Proceedings World Geothermal Congress 2010
Bali, Indonesia, 25-29 April 2010

Calculation of the Maximal Effect of Deep Lakes on Temperature Measurements: Some Estimations Derived From the Geothermal Gradient Determination

Veniamin T. Balobaev[1], Izzy M. Kutasov[2] and Lev V. Eppelbaum[3]

[1]Permafrost Institute, Siberian Branch of the Russian Academy of Sciences, Yakutsk *677018*, Russia

[2]Pajarito Enterprises, 640 Alta Vista, Suite 124, Santa Fe, New Mexico *87505*, USA

[3]Dept. of Geophysics and Planetary Sciences, Raymond and Beverly Sackler Faculty of Exact Sciences, Tel Aviv University, Ramat Aviv *69978*, Tel Aviv, Israel

levap@post.tau.ac.il

Keywords: Geothermal gradient, reduced temperature, Laplace equation, lake.

ABSTRACT

Understanding climate change processes on the basis of geothermal observations in boreholes is an important and highly complex problem. Many non-climatic effects could cause changes in ground surface temperatures. In this study, the effects of deep lakes on the borehole temperature profiles observed within or in the vicinity of the lakes were investigated. A method based on the utilization of a Laplace equation with non-uniform boundary conditions is proposed. The proposed method makes it possible to estimate the maximum effect of deep lakes on the borehole temperature profiles. This method also allows one to estimate the accuracy of the geothermal gradient determination.

1. INTRODUCTION

At present, many efforts are made to determine the trends in ground surface temperature history (GSTH) from geothermal surveys. In this case, accurate subsurface temperature measurements are needed to solve this inverse problem of estimating the unknown time dependent ground surface temperature (GST). The variations of the GST during the long-term climate changes resulted in disturbance anomalies in the temperature field of geological formations. Thus, the GSTH data could be evaluated by analyzing the present precise temperature-depth profiles. In the past, the effects of surface temperature variations on the temperature fields of formations have been widely discussed in the literature (Cermak, 1971; Lachenbruch and Marshall, 1986; Beltrami et al., 1992; Shen and Beck, 1992; Mareschal and Beltrami, 1992; Wang, 1992; Shen et al., 1992, 1995; Kukkonen et al., 1994; Bodri and Cermak, 1995, 1997; Harris and Chapman, 1995; Huang et al., 1996; Guillou-Frottier et al., 1998; Huang and Pollack, 1998; Huang et al., 2000; Pollack and Huang, 2000; Majorowicz and Safanda, 2005; Eppelbaum et al., 2006; Hamza et al., 2007; Hopcroft et al., 2007; Rath and Mottaghy, 2007; Gonz´alez-Rouco et al., 2008; Kooi, 2008; Mann et al., 2009; Woodbury et al., 2009).

2. PREVIOUS INVESTIGATIONS: A SHORT REVIEW

Earlier, the forward calculation approach (FCA) was used for the analysis and interpretation of borehole temperatures in terms of the GSTH (Eppelbaum et al., 2006). Three groups were formed based on geographical proximity. Fifteen borehole temperature profiles from Europe (5), Asia (4) and North America (6) were selected (Huang and Pollack, 1998). The objective of this study was the estimation of the warming rates in the 20th century using the FCA method and comparing this to results obtained using the few parameter estimation (FPE) technique (Huang et al., 1996; Huang and Pollack, 1998). It is reasonable to assume that for closely spaced boreholes, the values of the warming rates obtained by the two inversion methods should vary within narrow limits. The results of inversions (FCA) showed that for boreholes in North America, the current warming rates vary in the 0.41-2.45 K/100a range. The wide warming rate range of 0.33-2.48 K/100a was determined for boreholes in Europe. Interesting results were obtained for four boreholes in Asia (China) (Eppelbaum et al., 2006). In this case, the warming rate has a relatively narrow range (1.16-1.59 K/100a.). The warming rate estimated by the FPE technique (Huang and Pollack, 1998) varied in wide ranges: 0.38-2.49 K/100a in North America); 0.21-3.75 K/100a in Europe, and 0.30-2.53 K/100a in Asia. Thus, we can conclude that for boreholes in North America and Europe, both approaches provide practically the same ranges of warming rates. For Asian boreholes, the FCA approach gives a more consistent (narrow) range of warming rates.

