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Heat Flow Density in Estonia - Assessment of Palaeoclimatic and Hydrogeological Effects

A. Jõeleht[1] *and I.T. Kukkonen*[2]

[1] Institute of Geology, University of Tartu, Vanemuise 46, EE2400 Tartu, Estonia
(previously at: Geological Survey of Estonia, Pikk 67, EE0001 Tallinn, Estonia)
[2] Geological Survey of Finland, Betonimiehenkuja 4, FIN-02150 Espoo, Finland

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Abstract

New measurements of heat flow density have been made in six boreholes in northern and western Estonia. The mean heat flow values range from 20 to 40 mW m^{-2}. All holes display a vertical variation in apparent heat flow densities from 15 to 52 mW m^{-2}. Since most of the holes are shallow and therefore sensitive to surficial disturbances, the effects of palaeoclimatic ground temperature changes and heat transfer by groundwater flow were studied with numerical models. Forward modelling of palaeoclimatic effects was made with a homogeneous half-space model in which the assumed climate history covered glaciations during the past million years, the Holocene, and the recent climatic warming. The influence of the sea cover after deglaciation was also included. The palaeoclimatic models suggested that the measured vertical variation might be partly attributed to the palaeoclimatic effects, but when the corrections were applied to the measured data they did not entirely eliminate the vertical variation in heat flow density. This is probably due to thermal conductivity structures that deviate from the assumed half-space conditions and the palaeoclimatic ground temperature history used in the models. Fluid and heat transfer in the subsurface were simulated with finite difference techniques. The simulation results of a 2-dimensional model indicate that the thermal effect of regional flow systems is less than 5 mW m^{-2} in most of Estonia. Larger perturbations may occur in the southeast of the country, where the hydraulic gradient is higher.

Key words: geothermics, paleoclimatic, hydrogeology, drill holes, Estonia

1. *Introduction*

Geothermal information on Estonia, which lies on the southern slope of the Baltic (Fennoscandian) Shield, is based on heat-flow density (HFD) measurements provided by *Urban et al.* (1991). According to them, heat flow density is generally about 40 mW m^{-2}, but there are considerable local anomalies. Borehole depths range from 60 to 350 m. Most of the boreholes are in the Phanerozoic sediments; only five reach the Precambrian basement. The data used by *Urban et al.* (1991) were originally produced for hydrogeological purposes to characterize aquifer temperatures. Therefore, the depth interval between temperature readings in single boreholes varied from 2.5 to 50 m. Although *Urban et al.* (1991) did not report any thermal conductivity measurements in detail their data suggest that the average sampling interval was 20 m.

Geothermal logs are sensitive to convective and conductive disturbances such as shallow groundwater circulation and transient conductive disturbances created by palaeoclimatic changes in ground surface temperatures. These effects have not yet been discussed with reference to the area of Estonia.

We here present a palaeoclimatic ground temperature history for the study area compiled from various proxy and meteorological data. The data were used to calculate the palaeoclimatic disturbances to HFD data with half-space models.

The thermal effects of groundwater circulation in Phanerozoic sediments are discussed with the aid of numerical modelling. A 2-dimensional finite difference simulation of fluid and heat transfer in the subsurface is presented for a 230-km-long transect running from Haanja, southern Estonia, to Võsu, northern Estonia.

Finally we present new temperature, fluid resistivity and thermal conductivity data from six boreholes in northern and western Estonia, ranging from 110 to 810 m in depth.

2. *Palaeoclimatic effects on subsurface temperatures*

Climatic temperature changes at the ground surface propagate downwards to the subsurface, creating transient disturbances in the geothermal gradient. Due to the poor thermal diffusivity of rocks, the disturbances persist in the bedrock for a long time. To obtain an undisturbed steady-state heat flow density value, the measured values should therefore be corrected for ground surface temperature histories (e.g. *Beck,* 1977). Conversely, subsurface temperature logs can be applied to reconstruct the past ground surface temperature history (e.g. *Čermák et al.* 1993).

The surface temperature history of Estonia during the last million years (Table 1) was compiled from the literature. Deglaciation in the study area took place at 11 500 BP (*Donner* and *Raukas,* 1989). The periods of earlier glaciations and deglaciations were adapted from studies conducted in Finland (*Kukkonen,* 1987) using the oxygen-isotope curve of ocean sediments (*Shackleton* and *Opdyke,* 1976), which represents the variation in the ice volume in glaciers. The surface temperature under the ice sheets was assumed to represent pressure melting, that is, -1°C. The value is the same as in *Jessop* (1971) and *Kukkonen* (1987) for the Canadian and Baltic Shields, respectively.

Palaeontological (*Iversen*, 1944) and palaeobotanical (*Donner*, 1978) studies indicate that 8000-5000 years ago the air temperature was +2°C higher than it is today. The warmer period (+1 °C) continued until 2500 BP (*Aaby*, 1976). Little is known about "the Little Climatic Optimum" (700-1300 AD) and "the Little Ice Age" (1300-1700 AD) in the Shield area and so data on western Europe (*Bell* and *Walker*, 1992) were used. Regular meteorological temperature observations were started in Estonia in 1828. The air temperature is reported to have increased by 0.5°C since the end of the last century (*Jaagus*, 1994).

Table 1. Ground surface temperature changes relative to present values in Estonia. The data were compiled from various geological, proxy and meteorological sources (see text). The time of deglaciation (t_g) and the ground temperature difference between glaciation and the present (ΔT_g) were determined for each site on the basis of its deglaciation history. For northern and western Estonia, the values are 11500 years and -7 K, respectively. Outside the given t_i - t_{i+1} intervals the ground surface temperature was assumed to be equal to its present value.

ΔT_i	t_{i+1}	t_i
-0.2	95	295
-0.9	295	695
0.4	695	1295
0.4	2500	5000
0.9	5000	8000
ΔT_g	t_g	30000
"	40000	45000
"	60000	75000
"	126000	194000
"	240000	290000
"	340000	357000
"	429000	471000
"	479000	533000
"	573000	671000
"	680000	716000
"	732000	794000
"	803000	848000

The ground surface temperature is not the same as the air temperature, although these parameters are often linearly correlated (*Kukkonen*, 1987). The relation between air and surface temperatures was determined from data published by the Estonian Meteorological and Hydrological Institute (*Prokhorov*, 1970). Annual mean soil

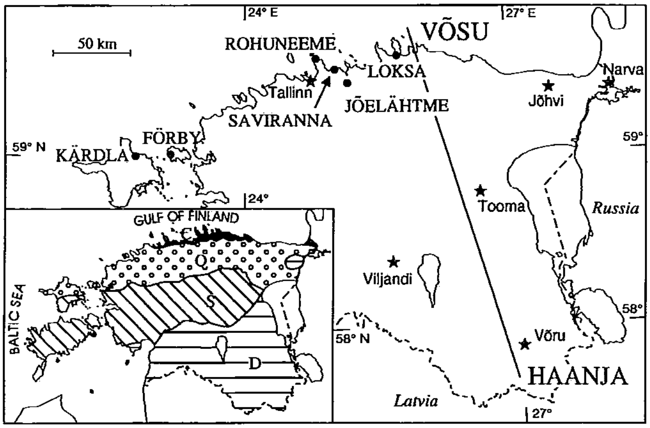


Fig. 1. Location of meteorological stations (stars) and new heat flow density measurement sites (dots; Table 3), the cross-section used for hydrogeological modelling and the main outcropping stratigraphic units in Estonia: Є , Cambrian, O , Ordovician, S , Silurian and D , Devonian.

temperatures at a depth of 40 cm at six meteorological stations (Fig. 1) and corresponding air temperatures at the same stations for the period 1952-1965 were used for regression analysis. The air temperatures range from 3.0 to 6.7°C and surface temperatures at a depth of 40 cm from 5.3 to 7.6°C. To avoid over-representation of some temperature intervals, the data were divided into 0.5°C air temperature intervals for which corresponding average ground temperature values were then calculated. The following regression line was obtained (Fig. 2):

$$T_S = 0.44\, T_a + 4.49, \tag{1}$$

where T_s and T_a are annual mean surface and air temperatures (°C), respectively. Eq. (1) was further used to convert palaeoclimatic air temperature estimates based on various geological, botanical and other proxy indicators to surface temperatures.

