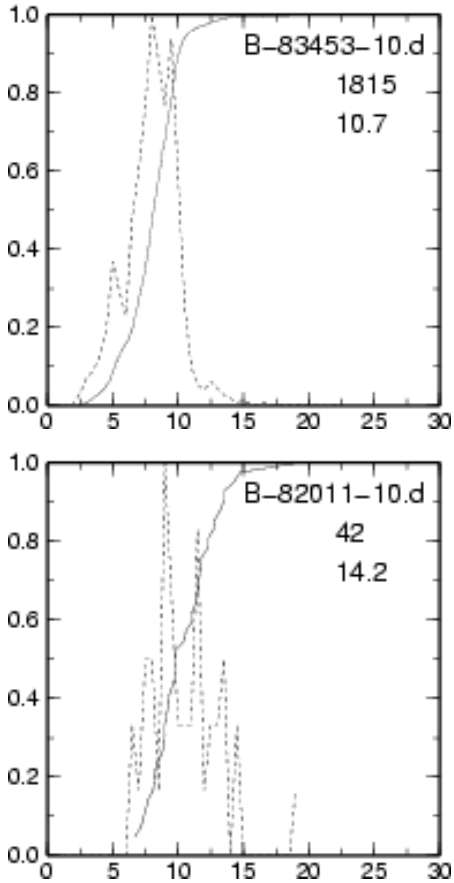

Figure 3. Example of the depth distribution of earthquakes in the area around two boreholes. The horizontal axis is the depth and the vertical axis shows fraction. The solid line is the cumulative depth distribution and the dashed line is the depth distribution. In the depth distribution the number of earthquakes within each 1/2 km depth interval is counted and the distribution is scaled with the number in the interval where most events occur. The labels within the frame refer to the location of the area under consideration, number of events and the depth of the 95% cumulative level.

The uncertainty of the depth estimate (apart from the possible error in absolute location) is based on the error estimate of the depth of each event, which is one standard deviation in the SIL catalogue. It is used to estimate depths where one can state with 68% confidence that all events are below or above the depth corresponding to the 95% depth level of the cumulative curve. The resulting interval around the 95% depth level is a conservative estimate of the error. The estimated error of the 95% depth level around the boreholes depends on the depth. From 8 to 13 km depth the proportional error increases from 6% to 16%. Below that the error is quite variable but on the average around 17%.

Rough overview of the distribution of earthquakes and the thickness of the seismogenic crust in Iceland based on the analysis of regular cells is shown on figure 4. Most of the earthquakes are along the Reykjanes peninsula, which is considered the landward extension of MAR, and SISZ and on the main fracture lines in the TFZ in northern Iceland. About half of the events in the dataset are at the Hengill triple junction, i.e. where the SISZ, WVZ and the seismic zone along the Reykjanes peninsula meet. In southern Iceland the thickness is greatest at the central volcanoes of Katla and Eyjafjallajökull and as mentioned above where there is a complex velocity structure. The thickness is similarly great, compared to the SISZ, off the southern coast and at the NE end of the WVZ. In northern Iceland the thickness is generally great particularity in the NE part of the seismic area.

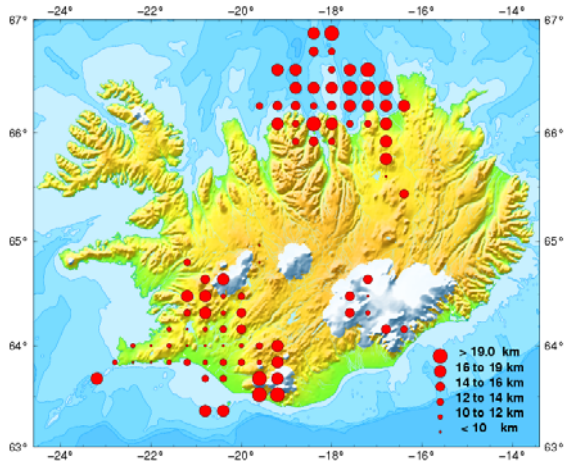

Figure 4. The thickness of the seismogenic crust within cells of approximately 17 by 19 km.