The results of temperature inversion using both techniques indicate that some non-climatic effects (vertical and horizontal water flows, steep topography, lakes, vertical variation in heat flow, lateral thermal conductivity contrasts, thermal conductivity anisotropy, deforestation, forest fires, mining, wetland drainage, agricultural development, urbanization, *etc*.) may have perturbed the borehole temperature profiles. The influence of these factors has been studied by many authors (e.g., Carslaw and Jaeger, 1959; Lachenbruch, 1965; Kappelmeyer and Haenel, 1974; Blackwell et al., 1980; Lewis and Wang, 1992; Majorowicz and Skinner, 1997; Powell et al., 1988; Guillou-Frottier et al., 1998; Lewis and Wang, 1998; Ichinose et al., 1999; Kohl, 1999; Safanda, 1999; Pollack and Huang, 2000; Cermak and Bodri, 2001; Gosselin and Mareschal, 2003; Gruber et al., 2004; Bodri and Cermak, 2005; Fan and Sailor, 2005; Mottaghy et al., 2005; Nitoiu and Beltrami, 2005; Allen et al., 2006; Taniguchi, 2006; Chouinard and Mareschal, 2007; Hamza et al., 2007; Safanda et al., 2007; Lind and Rind, 2008; Flanner, 2009; Yamano et al., 2009). Thee exact calculation of all of these factors is a complex physical and mathematical problem, which will be solved in the future using the method of successive approximations.

The temperature regime of sedimentary formations is influenced by many environmental and geological factors (local relief, sedimentation, erosion, lateral conductivity contrasts, underground water movement), past climate, and the heat flow from the Earth's interior (terrestrial heat flow). Most temperature surveys of this type are conducted in boreholes. In many cases, the drilling sites of boreholes are located within or outside of deep lakes (we employ the term “deep lake” to designate that the long term mean annual temperature of bottom sediments can be considered constant). The objective of this study was to evaluate to what extent the proximity of deep lakes can affect the temperature profiles of wellbores. In 1974, Balobayev and Shastkevich published the results of their analytical study, which can be used to determine the configuration of the steady state temperature field of formations beneath lakes of an arbitrary contour (Balobayev and Shastkevich, 1974). Taking into account that this publication is not easily accessible to researchers, a brief summation of its are presented below (Balobayev and Shastkevich, 1974). The authors assumed that the lake existed for an infinitely long period of time. In this case, the solution of the Laplace equation with nonuniform boundary conditions can be used to describe the steady state temperature field of formations beneath the lakes and estimate the maximum effect of lakes on borehole temperature profiles (due to the assumption that the lake existed for an infinitely long period of time).

3. APPROACHES USED IN CLIMATE RECONSTRUCTION; THEIR LIMITATIONS AND IMPEDIMENTS

We should note that all climate reconstruction methods are based on a one-dimensional heat conduction equation. It was assumed that a uniform boundary condition is applied on a plane surface, the formation is a laterally homogeneous medium, and the thermal properties depend only on depth. For this reason, any subsurface temperature variations arising from conditions that depart from this theoretical model have the potential to be incorrectly interpreted as climate change signatures (Pollack and Huang, 2000). Two examples are briefly presented to demonstrate the well selection procedures. In the study conducted by Guillou-Frottier et al. (1998), only 10 of 57 temperature profiles were selected for inversion of past ground surface temperatures. Nitoiu and Beltrami (2005) reported that only about 10 % of more than 10,000 borehole temperature logs worldwide (The International Heat Flow Commission global geothermal data set) are currently used for climate studies because a number of known non-climatic energy perturbations are superimposed on the climatic signal.

Therefore, extreme caution should be used in the selection of temperature-depth profiles for inferring ground surface temperature histories, according to the following guidelines:

1. The borehole should be sufficiently deep that the lower section of the temperature-depth profile allows a reliable determination of the geothermal gradient, presumably free of the effects of recent climate changes. Order of magnitude calculations indicate that surface temperature changes during the last few centuries penetrated to depths of ≈150 m,

2. The time elapsed between cessation of drilling and the temperature log reading should be at least an order of magnitude greater than the duration of drilling,

3. The temperature-depth profile should be free from the presence of any significant non-linear features in the bottom parts of the borehole, usually indicative of advection heat transfer by fluid movements, either in the surrounding formation or in the borehole itself,

4. The elevation changes at the site and in the vicinity of the borehole should be relatively small so that the topographic perturbation of the subsurface temperature field at shallow depths is not significant,

5. The lithologic sequences encountered in the borehole, should have relatively uniform thermal properties and should be sufficiently thick so that the gradient changes related to variations in thermal properties do not lead to systematic errors in the procedure employed for extracting the climate related signal.

Only 17 of 129 temperature logs were found to satisfy the above set of quality assurance conditions (Hamza et al., 2007). Corrections can be applied to correct borehole temperature profiles for the effect of topography (Lachenbruch, 1965; Blackwell et al., 1980; Safanda, 1994, 1999). However, this is rarely done because the amplitude of the climatic signals is often smaller than the uncertainty on these corrections (Chouinard and Mareschal, 2007). Safanda et al (2007) presented interesting results of repeated temperature logs from Czech, Slovenian and Portuguese borehole climate observatories within a time span of 8-20 years. The repeated logs revealed subsurface warming in all the boreholes amounting to 0.2-0.6 ^{o}C below 20 m depths. A warming rate of 0.05 ^{o}C/yr. was estimated at the Czech observatory located in a park within the campus of the Geophysical Institute in Prague. This warming rate is two times greater than the simulated value (using the surface air temperature as a forcing function). It was assumed that the subsurface temperature at the station was influenced by new structures built within the campus of the Geophysical Institute within the last 10-20 years and/or by other components of infrastructure built 40-50 years ago. The authors (Safanda et al., 2007) conducted a quantitative analysis of these effects by numerically solving the heat conduction equation in a 3D geothermal model of the borehole site. It was discovered that the mentioned anthropogenic structures influenced the temperature in the borehole quite strongly.

