

# Global and hemispheric temperature reconstruction from glacier length fluctuations

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**Abstract** Temperature reconstructions for recent centuries provide a historical context for the warming over the twentieth century. We reconstruct annual averaged surface temperatures of the past 400 years on hemispherical and global scale from glacier length fluctuations. We use the glacier length records of 308 glaciers. The reconstruction is a temperature proxy with decadal resolution that is completely independent of other temperature records. Temperatures are derived from glacier length changes using a linear response equation and an analytical glacier model that is calibrated on numerical model results. The global and hemispherical temperatures reconstructed from glacier length fluctuations are in good agreement with the instrumental record of the last century. Furthermore our results agree with existing multi-proxy reconstructions of temperature in the pre-instrumental period. The temperature record obtained from glacier fluctuations confirms the pronounced warming of the twentieth century, giving a global cumulative warming of  $0.94 \pm 0.31$  K over the period 1830–2000 and a cumulative warming of  $0.84 \pm 0.35$  K over the period 1600–2000.

**Keywords** Temperature · Glacier length · Proxy · Reconstruction · Global warming

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## 1 Introduction

Knowledge of the climate variability over the last centuries forms the key to understanding present day climate change. For recent times this information is provided by instrumental records (Jones and Moberg 2003; Brohan et al. 2006). However, before the mid 19th century, such instrumental records are mostly lacking. Temperature reconstruction then demands the use of natural climate archives, proxies, to reconstruct past climates on longer timescales. Focusing on temperature changes, many proxies have been used in studies made in the last decades. These proxies include those of biogenic nature, e.g. the widely used tree-rings (Briffa et al. 2001; Esper et al. 2002; Mann and Jones 2003) and corals (Lough 2004), reconstructions based on historical evidence (Pfister et al. 1999; Brázdil et al. 2005; Jones et al. 2009), as well as pure physical methods such as temperature reconstructions from ice cores (Mosley-Thompson et al. 2006; Overpeck et al. 1997), borehole temperatures (Huang et al. 2000) and groundwater composition (Alvarado et al. 2009). Biogenic and historical records mostly have a high resolution, resolving annual or even monthly variations, whereas the physical proxies have the advantage that they do not need a calibration on the instrumental record (Jones et al. 2009; Juckes et al. 2007). Recently multi-proxy composites were made by combining the results of different proxies (Mann et al. 2008; Huang 2004), which creates a more reliable temperature reconstruction. From this perspective it is useful to explore as many independent and reliable proxies as possible.

The observed worldwide glacier retreat over the 20th century is a strong indication for global warming (e.g. Dyurgerov and Meier 2000). Here we present a quantitative reconstruction of global and hemispheric temperatures from glacier length fluctuations for the period 1600–2000.

Glacier length fluctuations constitute a physical proxy for variations in temperature; they do not need calibration on the instrumental temperature record, and are fully independent of other reconstructions. Glaciers can be found on all continents and at virtual all latitudes, even in the tropics, and therefore the information on glacier length fluctuations potentially has good global coverage. However, as with other proxies (Huang et al. 2000; Mann et al. 2008), information on glacier fluctuations is more abundant in the Northern Hemisphere (NH) than in the Southern Hemisphere (SH). Glaciers respond slowly to changes in climate so they are a proxy with decadal resolution at best. The response of glaciers to changes in climate is dependent on the glacier geometry and its climatic setting (Oerlemans 2001). These differences have to be taken into account for each glacier individually when the variations in glacier length are interpreted.

First we discuss the improvements of the data set on glacier length fluctuations compared to the set used in the temperature reconstruction by Oerlemans (2005, Sect. 2). In Sect. 3 the adjustments to the methods for calculating an average temperature record from the glacier fluctuations are explained. In the next sections we present the reconstructed global (Sect. 4.1) and hemispheric (Sect. 4.2) temperatures. We compare them with the multi-proxy composite temperature reconstruction of Mann et al. (2008). In addition, we reconstruct regional European temperatures of the past four centuries. We compare the reconstructed temperatures to instrumental records and the temperature reconstruction based on the multi-proxy temperature reconstruction of Luterbacher et al. (2007) to test our assumption that variations in precipitation can be neglected on global and hemispherical scale (Sect. 4.3).

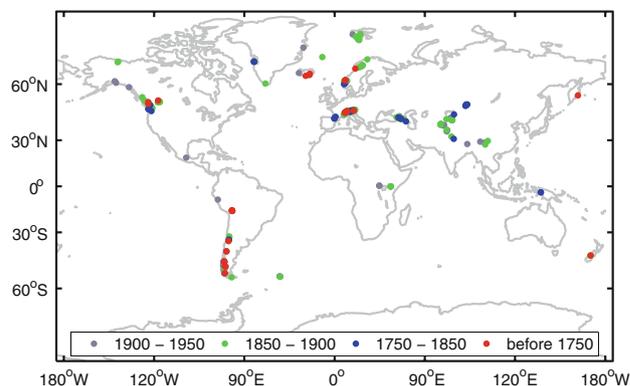
## 2 Data

### 2.1 Glacier length changes

For this study we have used an extension of the data on glacier length fluctuations as described in Oerlemans (2005). Where possible, we have revised and extended the previously used records. Moreover, we have added new records. More than half of the 169 records used in Oerlemans (2005) were of glaciers located in the European Alps. In the present data set, the number of glacier length records of glaciers outside the Alps has significantly increased: from 73 to 222. Only records that start prior to 1950 and extend over at least several decades were used. All records with less than five data points were excluded, as well as records of glacier length changes that are known to be strongly influenced by other causes than climate change, e.g. surges. However, some calving glaciers were included.

The data of the frontal position measurements available from the WGMS (<http://www.wgms.ch>, WGMS 2008 and earlier volumes) form the basis of our data set. Furthermore, there is a wealth of data on reconstructed glacier lengths in the literature. The measurements (often done by volunteers) and reconstructions of length changes involve time-consuming work for each of the glaciers individually. Glacier characteristics and complete references are given in the Supplementary Material.

Figure 1 shows the global distribution of the 308 glacier length records. Records can be found on all continents and are, not surprisingly, strongly clustered in the major mountainous regions. There is a striking paucity in data from glaciers in the Canadian and Russian Arctic, as well as from glaciers near the Antarctic Ice Sheet. On the other hand, there is a wealth of information on the glaciers in the Alps, Norway, and Iceland. For many glaciers, yearly measurements were available since the early 20th century or even since the end of the 19th century. In addition, there are historical sources, some of which go back to the 16th century (e.g. Zumbühl 1980; Zumbühl and Holzhauser 1988; Nussbaumer et al. 2007; Nesje et al. 2008). Continuous measurements are also available of glaciers in North West America, especially in the Canadian Rockies and Coast Mountains. Unfortunately, half of these records are not continued until the present day, but end between 1965 and 1985. In North America, additional historical documentation is less abundant than in Europe. In the Caucasus and Central Asia most records start in the second half of the 19th century, from first historical documents or the dating of Little Ice Age (LIA) moraines. Glaciers in mountain areas in the former Soviet Union, such as the Caucasus, Pamirs and Altai, are relatively well documented (e.g. Panov 1993; Kutuzov and Shahgedanova 2009). There is far less information of glaciers in the Himalayas, despite its large glacialization. The information on the arctic glaciers in Alaska, Greenland and Svalbard in the 19th century is a combination

