Physics of the Earth and Planetary Interiors xxx (2011) xxx–xxx

Contents lists available at ScienceDirect

Physics of the Earth and Planetary Interiors

journal homepage: www.elsevier.com/locate/pepi

Geothermal studies of the Outokumpu Deep Drill Hole, Finland: Vertical variation in heat flow and palaeoclimatic implications

Ilmo T. Kukkonen [a,*], Volker Rath [b], Liisa Kivekäs [a], Jan Šafanda [c], Vladimir Čermak [c]

[a] Geological Survey of Finland, P.O.B. 96, FI-02151 Espoo, Finland
[b] Dpto. Astrofísica y CC. de la Atmósfera, Facultad CC. Físicas, Universidad Complutense de Madrid, Madrid, Spain
[c] Institute of Geophysics, Czech Academy of Science, Bocni II, Prague, Czech Republic

A R T I C L E I N F O

Article history:
Received 4 August 2010
Received in revised form 13 June 2011
Accepted 13 June 2011
Available online xxxx
Edited by: M. Jellinek

Keywords:
Heat flow
Geothermal gradient
Thermal conductivity
Paleoclimatology
Fennoscandian Shield

A B S T R A C T

Detailed geothermal studies of deep drill holes provide insights to heat transfer processes in the crust, and allow separation of different factors involved, such as palaeoclimatic and structural conductive effects as well as advective fluid flow effects. We present high resolution geothermal results of the 2516 m deep Outokumpu Deep Drill Hole in eastern Finland drilled in 2004–2005 into a Palaeoproterozoic formation with metasedimentary rocks, ophiolite-derived altered ultramafic rocks and pegmatitic granite. The down-hole temperatures have been logged five times after end of drilling and extend to day 948 after drilling. The hole is completely cored (79% core coverage) and thermal conductivity measurements were done at 1 m intervals. The geothermal results on temperature gradient, thermal conductivity and heat flow density yield an exceptionally detailed data set and indicate a significant vertical variation in gradient and heat flow density. Heat flow density increases from about 28 – 32 mW m^{-2} in the uppermost 1000 m to 40–45 mW m^{-2} at depths exceeding 2000 m. The estimated undisturbed surface heat flow value is 42 mW m^{-2}. We present results on forward and inverse transient conductive models which suggest that the vertical variation in heat flow can mostly be attributed to a palaeoclimatic effect due to ground surface temperature (GST) variations during the last 100,000 years. The modeling suggests that the average GST was about −3 to −4 °C during the Weichselian glaciation. Holocene GST values are within ±2° from the present average GST in Outokumpu (5 °C). The topographic hydraulic heads and hydraulic conductivity of crystalline rocks are low which suggests that advective heat transfer in the formation is not significant. The slow replacement of fresh flushing water by saline formation fluids is observed in the hole, but it does not generate significant thermal disturbances in the logs. On the other hand, free sluggish thermal convection is present in the large diameter (22 cm) borehole, and temperature variations in the range of few mK to 0.01 K occur over times of minutes to tens of minutes. Theory suggests that convection cells are about as tall as the drill hole diameter, and thus the free convection is expected to generate only local thermal 'noise' not affecting the general geothermal results.

1. Introduction

A deep research borehole was drilled in 2004–2005 in eastern Finland by the Outokumpu Deep Drilling Project of the Geological Survey of Finland (GTK). The 2516 m deep hole 'Outokumpu R2500' was drilled into a Palaeoproterozoic metasedimentary, igneous and ophiolite-related sequence of rocks in a classical ore province with massive Cu–Co–Zn sulfide deposits (Fig. 1). The Outokumpu DDP (Deep Drilling Project) is carried out in an international framework, partly supported by International Continental Scientific Drilling Program (ICDP). The aims and results of the drilling project as a whole and the different subprojects are discussed in details in Kukkonen (2009, 2011). GTK used the company NEDRA (Jaroslavl, Russia) as the drilling contractor.

Geothermal studies are an essential component of the Outokumpu DDP. In geothermics, the near-surface disturbances should be avoided as far as possible, and therefore the depth reached in drilling and logging is always a value of itself. Only in a deep hole it is possible to directly recognize systematic vertical variations in temperature gradient and heat flow density due to long-term paleoclimatic disturbances, groundwater flow and structural effects. In the super-deep drilling project in the Kola Peninsula, Russia (Kremenetsky and Ovchinnikov, 1986) and the KTB holes in southern Germany (Clauser et al., 1997), an important argument in deciding the deep drilling sites was the low geothermal gradient suggested by shallow holes. However, the final hole temperatures and deep gradients measured in both drilling projects clearly

* Corresponding author. Tel.: +358 205502270; fax: +358 2015012.
E-mail address: ilmo.kukkonen@gtk.fi (I.T. Kukkonen).

0031-9201/$ - see front matter © 2011 Elsevier B.V. All rights reserved.
doi:10.1016/j.pepi.2011.06.002

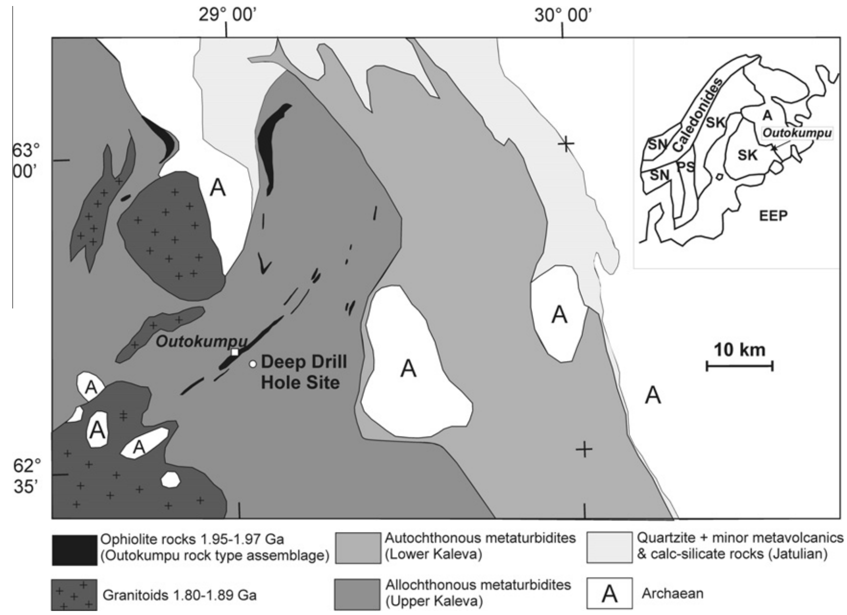

Fig. 1. Location of the Outokumpu Deep Drill Hole site on a simplified geological map of the Outokumpu area in eastern Finland. Inset shows the location of Outokumpu in the Fennoscandian Shield. (A) Archaean (>2.5 Ga); SK: Svecokarelian (2.5–1.75 Ga); PS: Post-Svecokarelian (1.7–1.2 Ga); SN: Sveconorwegian (1.8–0.9 Ga); EEP: East European Platform. Adapted from Huhma (1971), Papunen and Gorbunov (1985), Koistinen et al. (2001) and Peltonen et al. (2008).

exceeded the predictions. If such results would represent the common crustal conditions, they would have dramatic implications for correcting and using shallow (<1 km) heat flow data for calculating crustal and lithospheric temperatures. Ever since the geothermal results of the Kola hole became known in the late 1980s (Kremenetsky and Ovchinnikov, 1986), there has been a discussion on the existence and causes of vertical heat flow variation in the continental upper crust as well as need for and reporting of new experimental data (Kukkonen and Clauser, 1994; Clauser et al., 1997; Kukkonen et al., 1997, 1998; Popov et al., 1998, 1999a,b; Mareschal et al., 1999; Kukkonen and Jõeleht, 2003; Glaznev et al., 2004; Šafanda et al., 2004; Majorowicz and Šafanda, 2007; Mottaghy et al., 2005; Majorowicz and Wybraniec, 2010).

From the geothermal viewpoint, the Outokumpu Deep Drill Hole provides one of the deepest undisturbed temperature and heat flow density profiles in the uppermost 2.5 km. In the Fennoscandian Shield, there are not many boreholes reaching such depths with detailed geothermal observations. Holes down to about 1 km level are relatively common, but in Fennoscandia holes deeper than the Outokumpu hole are only the Kola super-deep hole in Russia (12.3 km; e.g., Kremenetsky and Ovchinnikov, 1986), the two Siljan holes (6 and 7 km; Balling et al., 1990) in central Sweden, and the recently finished Onetshkaya hole in Russian Karelia (3.5 km; Y.M. Erinchek, written communication, 2008). Detailed geothermal results have so far been published only on the Kola super-deep (Popov et al., 1999a,b). The Kola super-deep hole indicated that there is a significant increase in heat flow density in the uppermost 1–3 km of crust (Kremenetsky and Ovchinnikov, 1986). Similar results have been obtained in the KTB deep holes in Germany (Clauser et al., 1997), in the Udryn and Torun holes in Poland (Šafanda et al., 2004; Majorowicz and Šafanda, 2007), and in the central Kola Peninsula (Glaznev et al., 2004). As the common heat flow maps (e.g., Hurtig et al., 1992) are based to a great deal on holes shallower than 1 km, the vertical variation in heat flow is a crucial issue for applications of geothermal data in crustal studies.

Our motivations for the geothermal investigations in the Outokumpu hole are (1) to investigate whether there is a vertical variation in heat flow density, and if so, to determine the magnitude of the variation as well as to find out what are the factors causing such variation; (2) to study the effects of past climate on the subsurface thermal regime and to invert ground surface temperature histories covering the last 100,000 years; (3) to study to what degree the thermal conditions may be disturbed by advective heat transfer (fluid flow); and (4) whether there is free thermal convection of fluid in the large diameter drill hole affecting the high resolution temperature logs.

We present geothermal results based on nine temperature logs measured during and after drilling, and on more than 1900 thermal conductivity measurements of the drill core as well as theoretical modeling of the data.

2. The Outokumpu drilling site and structures intersected

The Outokumpu Deep Drilling site is located 2.5 km SE of the old Outokumpu mining town in eastern Finland at 62° 43′ 02.63″N, 29° 03′ 55.01″E (Fig. 1). The area is well-known for its massive and semi-massive Cu–Co–Zn sulfide deposits mined from about 1910 to 1984. The Outokumpu belt is one of the oldest (1.95 Ga) known ophiolite systems (Säntti et al., 2006; Peltonen et al., 2008). One of the aims of the deep hole was to reveal the geological nature of the strong seismic reflectors dominating the upper crust in the Outokumpu area (Kukkonen et al., 2006; Kukkonen, 2011). The deep hole was targeted to intersect one of the uppermost strong reflectors which proved to be packages of ophiolite-derived rock types (serpentinite, diopside–tremolite skarn, quartz rock, i.e., the typical host rocks of the Outokumpu type sulfide deposits) and black schist in a metasedimentary host rock (metaturbiditic mica schist). At deeper levels of the hole pegmatitic granite dominates the drilled section (Fig. 2).

The hole was drilled with continuous coring, and the final core coverage was 79%, with the main core losses occurring at zones of heavy fracturing and shearing. The most common rock types in the deep hole are mica schist and pegmatitic granite, which form together more than 90% of the drilled section. Rock types of the ophiolite-related Outokumpu assemblage form about 8% of the section (Fig. 2). The seismic data suggested that the structures are mostly

Please cite this article in press as: Kukkonen, I.T., et al. Geothermal studies of the Outokumpu Deep Drill Hole, Finland: Vertical variation in heat flow and palaeoclimatic implications. Phys. Earth Planet. In. (2011), doi:10.1016/j.pepi.2011.06.002

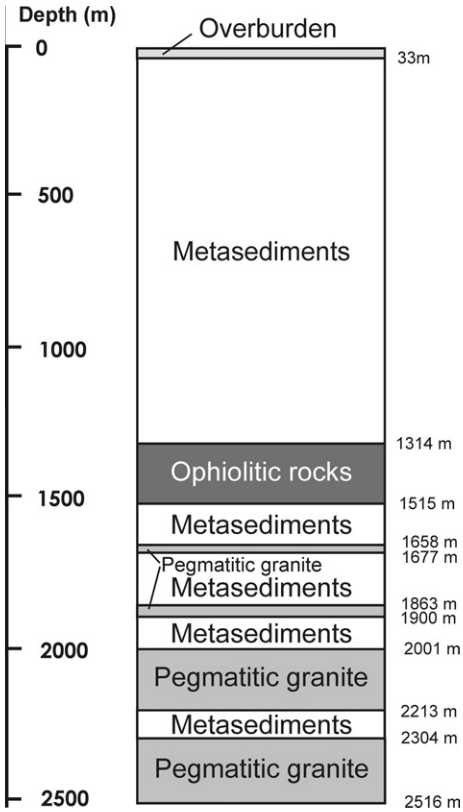

Fig. 2. The main lithological units of the Outokumpu Deep Drill Hole.

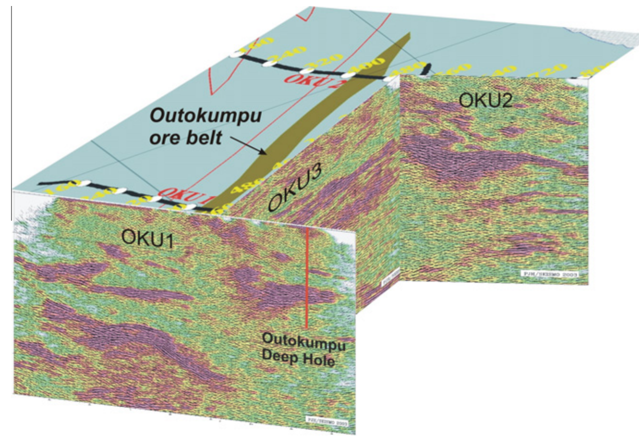

Fig. 3. 3D-presentation of high-resolution seismic lines in Outokumpu. The model represents migrated data by the FIRE project (Kukkonen et al., 2006; Heikkinen et al., 2007) and a lithological surface map. The seismic sections extend to 5 km depth. The amplitude scale of the reflectors has been color-coded for indicating higher reflection amplitudes. Layers of strong reflectivity (lilac tones) are associated to the ophiolite-derived rocks types and poorly reflective areas to mica schist and pegmatitic granite. The 2D forward geothermal model (Fig. 8) was constructed using the OKU1 seismic section and the crustal scale FIRE3 profile running on the same transect. Seismic data processing: P. Heikkinen, Institute of Seismology, University of Helsinki.

horizontal or dipping at very low angles at the drilling site which was confirmed by the drill core. 3D-visualization of the high-resolution seismic data further suggests that the structures are quite two-dimensional in the direction of the main ore belt (Fig. 3).

