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A new conceptual model of the Icelandic crust in the Hengill geothermal area based on the indirect electromagnetic geothermometry

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ABSTRACT

Application of the indirect electromagnetic geothermometer calibrated by available temperature well logs has enabled the construction of the three-dimensional temperature model of the Hengill geothermal area (Iceland) up to a depth of 20 km. Its analysis showed that the background temperature of the Icelandic crust above 20 km does not exceed 400 °C. It is overlapped by a network of interconnected high-temperature low resistive channels, which braid through the crust mainly at a level of 10–15 km and root to a depth greater than 20 km. Accordingly, the probable heat sources feeding the geothermal system are supposed to be the intrusions of the hot partially molten magma upwelling from the mantle through faults and fractures. The comparison between the vertical temperature cross-sections and the projections of the earthquake hypocenters showed that they all are located in the areas where temperature does not exceed 400 °C, which is a gabbro solidus in a silica-rich Icelandic crust. Joint analysis of the temperature model together with the resistivity and residual Bouguer gravity anomalies enabled us to explain the distribution of the earthquake hypocenters by different geothermal regimes in adjacent parts of the area and cooling of large massifs of the partially molten solidified magma beneath seismically active areas. Basing on the above inferences, we suggest a new self-consistent conceptual model of the Icelandic crust, which agrees with most of the previous geophysical results and provides an explanation for the facts that the previous models failed to explain.

1. Introduction

There are two main hypotheses on the structure of the unique Icelandic crust. The “thin and hot” model of the crust (Björnsson et al., 2005; Björnsson, 2008) is based on the implicit assumptions on the conductive heat flow and linear increase of the borehole temperature gradient in the lithologically uniform crust (Hermance and Grillot, 1974; Flóvenz, 1985; Flóvenz and Saemundsson, 1993; Tryggvason et al., 2002). According to it the maximal depth (5–6 km) of most of the seismic events corresponds to the brittle/ductile transition for the basalt (T = 650 °C), i.e., at the upper/lower crust boundary, while the lower boundary of the high electrical conductivity layers (10–15 km) detected by magnetotelluric (MT) soundings is attributed to the base of the crust.

The alternative hypothesis (so called “thick and cold” crustal model) is based mainly on the observation of a gradual increase of seismic P-wave velocities with depth and a stepwise rise from 7.2 to 7.7 km/s at about 22 km, which supposedly marks the Moho (Pavlenkova and Zverev, 1981; Bjarnason et al., 1993; Menke and Levin, 1994; Menke and Sparks, 1995; Menke et al., 1995, 1996; Foulger et al., 2003).

Unfortunately, these models offer no clue as to the following questions:

- What is the nature of highly conductive layers recognized by MT sounding at the depths of 1–3 km and 10–15 km (Björnsson et al., 2005)?
- Why has the drilling in the Krafla geothermal field penetrated rhyolitic magmas (~74% SiO_2, 1–2% H_2O) with a temperature of T = 1100 °C at a depth of 2.1 km and with a temperature T = 386 °C at a depth of 2.6 km (Elders and Fridleifsson, 2010)?
- Why is the continuous seismic activity in this region mainly confined to the superimposition of the zone constrained between the meridians 21.31° and 21.33° W and a band running beneath the second-order tectonic structure of Olkelduhals, instead of being maximal along SSW–NNE direction, as dictated by the position of the crustal accretion zone (Árnason et al., 2010)?
- Why do the earthquakes in the Icelandic crust occur at depths of 12–14 km (Stefansson et al., 1993) where the temperature must have been above solidus?

Analyzing the list of questions not answered by these models, we see that the choice of the conceptual crustal model of the region crucially depends on the ability to estimate the spatial distribution of temperature up to a depth of 20–25 km (instead of its estimates at some characteristic

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depths provided by available indirect geothermometers). Recently, a new approach has been proposed for estimating temperature in the Earth's interior from electromagnetic (EM) sounding data measured on the surface (so-called "indirect electromagnetic geothermometer") (Spichak and Zakharova, 2009b; Spichak et al., 2011c). This approach not only estimates temperature far below the borehole bottom, but also reconstructs deep temperature distributions from the ground EM data (see the review paper (Spichak and Zakharova, 2012) and references therein). Our research was aimed at answering the questions enumerated above basing on (1) estimation of the 3D temperature distribution in the Hengill geothermal area (Iceland) from the resistivity data up to the depth of 20 km, (2) identification of the heat sources of the geothermal system, and (3) analysis of the seismicity pattern.

2. Geology and volcanic activity in the region

The Earth's crust in Iceland is composed of volcanic rocks with inclusions of intrusive and effusive rocks (mainly oceanic-type flood basalts, tuffs, hyaloclastites, and some felsic rocks). The high-temperature Hengill area is a triple junction zone of intersection of the Western Volcanic Zone (WVZ), the Reykjanes Peninsula Rift (RPR), and the South Icelandic Seismic Zone (SISZ), which is located in the southwest of the island (Einarsson, 2008) (Fig. 1, upper panel). The Hengill volcanic complex comprises several interconnected geothermal fields located in different directions with respect to Mt. Hengill (marked by H in Fig. 1, lower panel): the Hveragerdi (Hv) area in the southeast, the Nesjavellir (Ne) area in the northeast, and Hellisheidi (He) area in the southwest (Arnorsson, 1995; Arnorsson et al., 2008; Zakharova and Spichak, 2012).

