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PERMAFROST AT THE ICE BASE OF RECENT PLEISTOCENE GLACIATIONS–INFERENCES FROM BOREHOLE TEMPERATURE PROFILES

Abstract: Paleo-temperature reconstruction from precise depth (>2.0 km) well temperature logs can offer information on whether the bed of an ice sheet was frozen. Inversion or upward extrapolation of the >2-km-deep geothermal profile is the only method by which temperature evolution at the base of long-disappeared ice sheets such as the Laurentide and Fennoscandian in the northern part of the Northern Hemisphere in North America and Europe can be inferred. It is obvious from the results from well temperature profiles that there were spatial variations in temperature at the base of the ice sheets during glaciations. This comes as no surprise, since modern-day measurements of temperature profiles through the ice of existing glaciers show a similarly large variability. Present bedrock temperatures measured beneath the central part of the Yukon Rusty glacier are near 0°C to –2°C while Greenland ice sheet base temperatures are –8 and –13°C. In case of very low paleo-temperatures derived from the interpretation of temperature profiles in the areas presently outside the current extent of glacial ice it can be shown that low temperature conditions under glacial ice could facilitate the existence of moderate (some 100–200 m) to thick (0.5 km–1 km) permafrost conditions. It is speculated here that, in many cases, paleo-glacial cold base ice could have existed right on top of paleo-permafrost in sediments just below. Such ice-bonded permafrost may have been frozen to glacial ice above, forming pillars which fixed glacial ice to permafrost below, thus limiting ice movement in such places and resulting in the –extended persistence of permafrost.

Key words: Paleoclimate at last glaciation, permafrost under glacial, paleo-permafrost, temperature profiles

Climate effect on geothermal gradient

Borehole temperature logs are influenced by the surface temperature, histories (glacial – interglacial, Post-glacial – Holocene, recent millennial and recent industrial warming – century/decadal time scale) etc. (see: Čermak 1971; Lachenbruch and Marshall 1986; Lachenbruch et al. 1988; Lachenbruch 1994 for review of the subject). Other than the influence of the recent warming period (RWP) preceded by the little ice age (LIA) there had been a warmer climate established long before "RWP". The Holocene Optimum (6–8 ka ago) typically had a warmer climate than RWP. Before that, Northern and Central Europe, Asia and northern North America were exposed to a harsh periglacial climate at the forefront of the Laurentide and Fennoscandian ice sheets. Deep borehole temperature logs done in equilibrium conditions reveal that positive temperature gradients gradually increase with depth to the values undisturbed by the glacial cycles (usually the signal can be seen down to 2 km; (see: Majorowicz and Wybraniec 2011)).

Deep down, to a depth of 2 km, perturbation of the heat flow Reflect warming since recent glaciations. Perturbations observed down at 0.1 km–0.3 km are mainly related to more recent "industrial" age climatic warming and recent global warming is observed also from inversions of temperature logs (Majorowicz 2010) on large regional, continental scale. This warming influences heat flow Q:

$$Q= k*grad(T) \qquad (1)$$

and thermal gradient grad (T):

$$grad (T) = \partial T/\partial z \qquad (2)$$

at fixed thermal conductivity of rock and or ice k, where T is temperature (see Beardsmore and Cull 2001 on review of heat flow measurement methods and Carlslaw and Jaeger 1959 for the theory of heat transfer). This phenomenon was originally observed by Hotchkiss and Ingersoll (1934). This perturbation affect geothermal gradient and heat flow data

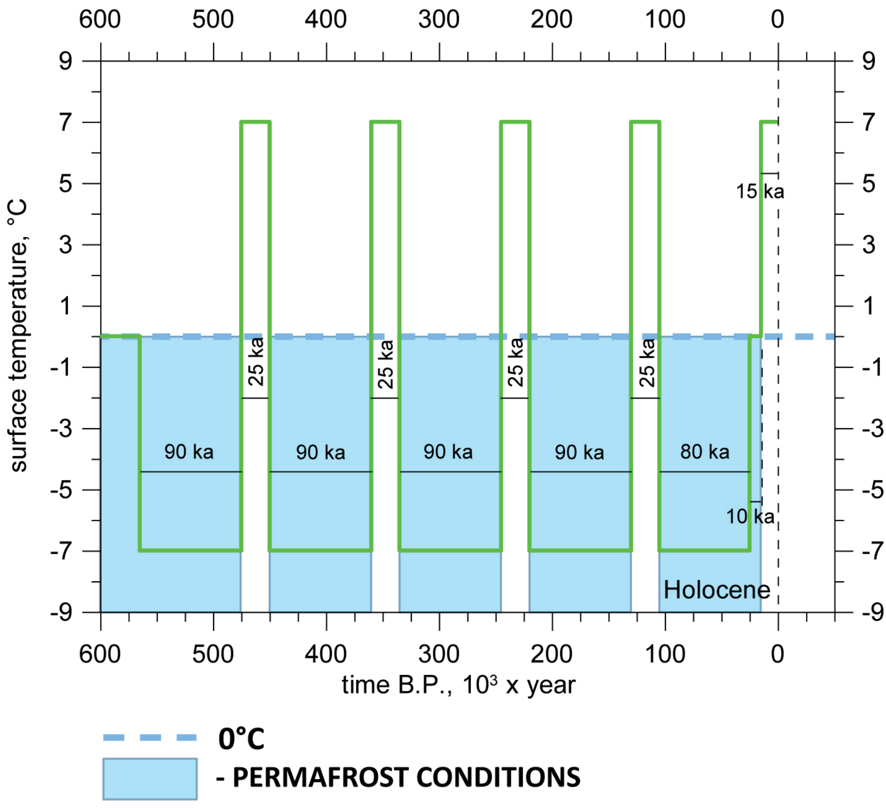

Fig. 1. Paleo-surface temperature history used in computation of the synthetic T-z profiles shown in Fig. 2. Temperature before the onset of the five glacial cycles considered was assumed to be 0°C (but the mean temperature of one glacial cycle is lower, –4°C, being the time-weighted average of –7 and +7°C). The area was covered by an ice sheet during the last glacial maximum (LGM) 25–15 ka ago, for which period we assumed 0°C. The glacial mean was therefore higher there, –6.2°C, instead of –7°C in an ice-free area (acc. to Majorowicz and Safanda 2008, modified)