The hole is cased to the depth of 39 m (casing inner diameter 324 mm), below which the hole is uncased and has a nominal diameter of 220 mm. The casing was cemented to the sediments and bedrock. Local tap water was used as drilling mud which was occasionally thickened with peat and bentonite to increase its viscosity at zones of unstable rocks to generate a mud-cake on the walls (especially at 520–850 m). Cementing was used at 75–274 m to stabilize the hole walls.

The drilling phase took 302 days (April 6, 2004–January 31, 2005). The temperature of the drilling mud, which was circulated back to the hole after removal of cuttings, was monitored daily. Also the total consumption of water on the drilling site was recorded (Ahonen et al., 2011).

3. Down-hole logging of *in-situ* temperature and laboratory measurements of thermal properties

Temperature logs were obtained in several stages during and after drilling (Table 1). During drilling breaks at every 500 m of accomplished depth the hole was logged by the NEDRA logging team with about 20 different geophysical down-hole tools, one of them temperature. This way four continuous temperature logs with reading interval of 0.2 m were obtained after shut-in times of 1–2 days. The applied temperature tool had a resolution of 0.01 K and it was calibrated by the contractors to be better than 0.1 K. At end of drilling the drill hole was flushed clean with tap water, and the temperature was logged after 7 days of shut-in time by the NEDRA team. During shut-in time during drilling breaks, the down-hole logs and fluid sampling were the only activities in the hole.

Post-drilling temperature logs were obtained first in August 31, 2005 (211 days shut-in time) by the geothermal team of the Geophysical Institute of the Czech Academy of Sciences. The used temperature probe was a memory logger probe (Antares GmbH) with a resolution of 0.001 K and an absolute accuracy of better than 0.05 K. The log was obtained by point-by-point readings at 5 m intervals. This log extended to only 1450 m depth. Thereafter

Table 1
Temperature logs of the Outokumpu Deep Drill Hole.

Date	Activity	Drilling depth reached (m)	Shut-in time (days)	Maximum logging depth (m)	Logger
April 6, 2004	Start of drilling	0			
May 24, 2004	Logging	550	1	550	NEDRA
July 15, 2004	Logging	997	1	997	NEDRA
September 18, 2004	Logging	1507	2	1507	NEDRA
November 11, 2004	Logging	2073	1	2073	NEDRA
January 31, 2005	End of drilling	2516			
February 7, 2005	Logging	2516	7	2516	NEDRA
August 31, 2005	Logging	2516	211	1418	GI-Prague
April 6, 2006	Logging	2516	433	2508	ICDP-OSG
September 20, 2006	Logging	2516	597	2507	ICDP-OSG
September 8, 2008	Logging	2516	948	2503	ICDP-OSG

Please cite this article in press as: Kukkonen, I.T., et al. Geothermal studies of the Outokumpu Deep Drill Hole, Finland: Vertical variation in heat flow and palaeoclimatic implications. Phys. Earth Planet. In. (2011), doi:10.1016/j.pepi.2011.06.002

ARTICLE IN PRESS

several temperature logs were measured by the logging team of the ICDP Operational Support Group in 2006 and 2008. These were done on April 6, 2006 (433 days shut-in time), September 20, 2006 (597 days) and September 5, 2008 (948 days). The used mud-parameter tool measures simultaneously temperature and fluid electrical resistivity. The continuous logs measured with reading interval of 0.1 m have a resolution of 0.01 K and an absolute accuracy of better than 0.1 K. The temperature logging history is summarized in Table 1.

Other activities in the hole during the post-drilling time included sampling of the drill hole water with a tube method in 2007 and 2008 (Ahonen et al., 2011), and hydraulic difference-flow measurements (in the uppermost 1290 m in August 2007; Sokolnicki and Heikkinen, 2008). These operations lasted less than 4 weeks altogether, and are considered to have had no detectable effects on the temperature logs.

In addition to the down-hole data, the water table level is monitored automatically by GTK in the deep hole as well as the ground temperatures and water levels in a shallow (17 m deep) hole in the Quaternary overburden (Hänninen et al., 2009). This data provided useful additional information on the hydrogeological condition of the hole and shows that the deep hole water is not in a hydraulic connection with the shallow sediments.

Thermal conductivity was measured on drill core samples at about 1 m intervals with the total number of samples being 1922. The laboratory measurement required drilling of a new specimen from the original 6–10 cm diameter cores, and finally 7 mm thick disks with 42 mm diameter were cut and polished. The disks were immersed in tap water for minimum duration of 2 days to saturate open pores and micro-fractures, and thermal conductivity was subsequently measured at room temperature with a divided bar instrument constructed at GTK (Kukkonen and Lindberg, 1998). The instrument applies quartz references (conductivity in direction of crystallographic *a*-axis) in the bar and the measurements are controlled by comparison with low-conductivity Duran glass and high conductivity quartz (conductivity in direction of crystallographic *c*-axis). The inaccuracies of the measurements are smaller than 5% (Kukkonen and Lindberg, 1998).

The specific heat capacity was measured with a calorimetric method using a subset of conductivity disks at about 100 m depth intervals. The samples were pre-heated in boiling water, and then immersed into room temperature water in a calorimeter. The measurements have inaccuracies below 5%, and the results were corrected to ambient temperature (Kukkonen and Lindberg, 1998). Bulk density for diffusivity determination was determined by the Archimedean principle by weighing the samples in water and air. The obtained accuracy is better than about 0.2% for samples with masses of ca. 25 g (typical for conductivity disks used).

4. Results: temperature gradient, thermal conductivity, heat production and heat flow density

The logs taken during drilling breaks show a very strong deviation from equilibrium temperatures. The ski-shaped logs differ by about 10–15° at maximum from undisturbed values, excluding the lowermost readings of each log, which were least disturbed by mud circulation (Fig. 4). The post-drilling logs show a rapid equilibration towards undisturbed formation temperatures, and the latest temperature log measured in September 2008 is already very close to equilibrium. In the uppermost parts of the hole where the disturbance is biggest, numerical modeling suggests that temperatures are already within ~0.1 °C of the equilibrium temperatures.

At end of drilling, the borehole was flushed clean and filled with fresh tap water. Subsequently, the electrical conductivity (representing fluid salinity) of the drilling fluid has gradually increased which suggests slow inflow of formation fluid from a restricted number of fractures (Fig. 5). However, the rate of fluid flow is very small, and does not seem to play a role in heat transfer that could be seen in the temperature logs.

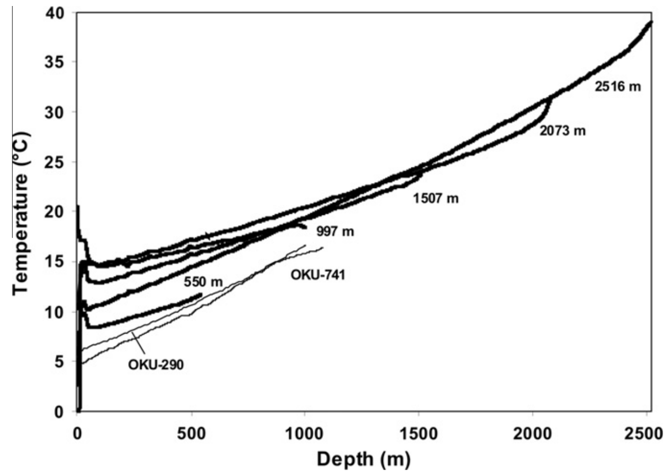

Fig. 4. Temperature logs obtained during drilling breaks and immediately after drilling. For dates of the logs and shut-in times, see Table 1. The logs 'OKU-290' (logged in 1963; Järvimäki and Puranen, 1979) and 'OKU-741' (logged in 1985; Kukkonen, 1988), are from other shallower holes of the Outokumpu area showing typical (undisturbed) formation temperatures in the uppermost 1 km.

Temperature gradient shows a distinct vertical increase from 13 – 14 mK m^{-1} at 50 m–1.3 km (in the upper mica schist unit), lower but more variable values of 10–13 mK m^{-1} at 1.3–1.5 km (in the ophiolite-derived rocks), and about 16 mK m^{-1} in the lower mica schist unit and pegmatitic granite at 1.5–2.2 km. In the lowermost part of the hole at 2.2–2.5 km temperature gradient shows lower but variable values in the range of 13–16 mK m^{-1} (Fig. 6 and Table 2).

Thermal conductivity shows small-scale variation with typical standard deviation of about 0.5 Wm^{-1} K^{-1} which we attribute to normal geological heterogeneity between small samples. However, when a moving average (20-m window) is calculated, thermal conductivity is very stable in the mica schist and pegmatitic granite (Fig. 6). In the mica schist average thermal conductivity is 2.5 Wm^{-1} K^{-1} (Table 3). The foliated rock is thermally anisotropic, but the (sub)horizontal orientation of foliation in the drilled section results in relatively low conductivity values as the conductivity is measured in a direction perpendicular to schistosity and foliation. In the pegmatitic granite, average conductivity is 3 Wm^{-1} K^{-1}. Higher but variable values of conductivity are associated to the Outokumpu assemblage of rock types with peaks up to 10 Wm^{-1} K^{-1} but the average is 3.35 Wm^{-1} K^{-1}. The generally elevated values are due to tremolite and diopside bearing skarn rocks and the peak values to the presence of talc in serpentinites as well as graphite in black schists enveloping the ophiolitic rocks.

Radiogenic heat production was determined according to Rybach (1973) from the neutron density log and the U, Th and K channels of the natural gamma ray spectrometric logs measured by NEDRA. The results show a practically constant heat production value in the mica schist (mean 1.7 μW m^{-3}), high values in the pegmatitic granite (mean 5.4 μW m^{-3}), and sharp variations in the altered ultrabasic rocks of the ophiolite-derived rock types. Namely, the serpentines have practically negligible heat production levels whereas the skarn rocks show values up to 6–7 μW m^{-3}. The average heat production of the ophiolitic rocks is 1.55 μW m^{-3} (Table 3).

When combined with temperature measurements, a very detailed heat flow density profile could be constructed (Fig. 6). Heat flow density was determined in two ways, first, as the product of the moving averages (20 m windows) of thermal conductivity

Please cite this article in press as: Kukkonen, I.T., et al. Geothermal studies of the Outokumpu Deep Drill Hole, Finland: Vertical variation in heat flow and palaeoclimatic implications. Phys. Earth Planet. In. (2011), doi:10.1016/j.pepi.2011.06.002

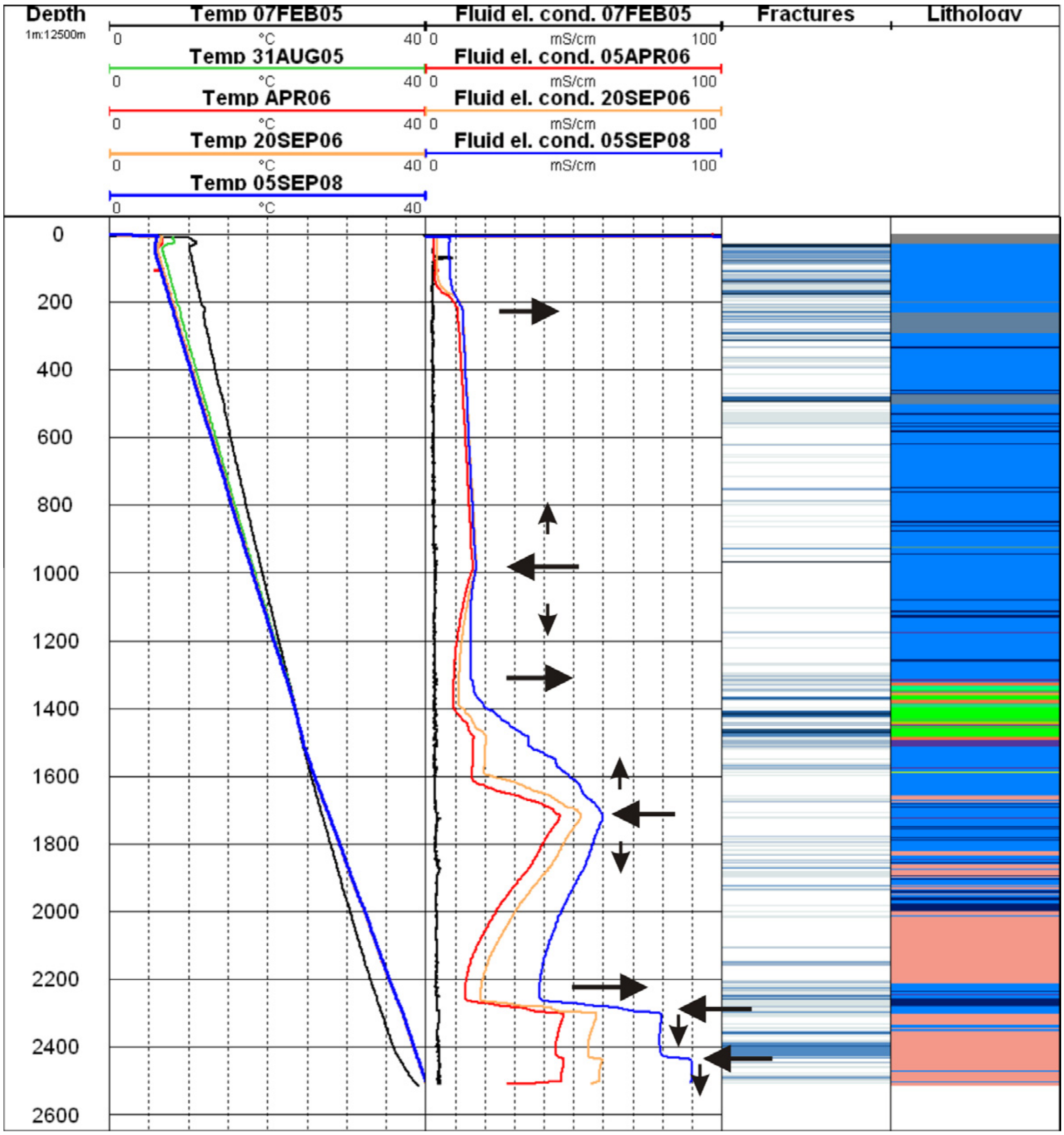

Fig. 5. Temperature and fluid electrical conductivity logs in the Outokumpu Deep Drill Hole in 2005–2008. Arrows pointing to the left indicate depths of saline formation fluid flowing into the hole, and arrows to the right out from the hole, respectively. Up and down pointing arrows indicate the flow in the hole. 'Fractures' column indicates interpreted fractures from sonic, galvanic and calliper logs (adapted from Tarvainen (2006)). 'Lithology' shows the rock types (blue: metasediments; green and orange: ophiolite-derived serpentinite and skarn rocks; pink: pegmatitic granite). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this paper.)

and temperature gradient, and second, as the product of average conductivity and gradient calculated in 100 m depth sections (Fig. 6 and Table 2). A systematic vertical increase in heat low density is observed. Heat flow increases from about 30 mW m^{-2} in the uppermost 1 km to about 40–45 mW m^{-2} at depths beneath 1.5 km.