When the Weichselian ice sheet had already retreated to the north, coastal areas were still covered by sea. Because of glacio-isostatic uplift of the lithosphere, the Baltic Sea retreated westwards and northwards, exposing new land areas. Depending on the location and present elevation above sea level, the time of retreat ranges in Estonia from 9000 years BP in more elevated areas to 1000 years BP in areas close to the present shore line.

Recent studies in the Baltic Proper and in the Gulf of Finland (*Kullenberg,* 1981; *Haapala* and *Alenius,* 1994) show that at depths exceeding 50 m the seawater temperature is close to +4 °C, with annual variations of a few tenths of a degree. In shallower depths, the water is effectively mixed and the sea bottom temperatures more or less follow the air temperature variations. Due to the lack of more accurate information, we assume that the present ground surface temperature and the sea surface temperature at the same location are equal (+6 °C) and that the sea bottom temperature varies linearly from 4°C (50 m) to the prevailing surface temperatures at zero sea depth.

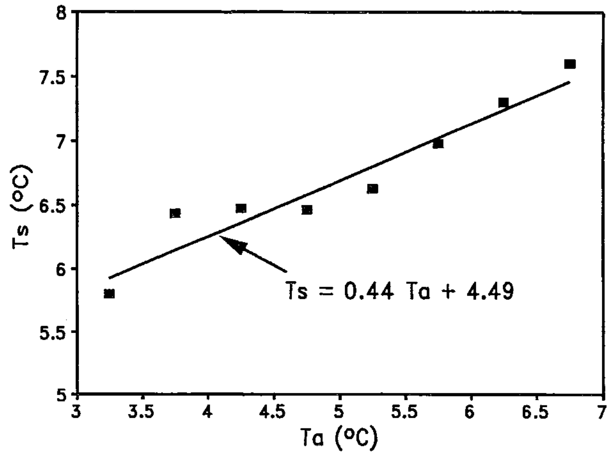


Fig. 2. Annual mean air temperature versus ground temperature in Estonia at 40 cm depth below the ground surface. Data from the six stations of the Estonian Meteorological and Hydrological Institute for 1952-1965. The points represent mean values of ground temperatures calculated from the original annual mean temperature data arranged into groups of 0.5 °C intervals of air temperature. The regression line corresponds to Eq. (1).

General palaeoclimatic ground surface temperature histories were constructed separately for northern and western Estonia. The temperature histories were approximated with stepwise changes (Table 1). The surface temperature variation was calculated as a function of sea depth and air temperature. Air temperatures were converted to ground temperatures with Eq. (1).

The subsurface thermal responses to the palaeoclimatic ground surface temperature histories were calculated with a homogeneous half-space model with constant diffusivity (s). The subsurface temperature $\Delta T(z,t)$ and gradient disturbances $\Delta\Gamma(z,t)$ due to several instantaneous changes in surface temperature ΔT_i during the time period $t_{i+1}<t<t_i$ are (*Carlslaw* and *Jaeger*, 1959; *Jaeger*, 1965; *Čermák*, 1976):

$$\Delta T(Z)_i = \Delta T_i \left(erf \frac{Z}{\sqrt{4st_i}} - erf \frac{Z}{\sqrt{4st_{i+1}}}\right) \tag{2}$$

and

$$\Delta\Gamma(Z)_i = \Delta T_i [(\pi st_i)^{-1/2} \exp(-z^2/4st_i) - (\pi st_{i+1})^{-1/2} \exp(-Z^2/4st_{i+1})]. \tag{3}$$

A thermal diffusivity of $1 \cdot 10^{-6}$ m^2 s^{-1} was used for all models.

In general, the results of models for different parts of Estonia (Fig. 3) are similar, since there are no significant differences between ground surface temperature histories. The results suggest that the palaeoclimatic effect reduces the geothermal gradient by 3-6 mK m^{-1} in the uppermost 200 m, which is the depth interval most affected by temperature changes during the Holocene.

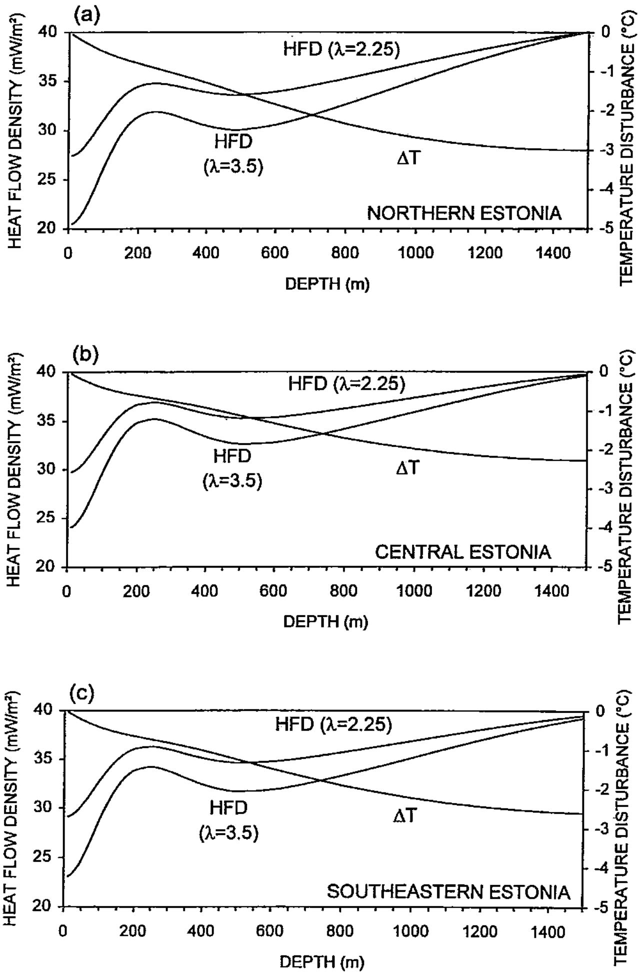


Fig. 3. Theoretical palaeoclimatic temperature disturbance and heat flow density *versus* depth curves calculated from the ground temperature history in Table 1. Calculations were made for (a) northern and western, (b) central and (c) southeastern Estonia. The HFD curves were calculated for sedimentary cover (thermal conductivity 2.25 W $m^{-1}K^{-1}$) and for the Precambrian basement (3.5 W $m^{-1}K^{-1}$) by assuming a basal heat flow density of 40 mW m^{-2}.

Deeper down the palaeoclimatic effect diminishes, and is mainly due to glaciations. The effects extend down to a depth of 1200 m, below which the gradient changes are less than 1 mK m^{-1}. Theoretical palaeoclimatic temperature distubance and heat flow-depth curves are also presented in Fig. 3. Due to the Holocene temperature changes, the palaeoclimatic effect is less than 3 mK m^{-1} at depths between 150 and 400 m. If we assume that the palaeoclimatic histories are correct, the result indicates that, even at shallow depths, heat flow density measurements may yield values close to those of undisturbed situations. Depending on conductivity, the deviation would not be more than 5-7 mW m^{-2}.

