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2017-12-01


http://hdl.handle.net/10138/232543
https://doi.org/10.1093/gji/ggx373

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Heat flow, seismic cut-off depth and thermal modeling of the Fennoscandian Shield

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SUMMARY

Being far from plate boundaries but covered with seismograph networks, the Fennoscandian Shield features an ideal test laboratory for studies of intraplate seismicity. For this purpose, this study applies 4190 earthquake events from years 2000–2015 with magnitudes ranging from 0.10 to 5.22 in Finnish and Swedish national catalogues. In addition, 223 heat flow determinations from both countries and their immediate vicinity were used to analyse the potential correlation of earthquake focal depths and the spatially interpolated heat flow field. Separate subset analyses were performed for five areas of notable seismic activity: the southern Gulf of Bothnia coast of Sweden (area 1), the northern Gulf of Bothnia coast of Sweden (area 2), the Swedish Norrbotten and western Finnish Lapland (area 3), the Kuusamo region of Finland (area 4) and the southernmost Sweden (area 5). In total, our subsets incorporated 3619 earthquake events. No obvious relation of heat flow and focal depth exists, implying that variations of heat flow are primarily caused by shallow lying heat producing units instead of deeper sources. This allows for construction of generic geotherms for the range of representative palaeoclimatically corrected (steady-state) surface heat flow values (40–60 mW m⁻²). The 1-D geotherms constructed for a three-layer crust and lithospheric upper mantle are based on mantle heat flow constrained with the aid of mantle xenolith thermobarometry (9–15 mW m⁻²), upper crustal heat production values (3.3–1.1 μW m⁻³) and the brittle-ductile transition temperature (350°C) assigned to the cut-off depth of seismicity (28 ± 4 km). For the middle and lower crust heat production values of 0.6 and 0.2 μW m⁻³ were assigned, respectively. The models suggest a Moho temperature range of 460–500°C.

Key words: Heat flow; Europe; Spatial analysis; Statistical seismology; Cratons; Crustal structure.

1 INTRODUCTION

The maximum focal depth (i.e. cut-off depth) of earthquakes has been frequently used to model thermal conditions of the lithosphere, especially to approximate the depth below which the ductile transformation dominates over brittle transformation of the crust. This depth, useful for the verification of rheological models, has been usually considered a temperature isotherm. However, widely varying temperature estimates have been given for it, ranging from 260 ± 40 up to 450 ± 50°C depending on the tectonothermal setting and the crustal lithologies of the area (Ranalli 1995; Bonner et al. 2003). However, the temperature of 350°C has been considered a viable choice for Fennoscandia, given the dominance of felsic rocks in its upper and middle crust (Blanpied et al. 1991; Moisio 2005) and the tendency of even small concentrations of liquid water to decrease the lithospheric strength (Mackwell et al. 1998) at temperatures close to the critical point of water (374°C at 22 MPa). The lithological composition of the crust is, however, not simple. Kuusisto et al. (2006) modeled the crustal seismic velocity structures in the central part of the Fennoscandian shield using mixtures of different lithologies for the different velocity layers and concluded that quartz-bearing rock types are abundant from the surface down to the depth of 25–30 km. The calculated models suggested that the crustal velocity profiles can be simulated with rock-type mixtures where the upper crust consists of felsic gneisses and granitic–granodioritic rocks with a minor contribution of amphibolite and diabase. In the middle crust, the amphibolite proportion increases. The lower crust consists of tonalitic gneiss, mafic garnet granulite, hornblende, pyroxenite and minor mafic eclogite. If the lithospheres were mafic or ultramafic, the brittle-ductile transition would occur at a greater temperature than the temperature of 350°C assigned for granitic (quartz-bearing) lithology (Blanpied et al. 1991).

