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Climate of the Past Discussions

North American regional climate reconstruction from Ground Surface Temperature Histories

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Abstract. Within the framework of the PAGES NAm2k project, 510 North American borehole temperature-depth profiles were analyzed to infer recent climate changes. To facilitate comparisons and to study the same time period, the profiles were truncated at 300 meters. Ground surface temperature histories for the last 500 years were obtained for a model describing the temporal ground surface temperature changes. The model consists of a series of 10 time-intervals of variable duration. The evaluation of the model is done by inversion of the transient temperature perturbations using singular value decomposition. The long-term surface temperature (T_0) and thermal gradient (Γ_0) were retrieved by linear regression for the bottommost 100 meters. In addition, a Monte-Carlo approach was used to find the range of solutions within an acceptable error difference between the forward-modelled history and the data. The results within 95% confidence interval suggest a warming between 1.0°C to 2.5°C during the last two centuries. A regional analysis of mean temperature changes over the last 500 years show that all regions experienced warming, but this warming is not spatially uniform and is more marked in northern regions.

1 Introduction

The energy imbalance between incoming and outgoing radiation in the upper atmosphere due to increased concentrations of greenhouse gases is well documented (e.g. Hansen et al., 2011; von Schuckmann et al., 2016). The redistribution of the excess energy between climate subsystems, the atmosphere, the oceans and the solid Earth, drives changes in global and regional scale climate. As the consequences of climate change are expected to be negative for natural ecosystems and society, it is necessary that the projected changes in climate be established with sufficient details and certainty to provide the framework for policy directives intended to mitigate, adapt and build resilience at the community scale. Although there are multiple measures of climate change, surface air temperature (SAT) is the most common indicator because of the availability of data over the post-industrial period and also because it represents, in one way or another, the thermal conditions near the ground where people live.

The great majority of information on the future character and dynamics of the climate system comes from experiments with general circulation models (GCMs). GCMs are useful tools to assess future climate scenarios under different Representative Concentration Pathways (RCPs). However, because of the limited resolution of GCMs, many climatically relevant processes operating at less than the GCM grid size-scale are parameterized differently among model teams, such that GCM's simulations for the same RCP yield a climate state with a wide range of variability. Thus, GCM's simulations must be compared with data to assess the validity of their climate change projections (PAGES 2k-PMIP3 group, 2015; Smith et al., 2015).

Since the availability of meteorological records is limited to the last 150 years, additional information can be obtained from climate-dependent natural phenomena to reconstruct long-term past climate changes (e.g. Masson-Delmotte et al., 2013). Some of these indicators include data extracted from paleoclimate archives, such as ice cores (e.g. Oeschger and Langway, 1989; Bauer et al., 2013; Thompson et al., 2013), tree rings (e.g. Douglass, 1919; Briffa et al., 1990; George and Ault, 2014), pollen (e.g. Davis et al., 2003; Viau et al., 2006, 2012; Jacques et al., 2015) or geothermal data measured in boreholes (e.g. Mareschal and Beltrami, 1992; Bodri and Cermak, 2007; González-Rouco et al., 2009).

However, these proxy indicators are responses to a complex dynamical system and do not represent a direct measure of climate variability. While they allow for the determination and comparison of past climate trends, each of these methods of paleoclimatic reconstruction has different resolution, advantages, disadvantages and uncertainties.

Furthermore, due to spatial and natural limitations, the significance of the global and regional climate reconstructions decreases as it extends back in time. Calibration disparities and different reconstruction methods among these proxies give rise to a diverse range of weaknesses and strengths, making each paleo-indicator better suitable for a specific timespan. From a large set of natural phenomena, those sensitive to temperature variations can be used as climate indicators to reproduce past temperature histories.

Collaborative efforts have been conducted under the '2k Network' of the Past Global Changes (PAGES) project to produce a global array of regional climate reconstructions for the past 2000 years using proxy data sets derived from different natural sources (2k Consortium, 2013). It is within this multidisciplinary framework that geothermal data measured in boreholes can contribute with low-frequency trends retrieved from anomalies of the underground thermal regime.

Temperature-depth profiles measured in boreholes have commonly been used to study the magnitude and spatial variability of the flow of heat from the interior of the Earth (Bullard, 1939; Benfield, 1939; Jaupart and Mareschal, 2015, and references therein). It has been known since the times of Fourier and Kelvin, that underground temperatures are affected by past surface conditions. Assuming a coupling between ground surface temperate (GST) and SAT, borehole temperature reconstructions can be used as climate indicators for hundreds to thousands of years before present. Lane (1923) and Hotchkiss and Ingersoll (1934) were the first to use temperature-depth profiles for paleoclimatie studies in an attempt to determine the timing of the last glacial retreat. It was only in the 1970s that studies to infer past climate from borehole temperature profiles (BTPs) became more systematic, leading to the field of borehole climatology (Cermak, 1971; Sass et al., 1971; Beck, 1977).

Following the work of Lachenbruch and Marshall (1986), and because of concern about climate change, paleoclimatic reconstructions from borehole temperature data have become widespread, and have yielded local, regional, and global analyses

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(see Lewis, 1992; Bodri and Cermak, 2007; González-Rouco et al., 2009). However, the majority of the data are from the northern hemisphere.

In North America, where the largest number of temperature profiles are available, several for local and regional analyses have been performed (e.g. Beltrami and Mareschal, 1992; Guillou-Frottier et al., 1998b; Chouinard et al., 2007), however, very few studies have addressed the entire North American continent.

In this paper, and within the framework of the PAGES NAm2k project, we aim to estimate regional trends in the GST change of the past 500 years in North America from a dataset containing almost twice the number of data and larger depth range ($> 300m$) as previous analyses. The dataset analyzed here contains 510 borehole temperature-depth profiles distributed over the North American continent.

2 Methodology

The thermal regime of Earth's subsurface is governed by the outflow of heat from the Earth's interior and by the temporal variations of the ground surface temperature. For a homogeneous subsurface with no internal heat sources and with no ground surface temperature variations, the temperature in the subsurface increases linearly with depth. This profile can be considered in a quasi steady-state relative to the timescale of recent climatic variations, since it depends solely on heat flux from Earth's interior, which varies at much greater timescales. Persistent temporal changes in ground surface temperature propagate into the subsurface and are recorded as transient perturbations to this geothermal quasi steady-state. Because of heat diffusion, the amplitude of the subsurface anomalies is proportional to the duration and magnitude of the ground surface temperature perturbations and decreases with time since their occurrence. Since these temperature fluctuations diffuse downward, only the low-frequency climate signals are preserved. To reconstruct the temporal evolution of the ground surface temperatures, the variation of the subsurface temperature as a function of depth is measured in boreholes following the procedure described in 2.4. The transient perturbation is then retrieved from the borehole temperature profile (BTP) and inverted as described in 2.3, reconstructing the temporal ground surface temperature changes.