collected in the northern latitudes of North America, Northern Europe and in Northern Asia (Jessop 1991; Kukkonen 1993; Kukkonen and Safanda 1996; Kukkonen et al. 2003; Kukkonen and Joeleht 2003; Gosnold et al. 2005; Demezhko et al. 2007; Chouinard and Mareschal 2009;Gosnold et al. 2011; Majorowicz and Wybraniec 2011). Jessop (1971) developed a correction for such perturbations of 1–10% in Canada and the northern

U.S. based on the assumption that the base of the ice in the Laurentian glaciation was fairly warm. Jessop (1971) assumed the base of the glacier was near the pressure point of melting (–1ºC). This appears to be confirmed in many cases in the deep well temperature profiles from the eastern part of the Canadian shield. Recent investigations (Chouinard and Mareschal 2009) using data from deep boreholes confirmed the temperature at the base of the ice sheet to be near the pressure point of melting. Very low heat flow in this area (<40 mWm^{-2}), has been reported by Blackwell and Richards

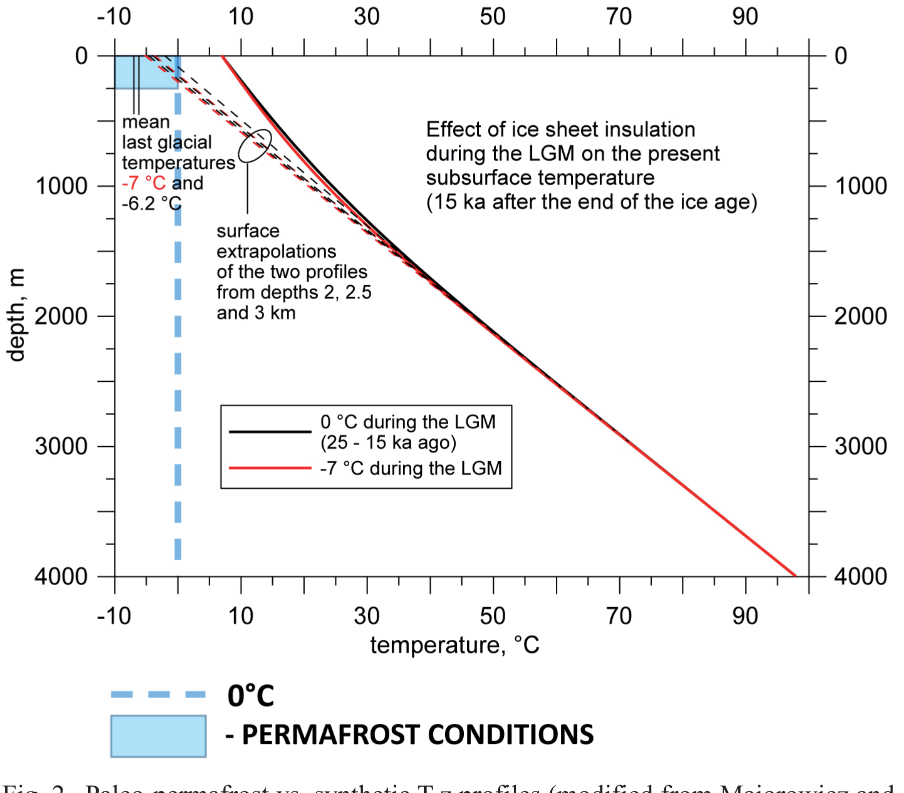

Fig. 2. Paleo-permafrost vs. synthetic T-z profiles (modified from Majorowicz and Safanda 2008) based on the ground surface temperature history shown for a homogeneous model with conductivity of 2 W/(m K), diffusivity of 0.9E–6 m^{2}s^{-1}, typical tectonically stabilized continental heat flow of 50 mWm^{-2}. They correspond to the time 15 ka after the end of the last ice age and differ due to different surface temperatures, –7°C or 0°C, during the LGM 25–15 ka ago. Also shown are upward steady-state extrapolations of the two profiles from depths of 2, 2.5 and 3 km, respectively. These all point to sub-zero temperature conditions and some 150 m-thick permafrost for the model assumed

(2004) and Davis and Davis (2010), who explain it as being accounted for by very low mantle heat flow and low <1mWm^{-3} radiogenic heat generation in the upper part of the granitic crust (Mareschal et al. 1999, 2000, 2004; Rolandon et al. 2002, 2003; Perry et al. 2004, 2006, 2010; Levy et al. 2010).

This assumption may be a simplification, as much higher variability of the ice base temperature in the Late Pleistocene has been reported in Northern America (Gosnold et al. 2011; Majorowicz et al. 2012), Eurasia (Safanda et al. 2004; Demezhko et al. 2007; Majorowicz and Wybraniec 2011). Present ice-base temperatures in cold base ice in Greenland have been measured at –8.4ºC for the log for GRIP and –13.2ºC for Dye-3 (Dahl-Jensen et al. 1998 and Demezhko et al. 2007).