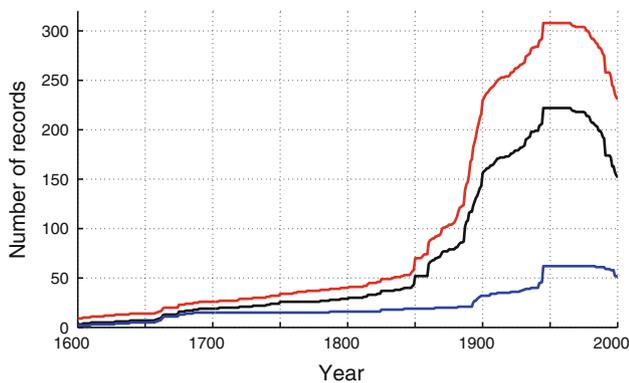


**Fig. 1** Distribution of records over the world; colours indicate the period of the first data point of the record. In many cases the distance between glaciers is so small that the dots overlap

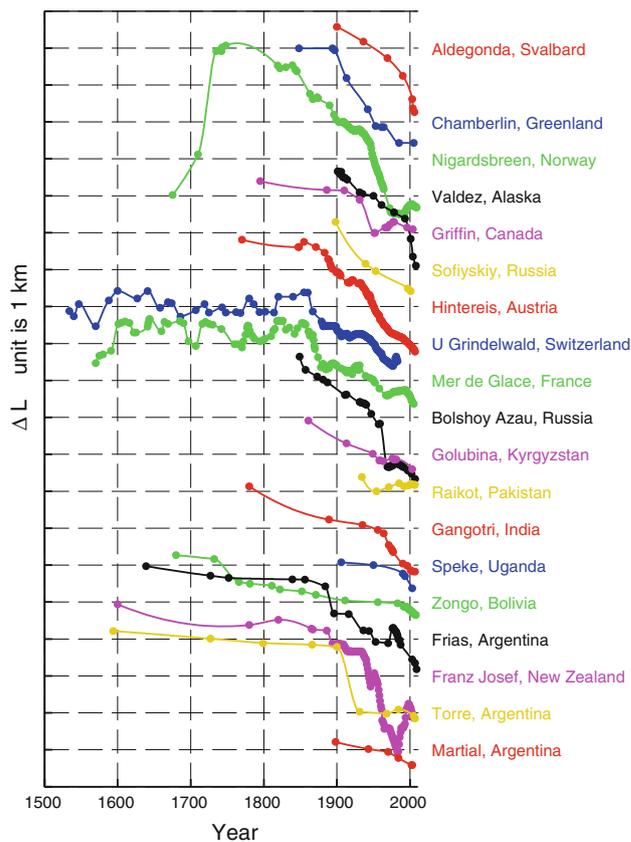
of historical sources and geological evidence (e.g. Yde and Knudsen, 2007; Rachlewicz et al. 2007). Also the tropics are represented in our data set, with records in the tropical Andes, central Africa, Himalayas and Indonesia. However, it should be noted that in central Africa no data prior to 1850 are available. Further south, the records are limited to the Central and Southern Andes and New Zealand. All three regions have a few detailed and long records as shown by the examples in Fig. 3. On the SH, almost all the data prior to 1850 rely on proxy-based dating of geomorphological features (Masiokas et al. 2009; Rabatel et al. 2008).

The maximum of 308 available records occurs between 1945 and 1965 (Fig. 2). In 1945 the first aerial photographs of the Patagonian Icefields are made (López et al. 2010), resulting in a marked increase in the number of available records from that year on. Going back in time, the number of available records shows a strong decrease in the period 1850–1900. In 1900 there are 230 records available, 156 of which are located outside the European Alps and 32 are on the SH. Only 70 records extend to 1850, of which 52 lie outside the Alps and 19 are located in the SH. In 1750 there are 34 records available (26 outside the Alps, 15 in SH). In 1650 these numbers are reduced to 14 records in total (7 outside the Alps, 5 on SH). In 1600 there are 9 remaining, of which 6 are in the Alps, 1 in Norway 1 in Southern Patagonia and 1 in New Zealand. After 1965 the number of records decreases as well. A few times the end of a record is due to the disappearance of the glacier, but more often more recent data are not available because measurements are not continued or not reported. In 1990 many of the records located in the former Soviet Union end, which leads to a notable drop of the available records. Of the 308 records, 231 continue up to at least 2000, and 181 records continue up to 2005.

Going back in time, not only the number of records decreases, but also the data points of the remaining records



**Fig. 2** The total number of records in time (red), the number of records outside the European Alps (black) and the number of records on the Southern Hemisphere (blue)



**Fig. 3** Examples of glacier length records from different parts of the world. Each dot represents a data point (for each of the records the references are given in the Supplementary Material). Data points are connected with Stineman interpolation

become scarce, as the examples in Fig. 3 show. Some length records have annual resolution in the 20th century, but before 1880 no record has annual data. This does not mean we do not have any information on the glacier length in between two measurements. Glacier advances usually leave marks in the landscape such as trim lines and moraines. In addition, glacier retreats or advances are limited to tens of metres per year (Appendix A). This limits the possible glacier stands between two measurements. The decrease in the number of data points going back in time leads nevertheless to a decrease in the resolution of the reconstructed temperature from decadal in the 20th century to about half a century in the 17th century (Appendix C). In order to be able to use the records for a temperature reconstruction, data points are connected using Stineman interpolation (Stineman 1980; Johannesson et al. 2009) to get yearly values for the entire record. Stineman interpolation creates no more maxima or minima than is required by the data. Moreover, the method performs very well when the density of data points varies strongly in time, as is the case with most glacier length records.

## 2.2 Additional glacier data

In order to assess the accuracy of the data on glacier length fluctuations, we include the method of data acquisition of each data point in the data set. Because of the large variety in methods, we use a bulk method of 4 categories: (1) direct measurements, (2) historical sources, (3) dendrochronological dating and (4) other dating methods. The category of direct measurements of the glacier terminus position includes field measurements, with or without GPS, (aerial) photography when designed for the purpose and satellite observations. The category of historical sources contains all data points derived from historical documents such as sketched maps from pioneers, pictures, paintings, written descriptions, etc. Glacier stands derived from geomorphical evidence, dated with dendrochronology are put in the third category. The fourth category includes geomorphical evidence dated with other, less accurate, methods, e.g. lichenometry and radiocarbon dating. All data points for which the method is unknown are assumed to have the same accuracy as the fourth category.