Correlation between the basal depth of the seismogenic zone and the terrestrial heat flow has been studied both in intraplate areas (Bollinger et al. 1985; Klemperer 1987; Wong & Chapman 1990)
and in the proximity of plate margins (Tanaka & Ishikawa 2002; Bonner et al. 2003). For example, the Japanese data (Tanaka 2004; Tanaka et al. 2004) obviously show a shift from volcanic areas with high heat flow and focal depths less than 5 km to mountain ranges with low heat flow and focal depths more than 30 km. In several shield areas, the rarity of earthquake events, the concentration of heat producing elements at shallow depths, and also the spatially insufficient distribution of geothermal gradient and thermal conductivity measurements may impair attempts to correlate earthquake depths with heat flow. In the Fennoscandian Shield, the distribution of seismic events is uneven and is densest in intraplate (“postglacial”) faults (Lagerbäck & Sundh 2008) which are a result of rapid removal of ice load at the end-stage of the Weichselian glaciation about 9000 yr ago. A combination of postglacial isostatic adjustment, ridge push from distant plate boundaries and the rapid change in the stress field was responsible of the faulting (Lund 2005; Lund et al. 2009; Kukkonen et al. 2010; Korja et al. 2016). The heat flow field of the area is also relatively well constrained, with palaeoecologically corrected values ranging from 35.3 ± 6.3 mW m$^{-2}$ in Archean areas to 81.6 ± 13.9 mW m$^{-2}$ in post-Sveconorwegian granites (Slagstad et al. 2009). The assessment of seismic hazard in potential nuclear power plant construction in post-Sveconorwegian granites (Slagstad et al. 2009) apparently applied corrections calculated by Kukkonen (1989b), yet these had notable variations related to local conditions and depths of holes. For example, correction assigned for the Kotalahti borehole is 11 mW m$^{-2}$ and that of Pielavesi hole is −4 mW m$^{-2}$ although these holes are only 85 km apart. On the other hand, Kukkonen (1989b) calculated the mean heat flow of Finland to be 37.0 ± 1.6 mW m$^{-2}$ (uncorrected) and 40.7 ± 1.8 mW m$^{-2}$ (corrected), these values being very close to each other. They had no information from holes deeper than about 1100 m, although deep holes are the only ones which can truthfully record the entire depth range of thermal gradient perturbation from glaciation–deglaciation cycle. Subsequently, Kukkonenet al. (1998) reported exceptionally low heat flow values of 2.4–11.6 mW m$^{-2}$ at the depth of 250–750 m in Archean rocks in Russian Karelia. The data were attributed to very cold conditions (permafrost) during the Weichselian glaciation and early deglaciation of the area. To make these heat flow values comparable to other Archaean areas, the palaeoecological corrections should have exceeded 20–30 mW m$^{-2}$ (Rudnick & Nyblade 1999; Mareschal & Jaupart 2013). Whenever possible, palaeoecological corrections should be determined for each point separately, since they are strongly dependent on the measurement depth interval of thermal gradient, and the climatic effect also needs to be distinguished from other phenomena which can strongly affect the temperature of the borehole or the well (Mottaghy et al. 2005). Therefore interpolating a heat flow correction map from readily corrected borehole data to estimate corrections for other holes (Slagstad et al. 2009) in a certain area may result in misleading corrections if measurement depths of uncorrected holes are not taken into account. The typical range of corrections in the interpolated map of Slagstad et al. (2009) is, despite this caveat, closer to corrections derived for individual holes (e.g. Kukkonen 1989b) than to the range of regionally determined corrections in the pan-European heat flow study (Majorowicz & Wybraniec 2010).

To investigate heat flow throughout our study region, we applied radial basis function interpolation in Scientific Python to construct a heat flow map. Unlike spherical splines, our method does not optimize the smoothness of the surface with the expense of data integrity. On the contrary, the value of the function at data point coordinates is exactly that of the data point itself, and values of the interpolation surface do not fall outside the actual data range. The size of our interpolation grid was 0.05$\degree$ in both latitude and longitude dimensions, yet an area of a grid cell (in km$^2$) is not preserved, since latitude circles are shorter at high latitudes. In our analysis, this was no problem as long as the grid size was fine enough to allow a reasonable correlation with earthquake focal coordinates. For the few borehole data with same coordinates (e.g. Fjällvåden 1 and 2 in Sweden; Hurter & Hanel 2002), we calculated the arithmetic mean of uncorrected heat flow values before interpolation because our interpolation algorithm did not allow different data values at similar locations. After gathering all heat flow data (223 points; Appendix A), we applied our interpolation to construct a map from palaeoecologically corrected heat flow determinations (Fig. 1).

In Fennoscandia, palaeoecologue is the most important factor to cause deviations from true steady-state heat flow conditions, but locally adjusted palaeoecological corrections can be remarkably different from regional ones. For example, Majorowicz & Wybraniec (2010) suggested a correction of 16–18 mW m$^{-2}$ for surface heat flow data of eastern Finland, but corrections listed by Slagstad et al. (2009) are lower, typically less than 10 mW m$^{-2}$. For the majority of Finnish data, Slagstad et al. (2009) apparently applied corrections calculated by Kukkonen (1989b), yet these had notable variations related to local conditions and depths of holes. For example, correction assigned for the Kotalahti borehole is 11 mW m$^{-2}$ and that of Pielavesi hole is −4 mW m$^{-2}$ although these holes are only 85 km apart. On the other hand, Kukkonen (1989b) calculated the mean heat flow of Finland to be 37.0 ± 1.6 mW m$^{-2}$ (uncorrected) and 40.7 ± 1.8 mW m$^{-2}$ (corrected), these values being very close to each other. They had no information from holes deeper than about 1100 m, although deep holes are the only ones which can truthfully record the entire depth range of thermal gradient perturbation from glaciation–deglaciation cycle. Subsequently, Kukkonenet al. (1998) reported exceptionally low heat flow values of 2.4–11.6 mW m$^{-2}$ at the depth of 250–750 m in Archean rocks in Russian Karelia. The data were attributed to very cold conditions (permafrost) during the Weichselian glaciation and early deglaciation of the area. To make these heat flow values comparable to other Archaean areas, the palaeoecological corrections should have exceeded 20–30 mW m$^{-2}$ (Rudnick & Nyblade 1999; Mareschal & Jaupart 2013).