Therefore, this perturbation may be much stronger than in Jessop (1971) at his assumed warm base glacial. The amplitude of such a temperature change will depend upon temperatures at the base of the glacier and the present ambient temperature. Also, the geothermal gradient in presently non-glaciated areas will be affected to a depth of approximately 1.5–2 km (Kukkonen 1993; Safanda et al. 2004; Majorowicz and Wybraniec 2011).

The amount of warming can vary by latitude and longitude based on the paleo-temperature and the present-day temperature (Demezhko et al. 2007).

Annually, the Northern Hemisphere has an average ambient temperature range of –20 ÷ +15ºC indicating a warming of up to 16ºC since the LGM (Last Glacial Maximum). It seems that the temperatures near the edge of the ice sheet may have actually been cool (from –4 to –15ºC (Safanda and Rajver 2001; Safanda et al. 2004; Demezhko et al. 2007)). This extent of warming has been confirmed by other supporting evidence, with pollen analyses in Manitoba, Canada indicating a warming of nearly 15ºC (Ritchie 1983), and climate modeling also indicated a warming of 15–20ºC (Schneider von Deimling et al. 2006).

It is possible that there has been a high warming amplitude (6ºC–15ºC) since the retreat of the Laurentide and Winsconsin ice sheets, and the lower end of this temperature range has been detected by the functional space inversion of precise measurements in the 2.4-km-deep well into Precambrian granites in north eastern Alberta, Canada (Majorowicz et al. 2012) and in Poland (Majorowicz et al. 2007; Majorowicz and Safanda 2008; Mottaghy et al. 2010).

Boreholes, temperature profiles and their role in predicting past surface temperatures

The surface climatic forcing due to a persistent increase in ground surface temperature (GST) after an event like the end of the glacial age (Fig. 1), or after the end of the little ice age (LIA), will diffuse by conduction into the subsurface and thereby impose transient "climate" signals on the steady-state geothermal gradient (Fig. 2). The interpretation of the GST history (GSTH) as related to a surface annual temperature history (SATH) is usually based on two basic assumptions: (1) The GST systematically couples with the SAT, (2) there is a constant offset between GST and SAT at each well site. These assumptions are a simplification, however, as such coupling can be disrupted during ground freezing and snow cover. However, it has been shown that, the long decadal century scale climatic changesand GST changes generally track SAT changes.

Time changes of the subsurface temperature can be solved in a one-dimensional geothermal model using the transient heat conduction equation:

$$C_v\ \partial T/\partial t = \partial[k(\partial T/\partial z)]/\partial z + A \qquad (3)$$

where T is the temperature, k is the thermal conductivity, C_v is the volumetric heat capacity, A is the rate of heat generation per unit volume, z is depth, and t is time.

The solution of the above equation for the conductive heat transfer is one with no freezing/thawing (Carlslaw and Jaeger 1959) for time t = t* and the initial condition t(z,t) = 0

is:

$$T(z) = \Delta\ 2^n\ \Gamma\ (0.5\ n + 1)\ i^n\ \mathrm{erfc}\ (z/(4\alpha\ t^*)^{-0.5}) \qquad (4)$$

where in erfc (b) is the nth integral of the error function of b, Γ (b) is the gamma function of argument b, Δ is surface temperature increase, α is diffusivity and t is time. The above solution gives the ground temperature after a warming event of duration t* with surface temperature change D. The model can be changed with adjustment of n. These functions control the model of GST change as follows; a step increase for n = 0, a parabolic increase for n = 1, and a linear change for n = 2.

Extrapolation of the linear portion of the thermal profile, which is controlled by deep heat flow and thermal conductivity k to the surface z_o yields the intercept temperature $T(z_o)$. The deviation of the measured temperature profile T(z) from the extrapolated linear profile, results in the temperature anomaly Δ (z, t) which in the simplest interpretation represents the response of the ground to a recent rise of mean annual SAT from a previous long-term value, or recent cooling in the case of a negative anomaly. The combination of subsequent warming, or cooling, events complicates the disturbing signal with depth even for simple models of the glacial-interglacial cycle (Fig. 1). The most important part of that history will be the latest transition from the cycle of the latest glaciations into the interglacial we are presently enjoying. This change in surface temperature will influence the shape of the temperature-depth profile. Here we show the temperature curve in the case of homogenous rock of constant thermal conductivity k (Fig. 2). Upward extrapolation of the >2-km-deep part of the thermal profile with constant geothermal gradient will give us an indication of the paleo-surface temperature at an age which can be calculated from the model (Majorowicz and Safanda 2008) (see example of such prediction Fig. 2). This temperature significantly determines the thickness of the permafrost below (Fig. 3) for a given constant regional thermal gradient.

In reality, the effects of freezing/thawing are taken into account in more complex conditions. The consumption or release of latent heat, *L,* in water/ice (334 kJ kg^{-1}) and GH (430 kJ kg^{-1}) accompanying either thawing or freezing plays a part. Additionally, the effects of interstitial ice and gas hydrates have been accounted for by Majorowicz et al. (2008), among others, using apparent heat capacity according to Carslaw and Jaeger (1959), when the volumetric heat capacity is increased in the depth sections of the model where the thawing and freezing occurs, i.e. where the temperature is within the thawing range between the temperature of solidus, and liquidus at the actual simulation time step (see: Safanda et al. 2004 for detail of the modeling in case of freezing/thawing).