5. Conductive vs. advective heat transfer

5.1. Flow systems between borehole and fractures

Generally, the Outokumpu temperature logs seem to represent a thermal regime characterized mainly by conductive heat transfer with practically no apparent indications of fluid flow disturbances in the hole. Nevertheless, there is slow flow of fluid into and out from the hole. This is indicated by the fluid electrical conductivity logs (Fig. 5). The fresh fluid used in final flushing of the hole at end of drilling had an electrical conductivity of about 2.5 mS/cm. The subsequent repeated logs of fluid electrical conductivity indicate gradual changes in borehole fluid salinity and suggest slow inflow of formation fluid from a restricted number of fractures (Fig. 5). The driving force of this flow is very likely the density difference between the fresh flushing water and the saline formation fluids. Density is about 1.00 g/cm^3 in the fresh and 1.04 g/cm^3 in the most saline fluids, respectively.

The fluid electrical conductivity seems to have reached a semi-stable condition at the uppermost parts of the hole, but in the lower parts the salinity keeps increasing even at the time of the latest log (September 2008, 948 days after drilling). A qualitative

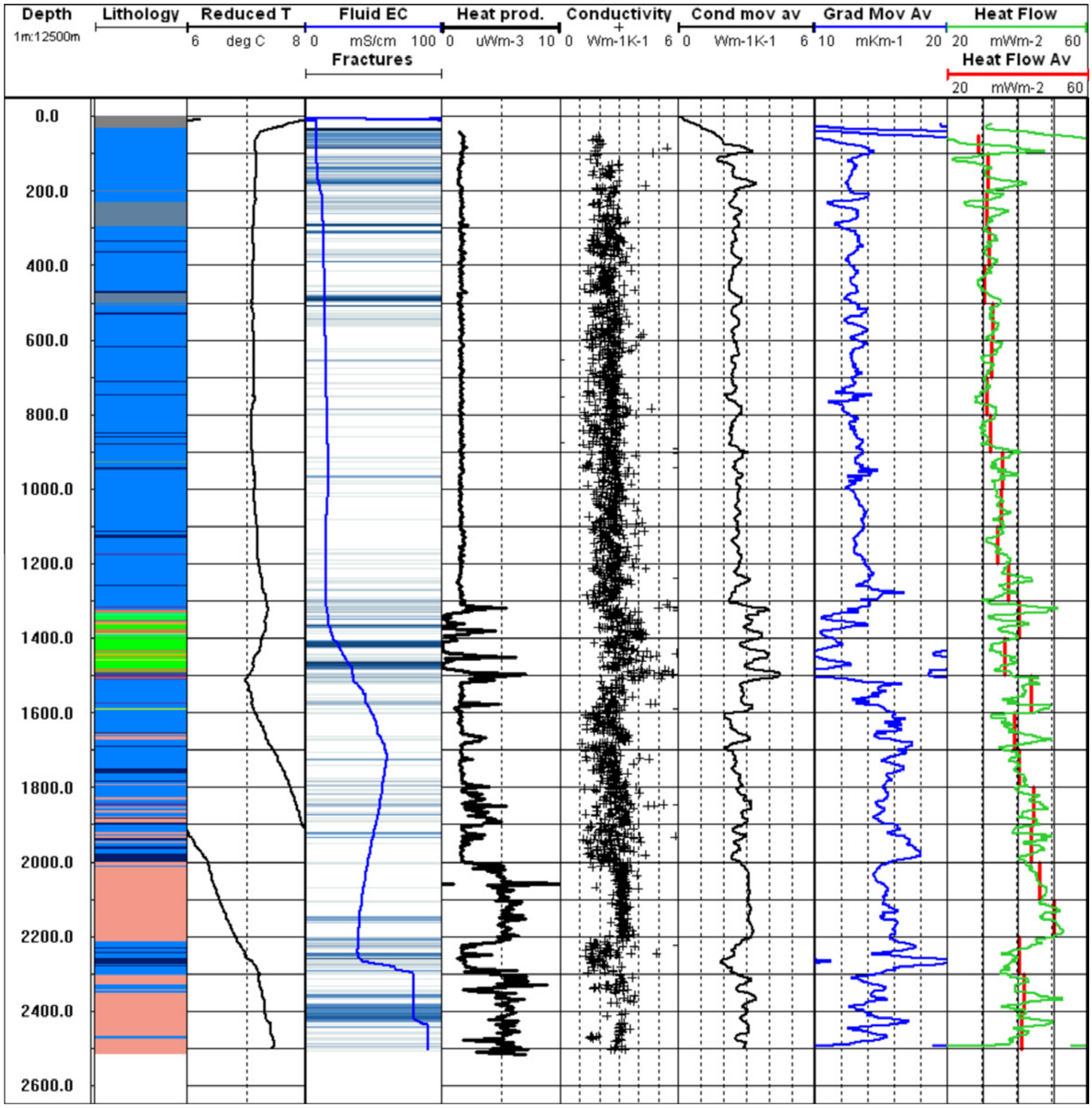

Fig. 6. Outokumpu Deep Drill Hole geothermal results. From left: lithological column (according to Västi, 2005; blue: metasediments; green and orange: ophiolite-derived serpentinite and skarn rocks; pink: pegmatitic granite), reduced temperature (log of September 2008, reduction gradient 13 mK m^{-1}), electrical conductivity of drill hole water with the hydraulically conductive fractures estimated from density, acoustic and galvanic logs ('Fractures', adapted from Tarvainen (2006)), radiogenic heat production rate calculated from density and gamma ray U–Th–K logs, thermal conductivity measured in the lab, moving average of thermal conductivity (20 m window), moving average of temperature gradient (20 m window) and calculated heat flow density (green line: heat flow from moving averages of conductivity and gradient; red line: average heat flow calculated at 100 m intervals). See also Table 2. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this paper.)

interpretation of the flow from and into fractures is presented in Fig. 5. At the depths of the most important in-flow points, the temporal fluid salinity changes show the development of mixing. At the time of the last log in 2008 the salinity seems to have already stabilized at 967 m, whereas at 1715 m the salinity is still slowly increasing, but at 2300 and 2435 m the salinity keeps increasing more or less linearly with time. This indicates the slow replacement of fresh drill hole water by the saline formation fluids, and also implies that the formation fluids become more saline with depth. The order of magnitude of flow rate into the hole can be estimated by calculating the dilution of the fluids using the depth ranges of the indicated flow systems along the hole (∼100–1000 m, Fig. 5), the corresponding fluid volumes and time past after the end of drilling. These calculations suggest that the flow rates of formation fluid into the hole are in the range of 0.4–0.04 ml/s.

The thermal effects of flow between the formation and borehole have been studied theoretically by Ramey (1962), Drury and Lewis (1983) and Drury (1984). At the depth of flow into the hole the temperature anomaly is very gradual, particularly if the fluid has no temperature contrast with the prevailing formation temperature. When flowing along the hole, the fluid heats or cools depending on the geothermal gradient, flow rate and direction, and the thermal properties of fluid and rock. As a result a deviation from undisturbed temperatures is developed. However, at the depth of flow out from the hole, a sudden change in temperature is observed.

The flow of fluid at the estimated rates of 0.4–0.04 ml/s in the 22 cm diameter Outokumpu hole would generate temperature anomalies (jumps) smaller than 2 mK (Drury, 1984; Eqs. (1)–(3)) at the in/outflow depths. In order to detect the flow in temperature logs would require anomalies clearly exceeding the temperature resolution (0.01 K). This would mean flow rates of the order of 4 ml/s which would generate temperature effects of about 0.02 K. Such anomalies could be detected in principle. However, the flow rate of 4 ml/s would be in contradiction with the slow time-dependent changes in fluid salinity.

At the depths of the major in/outflow indicated by the fluid electrical conductivity logs (Fig. 5) we do not observe any distinct changes in the temperature logs. The calculated gradient logs, on the other hand, seem to show local maxima and minima at about 230, 950–1000, 1700–1750, 2250, 2300 and 2430 m (Fig. 6), which may be attributed to flow effects. In addition, the gradient changes do not seem to correlate with local conductivity variations at these depths.

We conclude that although minor flow systems exist in the hole, they are driven by the density differences due to replacement of fresh drill hole water by saline formation fluids. The flow does not have any relevant effects on the thermal data. The same applies to the effects of free thermal convection of the drill hole fluid (see Section 6).

5.2. Flow in the formation

Flow in the formation is theoretically possible, but flow systems able to generate the detected vertical heat flow variation are not likely. In an earlier study, possible advective heat transfer effects were estimated for a similar type of a structure in Sukkulansalo located in the Outokumpu belt at a distance of about 20 km NE from the deep hole (Kukkonen and Šafanda, 1996). Hydraulic *in-situ* permeabilities of the rock types measured in the uppermost 1 km were in the range of 1×10^{-17} to 1×10^{-16} m^2 for mica gneiss, serpentinite, skarn rock and talc schist, and in the range of 1×10^{-14} to 1×10^{-15} m^2 for black schist (Ahonen et al., 1991; Ahonen and Blomqvist, 1994). Mainly due to the small topographic variation, and low hydraulic conductivities of the rock, the Peclet numbers of advective heat transfer estimated and modeled by Kukkonen and Šafanda (1996) with 2D analytical and numerical models were of the order of 10^{-4}–10^{-5}. In the Outokumpu deep hole, drill stem tests were carried out during drilling breaks after each 500 m of drilling in 40–70 m long test sections at hole bottom (Ahonen et al., 2011). The results indicated a relatively high value (0.7×10^{-12} m^2) at 480–580 m and a moderate value (5×10^{-14} m^2) at 957–997 m, whereas all deeper test sections were observed to be practically impermeable. A quantitative limit of impermeability is not given by Ahonen et al. (op. cit.), but is likely around 10^{-16}–10^{-17} m^2 in the applied test method. Unfortunately, the drill stem tests do not provide a continuous profile of permeability data with depth, and furthermore, the two uppermost test sections coincide with well-developed fracture systems detected in the drill core as well as in galvanic, sonic and caliper logs (Ahonen et al., 2011; Tarvainen, 2006, Fig. 5). Thus, these values probably overestimate the permeability of the formation. Anyway, the results imply a rapidly decreasing permeability with depth in the deep hole. In the uppermost 200 m fracturing is common in the deep hole (Fig. 5).

The possible contribution of advection was estimated with dimensional Peclet number analysis (van der Kamp and Bachu, 1989; see also Kukkonen (1995) for application of the method in Finnish conditions) assuming a hydraulic gradient of 4×10^{-3} determined from local topography and lake levels in a NW–SE profile across the deep hole site. The Peclet number (Pe) gives the ratio of heat transported by advection to that transported by conduction and can be calculated as (van der Kamp and Bachu, 1989):

$$Pe = \beta\kappa \frac{dh}{dl} DA/\alpha_m \quad (1)$$

where β is the ratio of volumetric heat capacities of the fluid and the medium (about 1.83), κ is the hydraulic conductivity, dh/dl is the hydraulic gradient, D is the thickness of the aquifer, A is the aspect ratio of the aquifer ($A = D/L$, where L is length) and α_m is the diffusivity of the medium (about 1×10^{-6} $m^2\,s^{-1}$). Peclet number values exceeding about 0.1–0.2 would indicate a detectable level of advective heat transfer. Assuming that the flow system (i.e., permeable rock) in Outokumpu would extend to a depth of 1 km and the flow distance were 10 km, the Peclet number reaches a value of 0.1 only if permeability exceeds 1.5×10^{-13} m^2. However, we do not consider such permeability values representative of the rocks in the Outokumpu belt. The chances of advective heat transfer effects in the formation are further reduced by the overall increase of fluid salinity (and density) with depth which tends to debilitate any potential flow systems. The heat transfer is therefore dominated by conduction. Similar results on advective heat transfer contribution were obtained by running coupled finite difference conductive–advective heat and fluid flow models (results not shown here).

6. Free convection of fluid in the borehole

When pulling out the memory-logger temperature probe from the hole in August 2005, it got temporarily stuck at the depth of 1390 m. The probe (resolution 1 mK, manufactured by Antares GmbH, Germany) continued logging temperatures for several days thereafter until the memory was full. Due to this incident, we obtained a high-resolution time series of temperature (Fig. 7), which clearly evidences continuous temperature variations with amplitude of few mK taking place in periods ranging from minutes to several hours. Instrument noise can be excluded as a reason for the variations and they must be attributed to natural phenomena in the borehole. Time dependent variations in borehole water table (and therefore temperatures recorded by a stationary probe) can be generated by several factors, namely tidal effects (Bredehoeft, 1967; Hsieh et al., 1987; Cutillo and Bredhoeft, 2010), barometric pressure variations (Spane, 2002) and free convection of fluid in the borehole (Hales, 1937; Čermak, 2009; Čermak et al., 2008a,b,c; Eppelbaum and Kutasov, 2011). The deep hole water table is automatically monitored with a pressure sensor at 20 m depth. In addition the water table is monitored in a 17 m deep well in the Quaternary sandy-silty sediments located at about 20 m from the deep hole. Atmospheric pressure is also monitored on the site (Hänninen et al., 2009). The water table variations in the deep hole and the shallow hole are not correlated which indicates that the shallow Quaternary aquifer is not hydraulically connected to the deep hole.

Tidal effects are present in the deep hole water table variations. Spectral analysis of data obtained (during a technically quiet period with no logging or fluid pumping activities) between November 11, 2007 and April 17, 2008 with 30 min recording intervals shows distinct periodicities at frequencies 1.00, 0.93, 1.93 and 2.00 cpd (cycles per day). They correspond accurately with the lunar-solar, principal lunar and principal solar tidal components (Munk and MacDonald, 1960). Groundwater table variations corrected for atmospheric pressure variations show peak-to-peak amplitude of 2–6 cm (Hänninen et al., 2009). Corresponding movement of the water column in the borehole can be expected to generate temperature variations smaller than 1 mK.

Barometric effects also generate water table variations (Spane, 2002). As they are driven by meteorological phenomena their characteristic periodicities are about one order of magnitude longer than the recorded temperature variations. As the tidal and barometric pressure effects do not explain the temperature variations, we investigate the possibility for free convection.