Furthermore, borehole climatology assumes that the ground surface temperature changes track long-term variations in surface air temperature. That is, it is assumed that ground surface and surface air temperature are coupled. This coupling has been confirmed by model simulations (e.g. González-Rouco et al., 2006; García-García et al., 2016), as well as data from continuous monitoring of air and ground temperature variations (Putnam and Chapman, 1996), and by comparing BTPs with meteorological records at nearby stations (Harris and Chapman, 1998). However, the relationship between surface air temperature and ground surface temperature can also be altered by transients effects in the surface conditions such as land use and associated hydrological, snow and vegetation cover changes (Lewis and Wang, 1998; Gosselin and Mareschal, 2003a; Bartlett et al., 2004). Thus, changes in ground surface temperature are not necessarily related to climate. Some of these perturbations of the surface environment can be observed at the time of measurement and should be considered prior to interpretation. When all non climatic effects have been ruled out, the interpretation of the perturbations of the temperature profiles allow us to reconstruct the past temperature changes at the surface.

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2.1 Temperature-depth equation

In order to interpret the temperature depth profiles, we must be able to describe quantitatively the thermal regime of subsurface and also how it is affected by changes in surface temperature. This requires the solution of the heat conservation equation for a continuous medium given by (Carslaw and Jaeger, 1959):

$$\frac{d}{dt}(\rho c_p T) - \boldsymbol{\nabla} \cdot (\lambda \boldsymbol{\nabla} T) = \dot{Q}_s \;, \tag{1}$$

where ρ is the density, c_p is the specific heat of the medium at constant pressure, λ is the thermal conductivity, $\boldsymbol{\nabla}$ is the vector differential operator and $\dot{Q}_s$ is the heat production rate per unit volume.

Because heat production rates in crustal rocks are small (on the order of 1 $\mathrm{mW\ m^{-2}}$) and the effect of heat production is negligible for holes that are only a few hundred meters deep, we have neglected heat production in this study.

Assuming that heat production can be neglected ($\dot{Q}_s \approx 0$), that there is no advection of heat ($\boldsymbol{v} \cdot \boldsymbol{\nabla} T = 0$) and that Earth is interpreted as a homogeneous half-space, the temperature at a depth z is given by the superposition of the steady-state profile and the transient perturbation due to time variations of surface temperature:

$$T(z) = T_0 + q_0 R(z) + T_t(z) \;, \tag{2}$$

where T_0 is the long-term surface temperature, q_0 is the quasi steady-state heat flux and $R(z)$ is the thermal depth defined as (Bullard, 1939):

$$R(z) = \int_0^z \frac{dz'}{\lambda(z')} \;, \tag{3}$$

where $\lambda(z')$ is the thermal conductivity at depth z'. For constant conductivity, equation 2 is written as:

$$T(z) = T_0 + \Gamma_0 z + T_t(z) \;, \tag{4}$$

where $\Gamma_0 = q_0/\lambda$ is the quasi steady-state temperature gradient.

If thermal conductivity can be assumed constant for the measured depth interval ($\lambda(z) = \lambda$), the transient component of temperature is calculated from the one dimensional heat conduction equation (Carslaw and Jaeger, 1959).

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} \;, \tag{5}$$

where $\kappa = \frac{\lambda}{\rho c_p}$ is the thermal diffusivity, also assumed constant ($\kappa \approx 10^{-6}\mathrm{m^2\ s^{-1}}$ or $\kappa \approx 31.6\mathrm{m^2 y^{-1}}$). Equation (5) must be solved with initial and boundary conditions: the temperature perturbation at the surface, $T(t, z = 0) = T_0(t)$, no perturbation for $z \to \infty$, $T(z = \infty, t) = 0$, and $T(z, t = 0) = 0$. The use of the one dimensional equation (5) is valid if the surface temperature variations have much larger spatial scale than their penetration depth (Clauser and Mareschal, 1995). Equation (5) also shows that the diffusivity determines the scaling relationship between time τ and depth L, scaling as $\tau \propto L^2/\kappa$. Periodic surface temperature variations propagate as a damped wave with skin depth $\delta = \sqrt{\kappa T/\pi}$. For standard values of κ for rocks, the amplitude of the wave associated with the annual temperature cycle is 10% of its surface value at 10m depth. For 100 year and 1000 year cycles, the amplitude of the wave is 10% its surface value at 100 and 300m respectively.

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2.2 Parametrization of the temperature anomaly

Assuming that Earth's underground thermal regime is at equilibrium and there are negligible diffusivity (κ) changes in the subsurface, the transient perturbation temperature $T_t(z) = T(z, t = 0)$ defined over a semi-infinite half-space with surface temperature $T(z = 0, t) = T_0(t)$ at time t before present is given by (Carslaw and Jaeger, 1959)

$$T_t(z) = \int_0^{\infty} \frac{z}{2\sqrt{\pi\kappa t}} \exp\left(\frac{-z^2}{4\kappa t}\right) T_0(t) t^{-\frac{3}{2}}\, dt\,. \tag{6}$$

For an instantaneously temperature change ΔT at time t before present, integrating the equation (6) yields (Carslaw and Jaeger, 1959)

$$T_t(z) = \Delta T\, \mathrm{erfc}\left(\frac{z}{2\sqrt{\kappa t}}\right)\,, \tag{7}$$

where erfc is the complementary error function:

$$\mathrm{erfc}(x) = 1 - \mathrm{erf}(x) = 1 - \frac{2}{\sqrt{\pi}} \int_0^{x} \exp(-u^2) du\,. \tag{8}$$

In order to approximate ground surface temperature changes, we assume that ground surface temperature can be replaced by its average value over time intervals of several years, so that the daily, annual, and solar activity cycles are removed.

Defining the contribution of ground temperature changes as ΔT_k during K time steps (i.e. ΔT_k for $t_{k-1} < t < t_k$ where $k = 1, ..., K$), the transient perturbation is the sum of the contributions for each time step:

$$T_t(z) = \sum_{k=1}^{K} \Delta T_k \left[\mathrm{erfc}\left(\frac{z}{2\sqrt{\kappa t_k}}\right) - \mathrm{erfc}\left(\frac{z}{2\sqrt{\kappa t_{k-1}}}\right)\right]\,. \tag{9}$$

Equation (9) gives the temperature anomaly $T_t(z)$ due to a sequence of ground surface temperature changes ΔT_k for K time intervals. The problem consists in determining the ground surface temperature history from the temperature versus depth anomaly, $T_t(z)$, at a given site. This is routinely done using inversion techniques.