Nitoiu and Beltrami (2005) attempted to correct borehole temperature data for the effects of deforestation. The authors simulated the ground surface temperature changes following deforestation by using a combined power exponential function describing the organic matter decay and recovery of the forest floor after a clear-cut (Covington, 1981). The presented examples demonstrate that application of this correction could allow the incorporation of many borehole data into the borehole climatology database (Nitoiu and Beltrami, 2005).

4. APPLIED EQUATIONS

Let us assume that the well site is located within or outside of a deep lake. As mentioned above, the long-term mean annual temperature of bottom sediments was considered to be constant (this problem was shortly outlined for the first time by Balobaev et al. (2008)). The bottom of the lake had a vertical coordinate of z = 0. The temperature regime of geological formations in this area (within and outside of the lake) is subjected to the thermal influence of the lake. The extent of this influence depends mainly on the lake’s dimensions, the current depth, the distance from the lake, and the difference between the long term mean annual temperature of bottom sediments and the long term mean annual temperature of surrounding lake formations (at z = 0). It was assumed that the lake existed for an infinitely long period of time. The designations shown below were used.

ρ, φ, z are cylindrical coordinates (ρ is the distance from the z axis, φ describes the angle from the positive xz-plane to the point, and z is depth). T_{is} is the long term mean annual temperature of bottom sediments, and T_{ot} is the long term mean annual temperature of surrounding lake formations at z = 0. Firstly, consider a lake of an arbitrary contour, as shown in Figure 1.

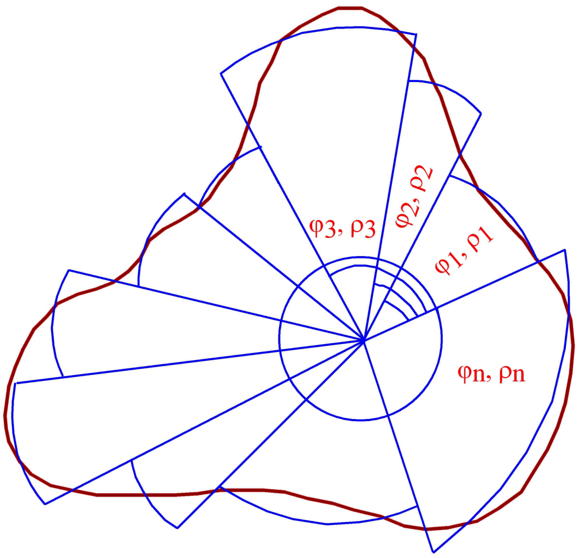

Figure 1: Division of an arbitrary contour lake into sectors (after Balobayev and Shastkevich, 1974).

The Laplace equation for the semi-infinite solid area is

$$\frac{\partial^2 T}{\partial \rho^2}+\frac{1}{r}\frac{\partial T}{\partial \rho}+\frac{1}{\rho^2}\frac{\partial^2 T}{\partial \varphi^2}+\frac{\partial^2 T}{\partial z^2}=0 \qquad (1)$$

where Γ is the regional geothermal gradient (outside the lake area).

The solution of the Laplace equation is possible by division of a lake of an arbitrary contour into sectors. However, the solution is expressed through a complex Poisson integral, and fairly elaborate and time-consuming computations are needed (Balobayev and Shastkevich, 1974). Let ρ_{max} be the maximum value of the set ρ_1, ρ_2, ρ_n. By introducing a safety factor (the maximum thermal effect of the lake on temperature profiles), we can assume that the lake has a circular shape with a radius R_i = ρ_{max}. Now the Laplace equation and boundary conditions are

$$\frac{\partial^2 T}{\partial \rho^2}+\frac{1}{r}\frac{\partial T}{\partial \rho}+\frac{\partial^2 T}{\partial z^2}=0 \qquad (2)$$

$$\left\{\begin{array}{ll} T(\rho, z=0)=T_{is} & \rho \le R_i \\ T(\rho, \varphi z=0)=T_{ot} & \rho > R_i \\ T(\rho=\infty, z)=T_{ot}+\Gamma z & \end{array}\right\}$$