In detail, the modelling suggests slight differences between northern and western coastal areas, the central Estonian upland and southeastern Estonia. The temperature perturbations in central Estonia are smaller than those in southeastern Estonia. In central Estonia, the present surface temperature is lower (< 5°C) than elsewhere (5.5°C in southeastern Estonia and 6°C in northern and western coastal areas) and surface temperature changes between the present and glaciations are relatively smaller. The disturbances in heat flow density do not, however, differ by more than 4-5 mW m^{-2} (Fig. 3).

The ice sheet withdraw later in northern and western Estonia than in the southwest, and so the modelling gave the largest temperature perturbations for these areas. Since the coastal areas were covered by sea after deglaciation, the calculated gradient changes depend on the elevations of the borehole sites. For instance, in the Rohuneeme and Förby holes, drilled very close (< 30 m) to the shoreline, the gradient perturbations are about 1 mK m^{-1} in the uppermost 300 m. This is more than in Jõelähtme and Saviranna, which are situated farther away from the shore, at elevations of 32 and 16 m, respectively (Fig. 4).

3. *Hydrogeological effects on subsurface temperatures*

Quaternary sediments cover the Phanerozoic deposits ranging from Vendian to Devonian on Precambrian basement. The basement depth increases rather monotonously from north to south, from 150 m on the shore of the Gulf of Finland to 600 m in southern Estonia (*Koistinen*, 1994). Aquifers suitable for water production are usually only a few tens of meters thick. The sedimentary rocks are very poorly consolidated and, with the exceptance of the Lower Cambrian clay formation, have rather high hydraulic conductivities (about 10^{-5}-10^{-4} m s^{-1}). These Lontova stage clays have much lower hydraulic conductivities (mean value $5 \cdot 10^{-14}$ m s^{-1}; *Pirrus* and *Saarse*, 1979). The hydraulic conductivities of other Phanerozoic sediments were determined from the specific yield data listed in *Tšeban* (1975) and *Arkhangelskiy* (1966):

$$k = 130q/(m \cdot 86400) \tag{4}$$

where k is hydraulic conductivity (m s^{-1}), q specific yield (l $s^{-1}m^{-1}$) and m the thickness (m) of the measured interval.

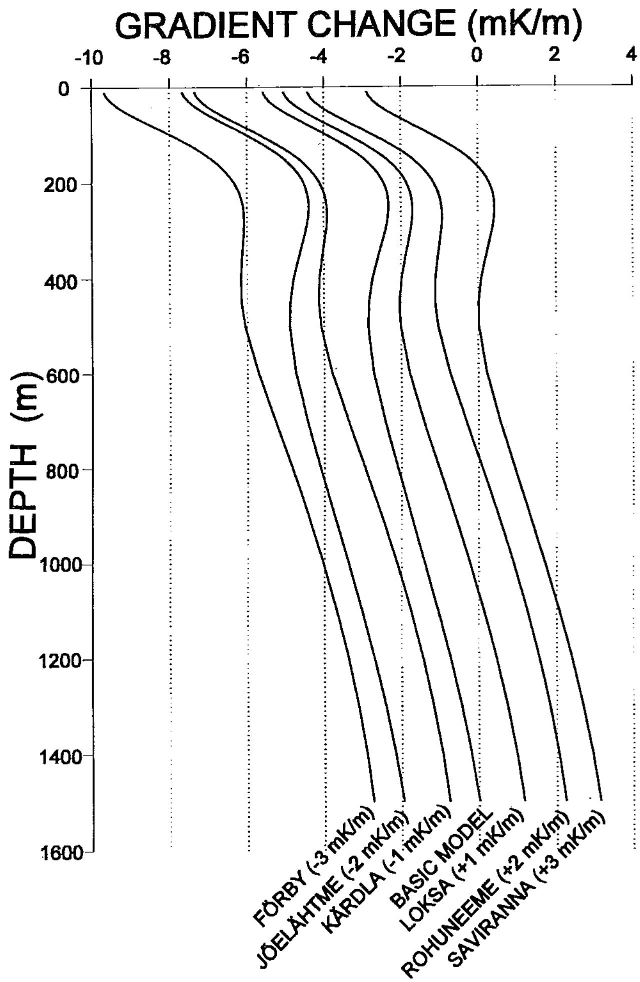


Fig. 4. Simulated temperature gradient changes produced by palaeoclimatic ground temperature variations at the sites of new geothermal measurements. For clarity, the curves were displaced by -3 - +3 mK m^{-1} from the general model for northern Estonia.

The calculated hydraulic conductivity values for the Cambrian-Vendian aquifer range from $6.8 \cdot 10^{-6}$ m s^{-1} to $1.1 \cdot 10^{-4}$ m s^{-1}, with a mean of $5.4 \cdot 10^{-5}$ m s^{-1}. For the Cambrian-Ordovician aquifer the mean is $4.8 \cdot 10^{-5}$ m s^{-1}, and for the Ordovician and Silurian strata $8.3 \cdot 10^{-5}$ and $9.6 \cdot 10^{-5}$ m s^{-1}, respectively. In the Ordovician and Silurian strata the hydraulic conductivity decreases downwards (Fig. 5) due to increased fracturing in the uppermost parts of these mainly calcareous sediment deposits. Further down, lithostatic pressure reduces the size of fracture apertures and conductivity can be attributed to other hydraulically connected porous textures. Other stratigraphic units have more or less constant hydraulic conductivity with depth (Fig. 5).

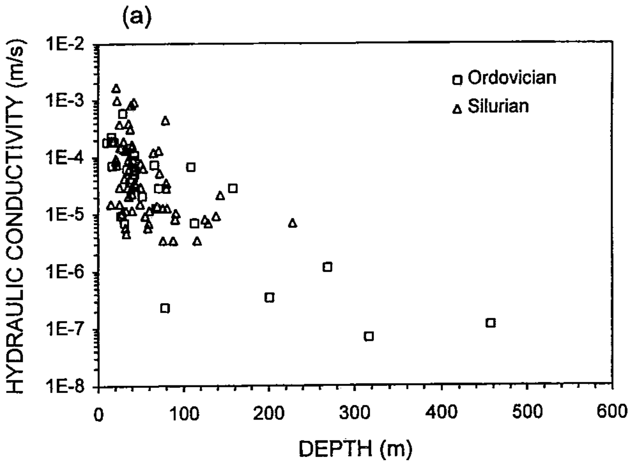
The area of Estonia is characterized by low topography, most of the country being less than 100 m above sea level. The regional hydraulic gradients estimated from topography range from 0.01 to 0.0001, depending on the assumed flow distances. Locally, there are higher elevations in the north and southeast of the country, and higher flow gradients may occur.

Since the hydraulic conductivities are high and the sediment beds are continuous over long distances without any major tectonic breaks, advective heat transfer may be large enough to affect subsurface temperatures. We have investigated the thermal effects of the Estonian hydrogeological system with numerical simulations of coupled heat and fluid transfer by assuming a porous medium. The numerical code SHEMAT (*Clauser,* 1988; *Clauser* and *Villinger,* 1990) was used for the simulations, which applies finite difference techniques to the numerical solutions.