2.3 Inversion

Combination of equations (2) and (9) yields a linear equation with the parameters T_0, Γ_0, and ΔT_k for each depth with temperature data. Thus, the inversion consists of solving the resulting system of linear equations. Obtaining the solution, however, is never straightforward because the system is “ill-conditioned", i.e., its solution is unstable (a small change in the data causes a very large change in the solution) and, for all practical purposes, the solution is non-unique. Different methods have been developed to solve inverse problems: the Backus-Gilbert method (Parker, 1977, 1994), singular value decomposition (SVD) (Lanczos, 1961; Jackson, 1972), Bayesian inversion (Tarantola and Valette, 1982), Tikhonov regularization (N and Y, 1977), and Monte-Carlo approaches (Mosegaard and Tarantola, 1995). One of the first applications of inversion to borehole temperature data was based on the Backus-Gilbert method (Vasseur et al., 1983); Shen and Beck (1991) proposed an algorithm

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based on the Bayesian approach while (Mareschal and Beltrami, 1992) used singular value decomposition. Because of the very small number of parameters, these methods of inversion are not computationally intensive. The Monte-Carlo approach, which has been used by Mareschal et al. (1999) and Kukkonen and Jõeleht (2003), explore the entire parameter space and requires significant computational resources. In this study, we have used singular value decomposition to find the ground surface temperature history because of its simplicity and then use a Monte-Carlo procedure to determine the range of model parameters that satisfy the data within some error bounds.

2.3.1 Subsurface temperature anomaly

In this study we determined the long-term surface temperature and quasi steady-state geothermal gradient by linear regression to the lowermost 100 meters of the measured temperature profile. This linear regression represents the geothermal quasi steady-state (eq. 2) from which the subsurface temperature anomalies are estimated. The anomaly $T_t(z)$ is obtained by subtracting this quasi-equilibrium thermal profile from the measured temperature profile. The least square regression also yields an estimate of the maximum error on slope and intercept estimates (95% confidence interval). These error bounds represent the upper and lower limits for the quasi steady-state temperature profile, hereafter referred to as the extremal geothermal steady-states. Figure 1 shows an example of a measured temperature profile and its estimate subsurface temperature anomaly, near Lynn Lake, Manitoba.

2.3.2 Singular value decomposition

After removal of the quasi steady-state component of the temperature profile, we are left with a system of linear equations between J temperature anomalies $T_t(z_j) = T'_j$ for each depth and the K parameters of the surface temperature history ΔT_k:

$$\begin{pmatrix} T'_1 \\ \vdots \\ T'_j \\ \vdots \\ T'_J \end{pmatrix} = \begin{pmatrix} A_{11} & \cdots & A_{1k} & \cdots & A_{1K} \\ \vdots & \ddots & \vdots & \ddots & \vdots \\ A_{j1} & \cdots & A_{jk} & \cdots & A_{jK} \\ \vdots & \ddots & \vdots & \ddots & \vdots \\ A_{J1} & \cdots & A_{Jk} & \cdots & A_{JK} \end{pmatrix} \begin{pmatrix} \Delta T_1 \\ \vdots \\ \Delta T_k \\ \vdots \\ \Delta T_K \end{pmatrix}, \tag{10}$$

where the A_{jk} are given by equation 9

$$A_{jk} = \mathrm{erfc}\left(\frac{z_j}{2\sqrt{\kappa t_k}}\right) - \mathrm{erfc}\left(\frac{z_j}{2\sqrt{\kappa t_{k-1}}}\right). \tag{11}$$

The number of equations J could be greater, equal, or less than the number of parameters K. In general, this number is larger than the number of parameters, but this does not ensure that the system 10 has a unique solution.

Writing formally, the matrix of equation (10)

$$\mathbf{\Theta} = \mathbf{A}\mathbf{x} \tag{12}$$

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where $\mathbf{\Theta}$ is the data vector, $\mathbf{A}$ is the rectangular $(J \times K)$ matrix containing the coefficients of the equations, and $\mathbf{x}$ is the vector of unknown coefficients.

SVD decomposes the matrix as (Lanczos, 1961):

$$\mathbf{A} = \mathbf{U\Lambda V}^{\top} \tag{13}$$

where $\mathbf{U}$ is an $(J \times J)$ orthonormal matrix in data space, $\mathbf{V}$ is an $(K \times K)$ orthonormal matrix in parameter space and $\mathbf{\Lambda}$ is a $J \times K$ rectangular matrix with only non-zero values, called "singular values" λ_l $(l = 1,..L)$ on the diagonal, with $L \leq \min(J,K)$. The singular values are the square root of the eigenvalues of the $J \times J$ symmetric matrix $(\mathbf{A}^{\top}\mathbf{A})$. If $L < J$, the system is overdetermined and if $L < K$, it is underdetermined. When the system is overdetermined and underdetermined, its general solution is given by:

$$\mathbf{X} = \mathbf{V\Lambda^{-1}U}^{\top}\mathbf{\Theta} \tag{14}$$

where Λ^{-1} is a $K \times J$ rectangular matrix with L elements $\frac{1}{\lambda_l}$ on the diagonal completed with zeros. This provides a solution which is usually not very meaningful (Mareschal and Beltrami, 1992) because it is unstable and dominated by noise. The instability of the solution comes the presence of very small singular values λ_l. In the case of borehole temperature profiles, the fifth largest singular value is 0.01 times the largest one, and the tenth is $< 10^{-8}$ times the largest one, that is, numerical noise. In order to stabilize the solution, we eliminate the part associated to the very small singular values. This is done by replacing with 0 the inverse of all the singular values less than a "*cut-off value*", typically on the order of 10^{-2}. This means that the actual solution is obtained as a linear combination of 4 orthogonal vectors in parameter space. Each vector represents a surface temperature history, and the vectors selected are those that have the largest impact on the data. By eliminating the small singular values, we choose to neglect the part of the solution that has little or no effect on the data, and therefore cannot be determined. In general, the selection of a cutoff value is done by trial and error, by increasing the number of singular values and inspecting the solution for signs of instabilities and loss of resolution, i.e. large non physically meaningful fluctuations or no useful information. For this study, we have always retained only the four largest singular values.

The choice of a proper parametrization is useful to reduce the number of parameters to be estimated. This can be achieved by increasing the duration of the ground surface temperature history model time intervals. For very long reconstructions a logarithmic distribution has been used (e.g. Mareschal et al., 1999). For the present study, we have used a model consisting of a series of 10 time intervals of varying duration. Their temporal length is smaller for the near (past 100 years) than for the remote past. The distribution used here is:

$$t_k = \{0,\ 25,\ 50,\ 75,\ 100,\ 150,\ 200,\ 250,\ 300,\ 400,\ 500\} \tag{15}$$

When doing regional averages, the GST histories are shifted in time to account for the date when they were logged (i.e. years before presentpresent is year of logging).