Based on our former experience (Čermak et al., 2008a,b) free convection is indeed present in large diameter boreholes. When a long fluid-filled column is subjected to a thermal gradient, free thermal convection may occur as soon as the temperature gradient exceeds a certain critical value. Such borehole fluid instability was originally described and theoretically treated by, e.g., Hales (1937)

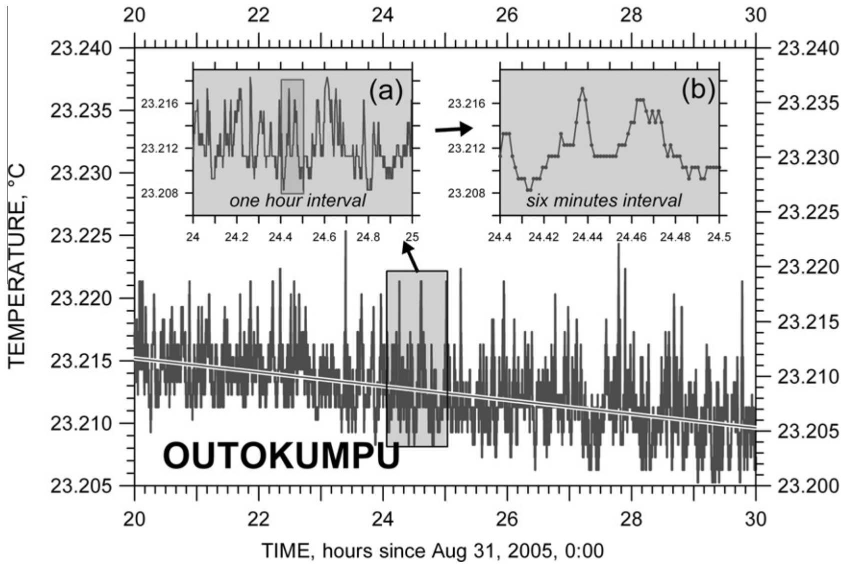

Fig. 7. Temperature recorded by a stationary temperature probe located at 1390 m. Sampling interval 5 s. Note the detailed oscillatory character of the temperature time variations in both inserted figures demonstrating the enlarged "1-hour" and/or "ten-minutes" time intervals. Slightly decreasing temperature corresponds to the unsettled probe temperature equilibration.

and reported to exist only in large diameter wells.The onset of free convection depends mainly on the borehole diameter and temperature gradient. The critical temperature gradient value G_c, above which convection takes place (Hales, 1937), can be calculated from

$$G_c = \frac{g\alpha T}{C_f} + \frac{16Bvk}{g\alpha d^4} \tag{2}$$

where g is the acceleration of gravity, T is absolute temperature, α is the coefficient of thermal expansion, k is the diffusivity, v the kinematic viscosity, C_f is the specific heat of fluid (4183 J K^{-1} kg^{-1} for water at 20 °C) and B is a constant which has the value 216 for a tube whose length is great compared with its diameter (d). For the Outokumpu hole (diameter 22 cm) the critical gradient is about 0.3 mK/m, which is about 50 times smaller than the measured geothermal gradient values. Therefore, free convection very probably takes place and the Outokumpu monitoring data supports this conclusion. The measured data reveals a complex stochastic structure of the microtemperature field. In analogy to the previous results based on theoretical considerations discussed in previous experiments (Čermak et al., 2008c) the height of convection cells is probably of the same order of size as the hole diameter; the characteristic period of the process amounts to 2.5–2.8 h. This is in good agreement with the weak periodicities observed in the temperature data. Although the borehole fluid becomes more saline and denser with increasing depth, the fluid density is practically constant over distances of convection cell heights. Therefore the fluid salinity has no effect on such small scale convection.

The free convection creates only local temperature variations, which do not jeopardize the overall geothermal results. However, it generates a certain background "noise" level that limits the temperature resolution that can be achieved in studying local temperature variations in a large diameter hole.

7. Palaeoclimatic effects: forward and inverse transient heat transfer models

7.1. Forward modeling

As there is no significant topography in the Outokumpu area, and the hydraulic conductivities (Tarvainen, 2006; Ahonen et al., 2011) are relatively low, forced convective heat transfer probably does not play a role in the vertical temperature gradient and heat flow variations. Therefore, the general vertical gradient variation is most probably due to palaeoclimatic ground temperature variations on the Outokumpu drilling site, but probably also affected by structural effects (i.e., conductivity and heat production). We have used the latest temperature log (September 2008) for forward and inverse modeling of the ground surface temperature (GST) history.

The forward modeling was done in 2D with a finite difference modeling code Processing Shemat (Clauser, 2003) and the model structure was constructed using reflection seismic data (Kukkonen et al., 2006, see Fig. 3). The model size is 10 km × 10 km (Fig. 8) and the basic discretization is 100 m × 100 m, but locally refined in the uppermost parts of the model and in the vicinity of the deep hole to 10–50 m cell size. Applied thermal parameter values are shown

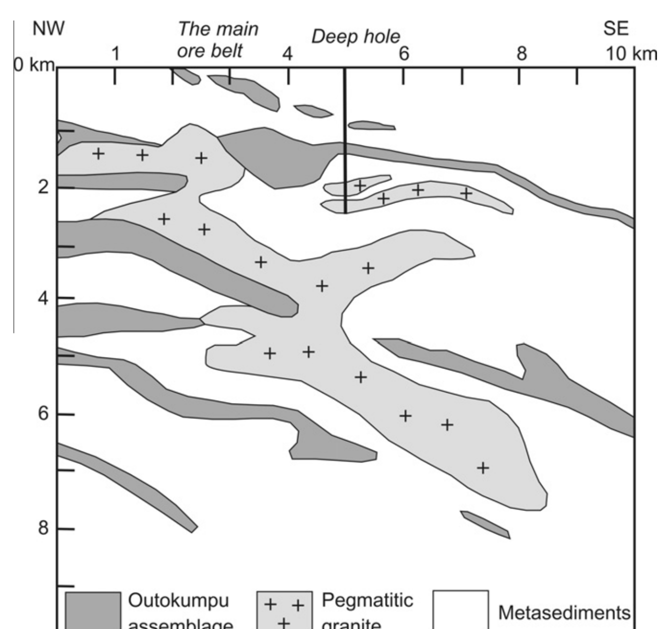

Fig. 8. 2D model of the Outokumpu structure used in forward conductive heat transfer modeling. The applied thermal parameter values are listed in Table 3.

Table 2
Mean thermal conductivity, temperature gradient and heat flow density of the Outokumpu Deep Drill Hole calculated at 100 m intervals.

Depth interval (m)	Conductivity ($Wm^{-1} K^{-1}$)	SD ($Wm^{-1} K^{-1}$)	Cond. mean error ($Wm^{-1} K^{-1}$)	*N* samples	*T* grad. ($mK m^{-1}$)	Grad mean error ($mK m^{-1}$)	Heat flow ($mW m^{-2}$)	Heat flow mean error ($mW m^{-2}$)
50–100	2.28	1.13	0.36	10	12.58	0.04	28.6	4.6
100–200	2.47	0.65	0.11	37	12.77	0.00	31.6	1.4
200–300	2.47	0.40	0.06	51	12.63	0.01	31.2	0.7
300–400	2.44	0.50	0.06	75	13.08	0.01	31.9	0.8
400–500	2.35	0.46	0.05	76	12.92	0.01	30.4	0.7
500–600	2.48	0.52	0.06	88	13.22	0.01	32.8	0.8
600–700	2.48	0.40	0.06	51	13.04	0.01	32.4	0.7
700–800	2.50	0.58	0.06	96	12.53	0.01	31.3	0.8
800–900	2.44	0.74	0.08	91	13.14	0.01	32.1	1.0
900–1000	2.64	0.39	0.05	51	13.44	0.01	35.5	0.8
1000–1100	2.62	0.49	0.05	94	13.51	0.01	35.4	0.7
1100–1200	2.60	0.53	0.05	94	13.16	0.00	34.3	0.7
1200–1300	2.61	0.40	0.06	51	14.31	0.01	37.3	0.8
1300–1400	3.33	1.57	0.16	98	12.10	0.02	40.3	2.0
1400–1500	3.39	0.95	0.10	90	10.67	0.02	36.2	1.1
1500–1600	3.02	0.93	0.10	94	14.47	0.02	43.7	1.4
1600–1700	2.44	0.60	0.06	94	15.88	0.01	38.7	1.0
1700–1800	2.53	0.52	0.05	91	15.89	0.01	40.3	0.9
1800–1900	2.95	0.76	0.08	96	15.02	0.01	44.3	1.2
1900–2000	2.56	0.73	0.07	103	17.09	0.01	43.8	1.3
2000–2100	3.09	0.37	0.04	90	14.94	0.01	46.1	0.6
2100–2200	3.18	0.21	0.03	66	15.75	0.01	50.0	0.4
2200–2300	2.33	0.87	0.10	72	17.33	0.02	40.3	1.8
2300–2400	2.91	0.54	0.10	31	14.35	0.01	41.7	1.4
2400–2504	2.83	0.53	0.09	32	14.49	0.02	41.0	1.4

Table 3
Average values of thermal parameters in the Outokumpu deep hole.

Rock type	Conductivity ($W m^{-1} K^{-1}$)	Volumetric heat capacity ($10^6 J m^{-3} K^{-1}$)	Heat production ($\mu W m^{-3}$)
Metasediments	2.50	1.89	1.70
Outokumpu assemblage[a]	3.35	2.11	1.55
Pegmatitic granite	3.00	1.78	5.40

[a] Ophiolite-derived altered ultramafic rock types serpentinite, skarn rock, quartz rock.

in Table 3. For volumetric heat capacity the average of $1.88 \times 10^6 J m^{-3} K^{-1}$ of the deep hole rocks was used. As the porosity of crystalline rocks is very small (<0.5%) we ignored the latent heat of freezing and thawing pore water. The GST variation was modified step by step until a satisfactory agreement with measured temperatures was achieved. Here we could use the earlier results by Kukkonen and Šafanda (1996) as first approximations of a possible GST history in the Outokumpu area. A measure of the two-dimensionality of the modeled structure is provided by the

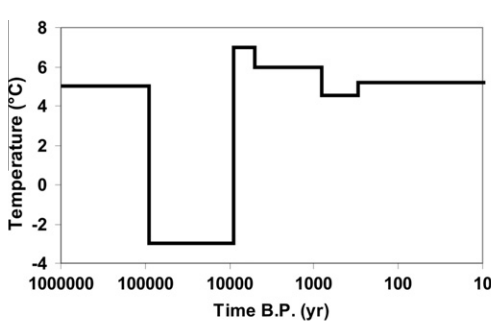

Fig. 9. Ground surface temperature variation used as the upper boundary condition in the 2D transient modeling.

simulated vertical and horizontal components of heat flow density. 2D thermal conductivity contrasts affect the heat flow vector and deflect it from the vertical due to channeling (or refraction) of heat flow to inclined high conductivity layers. Our thermal model suggests that the horizontal components of heat flow in the deep hole section are very local and smaller than 2 $mW m^{-2}$ which is equivalent to a deviation of the heat flow vector by less than 4° from vertical. This is due to the relatively small dips of the rock type layers (Fig. 8), which strongly constrain the refraction effects. It means that a 1D approximation could also be applied, which was actually done in the inverse modeling below.

In our final transient 2D model (Fig. 9) the Weichselian was preceded by a steady-state with GST of 5 °C, and during the Weichselian temperature is assumed to be −3 °C between 91 ka BP and 9 ka BP. During the subsequent "Holocene optimum" GST was 7 °C, decreasing to 6 °C at 5000 BP, further to 4.6 °C at 800 BP and finally increasing to 5.2 °C at 300 BP which was kept until the present.

The modeling is quite sensitive for the basal heat flow (in the final model 20.5 $mW m^{-2}$ at 10 km). Changing the basal heat flow by 0.5 $mW m^{-2}$ affects the temperature at 2.5 km by about 0.3°. The Outokumpu data is not simple to fit with a GST history. This is due to a relatively constant gradient down to about 1.3 km. To simulate this we need downward diffusing GST signals which compensate each other. Anyway, it is possible to find a GST history which works with the measured data and includes the main components of Holocene climate history (Fig. 9). The temperature fit is mostly within ±0.1 to 0.2 K (Fig. 10). Respectively, the simulated heat flow values (Fig. 11) are in a good agreement with values measured in the deep hole (Fig. 6).

7.2. Inverse modeling

1D inversions were performed with the smooth Tikhonov-type inversions described by Rath and Mottaghy (2007). For this purpose, a numerical model was constructed down to a depth of 5000 m. For this, logarithmic spatial and temporal meshes were used, with 700 and 600 cells, respectively. This logarithmic set up was found to be advantageous when modeling the propagation of a broad-band GSTH signal into the earth.

In order to associate temperatures and rock properties to grid cells, both temperature and the measured rock thermal conductivities had to be upscaled. For thermal conductivities this was accomplished by using the weighted harmonic mean of all measured values in a cell, assuming dominantly vertical heat conduction. Temperature, however, is available at a higher sampling rate (0.1 m). Therefore we chose a smoothing interpolation of temperatures for each cell (Garcia, 2010), where the smoothing parameter was determined by Generalized Cross Validation (Wahba, 1990). Residuals between observed and interpolated temperatures are shown in Fig. 12. As the residuals show a reasonably Gaussian distribution, we believe that the bias introduced by the upscaling procedure is small compared to other sources of error. The harmonic mean is less sensitive to high conductivities (say, >6 $\mathrm{W m^{-1} K^{-1}}$), which fortunately constitute only a very small fraction (<1%). Thermal diffusivity was estimated from conductivity based on the petrophysical studies which support a linear dependence (s [10^{-6} $\mathrm{m^2 s^{-1}}$] = 0.503 λ + 0.0839, R^2 = 0.94).

The model assumes a 1D layered structure with conductivity and heat production in the uppermost 2.5 km as determined from logs and laboratory measurements in the deep hole. Between 2.5 and 5 km, a constant thermal conductivity of 2.9 $\mathrm{Wm^{-1} K^{-1}}$ was assumed. For heat production, a value between pegmatite and metamorphites was obtained by multiplying the value for the pegmatites (H in Fig. 13) by a factor of 0.8, roughly corresponding to the proportions of pegmatites and metamorphites. The values of λ and heat production are somewhat arbitrary, because for this 1D model the conditions below 2500 m the borehole only add to the heat flow density at the bottom of the hole, equivalent to changing the (fictious) value of heat flow density a 5000 m.