As a glacier front is at least several hundreds of metres wide, even the most accurate measurement method inherently has an uncertainty of several metres in the denoted glacier length change. Larger uncertainties result from debris cover and limited resolution of observations, making it difficult to identify the position of the glacier front accurately. We assign a maximum uncertainty of 50 m to the direct measurements (first category, Hall et al. 2003). For the reconstruction of glacier fronts from historical documents one has to relate the glacier front to features in the landscape mentioned in the documents or depicted in paintings, sketches etc. The maximum error of the thus determined front positions is 150 m. Measurement methods of the third and fourth category not only have an inaccuracy in the position, but have an additional uncertainty in the timing of the measured frontal position (e.g. Villalba et al. 1990). We translate the uncertainty in the timing into an uncertainty in position. Taking into account a typical retreat rate of 10 m per year (Appendix A), we arrive at an uncertainty of 300 m for the third category and an uncertainty of 500 m for the fourth category.

Besides the length record, we need additional information on the glacier geometry and climatic setting to reconstruct the temperature anomalies (Sect. 3) Included are: the maximum, minimum and, if available, the median elevations; the length of the main flow line in the reference year 1950; the glacier area and position; and the average annual precipitation on the glacier. The data on the glacier geometry are mostly available from either the same source as the length changes, or from the World Glacier Inventory (National Snow and Ice Data Center 1999, update 2009, <http://nsidc.org/data/g01130.html>) and WGMS databases.

The length of the flow line in 1950 is calculated from the difference in glacier length between the year of measurement and 1950. If there is no measurement in 1950, the interpolated value is used. For the precipitation, we use the climatological annual precipitation at the mean altitude of the glacier. When in situ measurements are lacking, values are estimated from climatologies (e.g. Zuo and Oerlemans 1997a) or nearby weather stations. The majority of the weather stations are situated at lower elevations than the elevation of the glacier. Therefore, an estimate of the surplus precipitation at the glacier altitude is added to the measurements.

## 2.3 Temperature data for comparison

We compare our results with the University of East Anglia (Norwich, UK) Climatic Research Unit instrumental surface-air temperature data from 1850 to 2009 (Brohan et al. 2006, <http://www.cru.uea.ac.uk/cru/data/temperature>), the combined proxy records of Mann et al. (2008), and the reconstruction of European surface temperatures of Luterbacher et al. (2004, 2007), Xoplaki et al. (2005) [hereafter referred to as Luterbacher et al. (2007)]. In both reconstructions data on glacier fluctuations have not been included.

From the available instrumental temperature records, we use the combined land ocean HADCRUT3 global and hemispheric annual mean series. These series give the temperature anomaly relative to the mean temperature of the 1961–1990 reference period. Using the KNMI climate explorer (<http://www.climexp.knmi.nl>) we calculated the mean annual temperature anomaly over 25°W–40°E and 35°N–70°N for comparison with the reconstructed European temperature. In addition, we have calculated Scandinavian temperatures from the grid points between 5–20°E and 60–70°N and the average temperature over the Alps using the grid points between 5–15°E and 45–50°N.

Mann et al. (2008) reconstruct global and hemispheric temperature anomalies for the last two millennia from a multi-proxy composite. The majority of the records are tree-ring proxies, and in addition records from marine sediment, speleothem, lacustrine, ice core, coral and historical documentary series are used. Mann et al. (2008) thus combine most of the presently available temperature proxies. They describe the results of various proxy reconstructions resulting from different methods in averaging and the exclusions of a part of the available records. Here we use the “error-in-variables” (EIV) composite of all land and ocean records.

Luterbacher et al. (2007) reconstructed European seasonal temperatures for the period 1500–2000 on a 0.5° latitude/longitude grid extending from 25°W to 40°E and from 35°N to 70°N. The reconstruction is based on

instrumental series, documentary records of sea-ice and temperature and a few tree-ring records. From 1901 to 2000 the land-only CRU temperatures are used. We calculate European annual temperature anomalies by calculating seasonal temperature anomalies relative to the 1961–1990 seasonal mean and subsequently averaging over the 4 seasons per year and over the entire grid. In addition we have calculated Scandinavian temperatures from the grid points between 5–20°E and 60–70°N and the average temperature over the Alps using the grid points between 5–15°E and 45–50°N.

To conclude, we use the HISTALP temperature record for comparison of the temperature reconstructed from the records in the Alps. The HISTALP project (<http://www.zamg.ac.at/histalp>) provides temperature, precipitation, pressure, sunshine and cloudiness for the Greater Alpine Region (GAR) 4–19°E, 43–49°N (Auer et al. 2007). The GAR temperature record goes back to 1760 and consists of homogenized (historical) instrumental measurements in and around the European Alps (Böhm et al. 2010).

### 3 Theory and methods

#### 3.1 Glaciers and climate

The response of glaciers to a changing climate is dependent on the climate setting of the glacier and on its geometry. To describe the response, two main characteristics of a glacier are needed: its response time and its climate sensitivity. These are both conceptual quantities, derived from the response of a glacier in equilibrium state to a stepwise change in climatic forcing (Jóhannesson et al. 1989). As we are dealing with glacier length changes, the response time and climate sensitivity are defined in terms of change in glacier length and temperature. The climate sensitivity is a measure for the size of glacier length change going from one equilibrium state to another as a result of a change in the climatic conditions. The response time is a measure for the time needed to approach the new equilibrium length. It is defined as the time the glacier requires to reach  $(1 - e^{-1})$  of the final length change after a stepwise perturbation of the climatic forcing.

The most relevant meteorological parameters for changes in the climatic forcing are changes in temperature and precipitation (Oerlemans et al. 1999; Greuell and Smeets 2001). In principle, glacier length variations reflect fluctuations in both temperature and precipitation. However, experiments with mass balance models show that glaciers are far more sensitive to typical changes in temperature than to typical changes in precipitation (Giesen and Oerlemans 2010; Adhikari and Huybrechts 2009; Mackintosh et al. 2002; Oerlemans 2001). We neglect fluctuations in

precipitation and assume that fluctuations in glacier length are caused by fluctuations in temperature only (see Sects. 4.3, 5 for discussion). Furthermore, we assume that the temperature signal extracted from the glacier length fluctuations represents fluctuations in the annual averaged temperature. Glaciers in a cold and dry continental climate are mostly sensitive to changes in summer temperature, but for more maritime and tropical glaciers the melt season is longer, sometimes all year round. Those glaciers are thus more sensitive to fluctuations in the annual averaged temperature (Oerlemans and Reichert 2000).