As an example, Figure 2 shows the result of inversion of the subsurface temperature anomaly for the Fox mine site, and the results from the inversions of the two extremal geothermal steady-states.

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2.3.3 Forward model

GST histories can be forward-modelled using equation (9) to assess the fit of the SVD inversion with respect the initial anomaly profile. A Monte-Carlo procedure was applied (Mareschal et al., 1999; Kukkonen and Jõeleht, 2003; Chouinard et al., 2007) by randomly perturbing the model parameters to find the range of GST histories that fit the data within a maximum root mean square (RMS) error less or equal than the difference between the forward-modelled SVD reconstruction and the anomaly. Using the Monte-Carlo approach to invert the temperature profiles is particularly inefficient because it requires a very large number of simulations to explore the entire parameter space. It requires at least $10^7 - 10^8$ longer computational time as using the SVD inversion. However, this can be alleviated by using a-priori information or the result of an existing ground surface temperature history from inversion to reduce the region explored in parameter space. After the Monte-Carlo inversion, the mean and standard deviation of all the accepted models are estimated to show the trend of all the solutions with a same or better fit than the inversion for 4 singular values. For the present study, we halted the calculations after 500 models are accepted or after 5 million forward model comparisons.

This is illustrated in Figure 3 that shows the results of the Monte-Carlo inversion for the Fox mine temperature profile.

2.4 Data

We have compiled from different sources (Table 1) a set of temperature depth profiles for North America. Thousands of borehole temperature profiles have been measured in North America, but the majority of them are not suitable for climate reconstructions. For instance, bottom hole temperatures, commonly measured during oil exploration drilling, are not measured at equilibrium, and are affected by errors several times larger than the signals we want to detect. Water wells are usually too shallow to be useful and likely to be affected by water flow. Many holes were drilled for geothermal energy in the western US but are often perturbed by water circulation. For heat flow or climate studies, the most useful boreholes are those that have been drilled by mining companies for exploration or development purposes. Oil exploration wells cannot be used for several reasons: holes that are not put in production must be cemented and they are not accessible for steady-state measurements. In addition, oil-exploration boreholes have a large diameter and are susceptible to perturbations due to convection. Furthermore, sedimentary rocks are permeable and often affected by convection as well. Hence, their temperature profiles are not suitable for climate studies. Drilling perturbs the thermal regime of the subsurface around the drill site and some time is needed for thermal re-equilibration. As a rule of thumb, the time to return to equilibrium is $\sim$ 5-6 times the duration of drilling. The temperature in the hole is measured with a calibrated thermistor. The probe is lowered in the hole and measurements are made at fixed intervals along the length of the hole, which results in varying depth intervals as most boreholes are inclined. The sampling interval is usually 10m, sometimes 50 feet for US and old Canadian temperature logs. Continuous measurements can be obtained, but are not common because they require heavy equipment. Measurements made above the water table are rarely equilibrated; consequently, the upper 20 or 30m of the temperature logs must be discarded. This is also done in order to eliminate the annual temperature variation signal. In heat flow studies, core samples must be obtained to determine the underlying rock's thermal conductivity and heat production. Changes in thermal conductivity are thus included in the interpretation of these data.

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2.4.1 Data selection

Different criteria have been applied in selecting the temperature profiles. Temperature profiles must be at least 300 meters deep to contain the signal to allow for the reconstruction of the climate of the past 500 years. Profiles must include at least 10 measurements, and must include measurements in the uppermost 100m. Profiles that meet these conditions are then visually inspected to detect discontinuities, signs of water flow, or other perturbations that make them unsuitable for interpretation. The vertical temperature gradient profile amplifies the noise and usually provides a better diagnostic for the level of noise in the measurements. Although we have not established a quantitative criterion for selecting profiles based on the noise level, we have examined the vertical gradients to eliminate unsuitable profiles.

After selection process, we retained 510 profiles. These data will be available in a public database in Figshare (Jaume-Santero et al., 2016). Borehole locations are not uniformly distributed across the continent (Figure 4). Several regions are very poorly sampled because they are very difficult to access (Alaska and most of Canada, north of 56 ^{o}). Furthermore, in the northernmost regions, drill holes cannot be routinely logged because of permafrost. Temperature logging in frozen ground requires special equipment to be emplaced at the end of drilling and is very costly. The southern part of the Canadian Shield is the region most extensively sampled because of the mining activity and because the temperature profiles are less likely to be perturbed in the crystalline rocks of the Shield. In contrast, numerous drill holes are available in the south-western US, but most of them cannot be used because they are perturbed by water flow. The sedimentary cover in many regions of the US explains that no suitable holes have been found for many states, including Texas and Oklahoma and the south-eastern US. This very uneven distribution of suitable boreholes is demonstrated in Table 2 which shows the number of temperature profiles for each one of the regions defined for Pages2k (McKay, 2014).

3 Results & discussion

All 510 borehole temperature-depth profiles were inverted individually to reconstruct the GST histories for the past 500 years. The model consisted of a series of 10 temperature change intervals of varying temporal duration following the distribution (15). For the inversion, we used the singular value decomposition inversion with a cutoff of 0.03, retaining 4 singular values. We also used the Monte Carlo methodology to estimate the range of parameter values consistent with the data. The means of the GST's obtained by Monte-Carlo are similar to the solution by SVD inversion. With the condition that the RMS difference between model and data be no larger than the misfit for the SVD, the 2σ range of accepted models is no larger than $0.44°C$.

3.1 North-American ground surface temperature change

We have calculated the variation in ground surface temperature for North America by averaging all the Monte-Carlo inversions. The averaging was done on a yearly basis because the logging dates vary between boreholes from 1958 to 2014 (Figure 5).

Figure 5 shows the individual Monte-Carlo inversions together with their average. The individual inversions in Figure 5 exhibit a wide variability due to the large range of latitudes ($\sim 80°N$ to $\sim 18°N$) in the data set of GST reconstructions.

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Nevertheless, a clear warming transition is observed from the pre-industrial era (1500-1800) to the post-industrial era (1800-2000). The temperature difference between the pre-industrial mean (1500-1700) and the mean between the years (1961-1990) is $1.1°C$. Because of the marked warming of the past 50 years, the total change of the average ground surface temperature is $1.8°C$ between pre-industrial time and the year 2000.