2.2 Borehole data

There exist several thousands of boreholes with temperature logs in Iceland in the common database of Iceland GeoSurvey (ISOR) and the National Energy Authority (NEA). Most of them are drilled for geothermal purposes and are located within known geothermal fields. Only those that are outside the temperature disturbances of the geothermal fields are suitable for estimates of the background heat flow. These are probably around five hundred; most of them only 50-60m deep but several wells extends down to 1,0-1,5 km depth. In order to compare the background heat flow with maxiumum focal depth of earthquakes we have selected boreholes that are regarded to be outside areas affected by hydrothermal activity but have considerable numbers of earthquakes nearby.

Table 1 shows a list of the boreholes together with the estimated temperature gradient and the maximum focal depth of nearby earthquakes. Fig. 5 shows examples of the temperature logs from the boreholes. The error estimates in the table include an estimate of the error of the determination of the temperature gradient from the temperature logs but does not include any errors due to possible regional disturbances from hydrothermal activity. However it should be noted that suitable distance to any known hydrothermal field is a prerequisite for the selection of the boreholes used.

3. FOCAL DEPTH AND TEMPERATURE

When the focal depth and surface heat flow are compared it is a question how large circular area around the borehole should be included in the maximum focal depth estimate. If we use too small circle, the number of earthquakes behind the maximum focal depth estimate can be too low for reliable estimate. On the other hand, if too large circle is used and the bottom of the brittle layer is dipping, the result can be, that the estimated thickness of the brittle crust below the borehole is biased. It is probably most reasonable to use circle with radius of similar length as the depth to the

brittle-ductile boundary, i.e. to select 10 – 20 km circle. In the following presentation we use earthquakes with epicenter within 10 km radius from the borehole as the minimum circle, but if too few earthquakes are within this radius we use circles of 15 or 20 km.

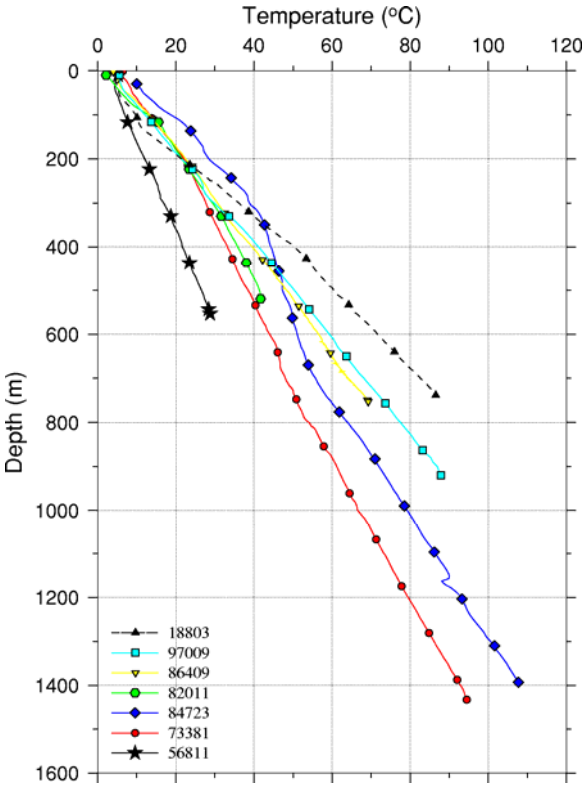

Figure 5. Temperature logs from the boreholes that are deeper than 500m and are used in our estimate of background temperature gradient.