A north-south transect running from Haanja to Võsu was chosen for modelling (Fig. 1), and general hydrogeological model was constructed from the geological and hydrogeological profiles of *Tšeban* (1975). The 2-dimensional model is 600 m deep and 230 km long. The discretization applied was 25 m in the vertical dimension and 5 km in the horizontal. The length of the model is justified by the need to investigate the existence of large-scale flow systems in the sediment aquifers as suggested by the hydraulic head variation maps of *Tšeban* (1975).

The thermal conductivity of the sedimentary rocks was assumed to be 2.25 W $m^{-1}K^{-1}$ except in the clays of the Lower Cambrian Lontova stage, which have a value of 1.5 W $m^{-1}K^{-1}$ (*Urban et al.,* 1991). Porosity was 10% and heat production 0.5 μW m^{-3} throughout. The thermal and hydraulic properties used in simulation are given in Table 2.

Topographic variation was not included in the model, but each upper boundary node was assigned a constant hydraulic head following the topography and the measured head variation (*Tšeban,* 1975).



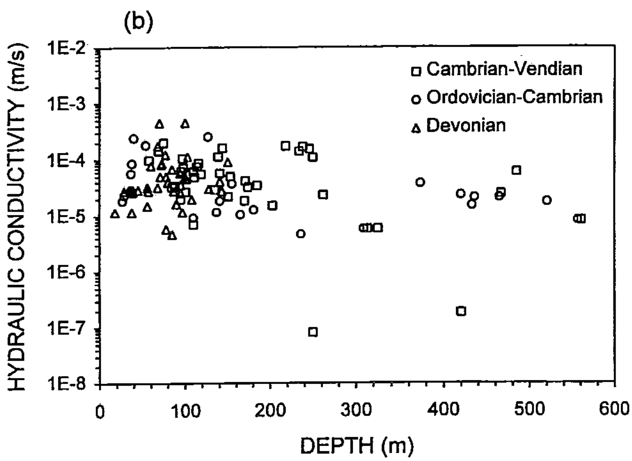


Fig. 5. Hydraulic conductivity versus depth estimated from the transmissivity data of *Tšeban* (1975) and *Arkhangelskiy* (1966). In (a) the Ordovician and Silurian strata the hydraulic conductivity decreases with depth, in (b) the Vendian-Cambrian, Cambrian-Ordovician and Devonian sediments there is no significant change with depth.

The following boundary conditions were set:

(1) Top: constant temperature (5 °C) and constant head (following topography) adapted from the maps of *Tšeban* (1975). The head varied from about 160 m a.s.l. in the Haanja area to zero on the shore of the Gulf of Finland (Fig. 6).

(2) Bottom: Constant heat flow density (40 mW m^{-2}) and no fluid flow.

(3) Lateral boundaries: No flow of heat or fluid was allowed.

Table 2. Hydraulic and thermal properties of different stratigraphic units used to simulate groundwater flow in the bedrock. Domain numbers refer to Fig. 6.

Stratigraphic unit / Depth	Hydraulic conductivity m/s	Thermal conductivity $Wm^{-1}K^{-1}$	Domain number
Devonian	$5 \cdot 10^{-5}$	2.25	1
Silurian			
0-100 m	$5 \cdot 10^{-5}$	2.25	2
100-200 m	$1 \cdot 10^{-5}$	2.25	3
Ordovician			
0-100 m	$5 \cdot 10^{-5}$	2.25	4
100-200 m	$1 \cdot 10^{-5}$	2.25	5
200-300 m	$5 \cdot 10^{-7}$	2.25	6
> 300 m	$1 \cdot 10^{-7}$	2.25	7
Cambrian-Ordovician	$5 \cdot 10^{-5}$	2.25	8
Lontova clays	$1 \cdot 10^{-14}$	1.5	9
Cambrian-Vendian	$5 \cdot 10^{-5}$	2.25	10
Precambrian	$1 \cdot 10^{-8}$	3.50	11

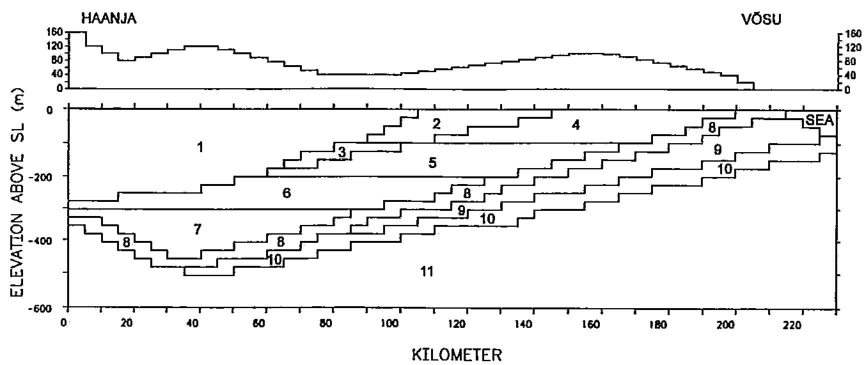


Fig. 6. The cross-section and hydraulic head variations on the transect used for simulating groundwater flow (Fig. 1). The domain numbers are as in Table 2.

The model results (Fig. 7) suggest that there are regional groundwater flow systems. The Cambrian-Vendian aquifer covered by Lontova clays recharges in southern Estonia and discharges to the Gulf of Finland. In other aquifers the groundwater flow directions follow the topography. The model further suggests that the thermal field is not markedly disturbed by groundwater flow, and that the thermal effects are less than 5 mW m^{-2} in most parts of the model. The highest heat flow density differences are confined to the uppermost 100-150 m. However, in the southern part of the transect, which has thick Devonian and Ordovician layers and where the hydraulic head reaches 160 m, the model suggests heat flow highs of up to 60-70 mW m^{-2} in the groundwater discharge areas. The situation would be different if the overlying strata had much lower hydraulic conductivities than the values applied here. We used hydraulic conductivity values measured in boreholes originally drilled and investigated for water production purposes, not regional groundwater flow studies. The high conductivity aquifers may therefore be over-represented in the data.

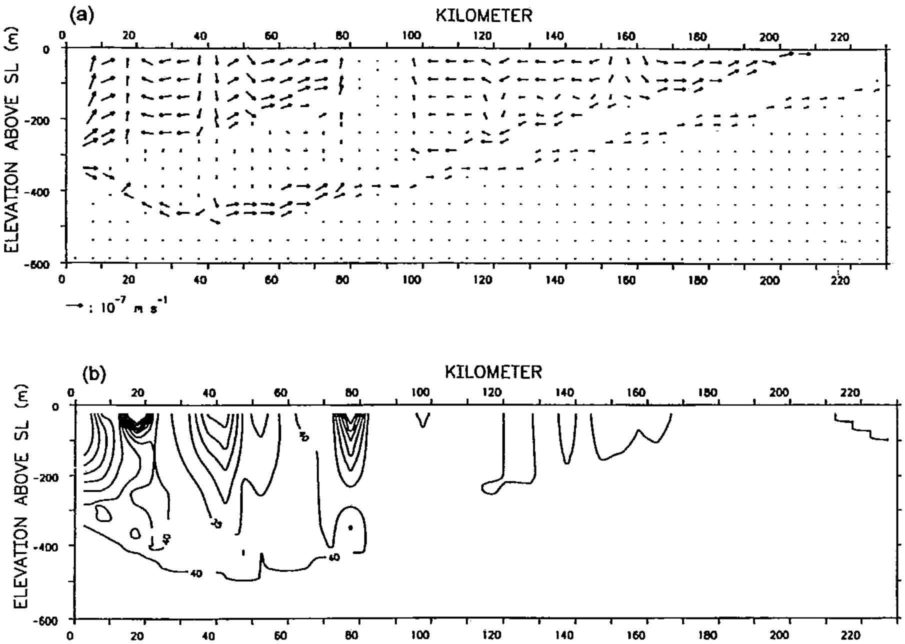


Fig. 7. The simulated (a) Darcy velocity and (b) heat flow density in the transect. The scale of Darcy velocity is logarithmic-exponential.