These results agree with findings of other ground surface temperature reconstructions (Huang et al., 2000; Harris and Chapman, 2001; Beltrami and Bourlon, 2004; Pollack and Smerdon, 2004). Furthermore they agree with instrumental data, CRUTEM4 (Jones et al., 2012; Morice et al., 2012), pollen and tree ring reconstructions (2k Consortium, 2013; Trouet et al., 2013). All of them presented as departures from the 1904-1980 temperature mean (Figure 6). However, the reconstructed GST warming signal for the past 200 years is greater than results from pollen reconstructions, coinciding with the findings of PAGES 2k-PMIP3 group (2015). Furthermore, multi-centennial temperature reconstructions for North America and the Northern Hemisphere, based on multiproxy records, showed trends similar to temperature-depth reconstructions: an unclear cold-warm trend followed by a clear increase in temperature for the past two centuries (Moberg et al., 2005; Mann et al., 2008; PAGES 2k-PMIP3 group, 2015). This warming has also been recorded by instrumental data for the last century (Hansen et al., 2010).

The Little Ice Age (LIA) is not resolved because the boreholes were truncated at 300 meters which is too shallow to allow for a clear LIA signal in most of the borehole profiles as can be shown with synthetic models (Mareschal and Beltrami, 1992) and was confirmed in several studies (Guillou-Frottier et al., 1998b; Chouinard et al., 2007; Pickler et al., 2016). Although, some profiles, such as the Fox Mine shown in Figure 2, may indeed show the LIA cooling, but the majority of them do not. In addition, because the LIA signal may vary both in time and in amplitude between regions, a marked signal cannot be expected from averaging weak and inconsistent signals.

3.2 Regional averages

The PAGES NAm2k working group divided the North American continent into seven subregions for paleoclimate studies (McKay, 2014). The distribution of boreholes between these regions is extremely uneven as shown in Table 2, with only 4 regions appearing adequately sampled (Central & Eastern Canada, Midwestern US, Arctic, and Pacific Northwest). Furthermore, the sampling in the Arctic and the Pacific northwest is very biased because all the boreholes are close to the coast (Figure 4). For the three other regions, the sampling is insufficient to obtain robust climate trends.

An increase of $\sim 1.8°C$ for the past 200 years is observed in the Arctic (Figure 7a). However, a wide variability is present. This variability is consistent with previous Arctic reconstructions (Chouinard et al., 2007) and suggests the need for smaller-scale regional analysis such as the pollen-based reconstructions of Gajewski (2015) and Viau and Gajewski (2009). Their findings illustrate that recent Arctic increases in temperature have exceeded natural climate variability, which is consistent with borehole GST reconstructions.

The region of the Pacific northwest (Western Canada & Northwestern US) shows an increase in temperature of $\sim 0.8°C$ with a 95% variability range of $\sim 3.4°C$ for the last two centuries(Figure 7b). This warming is consistent with previous findings (Majorowicz and Safanda, 2001).

An average warming of $\sim 1.1°C$ with a 95% variability range of $\sim 2.2°C$, for the past two centuries is observed for Central & Eastern Canada (Figure 7c), agreeing with previous studies (Guillou-Frottier et al., 1998a; Beltrami et al., 1992).

The Western US GST mean shows a small increase in temperature of $\sim 0.2°C \pm 1.8°C$ (Figure 7d). This could be the result of strong irrigation processes and water flow at the sampling locations.

The average reconstruction for the Midwestern US suggests a warming of $\sim 1.3°C \pm 2.0°C$ for the last 50 year average (Figure 7f). This recent warming has also been observed in previous GST reconstructions as well as SAT records (Skinner and Majorowicz, 1999) and could reflect the significant land use change in the region.

A warming of $\sim 1.0°C \pm 1.0°C$ has been reconstructed for the last 200 years in the Eastern United States (Figure 7e). However, due to the rejection of borehole profiles affected by topography and water flow, the number of reconstructions made is too small to describe with confidence climate trends of the region.

There is a warming trend of $\sim 3.0°C \pm 3.6°C$ until the mid 1960s in the Caribbean (Figure 7g). Due to the low number of profiles sampled in Mexico (0) $\&$ the Caribbean (4), it is not possible to obtain a robust reconstruction for this region.

3.3 Regional analysis

A North American regional analysis of GST changes is presented for different 50-year time intervals during the last 300 years, (Figure 8).

Trends between 1515 and 1714 are not shown because they did not yield significant information. However, a small (~0.5°C) cooling is observed in certain regions. Previous small scale regional analyses have reconstructed a LIA signal during this period (e.g. Beltrami and Mareschal, 1992; Chouinard et al., 2007). Furthermore, the regional variability of the cooling is consistent with previous studies, illustrating that not all regions of North America present a LIA signal (Gosselin and Mareschal, 2003b; Mann et al., 2009). However, due to the truncation at 300m of the temperature-depth profiles analyzed here, a clear LIA signal cannot be resolved.

Figure 8 indicates a warming trend of ~1-2°C in most parts of North America during the last 200 years. This is consistent with previous studies (Huang et al., 2000; Harris and Chapman, 2001; Beltrami et al., 2003). A cooling trend is observed in central California. Stevens et al. (2008) shows how this differs from the output of the ECHO-G model and postulates that it is the result of intensive irrigation in California's central valley, which could drive a regional cooling signal (Kueppers et al., 2007). A similar cooling signal is observed in British Columbia which might be associated with irrigation in the Fraser Valley. On the Canadian east coast, Newfoundland presents little to no changes with respect the long-term mean. This agrees with meteorological data for the region (Gullett and Skinner, 1992). The absence of temperature profiles along the Gulf coast and Mexico does not allow for any determination of climate trends. The southwestern US is also a region where the number of boreholes is not enough for reliable reconstructions. For these regions, multi-proxy approach would be necessary to improve the reconstruction of regional past climate in regions with an insufficient number of borehole profiles.

4 Conclusions

The average North American ground surface temperature change reconstructed from 510 boreholes deeper than 300 meters, suggests a warming of ~ 1.8°C for the last 200 years. However, these temperatures exhibit a wide range of spatial variability among all regions. For instance, reconstructed regional ground surface temperature changes for seven climate distinct regions, defined within the PAGES NAm2k project, suggest a warming range of $\sim 0.5°C$ to $\sim 2.0°C$ with a variability 2σ, no smaller than $1.0°C$. Furthermore, regional variations of GST yield a warming range of 1°C to 2°C between 1814 and 2014. These warming trends are consistent with multi-proxy reconstructions.

Although the number of borehole temperature profiles for North America has been notably increased in our study, it is still insufficient to guarantee a non spatial-biased regional analysis because their distribution is not sufficiently uniform. Nevertheless, despite spatial and natural limitations, subsurface thermal profiles obtained from boreholes provide robust long-term GST histories which could be used to improve climate multi-proxy-based reconstructions. Those enhanced reconstructions would bring out worthwhile information for a straightforward assessment of past climate GCM outputs.