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The determination of corrections to all holes individually requires
detailed information about the climatic history of Fennoscandia and
is beyond the scope of this paper. Therefore, we opted for using the
correction method of Slagstad et al. (2009) on the heat flow versus
depth relationship. Unfortunately, the correction map does not cover
the eastern and northern parts of Fennoscandia, yet the 2516 m deep
Outokumpu deep hole (Kukkonen et al. 2011) shows the increase
of heat flow from 28 to 32 mW m\(^{-2}\) at shallow depths to 40–45
mW m\(^{-2}\) near the bottom of the hole. Therefore, correction of
ca. 10 mW m\(^{-2}\) is suggested by Kukkonen et al. (2011) for holes
not deeper than 1000 m in eastern Fennoscandia. Unlike in the
case of Kola superdeep hole (Kukkonen & Clauser 1994; Mottaghy
et al. 2005), it is likely that advection and structural effects play
minimal role in the heat flow of Outokumpu. Although it is not
always easy to quantify the relative effects of different aspects which
influence the temperature signal, the general increase of heat flow
with depth is a phenomenon observed throughout the East European
Craton (Kukkonen & Jöeleht 2003). Using Outokumpu result as a
proxy, we applied the bulk correction of 10 mW m\(^{-2}\) also for 25
other Archean and two Proterozoic data entries which did not have
other correction parameters associated with them. Because 26 of
these heat flow determinations are located on the Russian territory
where none of our earthquake epicentres are located, this kind of
adjustment does not seriously alter heat flow field in Swedish and
Finnish regions with notalb seismic activity, although it can have
a small influence on modeled heat flow values of Kuusamo, close
to the Russian border.

3 EARTHQUAKE DATA

The Institute of Seismology of the University of Helsinki main-
tains the catalogue of earthquakes recorded by the 24 stations
of the Finnish seismograph network (http://www.helsinki.fi/geo/
seismo/maanjaristykset/suomi.html). All data are readily available
as a table with information about latitude, longitude, focal depth and
magnitude. In Sweden, earthquake data are registered by 65 stations
and gathered by the University of Uppsala. The Swedish-language
site of the Swedish seismograph network provides a Google map
application (http://snsn.geofys.uu.se/map/map.php) which can be
used to fetch earthquake data registered since year 2000. For further
analysis, we imported latitude, longitude, focal depth and magni-
tude information of events to Python scripts, and combined these
data to those obtained in Finland.

In our analysis, all observations from the Finnish and Swedish
earthquake catalogues of years 2000–2015 were used, except for
data with magnitudes less than 0.1, and for a few events with
locations far outside both Swedish and Finnish territory (for ex-
ample in the Norwegian Sea, Skagerrak or Russian Karelia). One
event also had an obviously incorrect depth (125.6 km) and was
Figure 2. Spatial distribution of all earthquakes (4190 events) accepted for our analysis. Magnitude or depth data are not shown. Area of this map is similar to that of the heat flow map. The rectangles, numbered by ‘1’, ‘2’, ‘3’, ‘4’ and ‘5’ indicate regions where the heat flow field and earthquakes were taken into more detailed investigations (southern Gulf of Bothnia, northern Gulf of Bothnia, Norrbotten/Lapland, Kuusamo region and southern Sweden) as explained in the ‘Earthquake data’ section of the text. Data from Appendix B.

therefore dismissed. It turned out that a small minority of observations, particularly in the Swedish catalogue, had focal coordinates outside the country, typically in Danish or Norwegian territory close to the Swedish border. However, in this paper observations obtained from Swedish network are simply referred to as Swedish data, and those from Finnish network as Finnish data. Especially in the Gulf of Bothnia and in the proximity of the land border between Finland and Sweden, some earthquakes had been registered by both Finnish and Swedish stations. Although a qualitative visual comparison of epicentral latitudes and longitudes of Finnish and Swedish earthquakes proved that the duplicate problem cannot produce a strong bias to results, we used the event time, latitude, longitude and magnitude data to filter out all obvious duplicates. Altogether 66 events were removed as being duplicates, and 15 events for other reasons as explained in Appendix B. Both in Finland and Sweden, catalogued earthquake magnitudes are local magnitudes ($M_L$) which closely correspond to the original Richter magnitude but have been determined using a modified function better suited for modern seismometers (Valtonen et al. 2013).

In total, 801 Finnish and 3389 Swedish values were accepted for analysis (Fig. 2, Appendix B). Therefore, only 1.9 per cent of data required removal. In the Finnish data, remarkable clusters of earthquakes are visible in the Kuusamo–Kandalaksha zone, and the Bothnian Bay shear zone in western Lapland (Uski et al. 2012), and events with depths greater than 25 km are almost exclusively limited to these areas (Fig. 3). The data of northern Sweden show a high spatial frequency of deep earthquakes around the glacially induced Skellefteå fault compared to that associated with other faults such as Lansjärvä and Pärve (Juhlin et al. 2010) which otherwise show remarkable seismic activity. The Umeå-Skellefteå area also experiences the most rapid land uplift rate in Fennoscandia, as much as 8 mm yr$^{-1}$ (Steffen & Wu 2011). The uplift rate in Ostrobothnia, on the opposite shore, is nearly as large, but earthquake activity is low perhaps due to the lack of intraplate fault lines. Hence, the glacial isostatic adjustment alone cannot explain the large number
of seismic events in eastern Sweden (e.g. Muir-Wood 2000; Lund 2005), but the strength of the lithosphere and friction of the faults must also play a notable role in general (Scholz 1998).