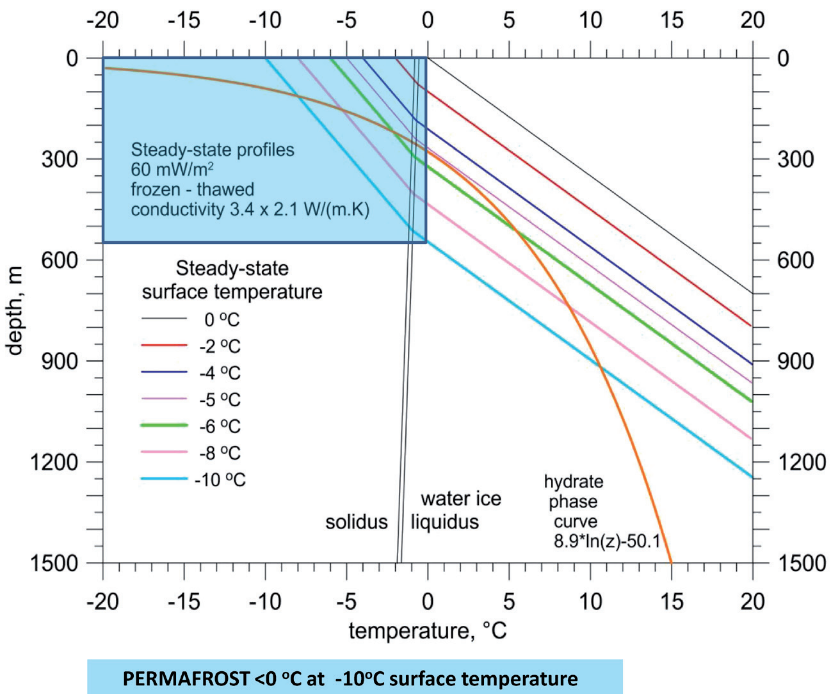

Fig. 3. Maximum thickness of paleo-permafrost for temperature-depth profiles constrained by 0- (–10) °C surface conditions. Theoretical steady-state profiles for the geothermal model in which heat flow is assumed 60mWm^{-2} (typical for thermal results in the Polish deep sedimentary basin in places like the Toruń-1 well location (Majorowicz and Safanda 2009; discussed further). Conductivity of the frozen and thawed rock is assumed to be 3.4 and 2.1Wm^{-1} K-1 respectively, modified from Majorowicz et al. (2012). The depth of permafrost for the –10°C sub-glacial ice surface will reach a thickness of 550 m (box area). The stability of hydrate (CH_4) is also shown to illustrate that warming and pressure after glacial retreat can result in destabilizing greenhouse gas CH_4 from gas hydrate within permafrost and/or below (more likely to occur) in sedimentary rocks under paleo-glacial

Temperature at the glacial ice base–paleo and present conditions

Paleo-temperature reconstruction from precise depth (>2.0 km) well temperature logs can offer information on whether the bed of an ice sheet was frozen. Inversion or upward extrapolation of the >2-km-deep geothermal profile is the only method that can be used to infer temperature evolution at the base of long-disappeared ice sheets such as the Laurentide (Rolandone et al. 2003) or Fennoscandian ice sheets (Kukkonen 1996; Safanda and Rajver 2001; Kukkonen and Jõeleht 2003; Kukkonen and Safanda 2003; Safanda et al. 2004; Demezhko et al. 2007). The sensitivity of downhole temperature-depth to past ground surface temperature variability decreases markedly as we go back in time. Therefore, short-period changes related to the glacial–interglacial cycle may safely be approximated by appropriate simple function chosen according to Majorowicz (2010), (Figs 4 and 7).

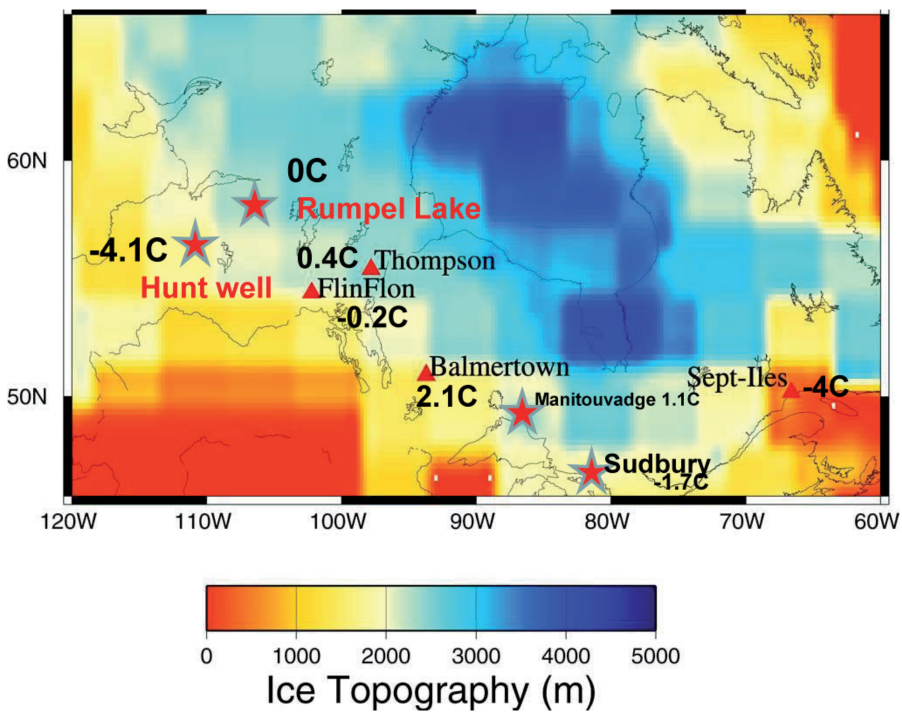

Fig. 4. Temperatures at the base of the Laurentide ice sheet in Canada. Minimum temperatures derived from precise well temperature logs (AOC-GRANITE-Hunt well and Rumpel lake wells (Majorowicz et al. 2012) vs. previously published results (Chouinard and Mareschal 2009; Perry et al. 2009)). Ice topography acc. to Peltier (2002). Note that the lowest paleo-temperatures are for three sites (Hunt well, Sudbury and Sept-Iles), where ice topography acc. to Peltier (2002) model is smaller than in high near melting point indication sites