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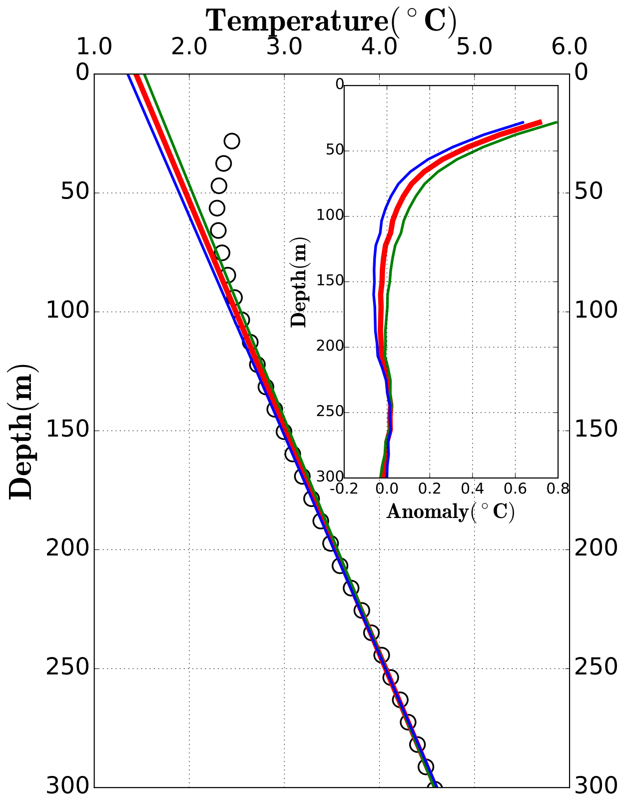

Figure 1. Temperature profile measured at Fox Mine (CA-9519), Lynn Lake, northern Manitoba, Canada. Main panel: Measurements are shown in circles $T(z)$, the red line represents the geothermal steady-state, obtained by linear regression of the lowermost 100 meters, and extrapolated to the surface ($z = 0$). Blue and green lines represent the 95% confidence interval from the linear regression. Inset: Transient perturbation or anomaly relative to the geothermal steady-state (red line) and the 95% confidence interval (blue and green lines). For this site, the geothermal steady-state is given by $\Gamma_o z + T_0 = (10.51 \pm 0.34) \times z + (1.44 \pm 0.19)$ (z in km).

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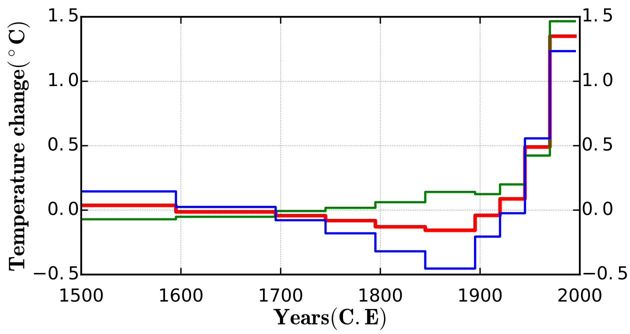

Figure 2. Ground surface temperature history for CA-9519 (Fox Mine, 1995). The red line represents the ground surface temperature history reconstructed from inversion. The blue and green lines are the GSTs for the anomalies estimated from the 95% uncertainty limits of the quasi steady-state profile.

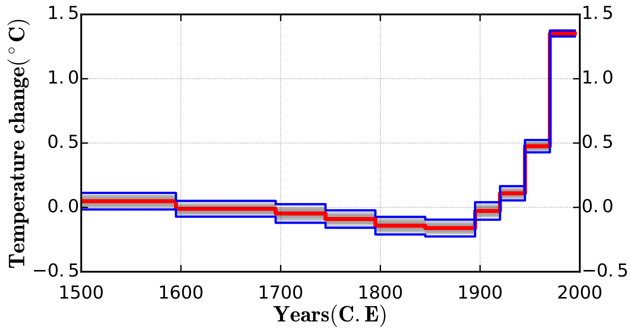

Figure 3. CA-9519 (Fox Mine, 1995) Mean ground surface temperature history (red) and 2σ uncertainty intervals (blue) from the Monte Carlo inversion. The grey lines represent all the perturbed models within an interval determined by the RMS misfit from the SVD inversion.

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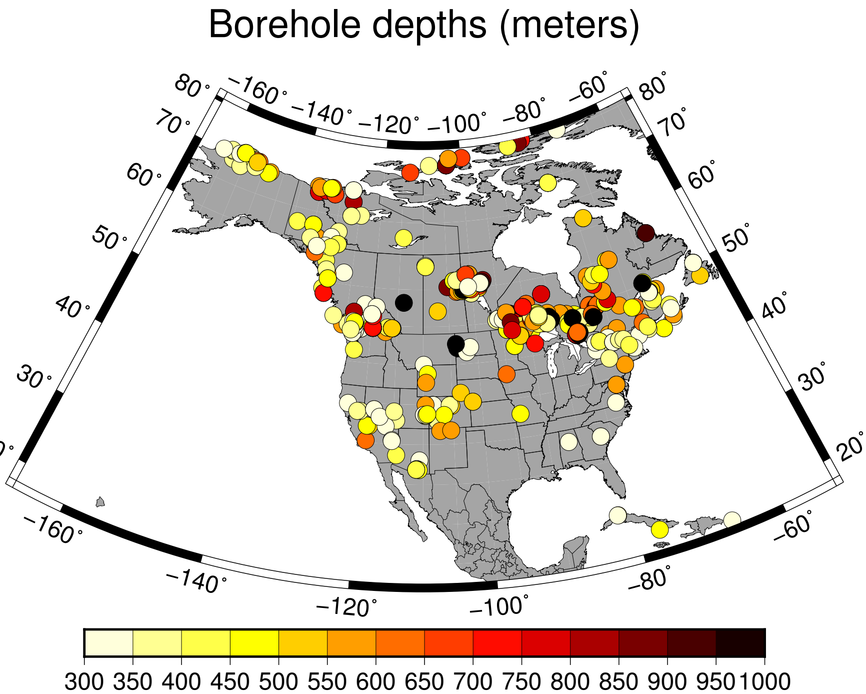

Figure 4. Location of the 510 selected boreholes. The colors represent the maximum depth of each borehole.