GSTH war parametrized by a logarithmically spaced series of step functions, between 100 kyrs B.P. and 10 years B.P., in order to take care of the decreasing resolution with depth. As is well known, GSTH inversion is an ill-posed problem. Regularization was done by a combination of zero and first order derivative matrices (see Rath and Mottaghy, 2007; Aster et al., 2005). Instead of adapting sophisticated methods of determining the regularization parameters, a simple decreasing schedule was adopted, starting

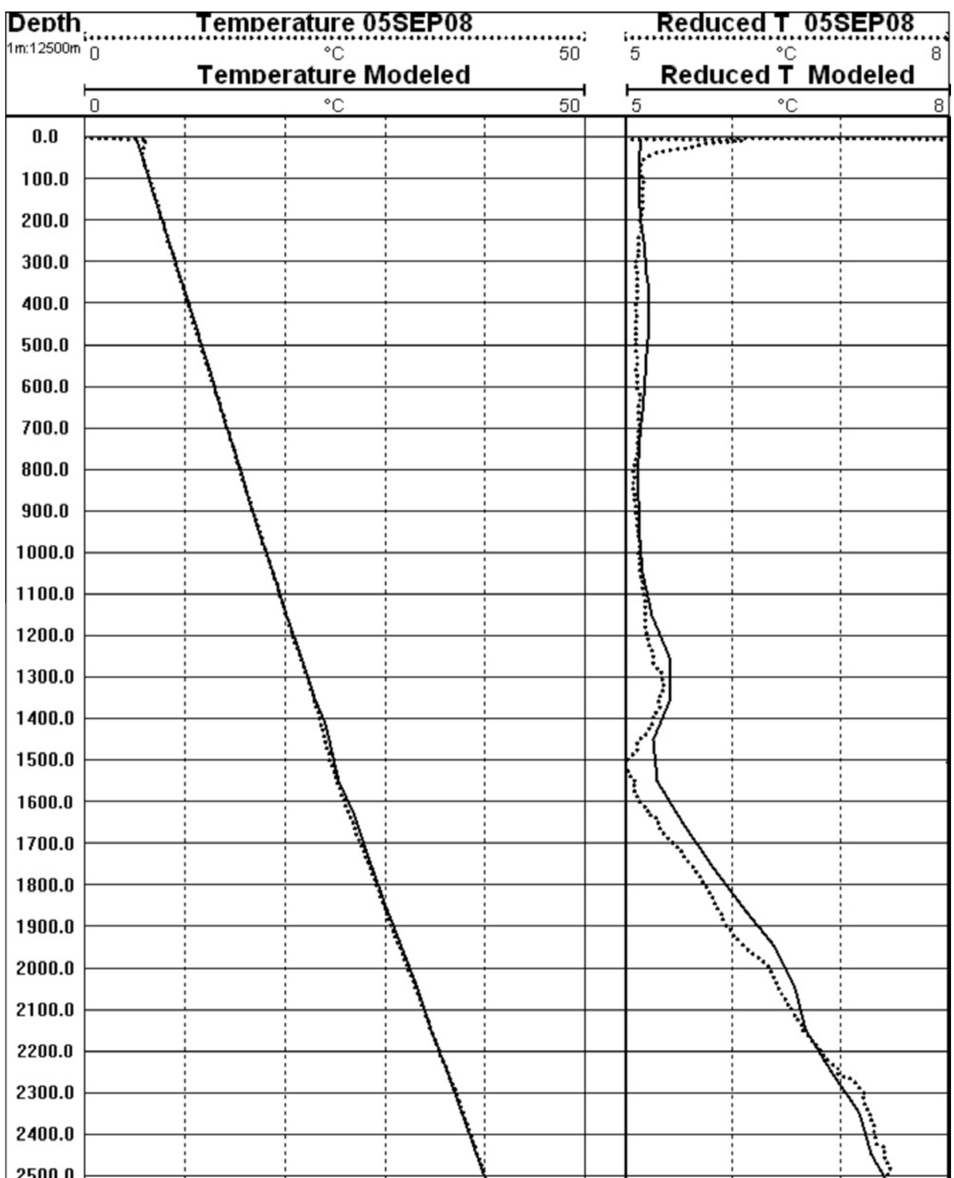

Fig. 10. Results of the 2D forward transient conduction model at the location of the Outokumpu deep hole. Dotted lines represent the measured temperatures (05 September 2008) and solid lines temperatures simulated with the model (Figs. 8 and 9). The right panel shows the temperatures reduced with a constant gradient of 13 $\mathrm{mK\,m^{-1}}$.

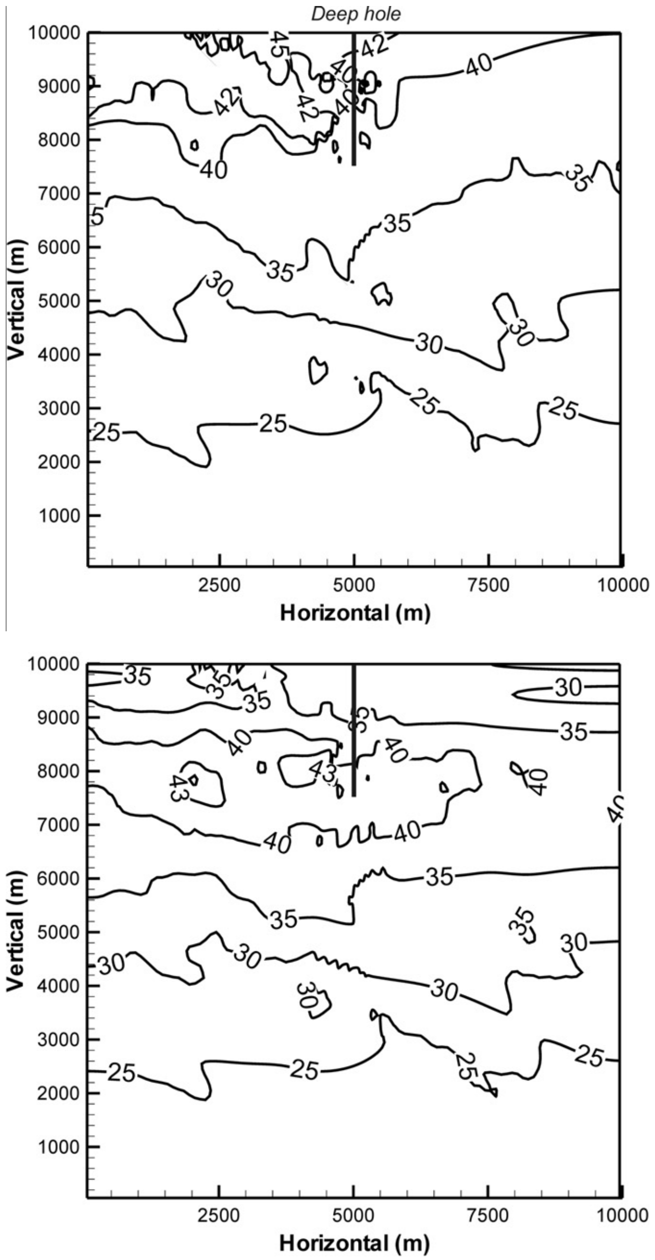

Fig. 11. Simulated steady-state (*top*) and present (*bottom*) values of vertical heat flow density (in units of mW m^{-2}) of the 2D forward transient modeling. The applied surface temperature is shown in Fig. 9 and modeled rock type domains in Fig. 8, and thermal parameters in Table 3.

with predetermined values and reducing them by a fixed factor at each iteration, until a target misfit is reached (see, e.g., Newman and Hoversten, 2000). For all inversions, a constant prior corresponding to the current surface temperature of 5 °C was assumed.

Inversion results thus obtained (Fig. 13) suggest the following: Nearly independent of the parameterization chosen, a rather stable GSTH for the last few thousand years was found. In particular, the results for the last two millennia are insensitive to the choice of the basal heat flow. In most cases, a reasonable fit could be found for depths <1500 m, with $|T_{\mathrm{obs}} - T_{\mathrm{mod}}| < 0.1$ K for the top 1000 m, and <0.2 K for $z < 1500$ m. As the influence of the post-glacial warming can be approximated by a linear trend (or biased heat flow) for short boreholes (Hartmann and Rath, 2005), it could be interpreted that the results are not biased by non-stationary conditions and may be interpreted independently. Temperatures of the LGM, however, are sensitive to the assumed heat flow density and heat production. Moreover, the data fit deteriorates strongly below 1500 m. Here, the residuals increase up to ±0.3 K even for the best cases. This is probably due to different effects: (1) fluid inflow may be important for depths >1500 m (see Fig. 5), (2) heterogeneity effects may produce temperature distributions not consistent with 1D conditions and (3) there may be a residual contribution of post-drilling equilibration of temperatures. All these effects are not easily coped with in the framework of 1D inversion, and must be left to future studies.

Systematic variation of basal heat flow and pegmatite heat production show that a similar fit can be obtained for many combinations of these two parameters. Plotting inverted models for parameter combinations along this elongated minimum of residual norms reveals that all these GSTHs are very similar. Fig. 14 shows contours of the

$$\mathrm{RMS} = \sqrt{N^{-1} \sum \sigma_i^{-1} (T_{i,obs} - T_{i,calc})^2} \tag{3}$$

and the GSTHs corresponding to the minimum. These results are very similar for basal HFD between 26 and 36 mW m^{-2} and pegmatite heat productions between 1 and 6 μW m^{-2}, showing a high degree of equivalence in the model. However, all these GSTHs produce practically the same fit (RMS ≈ 0.9).

A tentative interpretation of the reasonably well-resolved part of the GSTH can be given. The minimal temperatures for the LGM were 7–9 K lower than today, i.e., between −2 and −4 °C. Mainly due to the problems originating from the effect mentioned above, the post-glacial temperature rise is too smooth, as in this case the inversion is dominated by the smoothing regularization method. Generally, a warm Holocene (anomaly 2–3 K) was found, reaching its maximum near 4 kyrs B.P., broadly consistent with regional proxy reconstructions. There is no signature of a Medieval Warm Period, and the Little Ice Age appears rather flat with no clear indication of distinct minima. Finally, a strong recent warming of about 0.5 K is visible in the last 50 years. In comparing the present results with various palaeotemperature histories derived from proxies and meteororological data which commonly refer to the air temperature, we have to take into account that ground surface temperature in Finland is 1–2 K higher than air temperature (Kukkonen, 1987; linear dependence for annual average temperatures: T(ground) (°C) = 0.71 T(air) + 2.93). The difference is due to the insulating effect of the winter snow cover. Thus, the behavior observed in the inversion results is generally consistent with various proxy reconstructions given in Seppä and Poska (2004), Heikkilä and Seppä (2003) and Luterbacher et al. (2004).

During the 20th century the drilling site likely experienced several changes in the vegetation cover due to the Outokumpu town which started to grow quickly after the discovery of the ore deposit in 1910. Therefore, there have been significant changes in the environment due to clear cutting of forest and building of the railroad to the S side of the drilling site, and the warming trend of the last 50 years may be partly related to man-made effects. The present forest surrounding the site is about 30–50 years old and it was most probably clear-cut before that. The drilling site (area 0.6 ha) was clear-cut in the forest and covered with gravel in 2003. This removal of the vegetation is clearly reflected in the temperature curves in the uppermost 50 m.

8. Discussion

The geothermal results of the Outokumpu Deep Drill Hole indicate that there is a distinct vertical variation in temperature gradient and heat flow density. Due to the high resolution of data obtained both in the drill hole logging as well as in laboratory measurements, the conductivity, gradient and heat flow values could

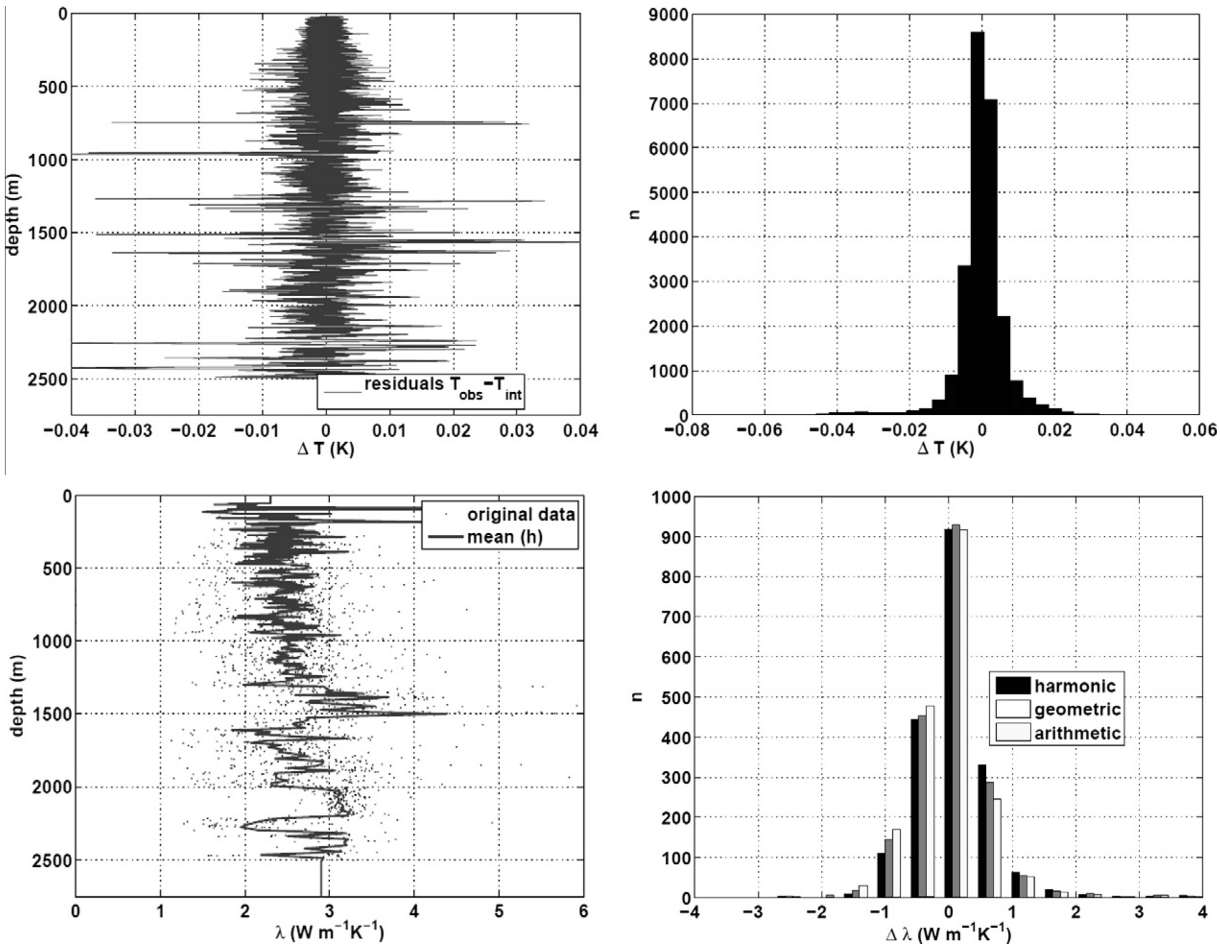

Fig. 12. Results of the upscaling of measured temperatures (upper panels) and thermal conductivities (lower panels) to the numerical grid. Elementary descriptive statistics lead to a very small average of −0.0002 K for temperature, and standard deviations for the residuals of 0.007 K. The situation is worse with thermal conductivity, where the upscaling process leads to considerable smoothing. For the harmonic mean upscaling the corresponding values are 0.10235 Wm^{-1} K^{-1} and 0.65983 Wm^{-1} K^{-1}, respectively.

be determined with very small determination errors (Table 2). The measured heat flow data in the uppermost 1 km differ by about 10 mW m^{-2} from the undisturbed values (about 42 mW m^{-2}, Fig. 11). The difference is significant and must be taken into account in applying heat flow data for estimating temperatures deeper in the crust. The result is in good agreement with the average vertical variation in heat flow density in Fennoscandia and the East European Platform (Kukkonen and Jõeleht, 2003). The vertical variation in heat flow density in the Outokumpu deep hole is very likely due to palaeoclimatic GST variations during the last 100,000 years. Structural effects, such as refraction of heat can be excluded, because the subhorizontal dip angles of the lithological contacts attenuate the effects of heat flow refraction to insignificant. Correspondingly, heat transfer by advection is also considered to be negligible.