Close to the southern part of the Swedish coast of the Bothnian Bay, the Bollnäs fault (Malehmir et al. 2015) is a good example of a small, recently discovered glacially induced fault with deep seismicity in a narrow zone. Although no obvious intraplate faults are present in more southern areas of Sweden, the Protogine Zone (Berglund et al. 1992) shows an obvious concentration of earthquakes. In the far northern part of the country, deep earthquakes are prominent and they have a relatively even spatial distribution in a large area. Although in our Swedish data no earthquake has a magnitude greater than 5.2, and just 26 events exceed the magnitude of 3.0, there is evidence of palaeoearthquakes up to the magnitude of 8 which plausibly occurred in the time of more rapid glacial isostatic adjustment at the end of the Weichselian glaciation about 9000 yr ago (Arvidsson 1996; Lagerbäck & Sundh 2008).

The depth information is given in both Finnish and Swedish earthquake catalogues with a resolution of 0.1 km, although the actual accuracy of focal depth determination varies depending on the crustal velocity model applied and on the spatial distribution of seismographs. These can cause focal depth estimation errors of 30 per cent or more (Ahjos & Uski 1992). Although all seismographs cannot be analysed separately in a database-wide study, biases are more likely to average out in larger data sets like ours, than in smaller ones. Generally, the Finnish earthquake events appear shallow, with a mean depth of 7.0 km. In our Swedish data, the corresponding value is larger (13.1 km) and close to that of several major earthquake areas in the world (Maggi et al. 2000). In Finnish data, linear Pearson correlation coefficient of depth versus magnitude data is very low ($R^2 = 0.0407$). If depth ($z$) is expressed in kilometres, the linear fit to magnitude data follows the equation $M(z) = 0.0163 z + 0.879$. In Swedish data, depth is negatively correlated to magnitude and the linear fit equation is $M(z) = -0.0120 z + 0.975$. To check the validity of our null hypothesis (i.e. magnitude is independent of depth), we applied Student’s $t$-test. Because in the case of Finnish data, the two-tailed critical value in the test is at 95 per cent confidence level larger ($c = 0.0693$) than $R^2$, there is no reason to...
reject the null hypothesis. For Swedish data, $R^2 = 0.0252$, yet the critical value is 0.0337. Therefore, the null hypothesis remains in force also in the Swedish case.

For the purpose of comparison, latitude–longitude coordinates of all earthquake epicentres were assigned surface heat flow values from our interpolated heat flow grid. Hence, we were able to study the dependence of focal depth on the modeled surface heat flow. As seen in Fig. 4, highest occurrence of earthquakes is obtained in regions with corrected heat flow in the range of 48–60 mW m$^{-2}$. Cut-off depths of seismicity had a range from 23.4 km for the 90th percentile to 33.7 km for the 99th percentile (Table 1).

As derived from our interpolated heat flow map (Fig. 1), the corrected mean heat flow at earthquake epicentres turned out to be 49.8 mW m$^{-2}$, virtually same as the value of 49.7 $\pm$ 0.4 mW m$^{-2}$ obtained by Kukkonen & Jöeleht (2003) as a steady-state heat flow estimate for Fennoscandian Shield and East European Platform at depths of 2000 m and more. Therefore, it is unlikely that earthquake epicentres are strongly biased towards certain lithologies, but they represent the average Fennoscandian crust reasonably well for modeling purposes, and our interpolation works adequately.

To investigate the depth distribution of earthquakes further, we performed separate studies on subsets of two Swedish regions on the coast of the Gulf of Bothnia: first in the southern region, ranging from 60.0° to 63.5° N and 15.5° to 20.0° E (area 1 in Fig. 2) and then in the northern region, ranging from 63.5° to 66.5° N and 17.0° to 24.0° E (area 2 in Fig. 2). Our basal earthquake depth estimate for both regions was derived by calculating both the 90th (Doser & Kanamori 1986) and the 95th and 99th percentiles of focal depths (Magistrale 2002). The 95th percentile turned out to be 26.6 km in the southern subset and 27.5 km in the northern subset (Table 1) where the corrected heat flow is lower, 50.8 mW m$^{-2}$ compared to 55.7 mW m$^{-2}$ in the southern subset (Fig. 5). The mean depth of events is 13.0 and 14.3 km, respectively. These two subsets included more than half (804 + 1489 = 2293, 54.7 per cent) of all data.

In addition to studying Swedish coastal regions with high concentrations of earthquakes, we also investigated the Swedish Norrbotten and western Finnish Lapland with more scattered earthquake locations (area 3 in Fig. 2, from 66.5° to 69.0° N and 17.5° to 25.0° E). In area 3, the total number of 864 earthquakes appeared to be concentrated on two distinct heat flow regions, which very roughly represented the older lithosphere with low heat flow and the younger lithosphere with high heat flow. However, the scatter of data in latitude–longitude dimensions was obviously larger than in cases of areas 1 and 2. In area 3, the seismic cut-off depth was 27.8 km (95th percentile) and the mean depth of seismic events was 10.6 km. The palaeoclimatically corrected mean heat flow was 45.0 mW m$^{-2}$ (Fig. 5).

Except for the western Lapland and the Kuusamo region, no other large concentrations of seismic events are located in the Finnish territory. In the Kuusamo region (area 4 in Fig. 2, from 65.0° to 67.0° N and 27.0° to 31.0° E), the total number of events appeared to be 139, palaeoclimatically corrected heat flow 34.9 mW m$^{-2}$, basal earthquake depth 24.0 km and mean earthquake depth 15.1 km. Excluding Scania, southern Sweden (area 5 in Fig. 2, from 56.0° to 60.0° N and 11.0° to 17.0° E) had a high corrected heat flow (61.9 mW m$^{-2}$), but the mean depth of the total of 323 earthquake events in the area (14.2 km) and the cut-off depths resembled those of other regions.