To relate the glacier length fluctuations to fluctuations in temperature, we follow the approach of Oerlemans (2005), who uses a linear first-order response equation:

$$\frac{dL'(t)}{dt} = -\frac{1}{\tau}(cT'(t) + L'(t)) \tag{1}$$

Here,  $t$  is time (a),  $\tau$  (a) is the response time of the glacier,  $c$  (m/K) the climate sensitivity,  $L'$  is length change (m) and  $T'$  is a temperature perturbation (K) (annual mean) with respect to a reference state. Oerlemans (2007) shows that for suitable values for  $\tau$  and  $c$  this relation performs very well. Rearranging the terms and replacing  $\frac{1}{c}$  by  $\gamma$  gives an expression for the temperature relative to the reference state in terms of glacier length change, response time and the inverse climate sensitivity  $\gamma$ :

$$T'(t) = -\gamma \left( L'(t) + \tau \frac{dL'(t)}{dt} \right). \tag{2}$$

#### 3.2 Calculation of $\gamma$ and $\tau$

Each glacier has a specific  $\gamma$  and  $\tau$  that characterize the response to changes in annual average temperature. In general,  $\gamma$  and  $\tau$  are dependent on the geometry of the glacier and its climatic setting. We calculate these values for each glacier using the expressions of Oerlemans (2001) that are based on an analytical glacier model:

$$\gamma = c_1 \frac{s}{\sqrt{P}} \tag{3}$$

and

$$\tau = c_2 \frac{1}{\beta s \sqrt{1 + 20s\sqrt{L}}} \tag{4}$$

Here,  $P$  ( $\text{m a}^{-1}$ ) is the mean yearly precipitation at the glacier,  $s$  is the mean slope of the glacier surface,  $\beta$  ( $\text{mwe a}^{-1} \text{m}^{-1}$ ) is the balance gradient, parametrized as  $\beta = 0.0006\sqrt{P}$  (Oerlemans 2005), and  $L$  (m) is the length of the flow line. The constants  $c_1$  and  $c_2$  are calibrated by comparing the results of numerical ice flow models with the results of the expressions derived from the analytical model (3, 4). We have to take the results of numerical models as ‘true values’ for the response time and climate sensitivity.

The response time and climate sensitivity are defined in terms of response from one equilibrium state to another, due to a stepwise change in the climatic forcing. In reality climate is never constant. Therefore glaciers hardly ever reach an equilibrium with the current climate. Moreover, climate changes not in a stepwise fashion from one constant state to another. This makes it impossible to derive the response time and the climate sensitivity of a glacier directly from observations (Oerlemans 2007).

For the calibration of the response time  $\tau$  we use the results of fifteen numerical models (Table 1). The calibration based on least squares gives a optimal value of 19.4 for  $c_2$ . The analytical model corresponds well with the results of the numerical models and there is a significant correlation of 0.83 between the response times calculated with the minimal model and the response times derived from the numerical models.

For the calibration of  $c_1$  we have the climate sensitivities ( $\frac{1}{\gamma}$ ) of fourteen glaciers, obtained with numerical models (Table 1). The best results, again based on least squares, are obtained with a value of 0.00204 for  $c_1$ . For  $\gamma$  there is a correlation between the values of numerical models and the values of the analytical model of 0.50. This correlation is only significant at the 90% confidence level (see Sect. 5 for discussion). We get different values for  $\tau$  and  $\gamma$  than Oerlemans (2005), who used the results of only 6 numerical models for the calibration of  $c_1$  and  $c_2$ .

### 3.3 Temperature reconstruction

Provided we know the precipitation, the average slope, and the absolute length of the glacier, we can reconstruct a

temperature anomaly record from a glacier length record using the Eqs. 2–4. We take the glacier state in 1950 as the reference state, because all records cover the year 1950.

Besides the length changes, Eq. 2 also contains the derivative of the length change. Our method, based on a simplified analytical glacier model, can not account for year-to-year variations in the rate of length change. It does not resolve glaciological details such as irregularities in the bed topography at the glacier tongue. To be able to apply the linear response Eq. 2 all records are smoothed (Oerlemans 2005; Oerlemans et al. 2007). For the smoothing we use the following weighted running average: for the calculation of the  $i$ th smoothed value, the weights  $w_j$  of the neighbouring values are given by

$$w_j = 10^{-\frac{|i-j|}{width}} \quad (5)$$

resulting in a smoothed value  $s_i$

$$s_i = \frac{\sum_{j=i-width}^{i+width} w_j v_j}{\sum_{j=i-width}^{i+width} w_j} \quad (6)$$

where  $v_j$  is the  $j$ th value of the record and  $width$  is the filter width. This width is set to 10 points (years), but it is decreasing when the beginning or end of the record is less than 10 points away. In this way the full record can still be used at the cost of less smoothing near its ends. The first and last point are not affected by the smoothing procedure, so the total length change remains the same after smoothing. We have tried different filter widths and different smoothing procedures, e.g. Gaussian filter, but the above mentioned method yields the optimal results in combination with the Stineman interpolation. From the interpolated

**Table 1** Values of climate sensitivity and response time obtained from the literature; if a range is given the number between brackets is used for calibration

Glacier	Response time (a)	Climate sens. (m/K)	Reference
Gr Aletsch	83	5436	Oerlemans (unpublished)
Argentière	27–45 (38)	4200	Huybrechts et al. (1989)
AX010	50–84 (60)	3616	Adhikari and Huybrechts (2009)
Brikdalsbreen	52–60 (56)	–	Laumann and Nesje (2009)
Franz Josef	20–27 (24)	3200	Oerlemans (1997a)
Hintereis	94 ± 15	4200	Greuell (1992)
Muragl	–	1700	Jouvet et al. (2008)
Nigardsbreen	63–73 (68)	6100	Oerlemans (1997b)
Pasterze	70–137 (80)	–	Zuo and Oerlemans (1997b)
Rhone	58	2600	Wallinga and van de Wal (1998)
Sofiyskiy	73–84 (77)	2000	De Smedt and Pattyn (2003)
Solheima	69	5500	Mackintosh (2000)
Storglaciären	125	3300–5400 (4500)	Brugger (2007)
Rabots	165–197 (179)	2600–4600 (3400)	Brugger (2007)
Untere Grindelwald	34–45 (38)	4500	Schmeits and Oerlemans (1997)
Vernagt	50	4667	Smith and Budd (1980)

and smoothed record we get a temperature anomaly  $T'(t)$  for each glacier length record using (2). As the length of the temperature reconstruction is the same as the length of the glacier length record, there is a wide variety in reconstruction length and covered period. To be able to compare the reconstructions from different records, the temperature anomaly of each record is expressed relative to the 1945–1965 mean, the period that is covered by all records.

### 3.4 Average of temperature records

As can be seen in Fig. 1, the records are clustered and thus do not cover large portions of the Earth's surface. Furthermore, the number of records varies in time (Fig. 2). This requires an averaging method that on the one hand takes the spatial distribution into consideration when assigning the temperature records of individual glaciers a certain weight, and, on the other hand, is flexible so it can change the assigned weights in time. For our method of estimating the uncertainty it is convenient to consider glaciers on an individual basis. We opt for a method that assigns a weight to each record based on the mutual distances between glaciers. For each year the mutual distances  $R_{ij}$  between the glaciers that have a record in that year are calculated. The weight of a particular record  $W_i$  in a year with  $N$  records is given by the ratio of the total of the squared distances of this glacier to the other  $(N - 1)$  glaciers, to the total of the squared mutual distances for all  $N$  glaciers

$$W_i = \frac{\sum_{j=1}^N R_{ij}^2}{\sum_{k=1}^N \sum_{l=1}^N R_{kl}^2} \quad (7)$$

We have compared the global average of the reconstructed temperature with the results obtained with other weight functions (see Appendix B for details). It appears that the temperature reconstruction is not sensitive to the weight function.

