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Advanced analysis of thermal data observed in subsurface wells unmasks the ancient climate

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Conventional methods of studying the ancient climate history are associated with statistical processing of accomplished meteorological data. These investigations have focused attention on meteorological records of air temperature, which can provide information on the only last 100–200 years. Number of the records is absolutely insufficient and their areal coverage is limited, some oldest meteorological stations may have been affected by local warming connected with urban and industrial growth. At the same time significant climate changes are accompanied by the corresponding variations in the Earth's surface (soil) temperature. This effect is based on the known physical law that temperature waves at the surface propagate downward into the subsurface with an amplitude attenuation and time delay increasing with depth. Earth's temperature profiles, measured by precise temperature logging $T(z)$ in boreholes to depth of about 80-300 meters, have a "memory" on what has happened on the surface during approximately several last centuries.

Knowledge of the past climate in archaeology is necessary not only for tracing some ancient events and more deep understanding some historical facts, but also for estimation of past harvests, analysis of some physical conditions of different constructions built in the past, and in many other fields (Eppelbaum, 2010; Eppelbaum et al., 2010).

The first attempts to recover the past ground surface temperature history (*GSTH*) from measured $T(z)$ profiles date back to the mid-1960s, however only after Lachenbruch et al. (1988) pointed out that the magnitude and timing of the ground surface warming in Alaska is consistent with models of the recent warming, the method became popular (Cermak et al., 1996).

Let us assume that t_x years ago from now the ground surface temperature started to increase (warming) or reduce (cooling). Prior to this moment the subsurface temperature is:

$$T_a(z, t = 0) = T_{0a} + \Gamma z, \tag{1}$$

where T_{0a} is the mean ground surface temperature at the moment of time $t = 0$ years; z is the vertical depth and Γ is the geothermal gradient. It is also assumed that the host medium is homogeneous with constant thermal properties. Now the current ($t = t_x$) subsurface temperature is (in case of warming):

$$T_c(z, t = t_x) = T_{0c} + f(z), \tag{2}$$

where T_{0c} is the current (at the time (date) of temperature logging) mean ground surface temperature; and $f(z)$ is a function of depth that could be obtained from the field data. In some cases the value of T_{0c} can be obtained by extrapolation of the function T_c to $z = 0$. However, in most cases, the value T_{0c} can be estimated by trial and error method: Assuming an interval of values for T_{0c}, calculating for each T_{0c} value of the temperature profiles $T_c$for various models of change in the ground surface temperature (*GST*) with time and, finally, finding a best match between calculated and field measured T_c profiles. In our study we found that a quadratic regression can be utilized to estimate the value of $T_{0c} = a_0$ (Kutasov et al., 2000):

$$T_c(z, t = t_x) = a_0 + a_1 z + a_2 z^2, \tag{3}$$

where a_0, a_1, and a_2 are the coefficients.

We will consider four different models (Eppelbaum et al., 2006). Apparently each of these models is more suitable (applicable) under concrete physical-geological conditions.

In the first model we assumed that t_{xC} years ago the *GST*value suddenly changed from T_0 to T_{0c}. The current temperature anomaly (the reduced temperature) is

$$T_R(z) = T_{0c} + f(z) - T_0 - \Gamma z \tag{4}$$

and the solution is

$$T_{RC} = T_R = \Delta T_0 \Phi^*(x) \left(\frac{z}{2\sqrt{at}} \right) , t = t_{xC}, \tag{5}$$

$$\Delta T_0 = T_{0c} - T_0, \tag{6}$$

where a is the thermal diffusivity of formation and $\Phi^*(x)$ is the complementary error function.

In the second model we assumed that t_{xL} years ago the *GST* value started gradually to change from T_0 to T_{0c}. We assumed that *GST* is a linear function of time and

$$T_{0c} = T_0 + \alpha_L t_{xL}, \tag{7}$$

where α_L is some coefficient.

The solution is

$$T_{RL} = T_R = \alpha_L t \left\{ \left(1 + \frac{z^2}{2at} \right) \Phi^* \left(\frac{z}{2\sqrt{at}} \right) - \frac{z}{\sqrt{\pi a t}} \exp \left(-\frac{z^2}{4at} \right) \right\}, \qquad t = t_{xL}, \tag{8}$$

In the third model we also assumed that t_{xS} years ago the *GST* value started gradually to change from T_0 to T_{0c}. We assumed that *GST* is a square root function of time and

$$T_{0c} = T_0 + \alpha_S \sqrt{t_{xS}}, \tag{9}$$

where α_S is a coefficient.

The solution is

$$T_{RS} = T_R = \alpha_S \sqrt{t} \left\{ \exp \left(-\frac{z^2}{4at} \right) - \frac{z\sqrt{\pi}}{2\sqrt{at}} \Phi^* \left(\frac{z}{2\sqrt{at}} \right) \right\}, \qquad t = t_{xS}. \tag{10}$$

And, finally, in the fourth model we assumed that the *GST* value exponentially increases with time and

$$T_{0c} = T_0 \exp(\alpha_E t_{xE}), \tag{11}$$

where α_E is a coefficient.

The solution is ($T_{RE} = T_R$)

$$T_{RE} = \frac{1}{2} \exp(\alpha_E t) \left[\exp \left(-z\sqrt{\frac{\alpha_E}{a}} \right) \Phi^* \left(\frac{z}{2\sqrt{at}} - \sqrt{\alpha_E t} \right) + \exp \left(z\sqrt{\frac{\alpha_E}{a}} \right) \Phi^* \left(\frac{z}{2\sqrt{at}} + \sqrt{\alpha_E t} \right) \right], \quad t = t_{xE} \tag{12}$$

At the same time not all boreholes are suitable for the thermal data processing for unmasking climate of the past. For deep wells ($>$ 120–300 m) the drilling process, due to lengthy period of drilling fluid circulation, greatly alters the temperature of formation immediately surrounding the well. As a result, the determination of formation temperature (with a specified absolute accuracy) at any depth requires a lengthy period of shut-in time. Therefore it is important to determine how long it takes before the error caused by mud circulation is small compared to the change arising from the change in surface temperature. We suggested two techniques, (Slider's method (e.g., Kutasov and Eppelbaum, 2007) and utilization of the γ–function (Kutasov, 1999)) enabling us to estimate the rate of temperature decline and the difference between the formation and shut-in temperatures (Kutasov and Eppelbaum, 2013).

The developed methodologies were successfully applied on the thermal borehole data from the northern America, Europe and Asia (Eppelbaum et al., 2014).

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