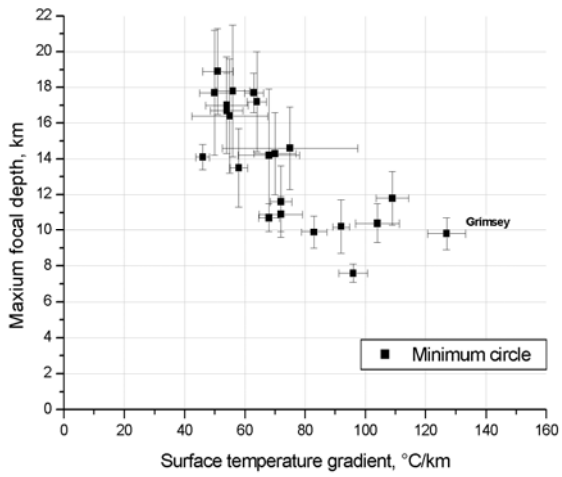

Figure 6. The figure shows the strong relation between the maximum focal depth of earthquakes and the near surface temperature gradient. The outlying data from Grímsey might indicate presence of nearby hydrothermal system or dip of the bottom of the seismogenic crust.

Figure 6 shows the maximum focal depth as function of the surface temperature gradient. A strong inverse correlation appears showing thinning of the brittle crust with higher surface temperature gradient. The point at Grímsey, a small island north of Iceland, shows relatively thick crust with respect to the high gradient. Only shallow holes exist in Grímsey and the there are know hydrothermal fields several km offshore

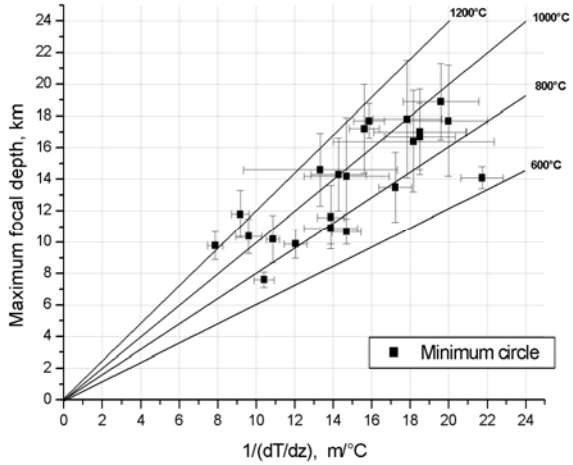

Figure 7. The maximum focal depth versus the inverse temperature gradient at the surface. The data used is from earthquakes within 10 km circle around the boreholes but data within 15 and 20 km are used where focal depth data for 10 or 15 km are insufficient. Most of our data falls close to or above the 800°C.

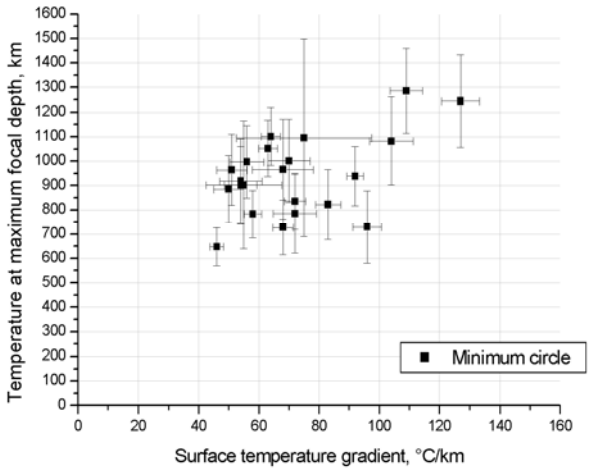

Figure 8. The temperature at the maximum focal depth obtained from linear extrapolation of surface gradients. A positive relation is observed instead of flat line corresponding to the isothermal surface of maximum focal depth.

the island. Therefore the high gradient in Grímsey might indicate presence of nearby hydrothermal activity and is possible not suitable for our analysis.

In order to visualize better the isothermal nature of the bottom of the brittle crust we plot the maximum focal depth as function of the inverse surface temperature gradient. On such a plot the datapoints should fall on a straight line through the origin and with gradient equal to the value of the isotherm. This is done on figure 7. The datapoints are somewhat scattered as would be expected for data of this nature. However, according to the figure most of our datapoints falls between the 800 and 1200°C isolines with average value of 950°C and standard deviation of 170°C. These are much higher values than are expected on basis of

previous work elsewhere in the world (Chen and Molnar, 1983).