The solution to Equation 2 is (Balobayev and Shastkevich, 1974) given in the following Equations:

$$T(\rho, z)=T_{ot}+\Gamma z+M(T_{is}-T_{ot}) \qquad (3)$$

$$\left.\begin{array}{l} M(\rho, z)= \\ 1-A_1[A_2\Pi(\alpha_1^2,k)+A_3\Pi(\alpha_2^2,k)] \end{array}\right\} \qquad (4)$$

$$\left.\begin{array}{l} A_1=\dfrac{z}{\pi\sqrt{z^2+(R_i+\rho)^2}}; \\ A_2=\dfrac{\sqrt{z^2+\rho^2}-R_i}{\sqrt{z^2+\rho^2}+\rho} \end{array}\right\} \qquad (5)$$

$$\left.\begin{array}{l} A_3=\dfrac{\sqrt{z^2+\rho^2}+R_i}{\sqrt{z^2+\rho^2}-\rho}; \\ \alpha_1^2=\dfrac{2\rho}{r+\rho}; \\ \alpha_2^2=-\dfrac{2\rho}{r-\rho} \end{array}\right\} \qquad 6)$$

$$\left.\begin{array}{l} r^2=\rho^2+z^2; \\ k^2=\dfrac{4\rho R_i}{z^2+(R_i+\rho)^2} \end{array}\right\} \qquad (7)$$

where $\Pi(\alpha_1^2,k)$ and $\Pi(\alpha_2^2,k)$ are the complete elliptical integrals of the third order (Abramowitz and Stegun, 1965).

For the center of the island (ρ = 0):

$$M(z) = 1 - \frac{z}{\sqrt{z^2+R_i^2}} \qquad (8)$$

The temperature gradient for a well drilled at the center of the lake (ρ = 0) can be determined from Equations 3 and 8:

$$\left.\begin{array}{l} \Gamma^* = \dfrac{\partial T(0,z)}{\partial z} = \\ \Gamma-(T_{is}-T_{ot})\left[\dfrac{R_i^2}{(z^2+R_i^2)^{3/2}}\right] \end{array}\right\} \qquad (9)$$

$$\left.\begin{array}{l} \Delta\Gamma = \Gamma^*-\Gamma = \\ -(T_{is}-T_{ot})\left[\dfrac{R_i^2}{(z^2+R_i^2)^{3/2}}\right] \end{array}\right\} \qquad (10)$$

For distances ρ > 0, Equations 11 and 12 were obtained from Equation 3:

$$\Gamma^*=\frac{dT(\rho,z)}{dz}=\Gamma+(T_{is}-T_{ot})\frac{dM}{dz} \qquad (11)$$

or

$$\Delta\Gamma=\Gamma^*-\Gamma=(T_{is}-T_{ot})\frac{dM}{dz} \qquad (12)$$

Introducing the reduced temperatures, $T_R(\rho, z)$ from Equation 3, we can write

$$T_R(\rho, z) = T(\rho, z) - T_{ot} - \Gamma z = M(T_{is} - T_{ot}) \quad (13)$$

4. EXAMPLE CALCULATION

Consider a 30 m deep lake with a radius of R_i = 100 m and T_{is} = 10 °C. The regional geothermal gradient is Γ = 0.0300 °C/m, and T_{ot} = 20 °C. The drilling site of a 3000 m wellbore is located 150 m from the center of the lake.

Table 1. The functions dM(z,0)/dz and M(z, ρ)

Z, m	-dM/dz, 10^{-6}1/m, $\rho = 0$	The function $M(z, \rho)$ — Distance from the center of the lake (ρ), m						
		0	50	100	150	200	300	400
20	9429	.8039	.7612	.3828	.0525	.0168	.0042	.0017
50	7155	.5528	.4937	.2815	.0950	.0370	.0100	.0091
100	3536	.2929	.2606	.1787	.0985	.0518	.0174	.0076
150	1707	.1679	.1538	.1189	.0805	.0512	.0212	.0101
200	894.4	.1056	.0991	.0827	.0628	.0450	.0222	.0116
250	512.3	.0715	.0683	.0598	.0488	.0379	.0215	.0122
300	316.2	.0513	.0496	.0449	.0384	.0315	.0198	.0122
400	142.7	.0299	.0292	.0275	.0249	.0219	.0159	.0111
500	75.43	.0194	.0119	.0184	.0172	.0157	.0125	.0095
600	44.43	.0136	.0135	.0131	.0125	.0117	.0099	.0080
700	28.28	.0100	.0100	.0098	.0094	.0090	.0079	.0066
800	19.08	.0077	.0077	.0076	.0073	.0071	.0064	.0056
900	13.47	.0061	.0061	.0060	.0059	.0057	.0052	.0047
1000	9.852	.0050	.0049	.0049	.0048	.0047	.0044	.0040
1100	7.421	.0041	.0041	.0041	.0040	.0039	.0037	.0034
1200	5.727	.0034	.0034	.0034	.0034	.0033	.0032	.0030
1400	3.617	.0025	.0025	.0025	.0025	.0025	.0024	.0023
1600	2.427	.0019	.0019	.0019	.0019	.0019	.0018	.0018
1800	1.707	.0015	.0015	.0015	.0015	.0015	.0015	.0014
2000	1.245	.0012	.0012	.0012	.0012	.0012	.0012	.0012
2200	.9362	.0010	.0010	.0010	.0010	.0010	.0010	.0010
2400	.7215	.0009	.0009	.0009	.0009	.0009	.0008	.0008
2600	.5677	.0007	.0007	.0007	.0007	.0007	.0007	.0007
2800	.4547	.0006	.0006	.0006	.0006	.0006	.0006	.0006
3000	.3698	.0006	.0006	.0006	.0006	.0006	.0006	.0005