In eastern Canada, where Laurentide glaciation was present in the past, all of the sites analyzed by Chouinard and Mareschal (2009) showed basal ice temperature to be near 0°C, except possibly Sept-Iles, Quebec and Sudbury, Ontario, where they were lower. According to Chouinard and Mareschal (2009), temperatures might have been lower, a few (2–4) degrees below 0°C, at Sept-Iles (Fig. 4). The two sites at Sudbury logged and reported by Perry et al. (2009) show the temperature at the base of the ice sheet was slightly below the melting point of ice (–1.5°C). For all the other sites in Ontario and Manitoba, the temperature at the base of the glacier did not fall much below 0°C during the LGM. It is likely that the difference observed in Eastern-Central Canada in surface heat flux had no influence on basal glacial temperatures, which are controlled by glacier dynamics, such as high velocity basal flows in the ice.

This is different in western Canada, in the Canadian sedimentary basin west of the Canadian Shield. Recently logged temperature-depth in the 2.35-km-deep Granite-Hunt well in northern Alberta AOC, (see location in Fig. 4 and Fig. 5), (Majorowicz et al. 2012) and the data inversion that was carried out, suggest a basal ice temperature significantly below the melting point of ice (–4.1°C), (Fig. 6). This is comparable to a result from Sept-Iles in the very eastern edge of Laurentide (see Fig. 4 and Fig. 5 for the locations). It is obvious from the above review of the results from several well-temperature profiles that the temperature at the base of the Laurentide ice sheet was spatially variable.

The deep-well temperature profile data also allowed an interpretation of the temperature at the base of ice in the Fennoscandian paleo glaciations. The lowest paleotemperature values were found in the area of north eastern Poland in the Suwałki massif where low heat flow / heat generating (40 mWm^{-2} and 0.1–0.2 mWm^{-3} respectively acc. to Majorowicz 1984) anorthosite crystalline rock is overlaid by some 0.8 km of high porosity sediments. In the area of Udryn and Krzemianka in the Suwałki massif, studies of temperature in several >2-km-deep boreholes (drilled into crystalline rocks in the course of ore prospecting) identified a decrease of temperature with depth in the uppermost 350–450 m (Fig. 7), with the minimum temperature being detected at a depth of around 400 m. This was linked to a postglacial event because the geothermal gradient is decreasing to the 400 m minimum and then increasing below. This was a unique situation not found anywhere else in sedimentary basins, and this phenomenon was first reported by Majorowicz (1976, 1984). Later, Michalski (1985) detected the presence of groundwater of cryogenic origin in the same area. These two discoveries lead Safanda et al. (2004) to speculate that the shape of the temperature curve must be related to a process of paleo-permafrost melting.

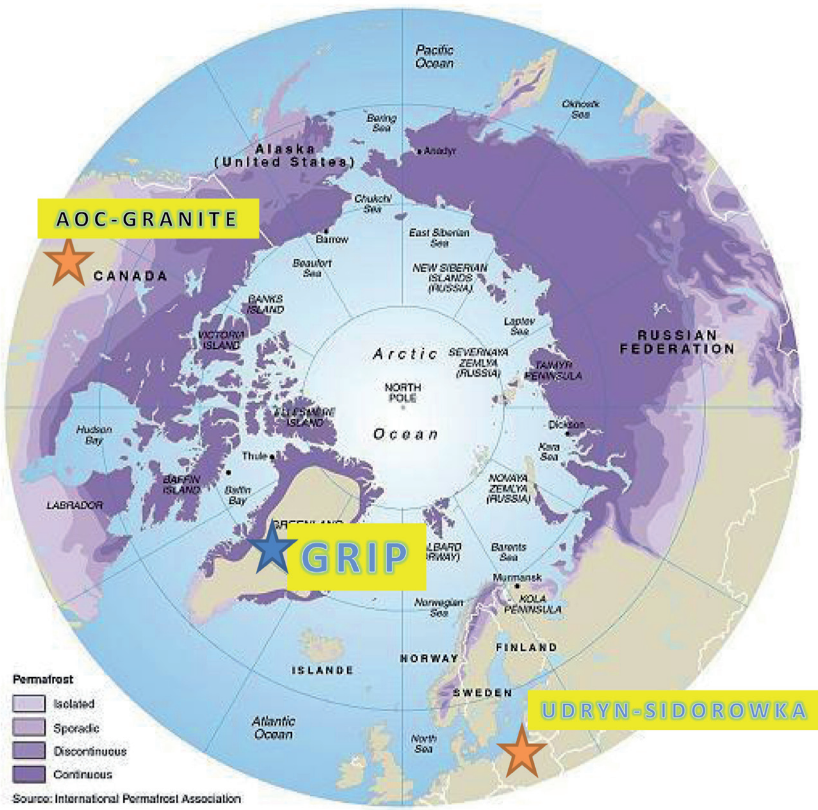

Fig. 5. Location of wells with >2-km-deep temperature profiles indicating thick paleo-permafrost below glacial ice (AOC-GRANITE (Hunt well) in northern Canada and UDRYN-SIDOROWKA in north-eastern Poland) and permafrost under present glacial ice bed in GRIP (also Dye-3 well) in Greenland against the map of present permafrost distribution from International Permafrost Association (http://ipa.arcticportal.org/resources/what-is-permafrost)