Figure 8 shows the relation between the extrapolated temperature at the bottom of the brittle crust and the surface gradient. If the bottom is an isothermal surface these parameters should not be dependent. However it appears that high surface temperature gradients are linked to higher temperature at the bottom of the brittle crust.

Therefore we conclude that most likely the surface temperature gradient cannot be extrapolated linearly downwards through the brittle crust but must decline with depth. Another possibility is that the maximum focal depth is systematically over-estimated but that is rather unlikely.

4. DISCUSSIONS

There are several possibilities to explain decreasing temperature gradient with depth in the brittle crust. Factors such as radiogenic heat production, increased thermal conductivity with depth, hydrothermal convection, intrusion activity into the lower brittle crust, recent erosion, exothermic chemical reactions and friction heat from earthquakes might affect the temperature gradient. Bödvarsson (1982) has estimated that contribution from earthquakes to the heat flow is negligible and Björnsson (1980) has shown the permeability of the crust is too low to allow considerable hydrothermal convection outside the fractured hydrothermal fields. The role of chemical reaction within the Icelandic crust is unknown but might as well be endothermic as exothermic. There is ongoing slow subsidence in Iceland as a result of accumulation of new lava in the volcanic zone. The subsiding crust is reheated and undergoes high temperature alteration reactions, which in many cases are endothermic. Furthermore, the devolatization processes that are ongoing in the lower crust are endothermic. As first approximation we neglect the contribution of chemical reaction to the heat flow, but this has to be looked into in the future. The radiogenic heat production rate in the Icelandic crust is very low and is unlikely to exceed 0.5 $\mu W/m^3$ (Flóvenz and Sæmundsson, 1993) and has therefore negligible effect on the downward extrapolation of the surface gradient. The value of the thermal conductivity is more questionable. At the surface the thermal conductivity is close to 1,7 +/- 0,1 W/m°C for the basaltic lava. However due to decreasing porosity with depth the thermal conductivity is supposed to increase to about 2,0 for basalts or gabbros with near zero porosity (Stefánsson, 1997). Data from Birch and Clark (1940) suggest that the thermal conductivity of gabbros and diabase is almost constant with temperature up to at least 200°C. In the following we will assume that the zero porosity is obtained at seismic layer 3 of the Icelandic crust, i.e. where the lower crust starts. The composition of the lower crust, from typically 5 km depth to the proposed Moho at 20-40 km depth is unknown but gravity data implies that the density is rather high (Menke, 1999, Kaban et al. 2002). Therefore it has been suggested that the lower crust consist of a mixture of crustal and mantel material (Kaban et al 2002). Since the mantle peridotite has thermal conductivity close to 3.0W/m°C and is almost constant with temperature above 200°C (Seipold, 2001) it is reasonable to expect increasing thermal conductivity with depth in the lower crust in Iceland.

We have therefore recalculated the temperature at the base of the brittle crust beneath our boreholes by assuming heat production rate of 0.5 $\mu W/m^3$, linearly increasing thermal conductivity from 1.7 W/m°C at the surface to 2.0 W/m°C at the top of seismic layer 3 and further linear increasing downwards to 3.0 W/m°C at the maximum focal depth of earthquakes. The depth to the top of seismic layer 3 is taken from a map in Flóvenz and Gunnarsson (1991). The results are shown on figures 9 and 10 as plots of the calculated temperature at the bottom of the seismogenic crust versus surface temperature gradient and maximum focal depth respectively. Both figures also show the results from linear extrapolation for comparison. The latter shows no significant correlation between the maximum focal depth and the temperature and the same applies to the relation with surface temperature gradient apart from 2 extreme values of the surface temperature gradient.