It is desired to know the magnitudes of the formation temperature perturbations (expressed through the reduced temperatures) caused by the lake. The results of calculations from Equations 4 and 8 are presented in Table 1 and Figure 2. It is noted that because the bottom of the lake has a coordinate z = 0, the actual depth is z^* = z + 30 m. In this case, T_{is} – T_{ot} = –10 °C, and the lake has a cooling effect on the temperature profiles. The values of $T_R(\rho, z)$ decrease with depth and practically can be neglected for radial distances of 550-600 m from the center of the lake, as can be seen in Figure 2.

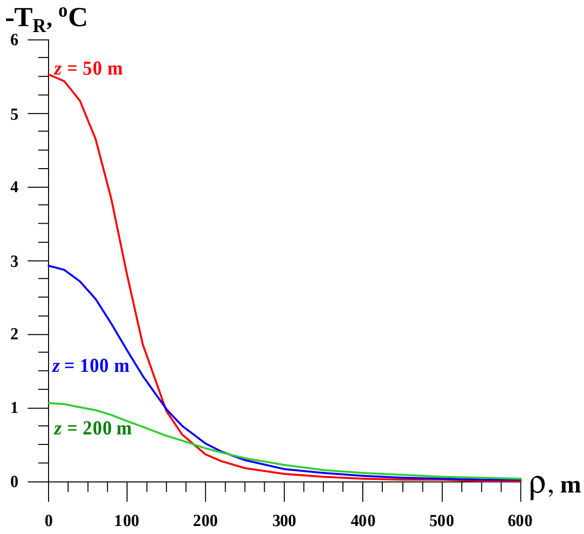

Figure 2: The reduced temperatures versus radial distance for three depths

No consider the determination of the maximum values of $T_R(\rho, z)$ and $\Delta\Gamma(\rho, z)$ for the 300-500 m section of the wellbore. The values of $T_R(\rho, z)$ can be estimated directly from Equation 13 and Table 1:

$$T_R(150{,}300) = -10^{o}\mathrm{C}\cdot 0.0384 = -0.384\,^{o}\mathrm{C}\,,$$

$$T_R(150{,}500) = -10^{o}\mathrm{C}\cdot 0.0172 = -0.172\,^{o}\mathrm{C}\,.$$

To determine the values of $\Delta\Gamma(\rho, z)$ for this case, the approximation of the function M by a quadratic polynomial is suggested. This polynomial is shown in Equation 14, and the results of the approximation are displayed in Table 2.

Table 2. The approximation of the function M by a quadratic polynomial (Equation 14)

z, m	M	M*	(M- M*)/M ·100, %
300	0.03842	0.03822	0.51
320	0.03505	0.03510	-0.15
340	0.03207	0.03221	-0.45
360	0.02941	0.02955	-0.49
380	0.02705	0.02712	-0.27
400	0.02494	0.02492	0.07
420	0.02304	0.02295	0.38
440	0.02135	0.02121	0.65
460	0.01982	0.01970	0.60
480	0.01844	0.01842	0.12
500	0.01720	0.01737	-0.97
The average squared deviation = 0.50%			

$$M^* = a_0 + a_1 z + a_2 z^2 \quad (14)$$

$$\frac{dM^*}{dz} = a_1 + 2a_2 z \quad (15)$$

$$\left\{ \begin{aligned} a_o &= 0.1126 \\ a_1 &= -0.3341 \cdot 10^{-3} \\ a_2 &= 0.2872 \cdot 10^{-6} \end{aligned} \right\}.$$

Then, the values of $\Delta\Gamma(\rho, z)$ can be determined from Equations 12 and 15:

$$\Delta\Gamma(150{,}300) = 1{,}6174\cdot10^{-3}\ ^{o}\mathrm{C/m};$$

$$\Delta\Gamma(150{,}500) = 0.4684\cdot10^{-3}\ ^{o}\mathrm{C/m}$$

The values of $T_R(\rho, z)$ and $\Delta\Gamma(\rho, z)$ are presented in Figure 3 for the 300-500m section of the well. In our example, the value of the regional geothermal gradient (Γ) is 0.03 °C/m.