The five regional subsets of data used in our analysis incorporate 3619 observations, 86.3 per cent of all earthquake events in Finnish
Table 1. Earthquake cut-off depths (km) corresponding to the 90th, 95th and 99th percentiles of data in our five subsets and mean values weighted using the number of data in each subset as a weight.

<table>
<thead>
<tr>
<th></th>
<th>1 (N = 804)</th>
<th>2 (N = 1489)</th>
<th>3 (N = 864)</th>
<th>4 (N = 139)</th>
<th>5 (N = 323)</th>
<th>Entire area (N = 4265)</th>
<th>Weighted mean</th>
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</thead>
<tbody>
<tr>
<td>90th</td>
<td>22.4</td>
<td>25.1</td>
<td>21.9</td>
<td>24.0</td>
<td>22.8</td>
<td>23.4</td>
<td>23.5 ± 1.4</td>
</tr>
<tr>
<td>95th</td>
<td>26.6</td>
<td>27.5</td>
<td>27.8</td>
<td>26.3</td>
<td>27.1</td>
<td>26.9</td>
<td>27.3 ± 0.5</td>
</tr>
<tr>
<td>99th</td>
<td>30.4</td>
<td>34.7</td>
<td>35.6</td>
<td>(31.1)</td>
<td>(31.1)</td>
<td>33.7</td>
<td>33.5 ± 2.1</td>
</tr>
</tbody>
</table>

Notes: Values for the entire study area of this paper, not limited to the five subsets, are also given. Parentheses indicate that less than five events with a greater focal depth than the cut-off depth have been registered in the area.

Figure 5. Palaeoclimatically corrected heat flow compared with the earthquake focal depth in five subsets of data: southern Gulf of Bothnia (area 1; red dots), northern Gulf of Bothnia (area 2; green dots), Norrbotten/Lapland (area 3; blue dots), Kuusamo (area 4; pink dots) and southern Sweden (area 5; cyan dots). Dotted lines, dashed lines and solid lines indicate the 90th, 95th and 99th percentiles of focal depths for all subsets, respectively. For interpretation of references to colour in this figure caption, the reader is referred to the online version of the paper.

and Swedish data sets combined. By weighting the obtained cut-off depths by the number of observations in these subsets, a mean estimate and standard deviations for the cut-off depth were calculated, being 23.5 ± 1.4 km for the 90th percentile, 27.3 ± 0.5 km for the 95th percentile and 33.5 ± 2.1 km for the 99th percentile of focal depths. The arithmetic mean of these values is 28.0 km and the standard deviation is 4.1 km. Using all 4190 earthquake events, including but not limited to all five subsets, results in cut-off depths which are within the standard deviations of data calculated from subsets only. As calculated from all 4190 events, the arithmetic mean cut-off depth was 28.0 km with a standard deviation of 4.3 km. The value of 31 km determined for the 99th percentile of focal depths by Kaikkonen et al. (2000), using Finnish data only, is therefore slightly smaller than our corresponding value of Finland and Sweden combined. Kaikkonen et al. (2000) also concluded that the wet crustal model for brittle-ductile transition works better than the dry one for the Finnish crust, although their assumed depth ranges for the frictional transition temperature of 350 °C is large (25–44 km). Despite the fact that a variety of lithologies in Fennoscandia, from granitic to ultramafic rocks, is represented in the range of epicentral coordinates, our data indicate no obvious relationship between heat flow and seismic cut-off depth. Our smallest data sets, namely those of areas 4 and 5, produce results relatively similar to those of the three larger data sets. Therefore, the entire range of seismogenic zone appears to be relatively well constrained also in the smaller data sets, and we suggest that 28 ± 4 km can be applied to approximate the depth of 350 °C isotherm in general models of the Fennoscandian lithosphere. The temperature corresponds to the change from elastic-brittle deformation to crystal plasticity of quartz, the rheologically weakest major rock-forming mineral in the crust (Scholz 1998).

Fig. 6 shows the depth histograms for all subsets in our study. Mean depths of seismic events within subsets range from 10.6 km in northern Sweden to 15.1 km in Kuusamo region, a variation almost similar to that observed in the cut-off depth. The large number of shallow events in northern Sweden is attributed to the plentiful small immature fault zones, while the seismic activity in areas 1 and 2 is concentrated on a smaller number of well-developed faults, leading
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Figure 6. Depth histograms of earthquakes in our subsets of data: southern Gulf of Bothnia subset (area 1), northern Gulf of Bothnia subset (area 2), Norrbotten/Lapland subset (area 3), Kuusamo subset (area 4) and southern Swedish subset (area 5).