The result has important implications for analyzing and correcting heat flow data in the Fennoscandian Shield (cf. Majorowicz and Wybraniec, 2010). The need of correcting heat flow data for palaeoclimatic effects has been known for a long time (e.g., Birch, 1948; Jessop, 1971; Beck, 1977), but the problem is that usually the GST history is not known sufficiently accurately to allow reliable forward corrections, and shallow holes (<1 km) do not provide unique solutions for the long-term climatic GST variations, but ignoring them, especially the glaciations, will bias the heat flow values significantly.

Earlier palaeoclimatic corrections for Weichselian and Holocene effects in heat flow values from about 1 km deep holes in the Outokumpu belt have given values of 6–7 mW m^{-2} (Kukkonen, 1987, 1989). These corrections assumed that the Weichselian GST was at the pressure melting temperature of a thick ice sheet, i.e., about −1 °C. Kukkonen and Šafanda (1996) modeled the GST history in the Outokumpu belt (holes about 1 km deep) with forward modeling and obtained an estimate of −2 °C for the Weichselian GST. The present forward and inverse results suggest −3 °C for the Weichselian GST and the correction for palaeoclimatic effects is about 10 mW m^{-2} for the average heat flow value in the uppermost 1 km. Our result suggests a correction smaller than the one applied by Majorowicz and Wybraniec (2010) for northern Europe. Their correction is +19 mW m^{-2} at surface and about 16 mW m^{-2} for an average correction in the uppermost 1 km, and the corrected heat flow value for eastern Finland is given as about 50 mW m^{-2} which is considerably higher than our present Outokumpu result of 42 mW m^{-2}. The difference is due to the much higher glacial–interglacial GST difference applied (14 K) and that Holocene effects were not included in the correction of Majorowicz and Wybraniec (2010). The Holocene effects, particularly the relatively long HCO warm period, would tend to reduce the overall correction in the uppermost 600 m. On the other hand, Slagstad et al. (2009) used a palaeoclimatic correction for the Fennoscandian area assuming glacier basal temperatures as −1 °C resulting in somewhat smaller corrections than the Outokumpu data suggest. The Outokumpu results emphasize the importance of experimental results from deep boreholes in constraining the necessary corrections of geothermal data.

Please cite this article in press as: Kukkonen, I.T., et al. Geothermal studies of the Outokumpu Deep Drill Hole, Finland: Vertical variation in heat flow and palaeoclimatic implications. Phys. Earth Planet. In. (2011), doi:10.1016/j.pepi.2011.06.002

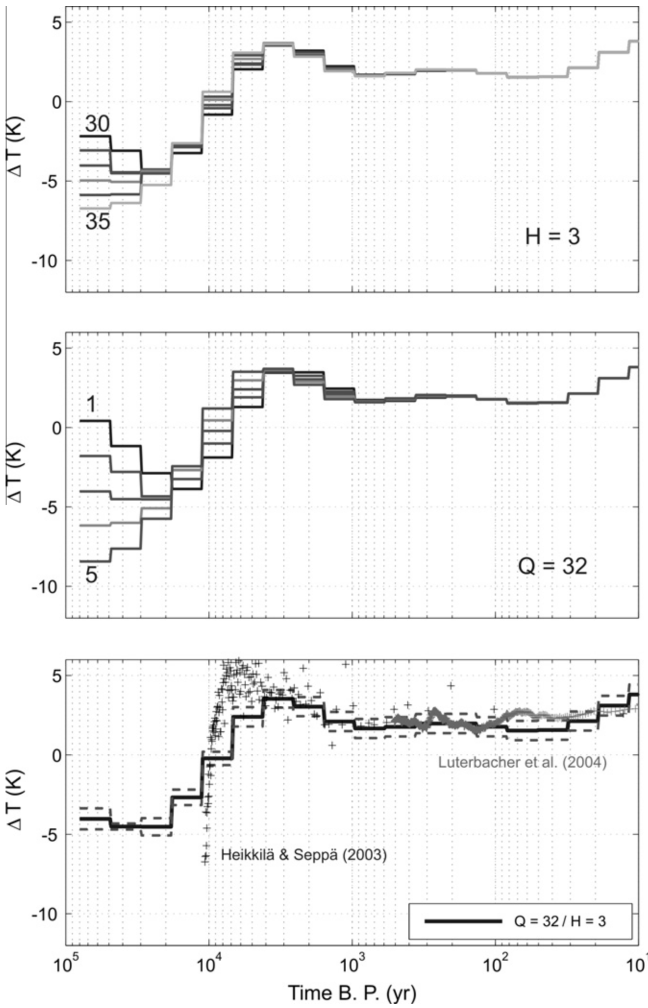

Fig. 13. *Top:* Results of smooth inversion for different values of basal heat flow values assumed at a depth of 5000 m. The model assumes a 1D layered structure with conductivity and heat production in the uppermost 2.5 km as determined from logs and laboratory measurements in the deep hole. Between at 2.5–5 km, a constant conductivity of 2.9 $Wm^{-1} K^{-1}$ and heat production were assumed. The latter was set to the value of the pegmatitic granite reduced by a factor of 0.8. *Center*: Basal heat flow is kept constant but the heat production of pegmatitic granite is varied at 1 $\mu W m^{-3}$ steps. Assuming that heat production 3 $\mu W m^{-3}$ is representative (cf. Table 3 and Fig. 6), we may conclude that the true value of basal heat flow is 32–34 $mW m^{-2}$. *Bottom:* "Optimum" GSTH with a basal HFD of 32 $mW m^{-2}$ and heat production of 3 $\mu W m^{-3}$ (solid line). Also shown are 2σ error bars (dashed), and the paleotemperature reconstructions of Luterbacher et al. (2004) and Heikkilä and Seppä (2003).

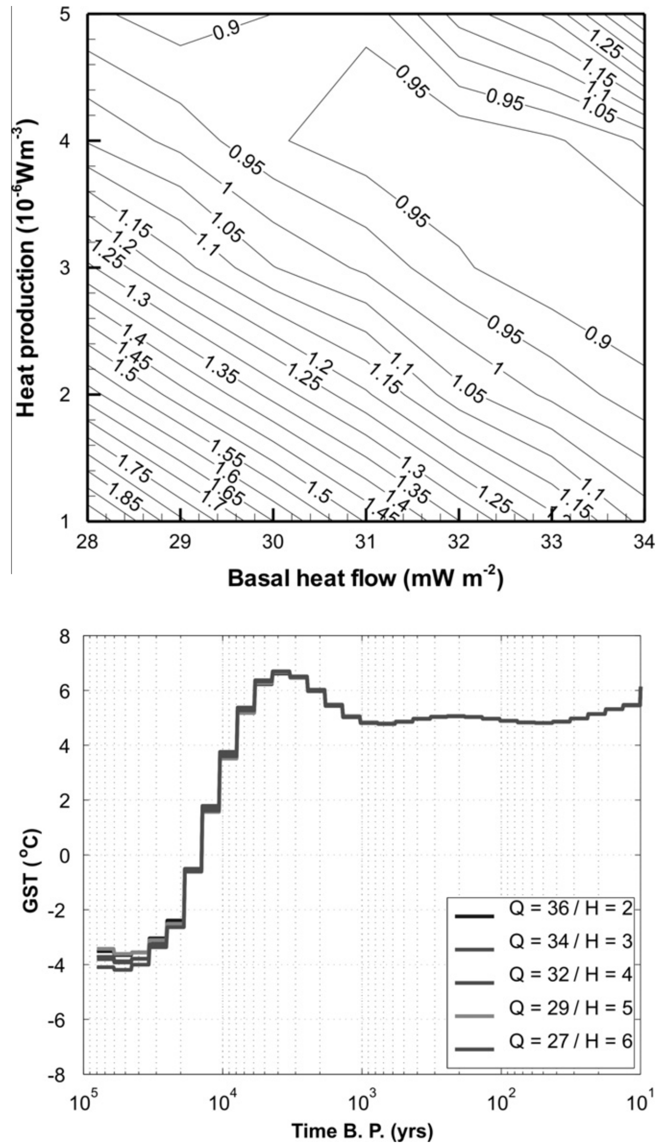

Fig. 14. Top: Root-mean-square values of inversion results for combinations of basal HFD and pegmatite heat production. The results indicate a valley of best fit clearly visible in the contours of RMS. All have a RMS value just below 0.9. *Bottom*: LGM temperatures of these best fit models seem to be well constrained near −4 °C.

Very low apparent heat flow values (2–12 $mW m^{-2}$) were reported in eastern Karelia, Russia, by Kukkonen et al. (1998) in a number of holes <750 m deep. The study site is located about 350 km to the E of Outokumpu. Advective fluid heat transfer and structural factors could be excluded and the low heat flow values were attributed to low Weichselian GST as low as −10 to −15 °C. The area was outside the Weichselian ice sheet for most of time and therefore experienced periglacial low temperatures. If the indicated GST values are representative they would imply a considerable difference in Weichselian GST values between Outokumpu and eastern Karelia.

The Outokumpu area was covered by ice sheets perhaps only during the Middle Weichselian (about 60,000 years B.P.) and Late Weichselian (about 20,000 years B.P.) glacial maxima (cf. Hubberten et al., 2004; Svendsen et al., 1999, 2004). Radiocarbon dating of mammoth fossils suggests that most of southeastern Fennoscandia (and Outokumpu) was probably ice-free during 37,000–26,000 years B.P. (Ukkonen et al., 1999; Lunkka et al., 2001). The isotopic composition of oxygen in woolly mammoth teeth enamel can be used as a proxy for palaeotemperatures, and it yields annual mean surface proxies in the range of −1 to −3 °C in southern and central Finland and NE Russia for Middle Weichselian (about 52,000–24,000 years B.P.) (Arppe and Karhu, 2006, 2010). The paleotemperatures obtained from mammoth fossils are in a good agreement with the Outokumpu geothermal result on Weichselian average GST.

When interpreting the geothermally derived GST values in terms of paleoclimate, it should also be kept in mind that the basal temperatures of a glacier are usually not spatially or temporally constant. The GST under a thick ice sheet with basal pressure melting (about −1 °C) can actually be higher than the GST in surrounding ice-free areas experiencing very cold periglacial climate (Paterson, 1994). On the other hand, areas under the ice divide where downward advection of ice dominate, may have basal temperatures well below freezing (e.g., in the Greenland ice sheet; Dahl-Jensen et al., 1998).

The forward and inverse modeling of the GST history suggests that the surface was about 8 K below present during the cold Weichselian climatic period (GST −3 to −4 °C). The downward attenuated GST signal does not provide high resolution for the Middle and Early Weichselian in a 2.5 km deep hole, and no details can be provided for these periods, although the inverse modeling (Fig. 13) could suggest trends there, too. The solution is, however, non-unique and depends on the deep heat flow value and the heat production at depths beneath the deep hole bottom. Selecting the heat flow and heat production values yielding the smallest RMS values in inversion the obtained GST history becomes very stable and suggests surface temperatures of −3 to −4 °C during the Weichselian (Fig. 14). Due to poor temporal resolution of the temperature history during the Weichselian the GST value should be regarded as an estimate of the average value.

The transient modeling provides also an estimate of the permafrost thickness in Outokumpu, which had its maximum at the end of the cold Weichselian period. The forward model indicates that permafrost would have extended to about 150 m at the Weichselian–Holocene transition at about 9000 years B.P. We can compare this result with other logging and core data and speculate that the maximum depth reached by permafrost would be reflected in the fracture frequency in the uppermost parts of the hole. Freezing of pore and fracture water causes cracking and opening of existing fractures, which would gradually increase the fracture volume in each permafrost cycle. The hydraulically conductive fractures interpreted by Tarvainen (2006) from the density, acoustic and galvanic down-hole logs are much more frequent in the uppermost 200–300 m than in deeper levels (Fig. 5). The near-surface fracturing is naturally contributed by other factors than only permafrost, such as slow relaxation of lithostatic pressure by erosion and weathering, but freeze cracking by permafrost is a conceivable process in the origin of the near-surface fracturing. Fracturing may also have been inherited from other earlier Quaternary cold climatic periods than only the Weichselian.

The small temperature gradient and heat flow variations in the uppermost 1 km requires that the Holocene variations in GST were within ±2 K from the present GST value. The Holocene GST history obtained by the smooth inversion (Fig. 13) is a very stable result particularly for the latest 2000 years which are not affected by uncertainties in the deep heat flow or heat production. The Holocene Climatic Optimum (HCO) is shifted towards more recent times in our result than conventional palaeoclimatic proxy data suggest and the maximum is seen around 4000 years B.P. The result differs from typical Holocene temperature reconstructions, and, e.g., proxies from pollen stratigraphy in southern Finland (Heikkilä and Seppä, 2003) show that the HCO temperature was at maximum about 6000 years B.P. This reflects the problems with the regularization parameters in the inversion which smoothes the very strong and steep signal of the Weichselian–Holocene transition and spreads it apparently to Early Holocene. The results for later times show only weak variations. Moreover, the temperatures obtained are approximately 2 K higher than paleotemperature reconstructions from nearby points as in Heikkilä and Seppä (2003), Seppä and Poska (2004) or Luterbacher et al. (2004). This reflects the difference between ground and air temperatures, as discussed above, and the distances between proxy study sites and Outokumpu. In Fig. 13, we also have plotted the European temperature reconstruction for the last 500 years, which were interpolated to the geographical coordinates of Outokumpu. Our results fit the general behavior quite well. However, there is larger disagreement in the last hundred years, which cloud be due to the history of land use change and deforestation in the industrial age. This is speculative, but deserves further investigations.

Assuming that the geological structure can be approximated as a 1D horizontally layered structure, and that heat transfer is conductive, Fourier's first law of heat conduction (i.e., heat flow density = conductivity × gradient) is in force. Therefore, thermal conductivity and measured temperature gradient should be inversely correlated. However, the simple inverse relationship is affected by transient effects (palaeoclimate) and heat production, and therefore it is not expected to hold in absolute values of conductivity and gradient in the scale of the complete borehole. Nevertheless, the inverse relationship can be assumed to exist locally representing small scale variations (in the scale of 10–100 m). Such a relationship is indeed observed in the Outokumpu hole data. It holds particularly in depth sections where there are strong thermal conductivity contrasts, such as (1) between the ophiolitic rocks and surrounding mica schist, (2) between serpentinites and skarn within the Outokumpu rock type assemblage, as well as (3) between the pegmatitic granite and mica schist (Fig. 6). The relationship is however, not completely systematic, and deviations can be seen in the upper parts of the hole. They may reflect small scale heterogeneities which are not sufficiently one-dimensional, or alternatively small fluid flow effects. For instance, at the depth of 230 m there is one of the hydraulically active fractures interpreted to cause flow out from the hole (Fig. 5) and the conductivity variations do not seem to correlate with gradient over the 200–300 m interval.