### 3.5 Estimate of uncertainty

The reconstructed temperatures from individual glacier length records have an uncertainty due to uncertainties in the data on length change, gaps in the glacier length records, and uncertainties in the calculated response time  $\tau$  and (inverse) climate sensitivity  $\gamma$ . We use a smooth bootstrapping procedure (Hesterberg et al. 2005) to evaluate the resulting uncertainty in the spatially averaged temperature reconstructions. We calculate the average of  $N$  reconstructed temperature records ( $N$  is 308 for the global average, 246 for the NH average, etc.). From these  $N$  records we then randomly draw with replacement 100

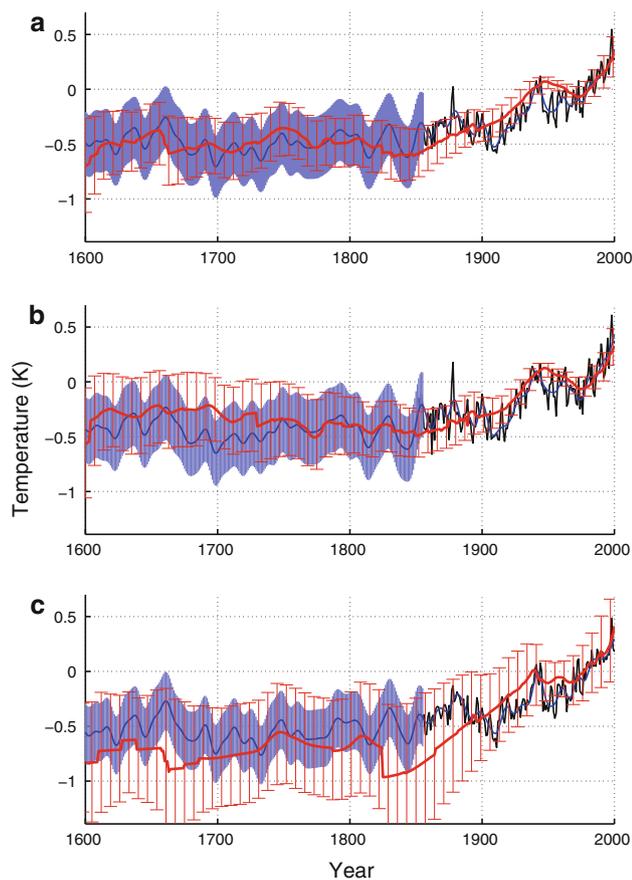
samples of containing again  $N$  records. For each of these samples the average is calculated and recalculated 10 times with perturbed values for each of the glacier length change data points, the response times and the climate sensitivities. In addition, interpolated gaps in the glacier lengths records are perturbed by adding autoregressive noise. The variance of the perturbations of the data points is dependent on the method of measurement. The variance of the perturbations of  $\tau$  and  $\gamma$  are derived from the goodness of fit between the analytical model and the literature values (Table 1). The autoregressive noise added to the records is calculated from an autoregressive model (Box and Jenkins 1976; Fisher 2002). Thus we have an ensemble of 1,100 temperature reconstructions. Twice the standard deviation of this ensemble, smoothed over 20 years, is taken to be the 95% confidence interval. See Supplementary Material for details of the uncertainty estimate.

## 4 Results

### 4.1 Global temperature anomaly

In Fig. 4a, we show the global temperature reconstructions from glacier length records, the multi-proxy global land and ocean temperature reconstruction from Mann et al. (2008) for the period 1600–2000, and the instrumental land and ocean record for the period 1850–2000. All records shown are temperature anomalies with respect to their 1961–1990 mean. The uncertainty bands give a 95% confidence interval. Mann et al. (2008) have calibrated their temperature reconstruction with the instrumental record. Therefore, this reconstruction has no uncertainty indicated for the period of the instrumental record (1850–2000). The temperature reconstruction from glacier length fluctuations has no annual resolution, inherent to the slow response of a glacier to perturbations of its mass balance. For the first period (1600–1700), the temperature record has a lower than decadal resolution, due to the lower resolution of the length records (Appendix C). Hence the temperature signal of an individual glacier length record is rather smooth in this period. The small irregularities in the global average before 1900 result from the addition to the sample of records that deviate from the average temperature. After 1900, these irregularities disappear because of the increase in the number of available records.

Our reconstruction shows a fairly constant global mean temperature for the period 1600–1830. From 1830 until 1940, temperatures continuously increased by  $0.61 \pm 0.26$  K. From 1940 to 1970, there is a temporal decrease of the global mean temperature of  $0.07 \pm 0.12$  K. After 1970, temperature is increasing leading to a cumulative warming from 1830 to 2000 of  $0.94 \pm 0.31$  K and of  $0.84 \pm 0.35$  K



**Fig. 4** Global (a), NH (b) and SH (c) temperature anomaly w.r.t. 1961–1990 mean from: HADCRUT3 instrumental record (black); Mann et al. (2008) multi-proxy with land and ocean records, shaded uncertainty (blue); glacier reconstruction of this study with 95% confidence interval bars (red)

for the period 1600–2000. The rate of temperature change over the period 1980–2000, with a linear trend of 0.16 K per decade, is the highest over the last 400 years. It is comparable with the temperature change over the period 1920–1940, which has a linear trend of 0.12 K per decade. Our reconstruction supports the conclusion of Mann et al. (2008) that the high global average temperatures of the 1990–2000 decade are unprecedented in at least the last four centuries.

For the period 1920–2000, the reconstructed temperature corresponds well with the instrumental land and ocean record, as both show the same amount of warming. However, in the reconstruction the temporal maximum in the 1940's is timed a few years later and the cooling afterwards is less abrupt than in the instrumental record. For the 19th century, there is a difference between the reconstruction and the instrumental record. The glacier records show a warming trend which starts between 1830–1840 and continues uninterrupted into the 20th century. The instrumental record shows a cooling trend at the end of the 19th century

and the start of the 20th century. Consequently, the instrumental record shows higher temperatures at the end of the 19th century and lower temperatures at the beginning of the 20th century. Before 1850, the reconstruction from the glacier records agrees very well with the multi-proxy record. Before 1650, the uncertainties of the glacier reconstruction increase rapidly as the number of available records becomes very small.

Comparison with the results of Oerlemans (2005, see Supplementary Material) shows that the present reconstruction with more data has a less pronounced warming in the period up to 1940, more in line with the instrumental record. In addition, the best estimate of the present reconstruction shows a cooler period in the mid-19th century than the best estimate of Oerlemans (2005). These differences are not caused by the difference in averaging method (Appendix B) and must thus result from the additional data. Also, the estimated uncertainty of the present reconstruction is larger, despite the increased number of records.

#### 4.2 Northern and Southern Hemisphere temperature anomaly

The reconstructed NH temperature anomalies (Fig. 4b) show results that are similar to the global reconstruction. The glacier reconstruction shows constant temperatures for the period 1600–1700, followed by a gradual cooling from 1700 to 1850. From 1850 to 1945, the NH mean temperature rises by  $0.61 \pm 0.26$  K. After a decline in temperature from 1945 to 1975 of  $0.18 \pm 0.11$  K, temperature rises until 2000. The cumulative temperature increase for the period 1850–2000 given by the reconstruction from glaciers is  $0.82 \pm 0.27$  K.