Safanda et al. (2004) presented a model of permafrost decay in the area of the Suwałki anorthosite intrusion. This modeling indicates the final decay of ice in permafrost layer at about (6.7 ka ago). The initial result of temperature logs from the Udryn wells has been confirmed by subsequent high-precision temperature profiles in thermally stable conditions in the Szypliszki well near the Udryn wells (Safanda et al. 2004), (see Fig. 5 in this paper for the location). The hypothesis about the possibility of permafrost preservation in the central part of the Suwałki anorthosite massif based on some additional ‘colder’ temperature profiles near Udryn led to the drilling

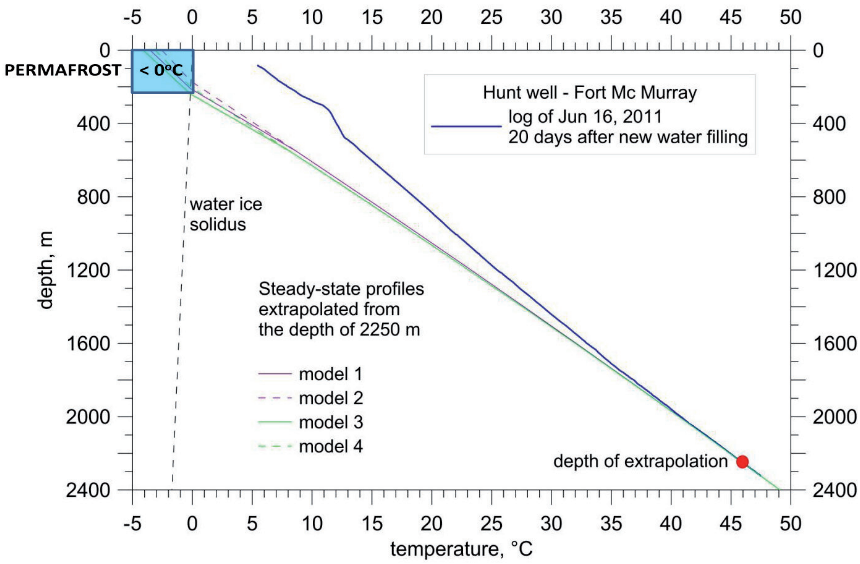

Fig. 6. Paleo-permafrost under the paleo-Laurentide ice sheet (retreated some 13 ka years from the site) inferred from precise equilibrium condition temperature depth temperature continuous logs in wells AOC GRANITE 7-32-89-10, Well (Alberta, Canada; 111. 272°W and 56.385°N) (log 6 of June 2011 in equilibrium condition; done under Helmholtz-Alberta Innitiative Theme 4 (HAI4) funding, Majorowicz et al. 2012). The four profiles correspond to 4 models of conductivity and heat production. The upward extrapolation starts at the depth of 2250 m, in the middle of the interval 2200–2300 m for which the gradient and heat flow was determined by linear fit of the measured curve. The heat flow is 56.7 mWm^{-2}. The two alternative values of heat production in the granite, 3E-6 or 4E-6 Wm^{-3} for the granite. The two alternative conductivity values considered for sediments, 2.4 and 2.7 W/(m K), were increased by a factor of 1.62 in the assumed permafrost zone close to the surface at temperatures below the water ice solidus ($T_{solidus}$ = –0.000715* depth, for zero salinity of groundwater). The frozen/melted conductivity ratio is assumed (3.4/2.1 = 1.62). Paleo-surface temperature was near –4°C at the time of the end of glaciation. Present temperature is near 5°C. Predicted thickness of paleo-permafrost during Laurentide glaciations is some 0.2 km

of a new well (Udryn Pig-1) which discovered temperatures near 0°C at 357 m (Szewczyk and Nawrocki 2011). 0°C is the upper temperature limit of permafrost as defined by the International Permafrost Association and it defines the top of preserved permafrost in Udryn Pig-1 established at a depth of 357 m. The other low paleo-temperatures < –1°C linked to LGM were reported for two other deep wells in Poland (Toruń 1; Majorowicz and Safanda 2009) and Czeszewo Ig-1 (Fig. 8). Temperatures well below zero during the Pleistocene were reported by Demezhko et al. (2007) for many