Heat transfer is dominantly conductive in the deep hole structure. This is demonstrated by the lack of thermally relevant fluid flow effects between fractures and the hole, as well as the inverse correlation of gradient and conductivity. Moreover, the low hydraulic conductivity of the crystalline rocks in general (Clauser, 1992; Manning and Ingebritsen, 1999), and in the Outokumpu area (Kukkonen and Šafanda, 1996; Tarvainen, 2006; Ahonen et al., 1991, 2011; Ahonen and Blomqvist, 1994) together with the small topographic hydraulic gradients suggest that the magnitude of heat transfer by fluid flow in the formation, if present, is very probably negligible. The downward increasing fluid salinity (and thus density) further tends to attenuate flow systems. In Outokumpu, the saline fluids recovered from the open borehole as well as pumped from fractures isolated by hydraulic packers also show vertical variation in their chemical composition (Ahonen et al., 2011). It speaks for minimal mixing and long residence times of groundwater in the fractured medium penetrated by the drill. Conductive heat transfer very likely dominates over advection.

During the Weichselian glaciation, the subglacial hydrogeological conditions may have resulted in significant recharge of glacial meltwater into the bedrock which may have produced advective heat transfer still affecting the present temperatures. Theoretical modelings and experimental data on subglacial hydrology and groundwater flow (Boulton et al., 1995, 1996; Person et al., 2007; Lemieux et al., 2008; Lemieux and Sudicky, 2010) have suggested that in northern Europe and North America considerable recharging of glacial melt water took place under the Weichselian and Laurentian ice sheets. In Canada, the recharge depth may have extended up to 2–3 km. The glacial hydrology may have considerably altered the groundwater flow systems, flow directions and affected the groundwater chemistry, particularly in the aquifers in sedimentary rocks, but also in crystalline shield rocks. These results depend strongly on the subglacial hydraulic heads and hydraulic conductivity of the bedrock and are therefore dependent on rock types, fracturing and local glacial conditions. The existence of eskers in the shield areas glaciated during the Weichselian glaciation in northern Europe is attributed to channelling of subglacial meltwaters at the interface between ice and low-permeability bedrock. On the other hand, there are no or only rare eskers in areas of sedimentary rocks which is attributed to the recharging of meltwaters into the permeable sedimentary rocks and aquifers in them (Boulton et al., 1995, 1996, 2009; Vaikmäe et al., 2001; Jansson et al., 2007; Jansson, 2010).

The order of magnitude of the thermal effect of the recharge can be estimated from the infiltration rates of meltwater into bedrock during the glaciation. The infiltration rates are not simple to estimate, but some values are available. Person et al. (2007) estimate from geochemical and environmental isotopic data that the volume of recharged meltwater in sedimentary rock aquifers in North America may amount up to 0.2–10% of the total amount of meltwater. Lemieux et al. (2008) estimated that the subglacial infiltration rate was on the average about 2 mm/year in the Canadian Shield area during the Wisconsinian glaciation. Using the value by Lemieux et al. (2008) the corresponding Peclet number for vertical (1D) filtration of water in a permeable layer can be calculated as (Bredehoeft and Papadopoulos, 1965):

$$Pe = V_D \rho_f C_f L/K \quad (4)$$

where V_D is the recharge flux (Darcy velocity), ρ_f is fluid density, C_f is fluid specific heat capacity, L is layer thickness and K is thermal conductivity. Assuming that the hydraulically permeable zone was 1–2 km thick in Outokumpu during glaciation and rock thermal conductivity is 2.5 $\mathrm{W m^{-1} K^{-1}}$, the (steady-state) Peclet number would be about 0.1–0.2. Such a disturbance in heat flow would be a small one, but possibly detectable under favorable conditions. However, the limited duration of ice cover (approximately 5000–10000 year during each glacial) and the fact that glaciation (and the infiltration) ended already about 10,000 year ago result in further reduction in the subglacial advective heat transfer effects on the present heat flow. It is possible that traces of a small advective palaeohydrological heat transfer component could be present in the Outokumpu temperature data, but the effect is very small and would be very difficult to distinguish from transient conductive effects by palaeoclimatic GST variations. We note that the subglacial hydrogeology of the Weichselian ice sheet would need more attention in future studies to allow more accurate estimation of the possible advective heat transfer effects by subglacial groundwater flow.

The critical gradient for the onset of free convection of fluid in the borehole is clearly exceeded in the Outokumpu hole, and convection was experimentally confirmed with time-dependent temperature records. The free convection is sluggish and generates time-dependent temperature variations with a maximum amplitude of about 0.01 K, and the convection cells are about as tall as the hole diameter (0.22 m). In single logs obtained with temperature resolution of 0.01 K at logging speeds of about 5–10 m/min and with reading intervals of 0.1–0.2 m, free convection can be expected to show as single anomalous readings which cannot be easily separated from instrumental noise. Moreover, the temperature variations are smoothed by the convolution of the temperature-depth profile depending on the time-constant of the logging tool and the applied logging speed in continuous loggings. As a result, we consider the free convection as a source of thermal noise, which does not harm the overall results or modeling.

9. Conclusions

The geothermal studies in the Outokumpu Deep Drill Hole revealed a distinct vertical variation in heat flow density. Heat flow density increases from about 28–32 $\mathrm{mW m^{-2}}$ in the uppermost 1000 m to 40–45 $\mathrm{mW m^{-2}}$ at depths exceeding 2000 m. The vertical variation can be attributed mostly to the cold climatic Weichselian period (about 91,000–9000 years B.P.). The forward and inverse modeling suggests that the GST was about −3 to −4 °C during Weichselian time. Inversion for Holocene GST history yields a relatively stable GST within ±2 K from the present value (5 °C). The Outokumpu results further suggest that the correction for palaeoclimatic effects in heat flow data measured in the uppermost 1 km in eastern Finland should be about 10 $\mathrm{mW m^{-2}}$. Heat transfer in the Outokumpu section is dominated by conduction and no relevant advective heat transfer effects by groundwater flow could be observed. In the hole, minor effects of fluid flow between fractures and the hole can be observed, which we attribute to replacement of fresh drilling fluid with saline formation fluids. Free convection takes place in the borehole and can be detected with very sensitive temperature recordings, but the convection produces temperature variation with very small amplitudes which can be treated as a source of thermal noise not affecting the heat flow results or modelings.

Acknowledgements

The present work benefited from the work of many people who contributed to the Outokumpu DDP. The authors are grateful to Kaj Västi (GTK, the project manager during the drilling phase), Erkki Ruokanen, Martti Damsten, Matti Kallunki and Eero Sandgren who were the GTK supervisors on the site during drilling, the NEDRA logging team, the ICDP-OSG logging team, especially Jochem Kueck, Christian Carnein and Karl Bohn. The authors also thank Jörg Franke (Bern) and Jürg Luterbacher (Giessen) for providing an updated version of their temperature reconstruction. ICDP is acknowledged for supporting the post-drilling logs. Constructive comments on the manuscript were provided by Mark Person. V.R. was funded by the Ramón y Cajal Program of the Spanish Ministry for Science and Education.

References

Ahonen, L., Blomqvist, R., Kurimo, M., Kuivamäki, A., 1991. Kalliopohjavesiolosuhteet Outokummun Sukkulansalossa: Vuoden 1990 tutkimukset (Deep bedrock hydrogeological condition in the Sukkulansalo area, Outokumpu: Results of 1990). Geological Survey of Finland, Nuclear Waste Disposal Research, Report YST-75, 42 pp (in Finnish with English abstract).

Ahonen, L., Blomqvist, R., 1994. Mode of occurrence of deep saline groundwater in Finland based on hydraulic measurements. In: Olofsson, B. (Ed.), Salt Groundwater in the Nordic Countries, Research Council of Norway, Oslo, NHP Report, vol. 35, pp. 49–59.

Ahonen, L., Kietäväinen, R., Kortelainen, N., Kukkonen, I.T., Pullinen, A., Toppi, T., Bomberg, M., Itävaara, M., Nousiainen, A., Nyyssönen, M., Öster, M., 2011. Hydrogeological characteristics of the Outokumpu Deep Drill Hole. In: Kukkonen, I.T. (Ed.), Outokumpu Deep Drilling Project 2003–2010, Geological Survey of Finland, Special Paper, vol. 51, pp. 151–168.

Arppe, L.M., Karhu, J.A., 2006. Implications for the Late Pleistocene climate in Finland and adjacent areas from the isotopic composition of mammoth skeletal remains. Palaeogeogr. Palaeoclim. Palaeoecol. 231, 322–330.

Arppe, L.M., Karhu, J.A., 2010. Oxygen isotope values of precipitation and the thermal climate in Europe during the middle to late Weichselian ice age. Quatern. Sci. Rev. 29, 1263–1275.

Aster, R., Borchers, B., Thurber, C., 2005. Parameter Estimation and Inverse Problems. Academic Press, San Diego, CA.

Beck, A.E., 1977. Climatically perturbed temperature gradients and their effect on regional and continental heat-flow means. Tectonophysics 41, 17–39.

Balling, N., Lind, G, Landström, O., Eriksson, K.G., Malmqvist, D., 1990. Thermal measurements from the deep Gravberg-1 well. R,D&D Report U(G) 1990/57, Swedish State Power Board, Vattenfall, 13 pp.

Birch, F., 1948. The effects of Pleistocene climatic variations upon geothermal gradients. Am. J. Sci. 246, 729–760.

Bredehoeft, J., 1967. Response of well-aquifer systems to earth tides. J. Geophys. Res. 72, 3075–3087.

Bredehoeft, J.D., Papadopoulos, I.S., 1965. Rates of vertical groundwater movements estimated from the Earth's thermal profile. Water Resour. Res. 1, 325–328.

Boulton, G.S., Caban, C.E., van Gijssel, K., 1995. Groundwater flow beneath ice sheets. Part I: Large scale patterns. Quatern. Sci. Rev. 14, 545–562.

Boulton, G.S., Caban, P.E., van Gijssel, K., Leijnse, A., Punkari, M., van Weert, F.H.A., 1996. The impact of glaciation on the groundwater regime of Northwest Europe. Global Planet. Change 12, 397–413.

Boulton, G.S., Hagdorn, M., Maillot, P.B., Zatsepin, S., 2009. Drainage beneath ice sheets: groundwater-channel coupling, and the origin of esker systems from former ice sheets. Quatern. Sci. Rev. 28, 621–638.

Čermak, V., 2009. Recurrence quantification analysis of borehole temperatures: evidence of fluid convection. J. Bifurc. Chaos 19 (3), 889–902.

Čermak, V., Šafanda, J., Bodri, L., 2008a. Precise temperature monitoring in boreholes: evidence for oscillatory convection? Part 1: Experiments and field data. Int. J. Earth Sci. (Geol. Rundsch.) 97, 365–373.