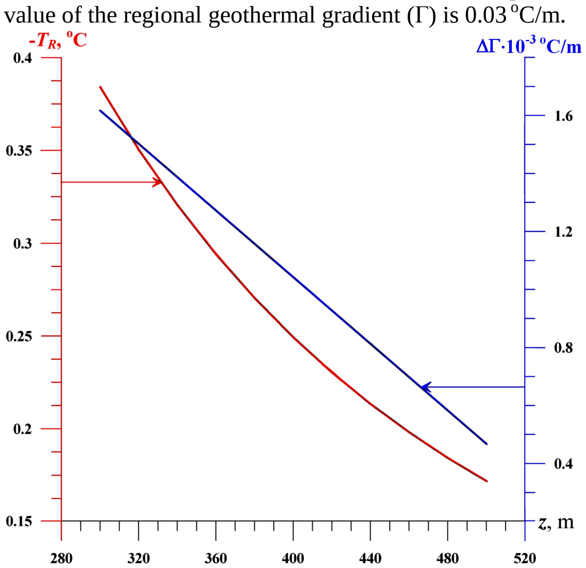

Figure 3. The reduced temperature and ΔΓ with depth for ρ = 150 m

Thus, the accuracy of the determined geothermal gradient is somewhere between

$$(0.4684\cdot10^{-3}/0.03)\cdot100\% = 1.56\%$$

and

$$(1.6174\cdot10^{-3}/0.03)\cdot100\% = 5.39\%\ .$$

For a wellbore located at the center of the lake, the values of T_R and $\Delta\Gamma$ are maximal. From Table 1 and Equations 12 and 13:.

$$T_R(0{,}300) = -10^{o}\mathrm{C}\cdot0.0513 = -0.513\ ^{o}\mathrm{C}\ ,$$

$$T_R(0{,}500) = -10^{o}\mathrm{C}\cdot0.01942 = -0.194\ ^{o}\mathrm{C}\ ,$$

$$\Delta\Gamma(0{,}300) = 3.16\cdot10^{-3}\ ^{o}\mathrm{C/m}\ ,$$

$$\Delta\Gamma(0{,}500) = 0.754\cdot10^{-3}\ ^{o}\mathrm{C/m}\ .$$

A commercially available software called Maple 7 (Waterloo Maple, 2001) was utilized to compute the function $M(\rho, z)$.

5. CONCLUSION

It is shown that borehole paleoclimate investigations are complicated by many disturbing factors, and the exact calculation of these effects is a complex physical-mathematical problem. The proposed method allows the estimation of the maximum effect of deep lakes on the borehole temperature profiles observed within and outside of the lakes. It was assumed that the lake existed for an infinitely long period of time. In this case, the solution of the Laplace equation with nonuniform boundary conditions can be used to describe the steady state temperature field of geological formations beneath the lakes and estimate the maximum effect (due to the assumption that the lake existed for an infinitely long period of time) of lakes on the borehole temperature profiles. The example calculations presented testify to what extent the proximity of a deep lake affects the borehole temperature profiles. The accuracy of the determination of the geothermal gradient was also estimated.

REFERENCES

Abramowitz, M., and Stegun, I.: *Handbook of Mathematical Functions*, Dover Publications Inc., N.Y., (1965).

Allen, M.R., Gillett, N.P., Kettleborough, J.A., Hegerl, G., Schnur, R., Stott, P.A., Boer, G., Covey, C., Delworth, T.L., Jones, G.S., Mitchell, J.F.B., and Barnett, T.P.: Quantifying anthropogenic influence on recent near-surface temperature change, *Survey of Geophysics*, **27**, (2006), 491-544.

Balobaev, V.T., Kutasov, I.M., and Eppelbaum, L.V.: Borehole paleoclimatology – the effect of deep lakes and "heat islands" on temperature profiles, *Climate of the Past*, No.4, (2008), 1-18.

Balobayev, V.T., and Shastkevich, Yu. G.: The estimation of the talik zones configuration and the steady temperature field of rocks beneath the lakes of arbitrary contour, *In: Lakes of the Siberia Cryolithozone* (in Russian), Nauka Publ., Novosibirsk, (1974), 116-127.

Beltrami, H., Jessop, A.M., and Mareschal, J.-C.: Ground temperature histories in eastern and central Canada from geothermal measurements: Evidence of climate change, *Palaeogeogr. Palaeoclimatol. Palaeoecol*. (*Global Planet. Change Sect.*), **98**, (1992), 167-183.

Blackwell, D.D., Steele, J.L., and Brott, C.A.: The terrain effect on terrestrial heat flow, *Journal of Geophysical Research*, **85**(B9), (1980), 4757-4772.

Bodri, L., and Cermak, V.: Climate change of the last millennium from borehole temperatures: results from the Czech Republic – Part I, *Global Planetary Change*, **11**, (1995), 111-125.

Bodri, L., and Cermak, V.: Climate change of the last two millennia inferred from borehole temperatures: results from the Czech Republic – Part II, *Global Planetary Change*, **14**, (1997), 163-173.

Bodri, L., and Cermak, V.: Borehole temperatures, climate change and the pre-observational surface air temperature mean: Allowance for hydraulic conditions, *Global Planetary Change*, **45**, (2005), 265-276.

Carslaw, H.S., and Jaeger, J.C.: *Conduction of Heat in Solids*, Oxford University Press, N.Y., (1959).

Cermak, V.: Underground temperature and inferred climatic temperature of the past millennium, *Palaeogeogr. Palaeoclimatol. Palaeoecol*., **10**, (1971), 1-19.