Attempts to give isotherm depth estimates for Precambrian areas using heat flow data or spectral analysis of magnetic sources suffer from the non-uniqueness of methods applied, and from the approximate nature of various parameters used in the models. Fortuitously, Moho depth can be credibly estimated using structural seismology (Kinck et al. 1993; Grad et al. 2009, 2014; Silvennoinen et al. 2014), and in addition to seismic cut-off depth, serves as a useful thermal boundary in models if its temperature or heat flow is known. On the crust, heat flow is highly affected by near-surface sources due to strong differentiation of radiogenic heat sources in the crust with high heat production values in felsic upper crustal rock types and low values in mafic lower crustal rock types. In their analysis of Finland only, Kukkonen & Järvimäki (1991) applied Moho depths from seismic wide-angle surveys and estimated Moho heat flow values using a layer model and surface heat flow determinations available. Heat production was approximated with the heat production versus seismic P velocity ($A-V_P$) relationship (Rybach & Buntebarth 1984). Their range of Moho heat flow was 16.7–37.3 mW m$^{-2}$ with a mean of 21.9 mW m$^{-2}$. Using superdeep boreholes only, Kremenetsky et al. (1989) suggested that heat flow from the mantle is 22–24 mW m$^{-2}$. $A-V_P$ relationships have been shown to be only approximate, and to provide only trends which cannot be used for detailed estimation of crustal heat production (e.g. Kukkonen & Peltoniemi 1998; Kukkonen & Lahtinen 2001). More robust estimates based on geothermobarometry of mantle xenoliths in eastern Finland have provided much lower values, $12 \pm 3$ mW m$^{-2}$ (Kukkonen & Peltonen 1999; Kukkonen & Lahtinen 2001; Kukkonen et al. 2003). These are not derived from crustal heat flow and heat production constraints, thus providing independent estimates for geotherms and mantle heat flow estimation. For other Precambrian regions, such as Trans-Hudson orogen and Slave and Kalahari cratons, xenolith-based mantle heat flow estimates are also available, and their values are in the range of 12–25 mW m$^{-2}$ (Mareschal & Jaupart 2013).

For central Fennoscandia, Jokinen & Kukkonen (1999a,b) and Kukkonen et al. (1999) applied a generic model for uncertainty analysis of forward and inverse geotherm calculations. They used a model with a 50 km thick crust composed of three layers with constant (temperature-independent) thermal conductivities and heat productions as follows: 3.00 Wm$^{-1}$K$^{-1}$ and 1.8 $\mu$W m$^{-3}$ (0–10 km; layer 1), 2.75 Wm$^{-1}$K$^{-1}$ and 0.6 $\mu$W m$^{-3}$ (10–30 km; layer 2) and 2.50 Wm$^{-1}$K$^{-1}$ and 0.2 $\mu$W m$^{-3}$ (30–50 km; layer 3). Our forward modeling partly followed the same guidelines, although we applied a temperature-dependent continuous function to represent thermal conductivity (Cermak & Haenel 1988; eqs 1 and 2) and we also required that the 350 °C isotherm stays within the seismic cut-off depth range of $28 \pm 4$ km. Our models consisted of three crustal layers (upper, middle and lower layers) and an upper-mantle layer. The upper crustal layer (layer 1) was fixed at 0–10 km, while the depth of 28 km represented the boundary of middle (layer 2) and lower (layer 3) crust. As surface conditions in all models, we applied a steady-state heat flow value (corresponding to the palaeoclimatically corrected heat flow). Our three models had different surface heat flow values (40, 50 and 60 mW m$^{-2}$) and upper crustal heat production constraints ($1.1, 3.3$ and $2.1 \mu$W m$^{-3}$), respectively. In all models, thermal conductivity at the surface was 2.70 Wm$^{-1}$K$^{-1}$ and temperature was $T_0 = 3$ °C, which represents the annual mean ground surface temperature in the study area. In addition, we applied the average Moho depth of Fennoscandia, namely 46 km.
Heat flow and seismic cut-off depth and thermal conductivity of the crust and mantle in Fennoscandia.

1. Introduction

The study of heat flow and thermal conductivity in the crust and mantle of Fennoscandia is crucial for understanding the Earth's interior and its evolution. The heat flow in the crust is primarily generated by radioactive elements, while the mantle heat flow is due to the residual heat from the formation of the Earth. The seismic cut-off depth is the depth below which seismic waves are not generated, and it is related to the upper limit of the mantle.

2. Methods

The heat flow data were obtained from various sources, including borehole measurements and geothermal surveys. The thermal conductivity of the crust was calculated using the temperature-depth data from deep drill holes. The seismic cut-off depth was determined using seismological data.

3. Results

The average heat flow in the crust of Fennoscandia is 14 mW m\(^{-2}\), with a standard deviation of 6 mW m\(^{-2}\). The average thermal conductivity of the crust is 2.6 W m\(^{-1}\) K\(^{-1}\), with a standard deviation of 0.2 W m\(^{-1}\) K\(^{-1}\). The seismic cut-off depth is 28 km, with a standard deviation of 4 km.

4. Discussion

The heat flow and thermal conductivity results are in good agreement with previous studies, and they confirm the high heat production in the crust and the low heat production in the mantle. The seismic cut-off depth is consistent with the seismic data, and it indicates the presence of a thick lithosphere in Fennoscandia.

5. Conclusion

The study provides valuable insights into the thermal structure of Fennoscandia and its implications for the Earth's evolution. The results can be used in geological and geophysical modeling, and they contribute to our understanding of the Earth's interior.

Table 2. Temperature and heat flow at layer boundaries in our models.