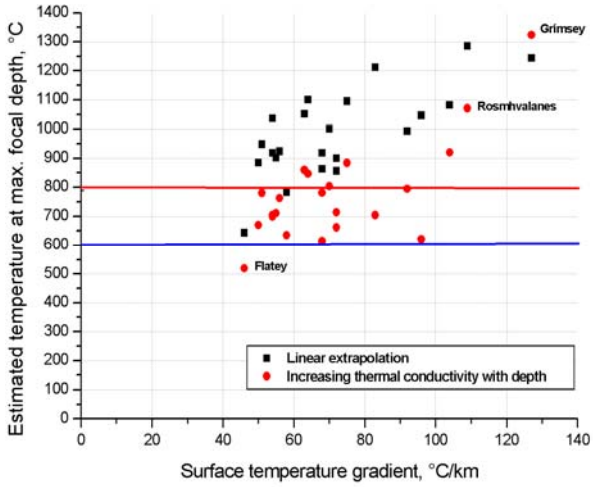

Figure 9. Estimated temperature at the base of the brittle crust versus surface temperature gradient. The black dots show the temperature when the near surface gradient is linearly extrapolated but the red dot when the thermal conductivity increases linearly with depth in two steps. For the latter most of the values falls into the 600-800°C interval.

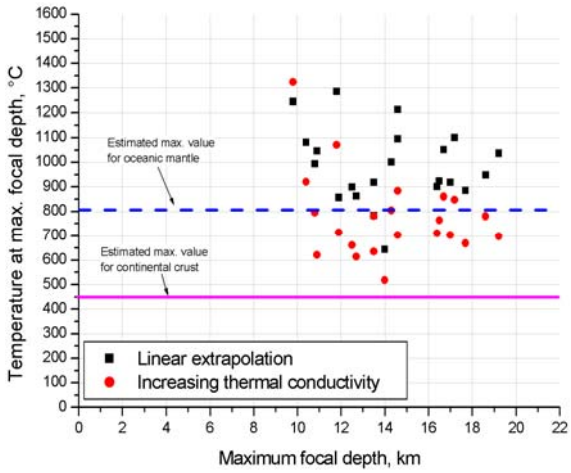

Figure 10. Estimated temperature at the base of the brittle crust versus the maximum focal depth. The black dots show the temperature when the near surface gradient is linearly extrapolated but the red dot when the thermal conductivity increases linearly with depth in two steps. As expected, no obvious correlation appears.

These two extreme values are from Grímsey and Rosmhvalanes, which are close high temperature fields in the neovolcanic zone north of Iceland and on the Reykjanes peninsula in SW Iceland respectively. These could both be

affected by too short distance to geothermal fields. In addition the minimum value from Flatey, an island sitting on thick sedimentary sequence in the Tjörnes fracture zone, is likely to be under-estimated by up to 10% because of high subsidence rate (Gunnarsson et al, 1984).

Both the figures 9 and 10 shows that the calculated temperature is in most cases between 600°C and 800°C. If we exclude the extreme values from Grímsey, Rosmhvalanes and Flatey the average value the temperature at the base of the seismogenic crust is 750°C with standard deviation of 100°C. This value is in good agreement with values for the oceanic lithosphere (Chen and Molnar, 1983) with mantle properties.

In addition to the possible increase in thermal conductivity with depth, the decrease of the temperature gradient with depth could be explained by heat production in the lower part of the brittle crust due to intrusion activity. From geological point of view such intrusion activity is most likely to be mostly restricted to the volcanic zones and their vicinity. However some modeling of this phenomenon is necessary to check if this is a possible explanation.

5 CONCLUSIONS

The main conclusions from our investigation are as follows:

1. The seismic activity in the Icelandic crust has a sharp lower limit, which we define, as the bottom of the seismogenic crust and marks the boundary between the brittle upper crust and the lower ductile crust.

2. Our data support the choice of Tryggvason et al (2002) for definition of the bottom of the seismogenic crust as being the level above which 95% of earthquakes occur within an area of radius 10-20km.

3. We have carefully selected boreholes with reliable heat flow data in areas where information on maximum focal depth exist. There is a clear relationship between the temperature gradient and the thickness of the brittle crust.

4. If the temperature gradients for these holes are extrapolated linearly to the bottom of the seismogenic crust, the temperature there would be within the range of 800 to 1200 °C.