4. *New measurements of heat flow density*

More than 150 holes have been drilled in the Estonian terrain to the Precambrian basement, but practically all are cemented or the casing has not been preserved. The number of open holes suitable for geothermal loggings is thus very small, and most of them are very shallow (< 100 m).

Six holes (Table 3, Fig. 1) were chosen for this study following two criteria: 1) they should be as deep as possible, and 2) they should reach the Precambrian basement.

The Precambrian basement was reached in all boreholes except that at Rohuneeme. At Loksa the hole was blocked at 132 m, and only 2 m of the Precambrian basement could be logged. At Jõelähtme, logging covered about 25 m of the Precambrian, but at Saviranna and Förby 60 m and 40 m, respectively. The deepest logging was made at Kärdla, where the drill hole penetrated about 300 m of the Precambrian basement under a meteorite impact crater (800 m below surface).

Temperature was measured with equipment constructed at the Geological Survey of Finland. The measuring element was a temperature-sensitive microcircuit (Analog Device 590H) capable of an absolute accuracy of 0.1°C and a resolution of 0.01°C. Temperature readings were taken at 2.5 m depth intervals.

Table 3. Sites of new heat flow density measurements (see also Fig. 1).

Hole	No.	Location		Elevation	Depth	Drilling
		N	E	(m)	(m)	year
Kärdla	K-1	58°59'	22°40'	6.0	815	1990
Förby	F369	59°00'	23°09'	1.0	306	1987
Jõelähtme	8-1	59°27'	25°10'	32.5	218	1991
Saviranna	F502	59°20'	25°03'	16.0	212	1979
Loksa	853	59°35'	25°44'	7.5	157	1989
Rohuneeme	6-2	59°34'	24°48'	2.8	116	1991

The electrical resistivity of drill hole fluid was logged with a two-wire cable and a resistivity meter. The reading interval was 5 m. Because of the casing structures at the borehole top, fluid resistivity could not be measured at Jõelähtme and Loksa. All drill hole loggings were made during August 1993.

The drill cores were sampled at 10 m intervals wherever possible. Thermal conductivity was measured at the Geological Survey of Finland with the divided bar method on rock disks cut perpendicular to the core samples. Vendian and Cambrian clays, sandstones and siltstones are very poorly consolidated and their conductivities could not be measured with the divided bar apparatus. The conductivity data from *Urban et al.* (1991) were used for these sections.

Heat flow densities were determined with the interval technique using mean conductivity and temperature gradient, but also with the Bullard technique (temperature-thermal resistance plots).

The geothermal gradient and heat flow density values for both apparent and palaeoclimatically corrected values are given in Table 4. The palaeoclimatic corrections were calculated for each borehole using the technique referred above. The local sea regression histories were taken into account in the borehole corrections, and are shown in Fig. 8. Bathymetric data were taken from *Donner* and *Raukas* (1992), *Hyvärinen et al.* (1992), *Kessel* and *Raukas* (1979), *Linkrus* (1962) and *Ratas* (1977).

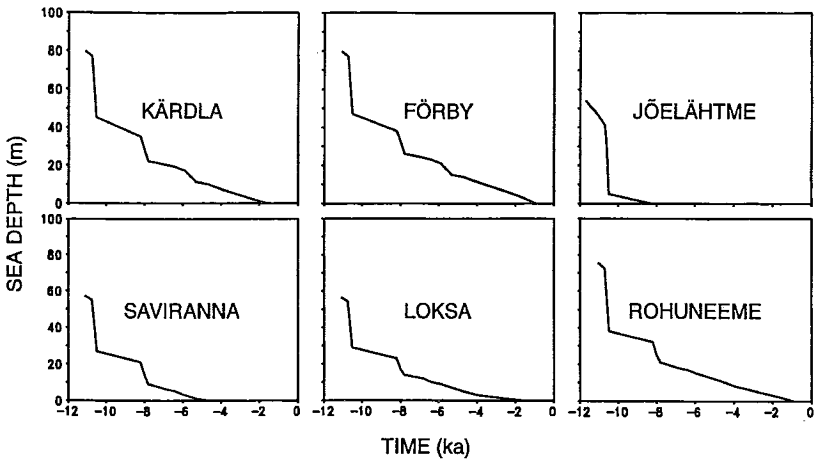


Fig. 8. Sea regression histories at the borehole sites.

Sections of drillholes obviously disturbed by water flow in the hole itself (see Appendix and e.g., *Drury et al.*, 1984) were not included in the heat flow density calculations. A detailed description of logging results is given in the Appendix.

Both apparent and palaeoclimatically corrected heat flow densities vary markedly with depth (Table 4). However, the mean heat flow density values are close to each other (34-40 and 40-48 mW m^{-2}, respectively), except in the Rohuneeme, where they are 20 and 30 mW m^{-2}.

5. *Discussion*

Palaeoclimatic models do not seem to eliminate the vertical variation in heat flow density nor do corrections reduce the standard deviation of the mean heat flow values (Table 4). There are two explanations for this: 1) the palaeoclimatic model is not quite correct, and/or 2) the heat transfer is not purely conductive. A homogeneous half-space is by no means the best model for a layered sedimentary environment, and the values of palaeoclimatic corrections should therefore be taken only as approximations.

Efforts should be made not only to include the geological structures more carefully in the models, but also to take many local climatic and artificial effects into account. The land-use history of a site may contain relevant information about ground surface temperature changes. For instance, deforestation and soil cultivation may have changed the ground temperatures by 1-2°C, which equals or even exceeds climatic warming during the last 200-400 years. The ground surface temperature may differ for several tenths of a degree depending on the present vegetation cover (forest or open fields). Vegetation types changed gradually during the Holocene.

Table 4. Apparent and palaeoclimatically corrected heat flow densities.

Hole	ΔZ	Γ	ΔΓ	λ	Q	Q'
Kärdla	32-130	2.6±0.1	-5.2	2.74±0.20	7.2	21.4
	130-275	18.0±0.1	-3.2	2.44±0.19	44.0	51.8
	275-375	12.3±0.1	-3.0	2.87±0.57	35.4	44.0
	375-495	11.4±0.0	-3.1	2.58±0.50	29.4	37.4
	495-810	12.6±0.3	-2.5	2.93±0.59	36.8	44.1
				mean of interval 275-810 m	33.9±3.2	41.9±3.1
Förby	130-159	14.4±0.3	-4.1	2.18	31.4	40.3
	159-170	28.3±1.3	-3.7	1.35	38.2	43.2
	170-264	15.4±0.2	-3.3	2.48	38.2	46.4
	264-306	14.7±0.3	-3.1	3.12±0.48	45.9	55.5
				mean	38.4±5.1	46.4±5.7
Jõelähtme	51-130	33.6±0.4	-4.3	1.25	42.0	47.4
	130-152	10.2±0.4	-3.2	1.72	17.5	23.0
	152-188	12.0±0.2	-2.8	2.53	30.4	37.4
	188-218	17.2±0.6	-2.5	2.60±0.23	44.7	51.2
				mean	33.7±10.8	39.8±10.9
Saviranna	40-93	38.9±0.7	-5.0	1.24	48.2	54.4
	93-116	14.7±0.3	-4.2	1.74	25.6	32.9
	116-148	13.0±0.2	-3.4	2.57	33.4	42.1
	148-212	15.0±0.1	-2.9	3.41±0.09	51.2	61.0
				mean	39.6±10.5	47.6±10.9
Loksa	32-53	36.1±1.1	-5.4	1.45	52.3	60.2
	53-82	14.9±0.7	-5.3	2.27	33.8	45.9
	82-94	16.2±1.6	-5.0	1.56	25.3	33.1
	94-130	17.5±0.4	-4.3	1.91	33.4	41.6
				mean	36.2±9.9	45.2±9.8
Rohuneeme	20-64	16.3±0.3	-6.1	1.25	20.4	28.0
	64-92	8.8±0.2	-5.2	1.72	15.1	24.1
	92-116	9.3±0.2	-4.7	2.57	23.9	36.0
				mean	19.8±3.6	29.4±5.0