<table>
<thead>
<tr>
<th>Temperature (°C)</th>
<th>Heat flow (mW m(^{-2}))</th>
<th>Thermal conductivity (W m(^{-1}) K(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Model 1</td>
<td>Model 2</td>
</tr>
<tr>
<td>Surface (0 km)</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>Boundary of layers 1 and 2 (10 km)</td>
<td>159</td>
<td>138</td>
</tr>
<tr>
<td>Boundary of layers 2 and 3 (28 km)</td>
<td>348</td>
<td>324</td>
</tr>
<tr>
<td>Moho (46 km)</td>
<td>494</td>
<td>468</td>
</tr>
<tr>
<td>LAB (250 km)</td>
<td>1583</td>
<td>1557</td>
</tr>
</tbody>
</table>

Notes: Heat flow was kept constant within each layer as explained in Section 4 of the text. Thermal conductivity was a continuous function of temperature in the crust, but constant in the mantle (between Moho and LAB). For corresponding geotherms, see Fig. 7.
Crustal heat production is also rather poorly known thermal parameter. It depends on the estimated concentrations of uranium (U), thorium (Th) and potassium (K), which are more abundant in accessory minerals than in main constituents of typical rocks (Jaupart & Mareschal 2003). In the Finnish rock geochemical database (Rasilainen et al. 2007), 6526 outcrops give a value of 1.33 ± 1.19 µW m⁻³, yet this is valid only in the immediate vicinity of surface, a problem typical of regional and global compilations. For Sweden, no corresponding data compilation has been published, but given the higher proportion of Proterozoic granitic rocks in the Swedish than in the Finnish lithosphere, we can assume the near-surface heat production in Sweden to be perhaps 0–20 per cent larger, although local variations are great. For example, the Sveconorwegian Bohus granite area features five-fold changes in thorium concentrations in scales ranging from tens of metres to kilometres (Landström et al. 1980). In general, near-surface heat production is more strongly dependent on rock type rather than on depth, and no obvious trend is visible.
in superdeep holes (e.g. Kremenentsky et al. 1989; Clauser et al. 1997).

Previous global estimates of the thickness of the upper crustal heat production layer range from 2 km to as much 16 km (Hasterok & Chapman 2011). Although simplistic models of linear heat flow–heat production relationship (Artemieva & Mooney 2001) suggest the heat producing layer, the ‘upper crust’, to be 8–10 km thick, this value, also referred to as ‘depth parameter’, is not related to any physical boundary. For Fennoscandian middle and lower crust, precise heat production constraints are not available, and for the time being, our method of using approximate values which fulfill boundary conditions is the only way to proceed. However, for southern Africa, Michaut et al. (2007) calculated a xenoliths-based heat production constraint of $0.34 \pm 0.08 \mu \text{W m}^{-3}$, corresponding to the depth range of 10–35 km. It appears that when heat production constraints in layers 2 (10–28 km) and 3 (28–46 km) of our models (0.6 and 0.2 $\mu \text{W m}^{-3}$) are weighted by depths of 18 and 7 km, we get a value of $0.49 \mu \text{W m}^{-3}$, which is close to the southern African estimate. However, the Moho heat flow, as determined from xenolith thermobarometry, is much larger (17–25 mW m$^{-2}$, Rudnick & Nyblade 1999) in southern Africa than in Fennoscandia, thus implying different conditions in the lithospheric mantle and asthenosphere.

Mantle heat flow has been directly determined from Kimberlite-hosted mantle xenoliths (Kukkonen & Peltonen 1999; Kukkonen et al. 2003), which provide an important lower boundary condition for geotherm calculations. The xenolith-based value is probably representative in a much larger area than the Eastern Finland kimberlite province, as suggested by the modeling of the effects of mantle heat flow variations on lithospheric isostasy and lack of major elevation contrasts in most of Finland, eastern Sweden, and NE Russia (Kukkonen 2014). It is conceivable that the thick lithosphere in Fennoscandian Shield is represented by relatively uniform and low-mantle heat flow. For a 250 km thick lithosphere used in our models, the characteristic time of heat diffusion is as much as 2 Ga, meaning that steady-state thermal models should be considered approximate (Michaut et al. 2007). In addition, the uncertainty in calculated temperatures increases quickly with depth. Kukkonen et al. (1999) carried out an uncertainty analysis for a simple four-layer lithosphere model, and the uncertainty of calculated temperature at 50 km may easily amount to $\pm 150 ^\circ \text{C}$ when the geotherm is calculated using the surface heat flow and ground surface temperatures as boundary conditions. The uncertainty can be reduced if independent data on the crustal and mantle temperature and heat flow can be incorporated in the modeling.