5. On the basis that the brittle ductile boundary is close to be an isotherm and the maximum temperature there is 800°C for the oceanic lithosphere we conclude that linear extrapolation of near surface temperature gradients is not valid down the this boundary in Iceland.

6. We suggest two possible explanations for the decrease in temperature gradient with depth, an increase in thermal conductivity and intrusion activity in the lower brittle crust.

7. Intrusion activity in the lower brittle part of the crust would have similar effect as radiogenic heat production, i.e. reduce the temperature gradient

8. The increase in thermal conductivity could be due to reduced porosity with depth in the upper crust and increasing content of mantle material in the lower part of the brittle crust.

9. By assuming linear increase in thermal conductivity from 1.7 W/m°C (average value for basalts at the surface) to 2.0 W/m°C (value for gabbro) at the top of the seismic layer 3 and further linear increase to 3.0 W/m°C (value for peridotite) at the bottom of the seismogenic crust we obtain average value of the isothermal surface of 745°C with standard deviation of 95°C.

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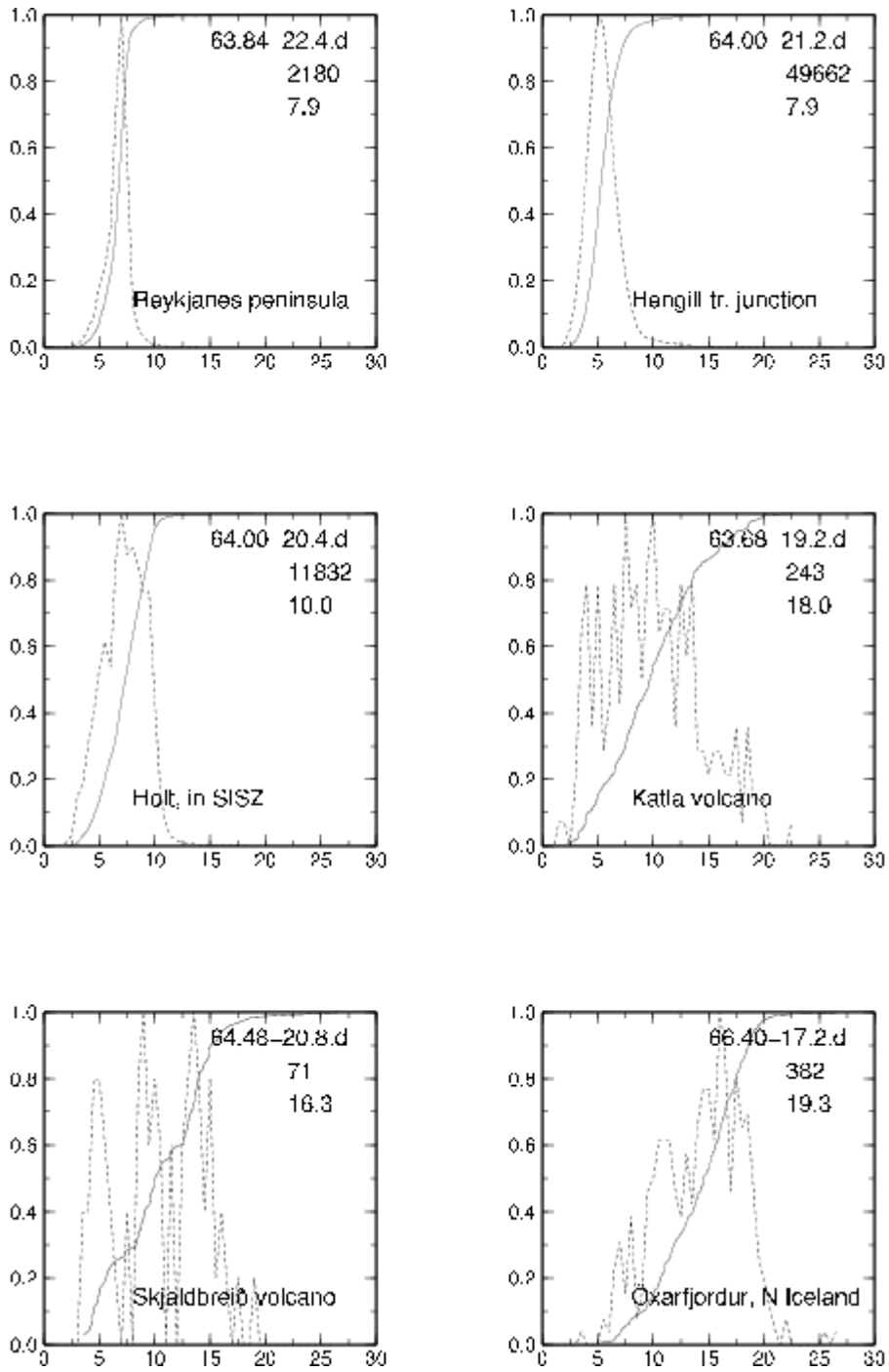
Place	Well no.	Lat	Long	Depth	dT/dz	error	Maxium focal depth within a circle or radius:		
				km	°C/km	%	10km	15km	20km
Hvítársíða*	29291	64,7162	20,997	54	75	0,3	14,6	14,4	16,7
Bær, Höfðaströnd	49471	65,9231	19,4378	60	50	0,1	NaN	NaN	17,7
Mýrarkot	49461	65,9335	19,4032	63	64	0,05	NaN	NaN	17,2
Ártún, Hofsósi	49501	65,8892	19,3946	104	54	0,13	NaN	NaN	17,0
Grímsey	52411	66,542	18,018	118	127	0,05	12,5	14,4	15,8
Flatey, Skjálfanda	56811	66,1631	17,8416	554	46	0,05	14,1	14,6	15,0