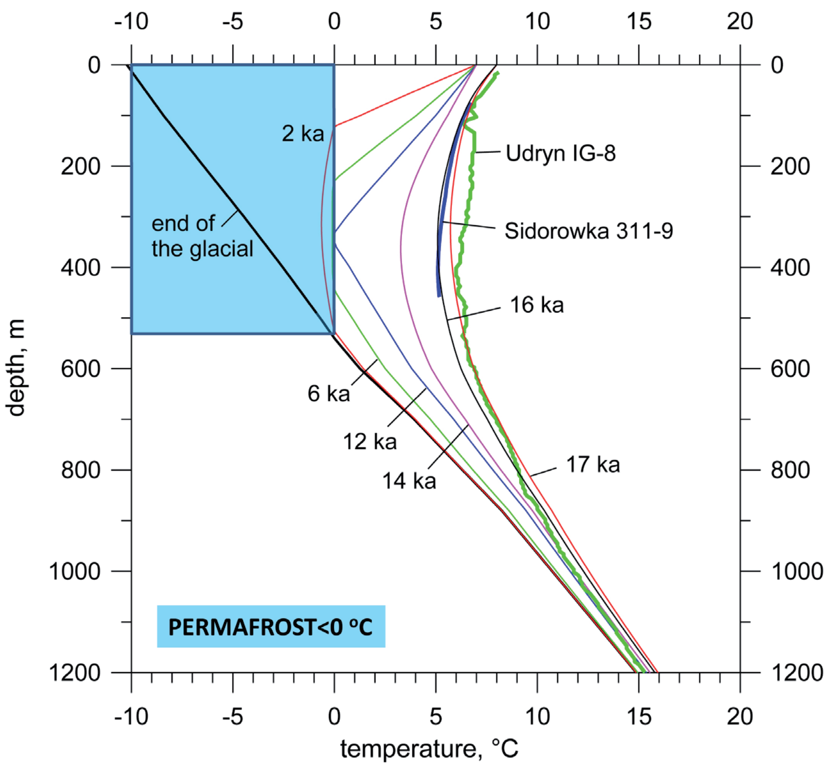

Fig. 7. The thickness of permafrost at the end of the glacial is 0.54 km (blue area in the electronic version and grey in the paper version) against the plot of change of temperature vs. time for the case of Udryn-Sidorowka in the Suwalski anorthosite massif (modified from Safanda et al. 2004). It is based on the following model: the glacial surface temperature of –10.23°C extrapolated to the surface from the depth of 2 km using the IG-8 profile; the conductivity and heat capacity model corresponds to Udryn site and 40% of unfrozen interstitial water; the Holocene temperature +7°C increased in the last 100 years of profiles 16 ka and 17 ka to +8°C

wells in Eurasia, including very low temperatures in Kola Peninsula and Karelia.

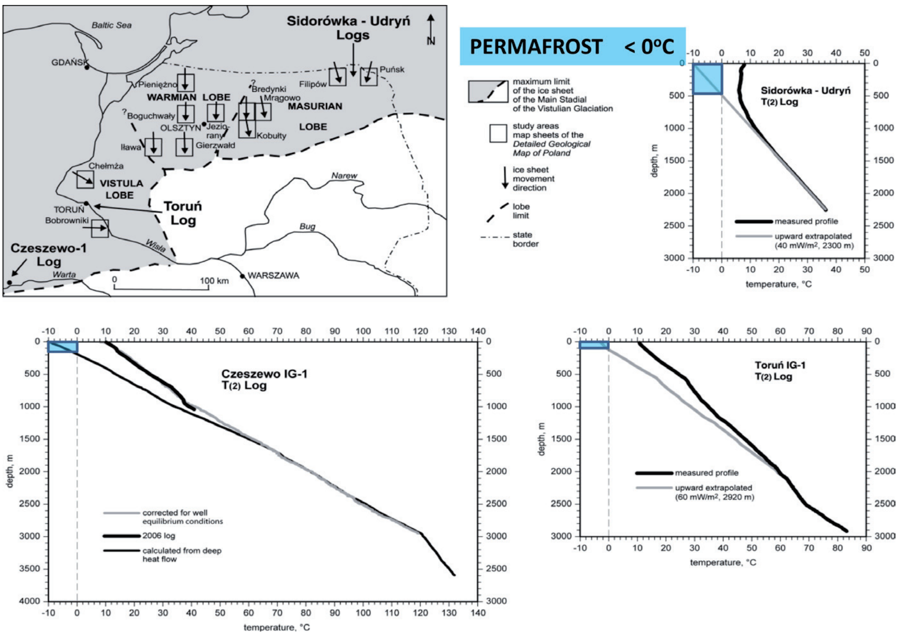

Fig. 8. Inference of paleo-permafrost from deep in Poland. Deep precise temperature logs in equilibrium borehole conditions vs. upward extrapolated temperature profiles pointing to negative surface paleo-temperatures at the time of the Pleistocene glaciation. The thickness of paleo-permafrost at the time is marked (squares). It was thickest under the Suwałki Massif in North-East Poland. Note that the thickest permafrost is in Sidorowka-Udryn (some 500 m vs. lower thickness in lowland wells of Torun-1 and Czeszewo (some 100–200 m), which can partly be explained by much lower paleo-temperature below the paleo ice bed (some –8°C in Sidorowka-Udryn), vs. –5°C in Torun-1, and much lower heat flow (40mWm^{-2} in Sidorowka-Udryn vs. 60 mWm^{-2} in Torun-1 and 80 mWm^{-2} in Czeszewo)

That the above found a large variability in paleo-temperatures under glaciated areas of northern Europe and northern America comes as no surprise, as modern-day measurements of temperature profiles through the ice of existing glaciers show a similarly large variability. Present bedrock temperatures measured beneath the central part of the Greenland ice sheet are –8 and –13°C (Dahl-Jensen et al. 1998), Fig. 9. Recently measured temperature profiles under glacial ice in many parts of the Arctic circle show a large variability in the temperature at the ice bed. These generally characterize two types of glacials: cold-based (such as the Greenland case GRIP and Dye-3 temperature profiles through ice (Dahl-Jensen et al. 1998; Ellsmere Island Lake Hazen region in Canada)) and warm-based (as with some of the Spitsbergen glaciers, e.g. Hans glacier and Yukon, Canada (Rusty glacier), etc. (see: Baranowski 1977 and Jania 1993 for a review of the subject)).