From 1600 to 1850, the temperature reconstruction from glacier length agrees very well with Mann et al. (2008) land and ocean NH reconstruction. Both Mann et al. (2008) and the instrumental record show a slight warming in the second half of the 19th century, followed by a cooling in the early 20th century. This fluctuation is not present in our reconstruction, that shows a constant warming since 1850 as in the global temperature reconstruction. In the instrumental record as well as in our reconstruction, the cooling in the middle of the 20th century is more pronounced on the NH than in the global average. To conclude, the recent strong warming is well reproduced by the temperature reconstruction from glacier length fluctuations up to 2000.

The reconstruction of SH temperature anomalies is shown in Fig. 4c. The reconstruction gives a SH temperature increase of  $1.27 \pm 0.76$  for the period 1850–2000. Despite of the much smaller number of records on the SH compared to the NH, is our SH temperature reconstruction in good agreement with the instrumental record between

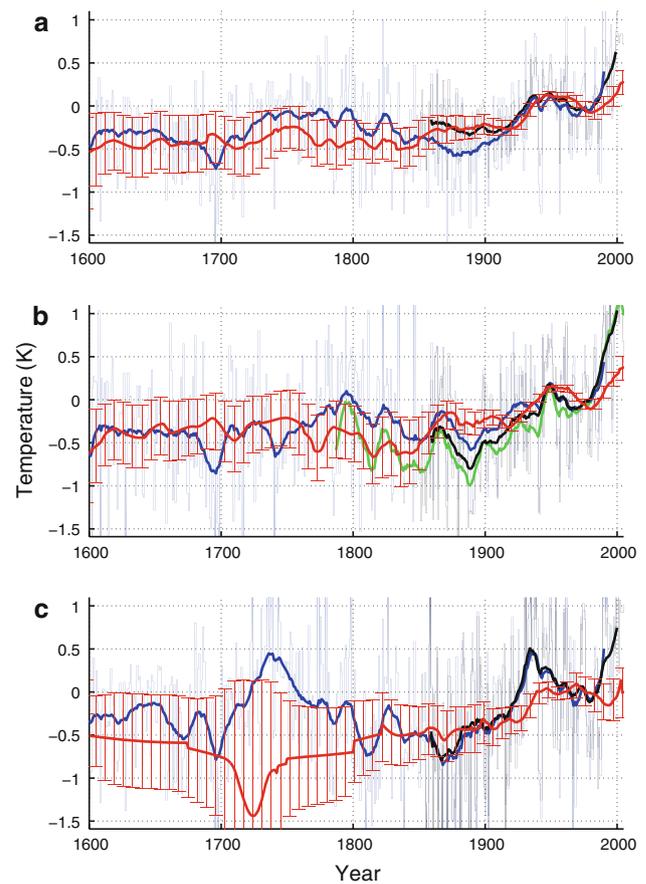
1900 and 2000. Both the instrumental records and our reconstruction show a less pronounced cooling in the middle of the 20th century than is observed on the NH, and strong warming over the last two decades of 0.13 K per decade. Between 1600 and 1825 our results are in good agreement with Mann et al. (2008). The larger error compared to the error in the NH temperature reconstruction in the period 1850–2000 is due to the smaller number of records (Fig. 2). Prior to 1850, the number of available records on the SH is comparable to that of the NH. The larger uncertainty in the reconstructed temperature of the SH for the period 1600–1850 results from larger uncertainty in glacier length data and sparser records. The clear jumps of the temperature in 1825 and around 1660 result from the addition of records to the sample.

#### 4.3 European temperature anomaly

Europe is very well represented in our data set, with ample records from different parts of the continent. Within the boundaries of 25°W–40°E and 35°N–70°N used in Luterbacher et al. (2007) we have 115 records. Most of the records (86) originate from the Alps in central Europe, but we also have records throughout Scandinavia (20), on Iceland (6) and in the Pyrenees (4) (Fig. 1). We use these records to test our assumption that the influence of variations in precipitation is of secondary importance for large-scale temperature reconstructions. In Fig. 5, we show the reconstructed temperature from two regions (Alps and Scandinavia) and the continental average. Along with our reconstruction, we have plotted the instrumental record HADCRUT3, the instrumental HISTALP record and the documentary reconstruction of Luterbacher et al. (2007).

The reconstructed temperature of the Alps (Fig. 5b) is in the second half of the 19th century higher than the instrumental records. This is probably due to glacier retreat that is attributed to a decrease in winter precipitation at the end of the LIA (Vincent et al. 2005). From the beginning of the 20th century until 1985 the reconstructed temperature is in agreement with the instrumental records. In the last two decennia the reconstructed temperature from glacier length changes does not reproduce the observed strong warming. The response in glacier length change could be delayed due to downwasting and consequent dynamical decoupling of the glacier tongue observed in the Alps (e.g. Paul et al. 2004). Prior to the start of the instrumental records, the temperature reconstructed from glaciers is in broad agreement with the reconstruction of Luterbacher et al. (2007).

The temperature reconstruction from Scandinavian glaciers (Fig. 5c) is also influenced by fluctuations in precipitation that are relatively large, due to the maritime climate of most of the glaciers in the data set. We reconstruct a cooling



**Fig. 5** Annual average European (a), European Alps (b) and Scandinavian (c) temperature w.r.t. 1961–1990 mean from: HADCRUT3 measurements with 15 years smoothed series (black); Luterbacher et al. (2007) with 15 years smoothed time series (blue); glacier reconstruction, with uncertainty bars (red). For the European Alps (b) also the 15 years smoothed HISTALP instrumental temperature record is shown (green)

in the 1990's due to glacier advances. The advances are explained by an increase in precipitation rather than a decrease in temperature (Andreassen et al. 2005). The strong warming in the 1930's is underestimated by the reconstruction from glaciers. The slight increase in precipitation during that period is likely to play only a minor role in this (Giesen 2009). The reconstruction of Luterbacher et al. (2007) and the temperature reconstructed from the glacier length variations widely diverge in the 18th century.

The reconstructed temperatures for the separate regions do not match the instrumental records. This is at least partly due to unaccounted variations in precipitation. The continental average over the temperature reconstructions from all European glacier length records is in good agreement with the European instrumental record (Fig. 5a). However, there seems to be a lag of 7 years in the warming period 1915–1940 and the cooling period around 1960. The warming in the last two decades 1980–2000 as

reconstructed from the glacier lengths is not as strong as in the instrumental record.

## 5 Discussion

The present glacier retreat is observed worldwide (WGMS 2008). All the temperature reconstructions presented in this article show a warming trend since the mid-19th century, resulting in a warming in the order of 0.5–1.0 K. Only a strong drying on a global scale could explain a worldwide retreat of the same magnitude. There is no independent evidence at all of such drying (Trenberth et al. 2007; Smith et al. 2009). Thus, only an increase in global temperature can explain the observed glacier retreat. In this study we have calculated the global and hemispherical temperature records that best explain the observed glacier fluctuations.