Háls, Fnjóskadal	56821	65,738	17,858	129	58	0,05	NaN	13,5	12,9
Stórólfshvoll	82011	63,7545	20,2071	575	68	0,15	14,2	13,6	12,6
Helluvað	82331	63,8442	20,4006	398	72	0,1	10,9	10,4	10,1
Lyngás	83453	63,8477	20,4245	145	68	0,05	10,7	10,4	10,1
Þykkvibær	84723	63,745	20,625	1400	72	0,05	11,6	11,9	11,8
Eyrarbakki	86409	63,865	21,1529	760	83	0,05	9,9	8,1	8,4
Sámsstaðir*.	896xxx	64,11	19,8	199	55	0,23	16,4	13,1	11,7
Nátthagi, Ölfusi	96751	63,9955	21,1025	183	96	0,05	7,6	8,0	8,3
Ingjaldsstaðir	59524	65,6926	17,518	60	70	0,1	NaN	NaN	14,3
Sandhóll	59522	66,1832	17,2146	58	56	0,1	17,8	19,1	18,7
Ystafell	59523	65,7901	17,5763	63	104	0,07	NaN	10,4	14,9
Heimaey	73381	63,45	16,7	1565	63	0,05	NaN	NaN	16,7
Rosmhvalanes	18803	64,03	22,66	767	109	0,05	NaN	NaN	11,8
Árskógsströnd*		65,92	18,4	60	54	0,1	16,7	14,7	14,9
Þorlákshöfn	97009	63,84	21,6	1096	92	0,03	10,2	9,9	9,0
Vík í Mýrdal	76002	63,62	19,1	1351	51	0,1	18,9	19,4	19,8

Table 1. Overview of the boreholes used. The asterix denotes places where the temperature gradient is an average value from several boreholes in the relevant area.

Appendix 1: Examples of depth frequency curves of earthquakes



Further examples of the depth distribution of earthquakes in selected areas in Iceland. The horizontal axis is the depth and the vertical axis shows fraction. The solid line is the cumulative depth distribution and the dashed line is the depth distribution. In the depth distribution the number of earthquakes within each 1/2 km depth interval is counted and the distribution is scaled with the number in the interval where most events occur. The labels within the frame refer to the location of the area under consideration, number of events and the depth of the 95% cumulative level.