Cermak, V., and Bodri, L.: Climate reconstruction from subsurface temperatures demonstrated on example of Cuba, *Physics of the Earth and Planetary Interiors*, **126**, (2001), 295-310.

Chouinard, C., and Mareschal, J.-C.: Selection of borehole temperature depth profiles for regional climate reconstructions, *Climate of the Past*, **3**, (2007), 297-313.

Covington, W.W.: Changes in the forest floor organic matter and nutrient content following clear cutting in northern hardwoods, *Ecology*, **62**, (1981), 41-48.

Eppelbaum, L.V., Kutasov, I.M., and Barak, G.: Ground surface temperatures histories inferred from 15 boreholes temperature profiles: comparison of two approaches, *Earth Sciences Research Journal*, **10**, No. 1, (2006), 25-34.

Fan, H., and Sailor, D.J.: Modeling the impacts of anthropogenic heating on the urban climate of Philadelphia: A comparison of implementations in two PBL schemes, *Atmospheric Environment*, **39**, (2005), 73–84.

Flanner, M.G.: Integrating anthropogenic heat flux with global climate models, *Geophysical Research Letters*, **36**, L02801, (2009), doi: 10.1029/2008GL036465.

Gonz´alez-Rouco, J.F., Beltrami, H., Zorita, E., and Stevens, M.B.: Borehole climatology: a discussion based on contributions from climate modeling, *Climate of the Past Discussions*, **4**, (2008), 1-80.

Gruber, S., King, L., Kohl, T., Herz, T., Haeberli, W., and Hoelzle, M.: Interpretation of geothermal profiles perturbed by topography: the Alpine permafrost boreholes at Stockohorn plateau, Switzerland, *Permafrost and Periglacial Processes*, **15**, (2004), 349-357.

Gosselin, C., and Mareschal, J.-C.: Variations in ground surface temperature histories in the Thompson Belt, Manitoba, Canada: environment and climate changes, *Global and Planetary Change*, **39**, (2003), 271-284.

Guillou-Frottier, L., Mareschal, J.-C., and Musset, J.: Ground surface temperature history in central Canada inferred from 10 selected borehole temperature profiles, *Journal of Geophysical Research*, **103**, No. B4, (1998), 7385-7397.

Hamza, V.M., Cavalcanti, A.S.B., and Benyosef, L.C.C.: Surface thermal perturbations of the recent past at low latitudes – inferences based on borehole temperature data from Eastern Brazil, *Climate of the Past*, **3**, (2007), 513-526.

Harris, R.N., and Chapman, S.D.: Climate change on the Colorado Plateau of eastern Utah inferred from borehole temperatures, *Journal of Geophysical Research*, **100**, (1995), 6367-6381.

Hopcroft, P.O., Gallagher, K., and Pain, C.C.: Inference of past climate from borehole temperature data using Bayesian Reversible Jump Markov chain Monte Carlo, *Geophysical Journal Intern*., **171**, (2007), 1430-1439.

Huang, S., Shen, P.Y., and Pollack, H.N.: Deriving century – long trends of surface temperature change from borehole temperatures, *Geophysical Research Letters*, **23**, (1996), 257-260.

Huang, S., Pollack, H.N., and Shen, P.Y.: Temperature trends over past five centuries reconstructed from borehole temperatures, *Nature*, **403**, (2000), 756-758.

Huang, S., and Pollack, H.N.: Global Borehole Temperature Database for Climate Reconstruction, IGBP PAGES/World Data Center-A for Paleoclimatology Data Contribution Series #1998-044. *NOAA/NGDC Paleoclimatology Program*, Boulder CO, USA, (1998). (see also http://www.geo.lsa.umich.edu/~climate).

Ichinose, T., Shimodozono, K., and Hanaki, K.: Impact of anthropogenic heat on urban climate in Tokyo, *Atmospheric Environment*, **33**, (1999), 3897–3909.

Kappelmeyer, O., and Haenel, R.: *Geothermic with Special Reference to Application*, Gebruder Borntrager, Berlin, (1974).

Kohl, T.: Transient thermal effects below complex topographies, *Tectonophysics*, **306**(3-4), (1999), 311-324.

Kooi, H.: Spatial variability in subsurface warming over the last three decades; insight from repeated borehole temperature measurements in The Netherlands, *Earth and Planetary Science Letters*, **270**, (2008), 86-94.

Kukkonen, I.T., Cermak, V., and Safanda J.: Subsurface temperature–depth profiles, anomalies due to climatic ground surface temperature changes or groundwater flow effects, *Global Planetary Change*, **9**, (1994), 221-232.

Lachenbruch, A.H.: Rapid Estimation of the Topographic Disturbance to Superficial Thermal Gradients, *Review of Geophysics*, **6**, (1965), 365-400.

Lachenbruch, A.H., and Marshall, B.V.: Changing climate: Geothermal evidence from permafrost in the Alaskan Arctic, *Science*, **234**, (1986), 689-696.

Lewis, T.J., and Wang, K.: Influence of terrain on bedrock temperatures, *Palaeogeogr. Palaeoclimatol. Palaeoecol*., **98**, (1992), 87-100.