Legend

ΔZ - depth interval (m);

Γ - geothermal gradient (mKm^{-1});

ΔΓ - palaeoclimatic correction (mKm^{-1});

λ - thermal conductivity ($Wm^{-1}K^{-1}$), values without standard deviation are from *Urban et al.* (1991) from nearby holes;

Q and Q' - apparent and palaeoclimatically corrected heat flow densities ($mW\ m^{-2}$).

Numerical simulation of groundwater circulation in the bedrock indicates that there are regional groundwater flow systems. Their existence can be mainly attributed to the high hydraulic conductivity of all sediments (excluding the Lontova clays). However, the small thickness of sedimentary rocks is responsible for small heat flow density disturbances. The thicker sedimentary cover in southern Estonia allows groundwater to flow at greater depth. These flow systems are driven by the greater topographic variation. The particularly low apparent heat flow values in western Estonia reported by *Urban et al.* (1991) can be attributed to a very local recharge of the aquifers, or even to flow in the holes. These postulates are supported by the fact that the holes are situated above or very close to major 10-20-km-wide fracture zones which affect both the Precambrian basement and the sedimentary cover (*Norman* and *Solovjova*, 1982, unpublished report in the Geological Fund of Estonia).

The low apparent heat flow density values measured under the Lontova clays at Jõelähtme, Saviranna, Loksa and Rohuneeme (see Appendix) may be due to the southward migration of seawater in these layers. Seawater intrusion is caused by extensive pumping of water from the Cambrian-Vendian aquifer during the last 50 years. Geothermal investigations may provide a valuable tool for the modelling and experimental testing of the above, as yet unproven, hypothesis.

6. *Conclusions*

According to the present forward modellings of palaeoclimatic effects, the greatest changes can be expected in northern and western Estonia, but the amplitude of perturbations depends on the location and elevation of the borehole site. In coastal areas, in the uppermost part of the cross-section (0-100 m), the apparent heat flow density has declined by 10-15 mW m^{-2} due to past climatic changes. In central and southeastern Estonia, the perturbations are smaller at these depths (7-10 mW m^{-2}). At depths of 150-400 m the palaeoclimatic effects are less than 5-7 mW m^{-2}.

Regional groundwater flow does not seem to cause significant thermal disturbances. Perturbations higher than 5 mW m^{-2} occur only in the uppermost 100 m and in areas where the hydraulic gradient exceeds 0.002.

Probably owing to the distinctly layered structure in the boreholes, and to the relatively scarce data on thermal conductivity, particularly from less consolidated sections of boreholes, the vertical heat flow density variations cannot be explained by either the present palaeoclimatic models or regional groundwater flow. Terrain effects, and groundwater flow effects such as sea water intrusion to Cambrian sediments also need to be studied in detail.

The heat flow density mean values presented here are in agreement with earlier data on geothermal conditions in Estonia. The average heat flow density in Estonia is 35-40 mW m^{-2}.

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APPENDIX

The following gives details of the logging results of the six boreholes measured in 1993.

Kärdla, Hiiumaa Island

K-1, drilled in 1990, is currently the deepest borehole in Estonia. It is situated in the centre of a meteorite crater formed about 455 Ma ago (*Puura* and *Suuroja,* 1992). This impact structure has a diameter of 4 km. The drillhole intersects post-impact Quaternary and Ordovician sediments, underlain by various allochthonous and autochthonous impact breccias (Fig. 9). At a depth of 512 m, it reaches the Precambrian basement, which is heavily fractured due to the impact. The shock effects decline with depth, but are still present at the borehole bottom.

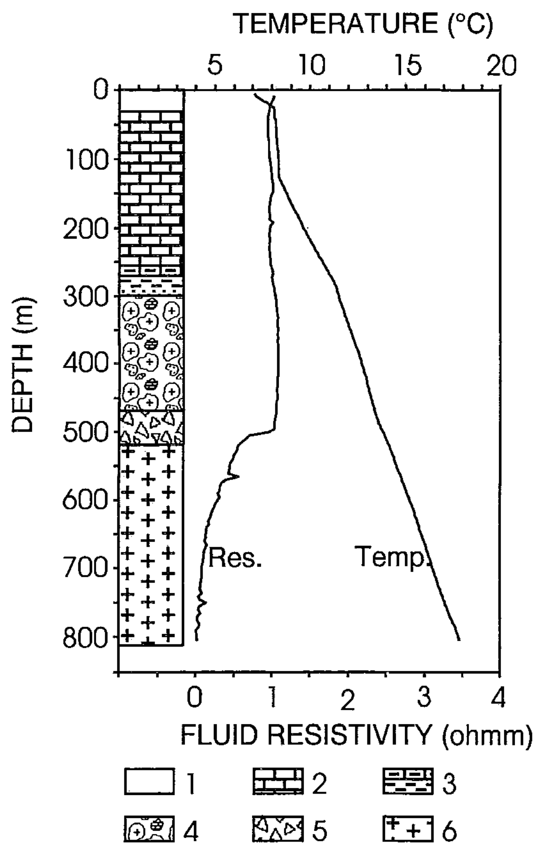

Fig. 9. Temperature and fluid resistivity at the Kärdla borehole. Cross-section: 1. overburden (Q), 2. limestone (O_2tt-O_3ad), 3. transition from sandstone to marl (O_2pl), 4. allochthonous impact breccia (O_2krd), 5. autochthonous impact breccia (O_2krd), 6. basement (PR_1).

The temperature log shows several intervals with distinctly differing gradients. The depths of gradient changes coincide with the main lithological and stratigraphic boundaries. The highest gradient was recorded at the interval 6-32 m (52 mK m^{-1}). This can be attributed partly to the low thermal conductivity of Quaternary glacial till and partly to annual temperature variations at these shallow depths. Below this zone, a very low gradient of 2.6 mK m^{-1} was measured to a depth of 130 m. Deeper down the gradient is more or less normal and varies from 10 to 19 mK m^{-1}.

Thermal conductivity was measured from core samples of the logged hole for the deeper part (> 360 m) and from the cores of the neighbouring hole, K-18, for the upper part. Hole K-18 is situated only 400 m to the northeast of hole K-1, and the sediment stratigraphy is similar to that in K-1. Therefore, we assume that the K-18 samples are representative of the hole K-1 section as well.

Heat flow density is not uniform with depth. The Bullard diagram (Fig. 10) shows several sections with different apparent heat flow values. The variations below 275 m may be due to the small water flows in a hole, or to biased conductivity values. Heat flow values at depths above 400 m in particular were based on thermal conductivities measured on K-18 samples. The mean heat flow density between 275 and 800 m is 42 mW m^{-2}.