As argued in Sect. 3.1, the mass balance of glaciers is more sensitive to changes in temperature than to variations in (winter) precipitation. The yearly variations in mass balance are explained by variations in precipitation as well as temperature, but the decadal variations in glacier mass balance result essentially from changes in temperature and the related dynamic adjustment of the glacier size. The glacier length signal reflects the integrated mass balance over decadal timescales. Hence we believe that variations in precipitation only have a limited contribution to the glacier length variations. Furthermore, in their estimate of the contribution of glaciers to global sea level rise over a period of 70 years, Van de Wal and Wild (2001) show that the projected variations in precipitation have virtually no influence on the calculated global glacier volume changes. Nevertheless, for individual glaciers and mountain regions, variations in precipitation contribute to the explanation of the observed glacier fluctuations (e.g. Steiner et al. 2008; Fischer 2010; Andreassen et al. 2005; Vincent 2002). Hence, on a regional scale, fluctuations in precipitation cannot be neglected in the climate reconstruction from glacier fluctuations. This is evident when we compare the reconstructed temperature of the European Alps and Scandinavia to the corresponding instrumental records (Sect. 4.3).

The reconstruction of European annual temperature anomalies from glacier length is in far better agreement with the instrumental temperature record than the reconstructions of the individual regions (Fig. 5). In the pre-instrumental period the continental average is also in better agreement with the temperature reconstruction of Luterbacher et al. (2007) than the regional reconstructions. It appears that on a continental scale, there is only a small influence of variations in precipitation on the derived climate signal. We expect that this is caused by the incoherency of variations of precipitation on a large spatial scale. The average correlation distance of annual

precipitation anomalies is about four times as small as the average correlation distance of temperature anomalies (Krajewski et al. 2000; Hansen and Lebedeff 1987). We believe this justifies our assumption that we can neglect the fluctuations of precipitation in the hemispheric and global climate reconstructions from observed glacier fluctuations.

The only geometric factor included in the analytical glacier model is the average slope, while assuming a constant glacier width. Detailed information on the glacier geometry is currently lacking for most of the glaciers used in this study. However, the climate sensitivity of a glacier is strongly dependent on its geometry (Oerlemans 2001; Brugger 2007). Furthermore, we assume a climate sensitivity that is constant in time, but in reality the climate sensitivity of a glacier changes with glacier advance or retreat (e.g. Mackintosh 2000). The poor fit of calibrated climate sensitivity of the analytical model to the results of numerical models suggests that for a good estimate of the climate sensitivity, more details of the glacier geometry should be taken into account. Analysis of the contributions of the four individual causes of uncertainty to the total uncertainty estimate, shows that the uncertainty in climate sensitivity is by far the largest source of uncertainty. We expect that including a more detailed glacier geometry could greatly improve the accuracy of the reconstructed temperature from individual records.

As described in Sect. 3.3, the amount of smoothing decreases towards the end of the records. This can result in overestimation of temperature changes near the ends of the record. This effect is most important for the glaciers with a large response time (see Eq. 2). For most records, this effect of less smoothing is less strong at the start of the records. About 80% of the records have more than 1 year between the first two length measurements. The interpolated values between the first and the second measurement give a smooth record. Near present time measurements are more abundant and less than 50% of the records have more than 1 year between the last two data points. So the overestimation of the temperature anomalies of individual records occurs mostly from 1968, when the first record ends, on to 2000, when 77 records have ended. This is represented by an increasing uncertainty from 1968 to 2000 (Figs. 4, 5). To prevent this we should cut off the ends of the smoothed record. However, we think there is more to gain from the information of the entire records than we lose in noise resulting from lack of smoothing near the ends.

## 6 Conclusions

We have shown that glacier length records can provide very useful information on past temperature fluctuations. They form a reliable proxy for fluctuations of the annual averaged

temperature on decadal time scales and on hemispherical, or larger, spatial scales, reproducing the instrumental record over the last century very well. Furthermore, the reconstruction of temperature fluctuations based on glacier length changes is fully independent of other proxies. Thus, it forms a valuable addition to existing proxies.

The reconstructed temperature shows a spatially coherent signal. This makes the global average insensitive to the method of averaging. Our reconstruction shows that the global mean temperature rose by  $0.94 \pm 0.31$  K over the period 1830–2000. The warming over the period 1850–2000 is  $0.82 \pm 0.27$  K and  $1.27 \pm 0.76$  K on the NH and SH, respectively. The high global averaged temperatures of the period between 1980 and 2000 are unprecedented in at least the last 400 years. In addition, the rate of temperature increase over the period 1980–2000 is the highest of the period 1600–2000. In these respects our reconstruction supports existing proxy-reconstructions. However, compared to the multi-proxy composite of Mann et al. (2008) and the instrumental record, the glacier records show a slightly colder second half of the nineteenth century. According to our reconstruction, the rise of the global temperature and the temperatures on both hemispheres started between 1830 and 1850 and continued uninterrupted into the 20th century.

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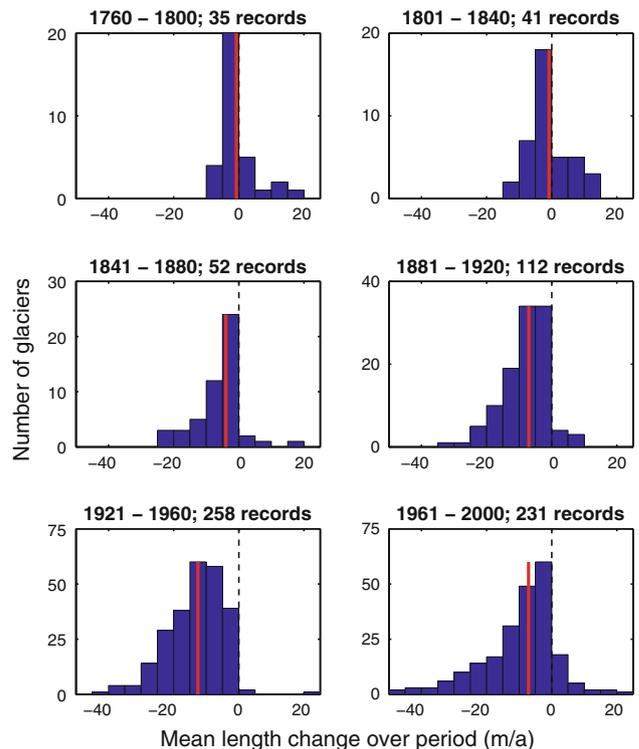
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**Appendix A: Mean glacier length change**

See Fig. 6.