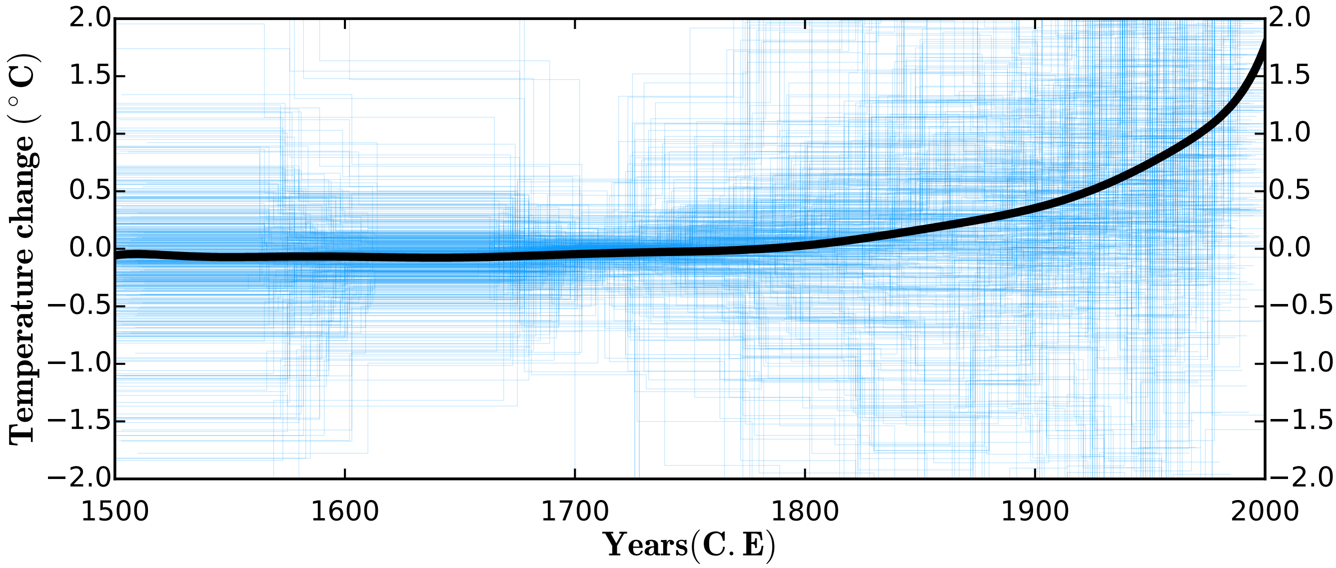

Figure 5. Mean North American ground surface temperature change (black). Shown in blue are the 510 ground surface temperature reconstructions inferred from the Monte Carlo inversion.

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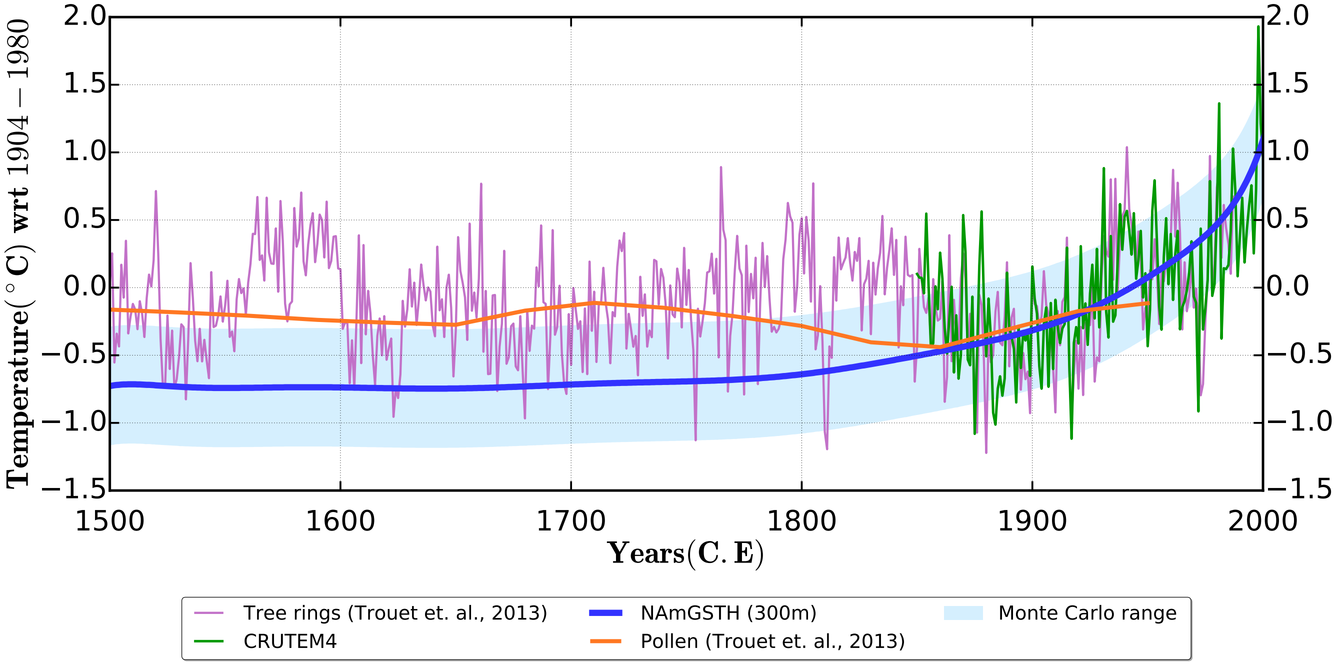

Figure 6. Mean North American ground surface temperature history (blue) and maximum temperature range of accepted models (∼0.44°C) obtained from the Monte Carlo method (blue shade). Also shown are proxy-based surface air temperature reconstruction for North America from 1500 to 2000 CE. All anomalies are displayed as departures from 1904-1980 mean.

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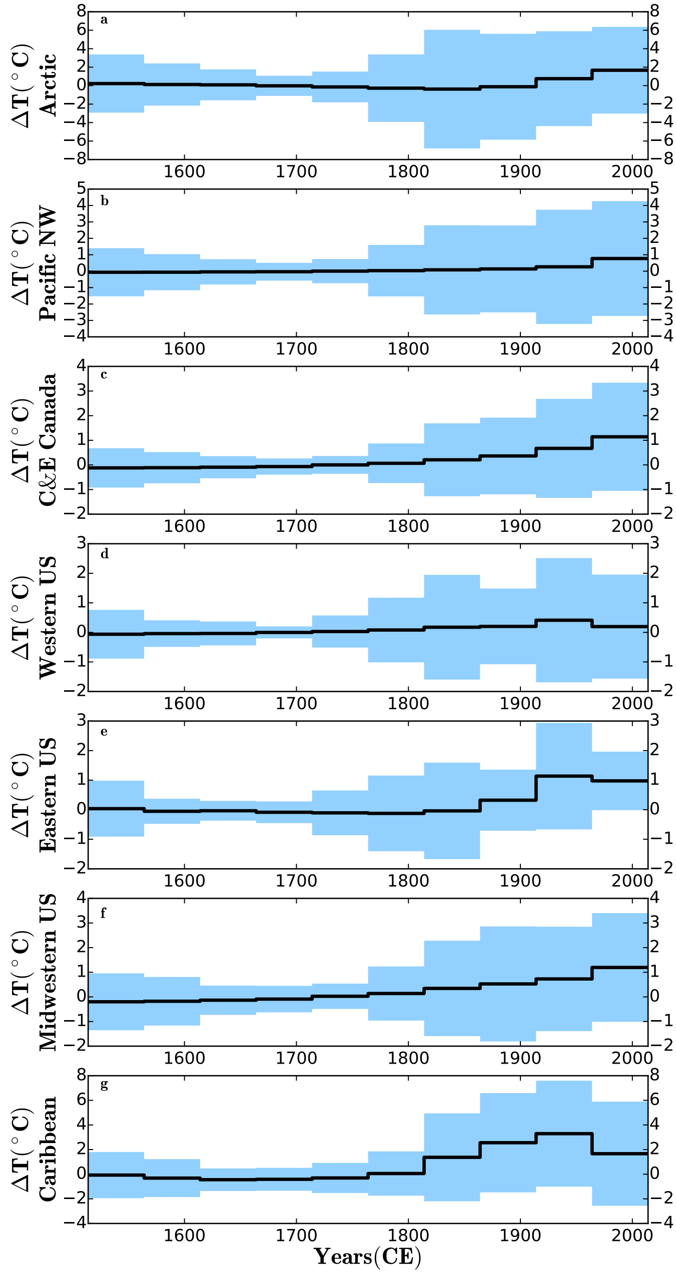