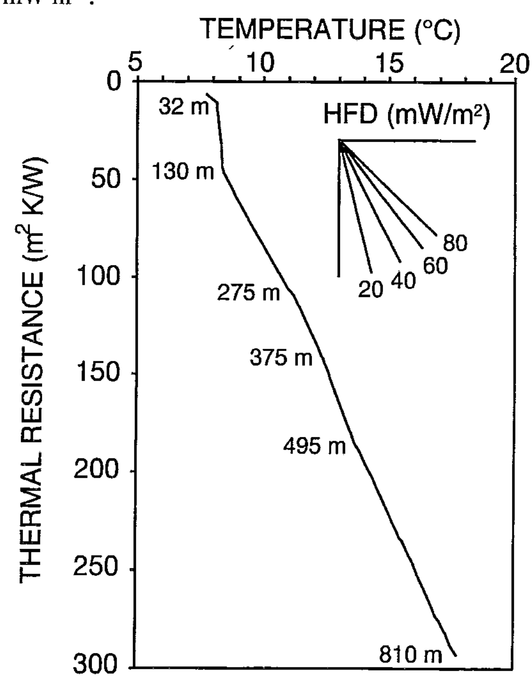

Fig. 10. Bullard diagram of the Kärdla borehole. Numbers next to curve refer to depths.

The variation does not, however, seem to follow the calculated palaeoclimatic variations (Fig. 4), at least those calculated with half-space models. The measured temperature log could of course be interpreted with a ground surface temperature history, but this would mean that the temperatures were at least 2°C lower 100-700 years ago than they are today. Such a long and cold period has not been reported.

Förby, Vormsi Island

The borehole (F369) is situated on the western shore of Vormsi, only about 20 m from the shoreline. This 306-m deep hole was drilled in 1987. The Precambrian basement was intersected for 42 m at depths of 264-306 m. On top of the basement there are Vendian, Cambrian and Ordovician sediments (Fig. 11).

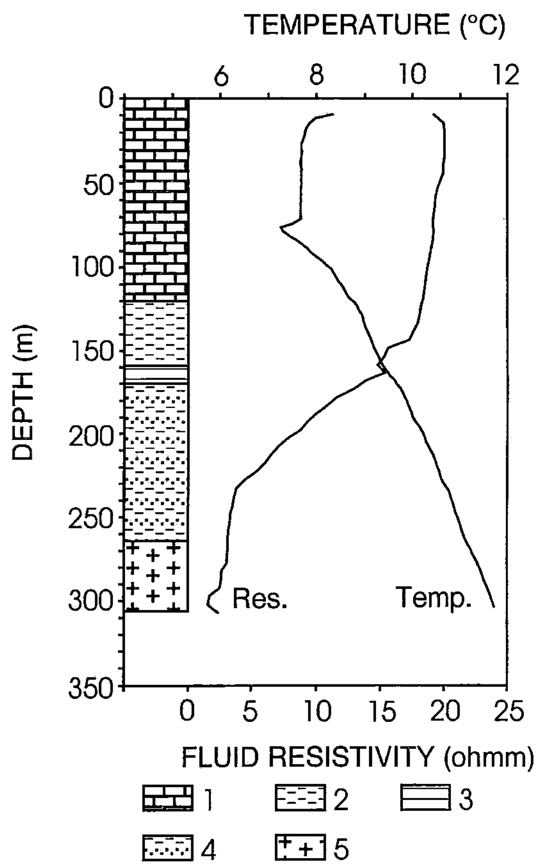

Fig. 11. Temperature and fluid resistivity at the Förby borehole. Cross-section: 1. limestone (O_2pkr-O_3mo), 2. siltstone (Є$_1$ts-O_1kl), 3. silty clay (Є$_1$lk), 4. intercalating beds of sandstone and siltstone (V-Є$_1$sr), 5. basement (PR_1).

In the upper part of the temperature log (< 130 m) the gradient is very disturbed and neither heat flow determinations nor palaeoclimatic corrections were made. The perturbations may be caused by water flowing in the hole. The low temperature peak at 65 m seems to be a product of cold water entering the hole. Above this depth, the temperature gradient is very low. It could also be attributed to water flowing in the hole, or to the existence of a horizontal heat flow component produced by the adjacent shoreline. If the sea bottom temperatures differ from the ground temperatures on shore, horizontal temperature gradients will be generated. These may produce very disturbed heat flow values (*Lewis* and *Wang,* 1992).

In the lower part of the hole (> 130 m), the temperature is less disturbed and gradient changes can be more or less attributed to thermal conductivity differences.

Thermal conductivity was measured on seven samples of Ordovician limestone and five samples of the Precambrian basement. In the middle part of the section, the data of *Urban et al.* (1991) on comparable stratigraphic units were used.

Below 130 m, both apparent and palaeoclimatically corrected heat flow densities increase with depth from 31 and 40 to 46 and 56 mW m^{-2}, respectively (Table 4). Average values are 38 and 46 mW m^{-2}.

Jõelähtme, northern Estonia

The Jõelähtme drillhole (8-1) penetrates Lower Ordovician, Cambrian and Vendian sediments down to the Precambrian basement (Fig. 12). The hole is 218 m deep and was drilled in 1991.

There do not seem to be any significant water flow effects on the temperature log. The increased gradient between 50 m and 130 m can be explained by the lower thermal conductivity of the Lontova clays (about 1.25 $Wm^{-1}K^{-1}$; *Urban et al.,* 1991) than of the Cambrian sands (about 2.5 W $m^{-1}K^{-1}$). In the sedimentary part below the clays, the temperature gradient does not vary very much (10-12 mK m^{-1}).

The ground surface temperature history at Jõelähtme is mainly dictated by glaciations and Holocene climatic changes. The influence of sea bottom temperatures is small, because the Baltic Ice Lake and the early Baltic Sea retreated from the hole site more than 8000 years ago (Fig. 8).

The apparent heat flow densities vary from 17 mW m^{-2} in Voronka siltstones to more than 40 mW m^{-2} in Lontova clays and Precambrian basement. This variation cannot be attributed to neither the regional groundwater flow discussed earlier or palaeoclimatic changes. The mean apparent and palaeoclimatically corrected heat flow densities are 34 mW m^{-2} and 40 mW m^{-2}, respectively.

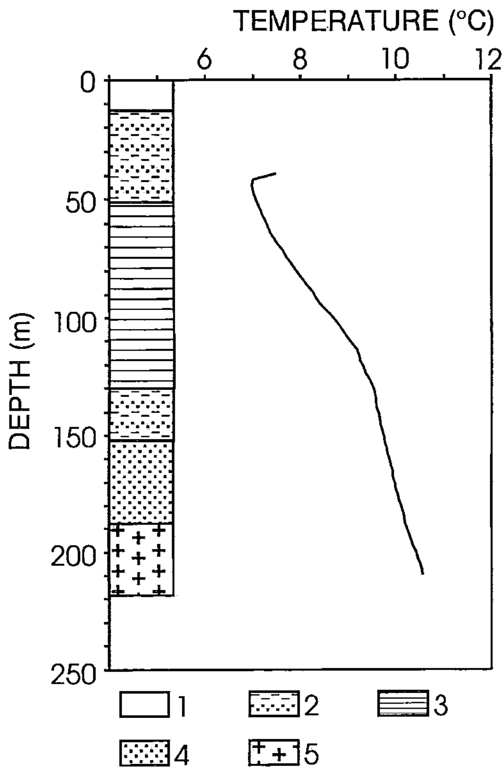

Fig. 12. Temperature at the Jõelähtme borehole. Cross-section: 1. overburden (Q), 2. intercalating beds of sandstone and siltstone ($Є_1$lk-O_1kl and V_2vr), 3. silty clay ($Є_1$ln), 4. sandstone (V_2gd), 5. basement (PR_1).