The accuracy of the depth estimation of earthquakes depends on various factors, such as the quality of the velocity model, the accuracy of picking the signals, the largest azimuthal gap from event to stations and distance to closest station. Especially, the shortest distance to station which produces a good quality S phase pick is a good constraint to earthquake depth. Gomberg et al. (1990) defined that to obtain an accurate depth estimate (error < 1.5 km), the closest station with S phase pick should lie closer than 1.4 times the depth from the source. In our study, the parameters which affect the accuracy of earthquake depth estimation are rather stable. The velocity models have been almost constant though different at different areas, and the azimuthal coverage of stations is good. The parameter with largest variation is distance to the nearest station. In our study areas 1, 2 and 5, the average distance to the nearest station is estimated to be 30 km. In area 3, the distance is 20 km and in area 4, 25 km, respectively. These estimates give the depths below which the accuracy of earthquake depth is at least 1.5 km. The threshold depth for areas 1, 2 and 5 is 21 km, for area 3, 14 km and for area 4, 18 km, respectively. Above the threshold depths, the accuracy is $> 1.5$ km, but it varies within study areas. The estimates apply best to sites with largest concentrations of earthquakes. For some smaller regions, independent depth accuracy estimates are available. For a local network in Kalajoki, partly in our area 2, Tiira et al. (2016) estimated accuracy of hypocentre depth to be 3–4 km without any analysis of accuracy variation with depth. Similar single value estimates were given by Lindblom (2011) for Pärvie fault in our area 3 (4 km) and by Uski et al. (2012) for Kuusamo region in our area 4 (also 4 km). It appears from Fig. 8 that deep earthquakes are rare in areas with corrected heat flow greater than 60 mW m$^{-2}$. Deepest events (49.6, 49.5, 45.8, 43.2 and 42.8 km) are not situated in areas with smallest palaeo-climatically corrected heat flow (less than 48 mW m$^{-2}$), and their locations are very scattered in general. A visual inspection of Fennoscandian Moho depth maps (Grad et al. 2009, 2014) is enough to reveal their closeness to Moho.

Results presented in this paper are heavily dependent on the representativeness of the assumed 350 $^\circ \text{C}$ temperature at the seismic cut-off depth. The cut-off temperature depends on the rheology of the rocks, which is mainly controlled by mineral composition and presence of volatiles. Traditionally, the crustal strength and rheological behaviour has been seen as a question whether the rocks are brittle or ductile. According to this view, the prevailing deformation mechanism simply depends on which of the mechanisms provides the smaller yield strength value. This model implies that earthquakes may occur down to the base of the brittle layer. The brittle behaviour is described by Byerlee’s law (Brace & Byerlee 1966) which is not dependent on rock type, but rather on vertical stress, type of faulting, friction and pore pressure. The ductile rheology which typically follows a power-law function is strongly affected by the rock composition and temperature. In quartz-bearing rocks, the brittle-ductile transition takes place at about 350 $^\circ \text{C}$, but if the rock is more mafic the transition is at higher temperatures, and the rheologically important minerals are feldspar or amphiboles.

Earthquakes are not simply controlled by the brittle-ductile transition of a homogeneous medium. Instead, they usually occur in pre-existing faults or zones of weakness. Slipping of the fault is controlled by the frictional stability of the fault. Frictional stability defines the conditions where earthquakes occur. In granitic (quartz-rich) rocks, this takes place at temperatures below $ca. 300–350 ^\circ \text{C}$. Above this temperature, the fault is in a stable regime where the yield is continuous. On the other hand, at low temperatures, the fault may also be in a stable regime due to flow in a cataclastic or poorly consolidated material. If the fault has a poorly developed gauge (e.g. recently formed faults), the low-temperature stability may reach close to the surface, but usually extends to depths of some kilometres. Thus, earthquakes occur in the unstable frictional regime of the fault, and may extend in depth from a few kilometres to the depth of the 300–350 $^\circ \text{C}$ isotherm (Scholz 1998). Depending on the properties of the frictional properties of the fault, prevailing stress and deformation rates, earthquakes may occur also deeper than the traditional brittle-ductile transition would indicate. Therefore, the seismic cut-off depth is not a sharply defined boundary, but a transition affected by the rock mineral composition and the frictional instability of faults and temperature.

6 CONCLUSIONS

Our result that surface heat flow values and earthquake cut-off depths do not correlate has important implications. Assuming that
in the intraplate conditions of Fennoscandian Shield, the earthquake mechanisms represent the same rheological properties in different areas, the crustal temperature must be relatively uniform at the cut-off depth. Then, the observed surface heat flow variations must be due to variations in radiogenic heat production in the crust, and particularly in the upper crust. U, Th and K are strongly differentiated in geological processes, particularly in partial melting. Melting processes have transported and concentrated heat producing elements in the upper crust, and depleted the middle and lower crustal layers. On the other hand, mantle heat flow has been determined from kimberlite-hosted mantle xenoliths, which provide an important lower boundary condition for geotherm calculations. Moreover, the xenolith-based mantle heat flow is probably representative in a much larger area than the Eastern Finland kimberlite province, as has been suggested by the modeling of the effects of mantle heat flow variations on lithosphere isostasy and lack of elevation contrasts in most of Finland, eastern Sweden, and northeastern Russia. It is conceivable that the thick lithosphere in Fennoscandian Shield is represented by relatively uniform and low-mantle heat flow.

ACKNOWLEDGEMENTS

Toni Veikkolainen acknowledges funding from the Jenny and Antti Wihuri Foundation, Finland. We thank Kari Moisio and an anonymous reviewer for their helpful comments which improved the manuscript.

REFERENCES


Figure 8. Epicentral locations of deep earthquakes (focal depth over 25 km) on a contour map of palaeoclimatically corrected heat flow (mW m$^{-2}$).


SUPPORTING INFORMATION
Supplementary data are available at GJI online.

APPENDICES
The heat flow data used in the study has been gathered in the MS Word file AppendixA.docx and the earthquake data in the MS Excel spreadsheet AppendixB.xlsx, both included in zip package “appendix.zip”. Both palaeoclimatically corrected and uncorrected heat flow information has been tabulated. In the seismic data table, the Finnish and Swedish data sets accepted for the analysis, as well as the rejected data from both countries, are provided in the form of spreadsheets with explanations.

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