Figure 7. Mean ground surface temperature histories (black), the shaded areas represent the 95% confidence interval associated with the climate variability of each area. a: Artic (78 sites), b: Pacific Northwest (78 sites), c: Central & Eastern Canada (220 sites), d: Western US (21 sites), e: Eastern US (9 sites), f: Midwestern US (100 sites), g: Caribbean (4 sites).

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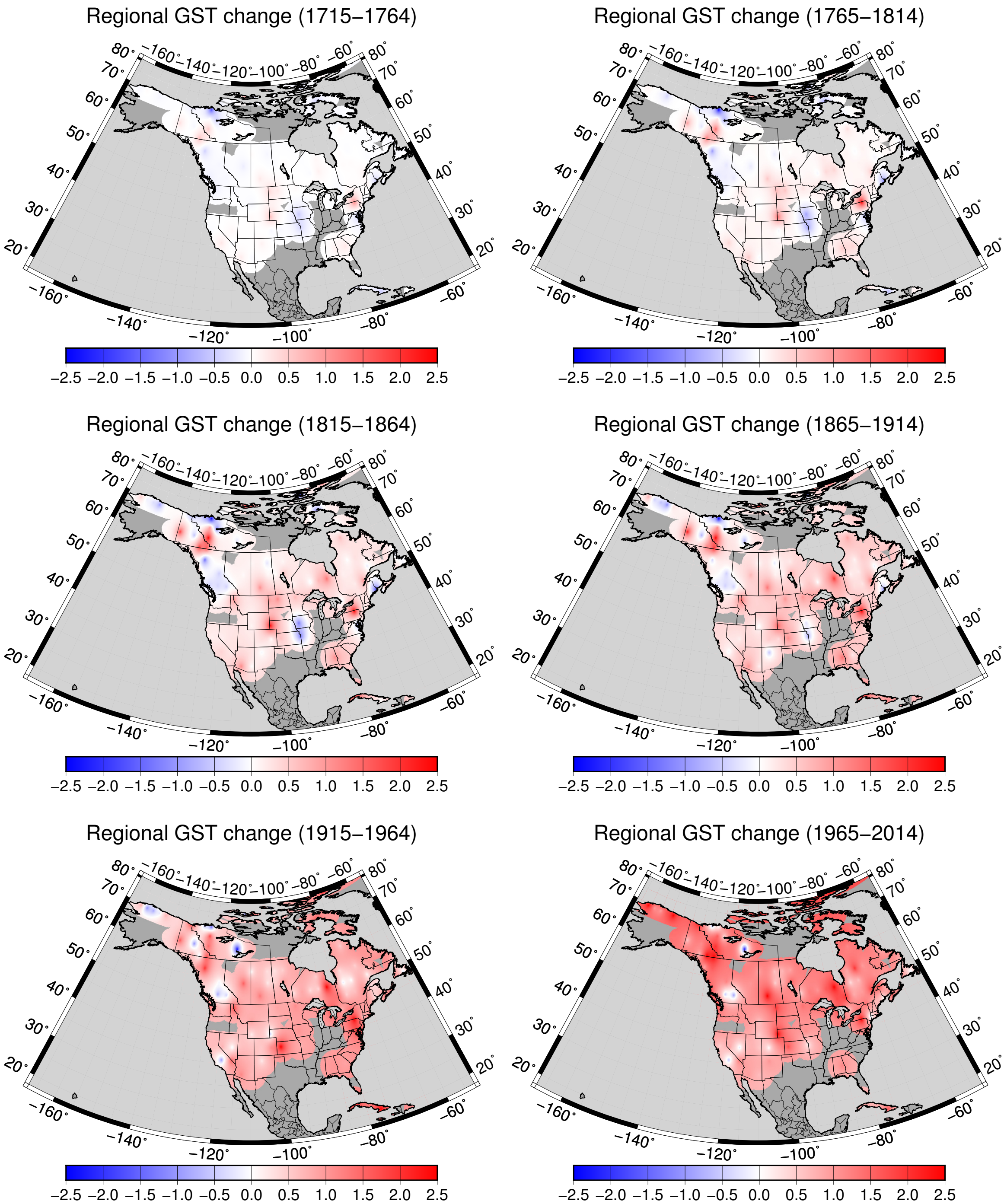

Figure 8. Spatial variability of the ground surface temperature variation from 1714 to 2014. Each panel shows a regionally interpolated mean ground surface temperature over 50 years. The surface has been masked for zones without at least one datum within a radius of 400 km. Ground surface temperature changes are presented as departures from long-term mean surface temperatures prior to 1500 CE.

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Table 1. Sources of the temperature-depth profiles.

Source name	Availability
University of Michigan	http://www.earth.lsa.umich.edu/
SMU Geothermal Lab	http://geothermal.smu.edu/
GEOTOP-IPGP heat flow database	http://www.geotop.ca/
USGS array	www.aoncadis.org/dataset/USGS_DOI_GTN-P/file.html
NOAA borehole datasets	Huang et al. (1999)
Canadian geothermal data compilation	Jessop et al. (2005)
Richard Scattolini, Ph.D. thesis	Scattolini (1978)

Data extracted from public databases and published papers. All rights belong to original publishers.

Table 2. Distribution of borehole between regions as defined for PAGES2k (McKay, 2014)

Region	Number of profiles
Arctic	78
Pacific NW	78
Central & Eastern Canada	220
Western US	21
Eastern US	9
Midwestern US	100
Caribean	4