Overall, the region and its immediate vicinity hosts four centers of volcanic activity: the Hengill area mentioned above, as well as the Grensdalur, Hromundartindur, and Husmuli areas. The Hengill volcanic complex comprises an active central volcano and a swarm of fractures trending north-northeastwards (Fig. 2). A secondary tectonic structural trend, perpendicular to the dominant NNE–SSW trend of the signs of crustal accretion, has developed in the zone connecting the centers of the Hengill and Grensdalur volcanic complexes and extending along the Olkelduhals line (Figs. 1 and 2). Attached to it are the eruption centers which migrate west-northwestwards (Foulger, 1988a), as well as the hot springs and fractures.

A vast high-temperature geothermal area that includes the Hengill and Gresdalur central volcanoes as well as the transversal tectonic structure between them is characterized by continuous microseismicity. There is a strong negative correlation between the seismicity and faulting observed at the surface: the earthquakes are clustered around the S–N or WNW–ESE azimuths but not in the SSW–NNE direction which dominates the surface geology. On the other hand, the seismicity spatially correlates with heat losses through the surface (Foulger, 1988b), which indicates that seismic activity in this region is associated with geothermal processes rather than with the plate boundary (Foulger, 1988a).

3. Resistivity data and temperature well logs

3.1. Resistivity data

In order to create an indirect EM geothermometer, we used magnetotelluric (MT) and a central-loop transient EM (TEM) data collected in this area (Fig. 3). Fig. 4 shows a typical example of 1D inversion of TEM sounding data provided in more than one hundred sites (Árnason et al., 2010). Besides, we used 1D resistivity profiles revealed from magnetotelluric data collected in Hengill area in the framework of the INTAS Project using Phoenix MTU instruments in the frequency range from $5 \cdot 10^{-4}$ Hz to 300 Hz at 50 sites with a remote reference point located 10 km apart. The distance between the MT sites and the ocean coast was sufficiently big at this frequency range, so the coast effect on the further MT data interpretation could be neglected (Beblo et al., 1983).

Fig. 5 indicates four examples of the apparent resistivity and phase curves (Árnason et al., 2010). According to these authors in the majority of the MT sites the xy- and yx-components manifest 1D behavior up to the periods of 1 s. So, taking into account that the background resistivity in the Hengill area varies in the range 15–100 Ωm (Árnason et al., 2010; Spichak et al., 2011a) it is reliable to extend the 1D approximation of the resistivity structure at least to the depth of 5 km.

Unlike the most popular approach to reducing of the so called "static shift" effect by correcting the MT curves with TEM data measured in the same locations (see, for instance, (Árnason et al., 2010)), we have used to this end all available TEM and MT data (not necessarily collected in the same sites). In this case, 1D resistivity profiles determined from the TEM data were used up to the depth of 1 km while 1D resistivity profiles determined from the MT data — from 1 km to 3 km. This depth limit was justified by two factors: on the one hand, it exceeds the maximal depth of the temperature logs available in this area but, on the other hand, it is less than 1D dimensionality depth limit estimated above. So, this procedure enabled us to avoid the effect of the near-surface geological noise and to enlarge the available database, which, in turn, increased the accuracy of the resistivity reconstruction.

3.2. Temperature well logs

Both the high- and relatively low-temperature geotherms recorded in 20 boreholes with different depths and drilled in geologically different Hellisheidi and Nesjavellir geothermal fields were used (see Fig. 3 for their locations and Fig. 6 for geotherms). The temperature gradient in the boreholes drilled to a depth of 1–2 km varies from 84 ± 9 °C/km in the low-temperature regions of the transform zone to 138 ± 15 °C/km in the geothermal areas (Foulger, 1995).

Both the geothermal areas of Hellisheidi and Nesjavellir show common characteristics in their measured geothermal gradient as they both are dissected by a volcanic fissure swarm with predominant SSW–NNE trend (Fig. 2). The southeastern part of the block of boreholes in the Hellisheidi area is mainly controlled by cooling caused by intrusion of cold groundwater to a depth of 1.4–2 km (Franzson et al., 2010). Correspondingly, in most of the temperature logs from this block, the initial temperature rise is followed by a decrease at a depth below 1–1.5 km (Fig. 6a). Within the Nesjavellir geothermal field (Fig. 6b) there is a difference between the temperature patterns of the southwestern "cooled" part of the region and its eastern "hot" part, probably, caused by intrusion of the overheated vapor through one of the fractures associated with heating in the eastern part of the area.

4. Application of the indirect electromagnetic geothermometer

The application of the indirect EM geothermometer comprises 3 stages (Spichak and Zakharova, 2009b; Spichak et al., 2011c): (1) the geothermometer calibration, i.e. the training of an artificial neural network (ANN) "with a teacher" to provide a correspondence between the resistivity/conductivity values and the known temperature values; (2) the testing of the geothermometer by comparing the forecast temperatures with available temperature well logs not used during its calibration; and, finally, (3) ANN forecasting of the temperature for the whole study area.

In Sections 4.1 and 4.2 we will consider steps (1) and (2), accordingly, while step (3) will be considered in Section 5.

4.1. Calibration of the geothermometer

Before training of the ANN by correspondence of the resistivity and temperature data we had first to assign the resistivity values in