Lewis, T.J., and Wang, K.: Geothermal evidence for deforestation induced warming: implications for the climatic impact of land development, *Geophysical Research Letters*, **25**, (1998), 535-538.

Lind, J.L., and Rind, D.H., How natural and anthropogenic influences alter global and regional surface temperatures: 1889 to 2006, *Geophysical Research Letters*, **35**, L18701, (2008), doi: 10.1029/2008GL034864

Majorowicz, J.A., and Skinner, W.P.: Potential causes of differences between ground and surface air temperature warming across different ecozones in Alberta, Canada, *Global Planetary Change*, **15**, (1997), 79-91.

Majorowicz, J., and Safanda, J.: Measured versus simulated transients of temperature logs – a test of borehole climatology, *Journal of Geophysics and Engineering*, **2**, (2005), 291-298.

Mann, M.E., Schmidt, G.A., Miller, S.K., and LeGrande, A.M.: Potential biases in inferring Holocene temperature trends from long-term borehole information, *Geophysical Research Letters*, **36**, L05708, (2009), doi: 10.1029/2008GL036354

Mareschal, J.-C., and Beltrami, H.: Evidence for recent warming from perturbed geothermal gradients: examples from Eastern Canada, *Climate Dynamics*, **6**, No. 3-4, (1992), 135-143.

Mottaghy, D., Schellschmidt, R., Popov, Y.A., Clauser, C., Kukkonen, L.T., Nover, G., Milanovsky, S., and Romushkevich, R.A.: New heat flow data from immediate vicinity of the Kola super-deep borehole: Vertical variation in heat flow confirmed and attributed to advection, *Tectonophysics*, **401**, (2005), 119-142.

Nitoiu, D., and Beltrami, H.: Subsurface thermal effects of land use changes, *Journal of Geophysical Research*, **110**, (2005), F01005.

Pollack, H.N., and Huang, S.: Climate reconstruction from subsurface temperatures, *Annual Review of Earth Planetary Sciences*, **28**, (2000), 339-365.

Powell, W.G., Chapman, D.S., Balling, N., and Beck, A.E.: Continental Heat-Flow Density, In: Haenel, R., Rybach, L., and Stegena, L. (Eds.) *Handbook of Terrestrial Heat-Flow Density Determination*, Kluwer Academic Publishers, Dordrecht / Boston / London, (1988), 167-222.

Rath, V., and Mottaghy, D.: Smooth inversion for ground surface temperature histories: estimating the optimum regularization parameter by generalized cross-validation, *Geophysical Journal Intern*., **171**, (2007), 1440–1448.

Safanda, J.: Effects of topography and climatic changes on the temperature in borehole GFU-1, Prague, *Tectonophysics*, **239**, (1994), 187-197.

Safanda, J.: Ground surface temperature as a function of slope angle and slope orientation and its effect on the subsurface temperature field, *Tectonophysics*, **306** (3-4), (1999), 367-376.

Safanda, J., Rajver, D., Correia, A., and Dedecek, P.: Repeated temperature logs from Czech, Slovenian and Portuguese borehole climate observatories, *Climate of the Past*, **3**, (2007), 453-462.

Shen, P.Y., and Beck, A.E.: Paleoclimate change and heat flow density inferred from temperature data in the Superior Province of the Canadian Shield, *Palaeogeogr*. *Palaeoclimatol. Palaeoecol*. (*Global Planetary Change Sect*.), **98**, (1992), 143-165.

Shen, P.Y., Wang, K., Beltrami, H., and Mareschal, J.-C.: A comparative study of inverse methods for estimating climatic history from borehole temperature data, *Global Planetary Change*, **98**, (1992), 113-127.

Shen, P.Y., Pollack, H.N., Huang, S., and Wang, K.: Effects of subsurface heterogeneity on the inference of climate change from borehole temperature data: model studies and field examples from Canada, *Journal of Geophysical Research*, **100**, (1995), 6383-6396.

Taniguchi, M.: Anthropogenic effects on subsurface temperature in Bangkok, *Climate of the Past Discussions*, **2**, (2006), 832-846.

Wang, K.: Estimation of ground surface temperatures from borehole temperature data, *Journal of Geophysical Research*, **97**, (1992), 2095-2106.

Waterloo Maple: *Maple 7 Learning Guide*, Waterloo Maple Inc., Waterloo, Canada, (2001).

Woodbury, A.D., Bhuiyan, A.K.M.H., Hanesiak, J., and Akinremi, O.O.: Observations of northern latitude ground-surface and surface-air temperatures, *Geophysical Research Letters*, **36**, L07703, (2009), doi: 10.1029/2009GL037400.

Yamano, M., Goto, S., Miyakoshi, A., Hamamoto, H., Lubis, R.F., Monyrath, V., and Taniguchi, M.: Reconstruction of the thermal environment evolution in urban areas from underground temperature distribution, *Science of the total Environment*, **407**, (2009), 3120-3128.