Saviranna, northern Estonia

The hole (F502) is 212 m deep and reaches 64 m to the Precambrian basement, which is represented by granites of the Naissaare intrusion (Fig. 13). The hole is in a cliff near the coast.

The shape of the temperature curve is very similar to the Jõelähtme temperature log. This is to be expected, because the distance between the two holes is less than 15 km and their stratigraphies are similar. The differences in palaeoclimatic models are connected with hole site elevations and the withdrawal of the sea, which took place about 5000 years ago.

The heat flow density variations are very similar to those in the Jõelähtme hole. The higher mean heat flow density values (40 and 48 mW m^{-2}) can be attributed to the heat production of granite, which is higher than that of the gneissic basement rocks at Jõelähtme.

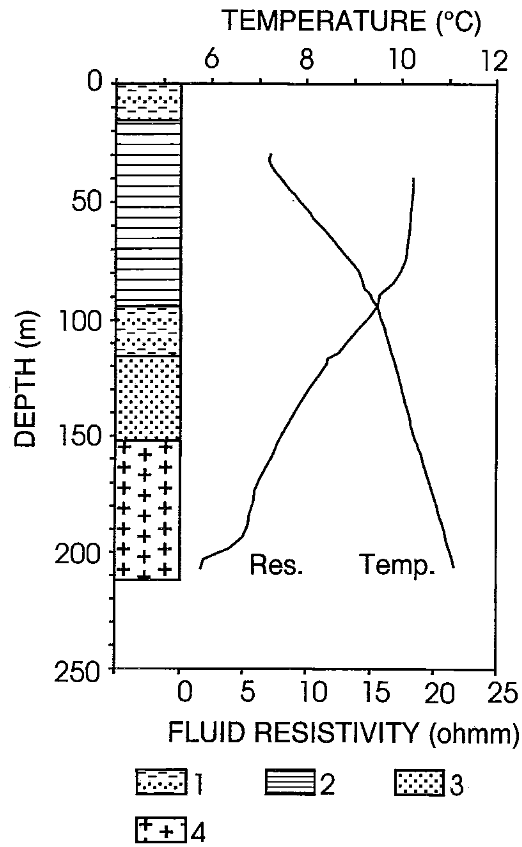

Fig. 13. Temperature and fluid resistivity at the Saviranna borehole. Cross-section: 1. intercalating beds of sandstone and siltstone ($Є_1$lk and V_2vr), 2. silty clay ($Є_1$ln), 3. sandstone (V_2gd), 4. basement (PR_1).

Loksa, northern Estonia

The 157-m deep hole (853) was drilled in 1989. In 1993 we only reached a depth of 132 m, where there was a blockage. The stratigraphy is represented by Quaternary, Cambrian and Vendian sediments and the Precambrian basement.

The temperature gradient changes (Fig. 14) correlate with the stratigraphic units, but the available thermal conductivity data do not support an explanation based exclusively on conductivity variations. However, the conductivities measured on other holes in the vicinity (*Urban et al.*, 1991) may not be representative.

As in previous holes, the heat flow density varies with depth. Below the Lontova clays flow values are low. The palaeoclimatic models do not explain this variation. The mean heat flow density values are 36 and 45 mW m^{-2}.

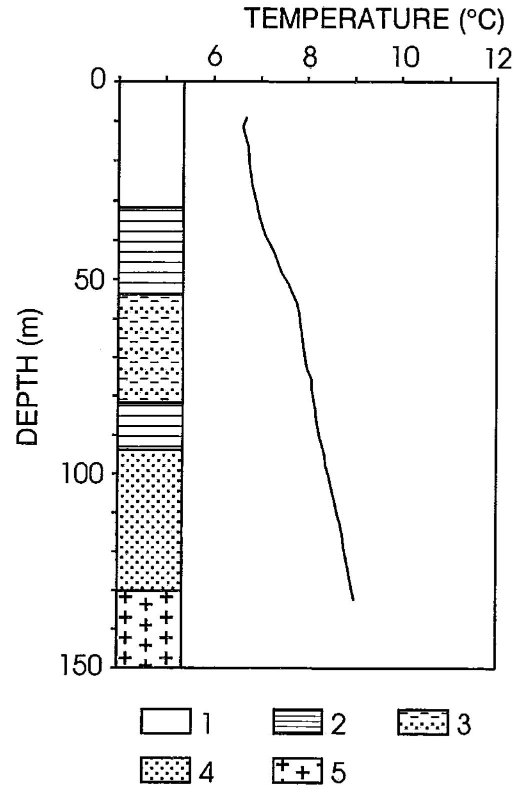

Fig. 14. Temperature at the Loksa borehole. Cross-section: 1. overburden (Q), 2. silty clay (Є$_1$ln and V$_2$kt), 3. intercalating beds of sandstone and siltstone (V$_2$vr), 4. sandstone (V$_2$gd), 5. basement (PR$_1$).

Rohuneeme, northern Estonia

The Rohuneeme hole (6-2) was drilled in 1991 only 25 m from the sea shore. The stratigraphy of the intersection is represented by Cambrian and Vendian sediments (Fig. 15).

On the temperature log the main gradient change at 65 m depth is related to the boundary of Lontova clays and Voronka siltstones, and can be at least partly attributed to the conductivity contrast between these stratigraphic units.

The apparent heat flow density varies from 15 to 24 mW m^{-2}. As discussed above for Förby the proximity of the shoreline may have created non-vertical heat flow components resulting in low apparent heat flow density at these very shallow depths.

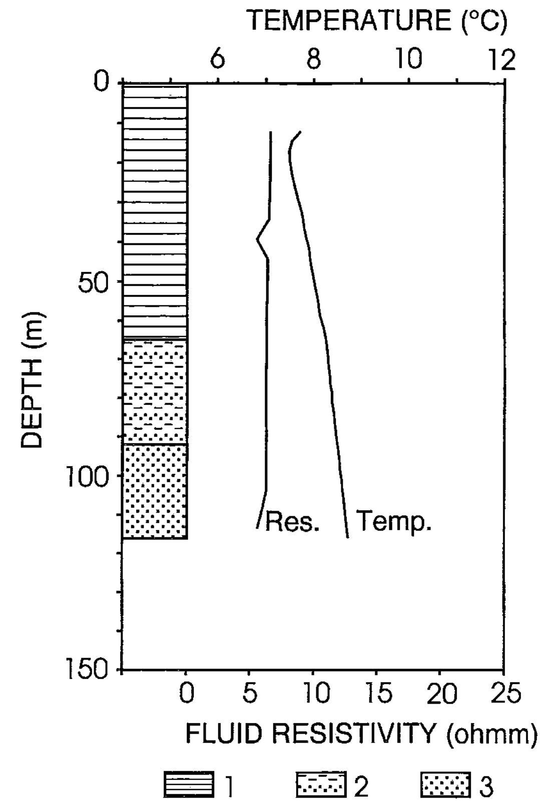

Fig. 15. Temperature and fluid resistivity at the Rohuneeme borehole. Cross-section: 1. silty clay ($Є_1$ln), 2. intercalating beds of sandstone and siltstone (V_2vr), 3. sandstone (V_2gd).