Čermak, V., Bodri, L., Šafanda, J., 2008b. Precise temperature monitoring in boreholes: evidence for oscillatory convection? Part II: theory and interpretation. Int. J. Earth Sci. (Geol. Rundsch.) 97, 375–384.
Čermak, V., Šafanda, J., Kresl, M., 2008c. Intra-hole fluid convection: high-resolution temperature time monitoring. J. Hydrol. 348, 464–479.
Clauser, C., 1992. Permeability of crystalline rocks. Eos Trans. AGU 73, 233, 237.
Clauser, C. (Ed.), 2003. Numerical Simulation of Reactive Flow in Hot Aquifers. SHEMAT and Processing SHEMAT. Springer, Berlin, 332 pp.
Clauser, C., Giese, P., Huenges, E., Kohl, T., Lehmann, H., Rybach, L., Šafanda, J., Wilhelm, H., Windloff, K., Zoth, G., 1997. The thermal regime of the crystalline continental crust: implications from the KTB. J. Geophys. Res. 102, 18417–18441.
Cutillo, P.A., Bredhoeft, J.D., 2010. Estimating aquifer properties from the water level response to Earth tides. Groundwater. doi:10.1111/j.1745-6584.2010.00778.x.
Dahl-Jensen, D., Mosegaard, K., Gundestrip, N., Clow, G.D., Johnsen, S.J., Hansen, A.W., Balling, N., 1998. Past temperatures directly from the Greenland ice sheet. Science 282, 268–271.
Drury, M.J., 1984. Perturbations to temperature gradients by water flow in crystalline formations. Tectonophysics 102, 19–32.
Drury, M.J., Lewis, T.J., 1983. Water movement within Lac du Bonnet Batholith as revealed by detailed thermal studies of three closely-spaced boreholes. Tectonophysics 95, 337–351.
Eppelbaum, L.V., Kutasov, I.M., 2011. Estimation of the effect of thermal convection and casing on the temperature regime of boreholes: a review. J. Geophys. Eng. 8, R1–R10. doi:10.1088/1742-2132/8/1/R01.
Garcia, D., 2010. Robust smoothing of gridded data in one and higher dimensions with missing values. Comput. Stat. Data Anal. 54, 1167–1178.
Glaznev, V.N., Kukkonen, I.T., Raevskii, A.B., Jokinen, J., 2004. New data on thermal flow in the central part of the Kola Peninsula. Doklady Earth Sci. 396, 512–514.
Hales, A.L., 1937. Convection currents in geysers. Mon. Not. R. Astron. Soc. Geophys. Suppl. 4, 122–131.
Hänninen, P., Äikää, O., Sutinen R., Kukkonen, I., 2009. Borehole monitoring of water table and temperature in bedrock and Quaternary sediments in Outokumpu. In: Kukkonen, Ilmo T. (Ed.), 2009. Outokumpu Deep Drilling Project, Third International Workshop, Espoo, Finland, November 12–13, 2009, Programme and Abstracts. Geological Survey of Finland, Southern Finland Office, Marine Geology and Geophysics, Report Q10.2/2009/61, pp. 57–60.
Hartmann, A., Rath, V., 2005. Uncertainties and shortcomings of ground surface temperature histories derived from inversion of temperature logs. J. Geophys. Eng. 2, 299–311.
Heikkilä, M., Seppä, H., 2003. A 11,000 yr palaeotemperature reconstruction from the southern boreal zone in Finland. Quatern. Sci. Rev. 22, 541–554.
Heikkinen, P.J, Koivisto, E., Kukkonen, I.T., 2007. FIRE high resolution reflection seismic survey in Outokumpu. In: Kukkonen, I.T. (Ed.), Outokumpu Deep Drilling Project, Second International Workshop, May 21–22, 2007, Espoo, Finland. Programme and Extended Abstracts. Geological Survey of Finland, Southern Finland Office, Marine Geology and Geophysics, Report Q10.2/2007/29, pp. 17–20.
Hsieh, P.A., Bredehoeft, J.D., Farr, J.M., 1987. Determination of aquifer transmissivity from Earth tide analysis. Water Resour. Res. 23, 1824–1832.
Hubberten, H.W., Andreev, A., Astakhov, V.I., Demidov, I., Dowdeswell, J.A., Henriksen, M., Hjort, C., Houmark-Nielsen, M., Jakobsson, M., Kuzmina, S., Larsen, E., Lunkka, J.P., Lys, A., Mangerud, J., Möller, P., Saarnisto, M., Schirrmeister, L., Sher, A.V., Siegert, C., Siegert, M.J., Svendsen, J.I., 2004. The periglacial climate and environment in northern Eurasia during the Last Glaciation. Quatern. Sci. Rev. 23, 1333–1357.
Huhma, A., 1971. Geological map of Finland 1:100,000, Pre-Quaternary rocks, Sheet 4222 – Outokumpu, Geol. Surv., Espoo, Finland.
Hurtig, E., Čermak, V., Haenel, R., Zui, V. (Eds.), 1992. Geothermal Atlas of Europe. Hermann Haack, Gotha, 156 pp (+ maps).
Jansson, P., 2010. Ice sheet hydrology from observations. Swedish Nuclear Fuel and Waste Management Co., Technical Report TR-10-68, 59 pp.
Jansson, P., Näslund, O., Rodhe, L., 2007. Ice sheet hydrology – a review. Swedish Nuclear Fuel and Waste Management Co., Technical Report TR-06-34, 145 pp.
Järvimäki, P., Puranen, M., 1979. Terrestrial heat flow in Finland. In: Čermak, V., Rybach, L. (Eds.), Terrestrial Heat Flow in Europe. Springer, Berlin, pp. 172–178.
Jessop, A.M., 1971. The distribution of glacial perturbation of heat flow in Canada. Can. J. Earth Sci. 8, 162–166.
Koistinen, T., Stephens, M.B., Bogatchev, V., Nordgulen, Ø., Wennerström, M., Korhonen, J., 2001, Geological map of the Fennoscandian Shield, scale 1:2,000,000: Geol. Surv. Finland, Norway and Sweden and the North-West Dept. of Nat. Res. of Russia.
Kremenetsky, A.A., Ovchinnikov, L.N., 1986. The Precambrian continental crust: its structure, composition and evolution as revealed by deep drilling in the USSR. Precambr. Res. 33, 11–43.
Kukkonen, I.T., 1987. Vertical variation of apparent and palaeoclimatically corrected heat flow densities in the central Baltic Shield. J. Geodyn. 8, 33–53.
Kukkonen, I.T., 1988. Terrestrial heat flow and groundwater circulation in the bedrock in the central Baltic Shield. Tectonophysics 156, 59–74.
Kukkonen, I.T., 1989. Terrestrial heat flow and radiogenic heat production in Finland, the central Baltic Shield. Tectonophysics 164, 219–230.
Kukkonen, I.T., 1995. Thermal aspects of groundwater circulation in bedrock and its effect on crustal geothermal modelling in Finland, the central Fennoscandian Shield. In: Balling, N., Decker, E.R. (Eds.), Heat Flow and Thermal Regimes of Continental Lithosphere, Tectonophysics, vol. 244, pp. 119–136.
Kukkonen, I.T. (Ed.), 2009. Outokumpu Deep Drilling Project, Third International Workshop, Espoo, Finland, November 12–13, 2009, Programme and Abstracts. Geological Survey of Finland, Southern Finland Office, Marine Geology and Geophysics, Report Q10.2/2009/61, 92 pp.
Kukkonen, I.T. (Ed.), 2011. Outokumpu Deep Drilling Project 2003-2010. Geological Survey of Finland, Special Paper, 51, 252 pp.
Kukkonen, I.T., Clauser, C., 1994. Simulation of heat transfer at the Kola deep-hole site: implication for advection, heat refraction and paleoclimatic effects. Geophys. J. Int. 116, 409–420.
Kukkonen, I.T., Jõeleht, A., 2003. Weichselian temperatures from geothermal heat flow data. J. Geophys. Res. 108, 2163. doi:10.1029/2001JB001579.
Kukkonen, I., Lindberg, A., 1998. Thermal properties of rocks at the investigation sites: measured and calculated thermal conductivity, specific heat capacity and thermal diffusivity. Posiva Oy, Work. Rep., 98-09e, 29 pp.
Kukkonen, I.T., Šafanda, J., 1996. Palaeoclimate and structure: the most important factors controlling subsurface temperatures in crystalline rocks. A case history from Outokumpu, eastern Finland. Geophys. J. Int. 126, 101–112.
Kukkonen, I.T., Golovanova, I.V., Khachay, Yu.V., Druzhinin, V.S., Kosarev, A.M., Schapov, V.A., 1997. Low geothermal heat flow of the Urals fold belt – implication of low heat production, fluid circulation or palaeoclimate? Tectonophysics 276, 63–85.
Kukkonen, I.T., Gosnold, W.D., Šafanda, J., 1998. Anomalously low heat flow density in eastern Karelia, Baltic shield: a possible palaeoclimatic signature. Tectonophysics 291 (1–4), 235–249.
Kukkonen, I.T., Heikkinen, P., Ekdahl, E., Hjelt, S.-E., Yliniemi, J., Jalkanen, E., FIRE Working Group, 2006. Acquisition and geophysical characteristics of reflection seismic data on FIRE transects, Fennoscandian Shield. In: Kukkonen, I.T., Lahtinen, R. (Eds.), Finnish Reflection Experiment 2001–2005. Geological Survey of Finland, Special Paper 43, pp. 13–43 (+11 appendices).
Lemieux, J.-M., Sudicky, E.A., 2010. Simulation of groundwater age evolution during the Wisconsinian glaciation over the Canadian landscape. Environ. Fluid Mech. 10, 91–102.
Lemieux, J.-M., Sudicky, E.A., Peltier, W.R., Tarasov, L., 2008. Dynamics of groundwater recharge and seepage over the Canadian landscape during the Wisconsinian glaciation. J. Geophys. Res. 113, F01011. doi:10.1029/2007/JF000838.
Lunkka, J.P., Saarnisto, M., Gey, V., Demidov, I., Kiselova, V., 2001. Extent and age of the Last Glacial Maximum in the southeastern sector of the Scandinavian Ice Sheet. Global Planet. Change 31, 407–425.
Luterbacher, J., Dietrich, D., Xoplaki, E., Grosjean, M., Wanner, H., 2004. European seasonal and annual temperature variability, trends and extremes since 1500. Science 303, 1499–1503 (with supporting online material).
Majorowicz, J., Šafanda, J., 2007. Heat flow variation with depth in Poland: evidence from equilibrium temperature logs in 2.9-km-deep well Torun-1. Int. J. Earth Sci. (Geol. Rundsch.). doi:10.1007/s00531-007-0210-2.
Majorowicz, J., Wybraniec, S., 2010. New terrestrial heat flow map of Europe after regional paleoclimatic correction application. Int. J. Earth Sci.. doi:10.1007/s00531-010-0526-1.
Manning, C., Ingebritsen, S.E., 1999. Permeability of the continental crust: implications from geothermal data and metamorphic systems. Rev. Geophys. 37, 127–150.
Mareschal, J.-C., Rolandone, F., Bienfait, G., 1999. Heat flow variations in a deep borehole near Sept-Iles, Québec, Canada: palaeoclimatic interpretation and implications for regional heat flow estimates. Geophys. Res. Lett. 26, 2049–2052.
Mottaghy, D., Popov, Y.A., Schellschmidt, R., Clauser, C., Kukkonen, I.T., Nover, G., Milanovsky, S., Romushkevich, R.A., 2005. New heat flow data from the immediate vicinity of the Kola superdeep borehole: vertical variation in heat flow density confirmed and attributed to advection. Tectonophysics 401 (1–2), 119–142.
Munk, W.H., MacDonald, G.J.F., 1960. The Rotation of the Earth: A Geophysical Discussion. Cambridge University Press, London, 323 pp.
Newman, G.A., Hoversten, G.M., 2000. Solution strategies for two- and three-dimensional electromagnetic inverse problems. Inverse Prob. 16, 1357–1375.
Papunen, H., Gorbunov, G.I. (Eds.), 1985. Nickel deposits of the Baltic Shield and Scandinavian Caledonides. Geol. Surv. Finland, Bull. 333.
Paterson, W.S.B., 1994. The Physics of Glaciers, third ed. Elsevier, Oxford, 480 pp.
Peltonen, P., Kontinen, A., Huhma, H., Kuronen, U., 2008. Outokumpu revisited: new mineral deposit model for the mantle peridotite-associated Cu–Co–Zn–Ni–Ag–Au sulphide deposits. Ore Geol. Rev. 33, 559–617.
Person, M., McIntosh, J., Bense, V., Remenda, V.H., 2007. Pleistocene hydrology of North America: the role of ice sheets in reorganizing groundwater flow systems. Rev. Geophys. 45, 3007. doi:10.1029/2006RG000206, 28 pp.
Popov, Yu. A., Pimenov, V.P., Pevzner, L.A., Romushkevich, R.A., Popov, E. Yu., 1998. Geothermal characteristics of the Vorotilovo deep borehole drilled into the Puchezh-Katunk impact structure. Tectonophysics 291, 205–223.
Popov, Y.A., Pevzner, S.L., Pimenov, V.P., Romushkevich, R.A., 1999a. New geothermal data from the Kola superdeep well SG-3. Tectonophysics 306, 345–366.
Popov, Yu. A., Romushkevich, R.A., Popov, E. Yu., Bashta, K.G., 1999b. Geothermicheskiy characteristikiy razreza SG-4 (in Russian) (Geothermal characteristics of profile SG-4). In: Khakhaev, B.N., Morozov, A.F., Gorbachev, V.I., Pevzner, L.A., Bashta, K.G., Dokuchaev, A.J., Kenig, V.V., Mezelovskyi, N.V., Rapoport, M.S., (Eds.), Resultatiy Bureniya i Issledovannih Uralskoy Sverghlubukoy Shkvashiny (SG-4) (Results of drilling and investigations of the Ural super-deep borehole (SG-4)), Ministerstvo Prirodnyih RF, Jaroslavl, pp. 77–88.

Ramey, H.J., 1962. Well-bore heat transmission. J. Pet. Technol. 14, 427–435.
Rath, V., Mottaghy, D., 2007. Smooth inversion for ground surface temperature histories: estimating the optimum regularization parameter by generalized cross-validation. Geophys. J. Int. 171, 1440–1448.
Rybach, L., 1973. Warmeproduktionsbestimmungen an Gesteinen der Schweizer Alpen. Beitr. Geol. Schweiz, Geotech. Ser., vol. 51, 43 pp.
Šafanda, J., Szewzyk, J., Majorowicz, J.A., 2004. Geothermal evidence of very low glacial temperatures on a rim of the Fennoscandian ice sheet. Geophys. Res. Lett. 31, L07211. doi:10.1029/2004GL019547.
Säntti, J., Kontinen, A., Sorjonen-Ward, P., Johanson, B., Pakkanen, L., 2006. Metamorphism and chromite in serpentinized and carbonate-silica-altered peridotites of the Paleoproterozoic Outokumpu-Jormua ophiolite belt, eastern Finland. Int. Geol. Rev. 48, 494–546.
Seppä, H., Poska, A., 2004. Holocene annual mean temperature changes in Estonia and their relationship to solar insolation and atmospheric circulation patterns. Quatern. Res. 61, 22–31.
Slagstad, T., Balling, N., Elvebakk, H., Midttømme, K., Olesen, O., Olsen, L., Pascal, C., 2009. Heat-flow measurements in Late Palaeoproterozoic to Permian geological provinces in south and central Norway and a new heat-flow map of Fennoscandia and the Norwegian–Greenland Sea. Tectonophysics 473, 341–361.
Sokolnicki, M., Heikkinen, P., 2008. Difference Flow Measurements in Outokumpu Deep Hole OKUR2500. Posiva Oy, Finland, Work. Rep., 2008-87, 19 pp.
Spane, F.A., 2002. Considering barometric pressure in groundwater flow investigations. Water Resour. Res. 38 (6). doi:10.1029/2001WR000701.
Svendsen, J.I., Astakhov, V.I., Bolshianov, D.Yu., Demidov, I., Dowdeswell, J.A., Gataullin, V., Hjort, C., Hubberten, H.W., Larsen, E., Mangerud, J., Melles, M., Möller, P., Saarnisto, M., Siegert, M.J., 1999. Maximum extent of the Eurasian ice sheets in the Barents and Kara Sea region during the Weichselian. Boreas 28, 234–242.
Svendsen, J.I., Alexanderson, H., Astakhov, V.I., Demidov, I., Dowdeswell, J.A., Funder, S., Gataullin, V., Henriksen, M., Hjort, C., Houmark-Nielsen, M., Hubberten, H.W., Ingolfsson, O., Jakobsson, M., Kjær, K.H., Larsen, E., Lokrantz, H., Lunkka, J.-P., Lys, A., Mangerud, J., Matioushkov, A., Murray, A., Möller, P., Niessen, F., Nikolskaya, O., Polyak, L., Saarnisto, M., Siegert, C., Siegert, M.J., Spielhagen, R.F., Stein, R., 2004. Late Quaternary ice sheet history of northern Eurasia. Quatern. Sci. Rev. 23, 1229–1271.
Tarvainen, A.-M., 2006. Identification of water-bearing structures in the Outokumpu Deep Drill hole by geophysical well logging. Master's thesis (technology), Helsinki University of Technology, Department of Civil and Environmental Engineering, 93 p (in Finnish with English abstract).
Ukkonen, P., Lunkka, J.P., Jungner, H., Donner, J., 1999. New radiocarbon dates from Finnish mammoths indicating large ice free areas in Fennoscandia during the Middle Weichselian. J. Quat. Sci. 14 (7), 711–714.
Vaikmäe, R., Vallner, L., Loosli, H.H., Blaser, E.C., Juillard-Tardent, M., 2001. Palaeogroundwater of glacial origin in the Cambrian–Vendian aquifer of northern Estonia. Geol. Soc. London Special Publ. 189, 17–27.
Västi, K., 2005. Outokumpu deep drill core: geological report. Geological Survey of Finland, Report M 52.5/4222/04/R2500, 24.03.2006, 13 pp.
Wahba, G., 1990. Spline Models for Observational Data. Society for Industrial Mathematics, Philadelphia.
Van der Kamp, G., Bachu, S., 1989. Use of dimensional analysis in the study of thermal effects of various hydrogeological regimes. In: Beck, A.E., Garven, G., Stegena, L. (Eds.), Hydrogeological Regimes and their Subsurface Thermal Effects. Am. Geophys. Union Geophys. Monogr., vol. 47, pp. 23–28.