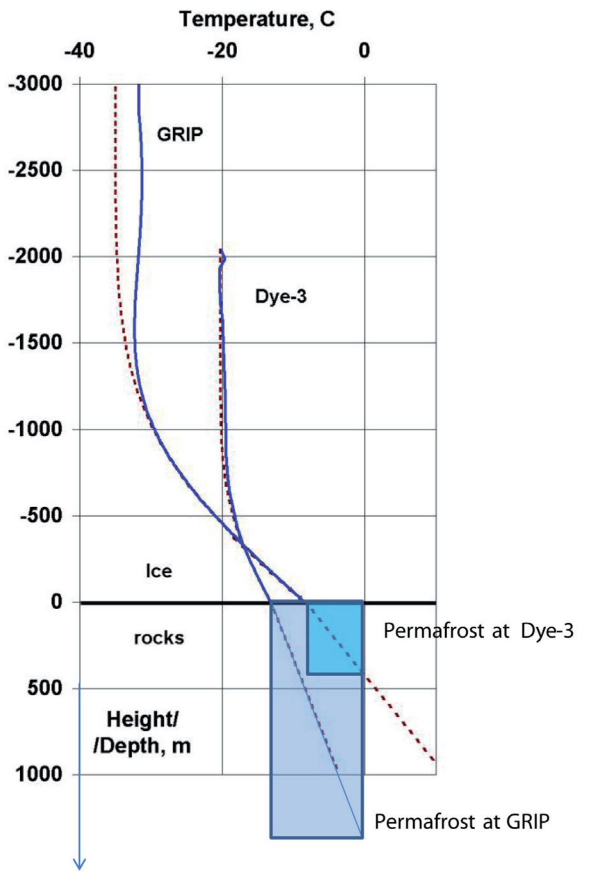

Fig. 9. Thickness of below glacial ice base permafrost (the blue area in the electronic version and grey in the paper version) determined from measured temperature profiles (solid lines, comment: Measured data were taken from (Dahl-Jensen et al., 1998 – Dye-3) and (GRIP Temperature Profile: http:// www.nsidc.org/ data/gispgrip/data/grip/) and calculated temperature-depth profiles in the Greenland Ice Sheet (dashed line from Demezhko et al. 2007)

Glacial ice-base temperature and permafrost below–paleo and present

According to the definition from the International Permafrost Association (http://ipa.arcticportal.org/resources/what-is-permafrost) permafrost is defined as ground (soil or rock and included ice or organic material) that remains at or below 0°C for at least two consecutive years. According to the above definition, sediments or crystalline rocks below 0°C are in permafrost conditions. Such conditions can occur below the present glacial bed or below the base of glacial ice at the time of Pleistocene glaciations (Laurentide, Fennoscandian, ...). In the case where these rocks consist of porous sediment filled with non-mineralized water, the melting point of 0°C will be the upper boundary of ice-bearing permafrost. In the case where water fills crystalline rocks with secondary porosity (cracks, fissures, fractures) we will also have an ice-bearing permafrost layer. Its thickness will depend on various conditions, including surface temperature, geothermal gradient and the pressure-dependent melting-point function for water with varying mineralization. Under constant upper surface-temperature conditions, the higher the temperature gradient, the thinner the permafrost (Fig. 3). In the case of saline waters ice, ice-bearing permafrost conditions will depend on the melting point of saline water. In order to predict the thickness of presently existing permafrost or paleo-permafrost under glacial ice or paleo-glacial ice respectively, we need at least to know the temperature at the base of glacial ice and the thermal gradient of the rock below. Changes in permafrost thickness are the result of a dynamic process for which climatic change is crucial. Its dynamic changes are monitored throughout permafrost areas as part of an international collaborative effort (Romanovsky et al. 2011).

We can estimate the thickness of permafrost (<0°C) for present conditions if the temperature at the base of glacial ice has been measured for the presently existing condition and the regional geothermal gradient is known. One such example is the case of Greenland (Fig. 9). Here, present permafrost can reach depths of 400 m below glacial ice in the case of Dye-3 and >1000 m in the case of the Grip.

In the case of paleo-conditions at the end of the LGM (the paleo-conditions for the areas where glacial ice retreated 9–14 k years ago) we can use the paleo-temperature at ice base derived from inversions of temperature-depth profiles, or upward extrapolation of the deep temperature profiles in wells

>2 km and the known regional geothermal gradient. This allows us to predict paleo-permafrost thickness. Such exercises have been carried out for the western Canadian deep well (Fig. 6) and for the Polish deep wells (Figs 7 and 8). The predicted thickness of paleo-permafrost varies and it is between some 100 m and 500 m for the cases analyzed.

It can be speculated that thick ice-bearing permafrost possibly occurring in porous sediments like in the western Canadian well (>200 m) and in Sidorowka-Udryn in Poland (>500 m) under the Paleo-glacial ice bed, could be bonded to the ice above and be a factor in glacier dynamics around these places.

Conclusions

1. Inversion or upward extrapolation of the >2-km-deep geothermal profile is the only physical (non-proxy) method by which temperature at the base of long-disappeared ice sheets such as the Laurentide and Fennoscandian in the northern part of the Northern Hemisphere in North America and Europe can be inferred.
2. The results inferred from well-temperature profiles show that there were spatial variations in the temperature at the base of the ice sheets during glaciations. This comes as no surprise, as modern-day measurements of temperature profiles through ice of existing glaciers show a similarly large variability.
3. Inferred from selected deep-well temperature logs (wells >2 km deep), low paleo-temperatures in areas presently outside the current extent of glacial ice, and geothermal conditions in sedimentary rocks under present "cold" glacial ice facilitate the existence of moderate (some 100–200 m) to thick (0.5 km–1 km) permafrost conditions.

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