**Appendix B: Average weight function**

We test the sensitivity of the temperature reconstruction to the chosen spatial averaging method. We calculate the



**Fig. 6** Mean rate of length change ( $m a^{-1}$ ) of the available glacier records for 6 forty-year periods between 1760 and 2000. The median rate of length change of all glaciers in that period is given by a red line. Note the changing number of glacier length records for the different periods

reconstructed global average temperature with three different weight methods:

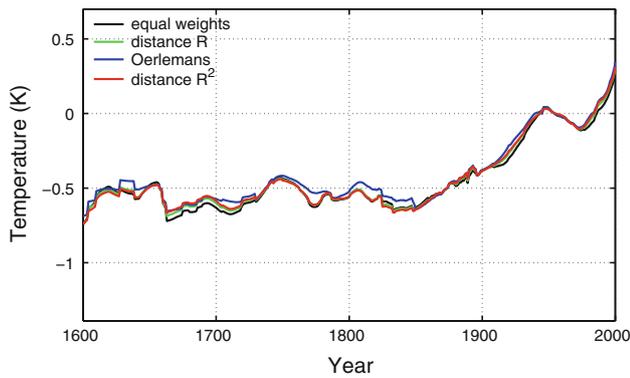
1. equal weights for all records
2. the weights of Oerlemans (2005)
3. weights of individual records based of the distance between the glaciers instead of the squared distance

and compare these results with the average obtained from the method as described in Sect. 3.4 The resulting global average temperature anomalies (relative to 1950) are shown in Fig. 7.

In the case of equal weights for all records, the average temperature anomaly in a certain year is the arithmetic mean of all available records in that year.

Oerlemans (2005) defined 4 regions on the Northern Hemisphere:

- North America, containing the glaciers of the North-western USA, the Canadian Rockies and Alaska,
- North Atlantic, including the glaciers of Greenland, Iceland, Scandinavia and Svalbard,
- the Alps and Pyrenees,
- Asia, including the Caucasus, Tien Shan, Himalyas, Altai and Kamchatka.



**Fig. 7** Average global temperature calculated with different weights: equal weights to all records (*black*); 5 weighted samples as in Oerlemans (2005, *blue*); the weights based on  $R$  (*green*); the weights based on  $R^2$  as used in this study (*red*)

For each region the average temperature is calculated without further weighting of individual records. To calculate the Northern Hemisphere average, the regions 1–4 are given different weights: 0.2, 0.3, 0.2, and 0.3, respectively. The Southern Hemisphere is not subdivided in regions, the average is the arithmetic mean of the available records. The global mean is given by the arithmetic mean of both hemispheres.

The last weighting procedure considered is similar to that used in the reconstruction (as described in Sect. 3.4). The difference is that the weight is not depended on the squared distance between the glaciers, but on the distance

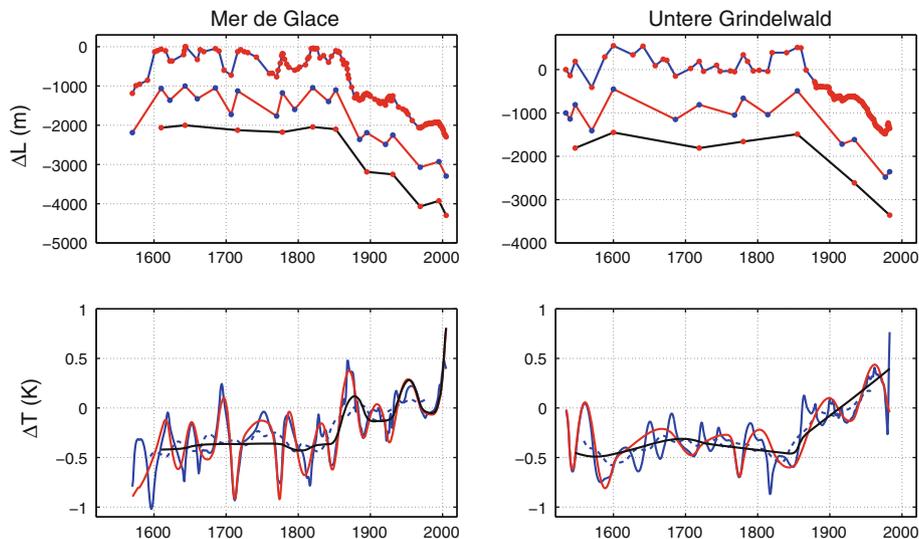
itself. Hence, the influence of the spatial distribution on the attributed weights is less strong.

The different averaging procedures result in a different weight for specific regions. For example, the cumulative weight of the records on the Southern Hemisphere is 50% in the method of Oerlemans (2005), independent on the number of records. It varies between a 32 and 53% in the average of the mutual distance methods. When we give equal weights to all records, it is less than 25% for the period 1850–2000, but more than 60% for the period 1660–1700.

The differences between the temperature reconstructions resulting from different weights methods are small, much smaller than the uncertainty in the reconstruction due to the causes discussed in Sect. 3.5 and shown in Fig. 4a. Thus, the temperature reconstruction is not sensitive to the kind of average taken, which suggests a strongly coherent temperature signal over the globe (Fig. 7).

### Appendix C: Records with sparse data

As shown in Fig. 3 most of the records have sparse data points before 1900 (if any data points at all). Very often these points are maximum stands derived from dated end moraines. This obviously influences the resolution of the temperature reconstruction and could cause a bias to lower temperatures. Here we have included an experiment with the two high resolution records of Mer de Glace (Nussbaumer et al. 2007) and Untere Grindelwald (Zumbühl



**Fig. 8** Length fluctuation records of Mer de Glace (Nussbaumer et al. 2007) and Untere Grindelwald (Zumbühl 1980) and the reconstructed temperatures. The *upper panel* shows the original length record (*blue*) and 2 derived low resolution records: (1) when only the minimum and maximum stands are taken into account (*red*) and (2) when only maximum stands are taken into account (*black*).

The derived records have been shifted for better visibility. In the *lower panel* the reconstructed temperatures from the original record (*blue*) with 51 years running average (*dashed blue*), the record with minimum and maximum stands (*red*) and the record with only maximum glacier stands (*black*) are shown

1980), shown in Fig. 8 to get a qualitative idea of the influence of sparse data on the temperature reconstruction. We have plotted the original record and two stripped versions: one using only the data points of maximum and minimum glacier stands (min-max record) and one using only the maximum glacier stands before 1900 (max-only record). The number of data points decreases correspondingly. For Mer de Glace the original record has 153 data points covering the period from 1570 to 2005, the record with only minimum and maximum glacier stands has 21 data points and the maximum-only record has 12 data points. The length record of Untere Grindelwald glacier has 128 data points in the period from 1534 to 1983, the minimum–maximum record has 15 data points and the maximum-only record has 7 data points covering the period 1547–1983.

The reconstructed temperatures of these 6 records are also shown. In the case of Mer de Glace, the temperature reconstructed from the min–max record is very similar to the original temperature reconstruction. In the min–max record of Untere Grindelwald glacier not all minimum and maximum glacier stands are used, because then the min–max record would resemble the original very much. It only has 7 points from 1600 to  $\pm 1850$ . Therefore, the reconstructed temperature misses some of the variations. After 1850 it is again very similar to the temperature reconstructed from the original record with almost annual resolution. The temperature reconstructed from the max-only record is in both cases close to the 51 years running average of the temperature reconstruction from original record.

These experiments show that we do not need glacier length records with annual resolution to reconstruct temperatures with a decadal resolution. Gaps of several decades can be bridged with interpolation as long as the extremes are known. The temperature derived from records that exist of maximum glacier stands does not necessarily have a cold bias. Records that are based on dated moraines are similar to this maximum-only records. In general, records with large gaps (up to 100 years) give us relevant climatic information, but with a low ( $\pm 50$ -year) resolution.

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