5 CLIMATE RECONSTRUCTIONS DERIVED FROM GLOBAL GLACIER LENGTH RECORDS INCLUDING A CASE-STUDY FOR EUROPEAN GLACIERS

Abstract – Glaciers have fluctuated in historic times and the length fluctuations of many glaciers are known. From these glacier length records, a climate reconstruction described in terms of a reconstruction of the equilibrium line altitude (ELA) or the mass balance can be extracted. In order to derive a climate signal from numerous glacier length records, a model is needed that takes into account the main characteristics of a glacier, but uses little information about the glacier itself. Therefore, a simple analytical model was developed based on the assumption that the change in glacier length can be described by a linear response equation. The model takes into account the geometry of the glacier, the length response time and the mass balance - surface elevation feedback. The model was tested on seventeen European glacier length records. The results revealed that the ELA of these glaciers increased on average 54 m between 1920 and 1950. The results were compared to mass balance reconstructions calculated with a numerical flowline model and derived from historical temperature and precipitation records. The findings indicate that the analytical model is useful to gain information about historical mass balance rates and ELAs. Then, we derived historic fluctuations in the ELA on the basis of nineteen glacier length records from different parts of the world. The results show that all glaciers of this global data set experienced an increase in the ELA between 1900 and 1960. Between 1910 and 1959, the average increase was 33±8 m. This implies that during the first half of the twentieth century, the climate was warmer or drier than before. The ELAs decreased to lower elevations after around 1960 up to 1980, when most of our ELA reconstructions end. These results can be translated into a global temperature increase of 0.8±0.2 K and a sea level rise of 0.3 mm a⁻¹ for the period 1910–1959.

5.1 INTRODUCTION

Glacier length records contain information on how climate has changed. This information often complements historical meteorological data, as glacier length records generally extend further back in time. Besides, glacier records are often from remote areas and higher altitudes, for which meteorological data are scarce (IPCC, 2001). Hence, glacier length records form an alternative method for climate reconstruction for periods and locations for which instrumental or proxy indicators are inadequate or contradicting. Extracting climatic information from glacier length records implies inverse modelling. Normally, the climate signal extracted from length records is represented as a mass-balance history or as a reconstruction of the equilibrium line altitude (ELA). ELA reconstructions resemble transient climate changes more properly than glacier length changes, as the glacier length is subject to the length response time and the sensitivity of a glacier.

Callendar (1950) was probably the first who attempted to extract a climate signal from the dimensions of a glacier. He presented a relation between the height of the firn line and the glacier length, including the width at the glacier snout, the glacier width and slope at the firn line altitude and a constant ratio between the accumulation and ablation area. Other simple methods to reconstruct the ELA are (I) the median elevation of a glacier (Manley, 1959), which is the elevation midway between the glacier snout and the base of the headwall, (II) the THAR (Toe-to-Headwall Altitude Ratio), which is a fraction of the height range of the glacier (Meierding, 1982), and (III) the ratio of the accumulation area to the total area (AAR) (Porter, 1975). Benn and Lehmkuhl (2000) discussed the applicability of these and other commonly used methods for different glacier types. Haeberli and Hoelzle (1995) developed a simple parameterisation scheme, build on four geometric parameters (glacier length and area, minimum and maximum elevation). They reconstructed changes in the mass balance of glaciers in the European Alps.

However, none of the methods above considers the response time of a glacier. Nye (1965) was the first who used a numerical method to infer the mass-balance history of a glacier from its length fluctuations. Oerlemans (1997), Wallinga and Van de Wal (1998) and Mackintosh and Dugmore (2000) used a numerical flowline model and the procedure of dynamic calibration to derive a mass-balance history. The advantage of a numerical flowline model is that the geometry, the climate sensitivity and the response time of each particular glacier are taken into account. However, numerical flowline models need lots of information, which is not always available. Therefore, they cannot be applied to a large number of glacier records.
The aim of this research was to develop a simple analytical model that is applicable to many glacier length records for deriving the mass-balance history of a glacier, and to use this model for the climatic interpretation of worldwide glacier length changes. The model should take into account the main characteristics of a glacier, including the response time. Nevertheless, it should only use a limited amount of information about the glacier. The model that we developed is based on the assumption that the change in length can be described by a linear response equation. Its advantage is that it takes the response time and the geometry of the glacier as well as the mass balance–surface height feedback into account. The model is described in Section 5.2.

As accurate model input data, for instance on the glacier geometry or the mass-balance regime, is not available for some glaciers, we investigated
the effects of the uncertainties in the input data on the ELA reconstruction. This sensitivity study is presented in Section 5.3.

As a start, we performed a case study and tested the analytical model for seventeen European glaciers, the length records of which are shown in Figure 5.1. At the top of the graph, glaciers from the western Alps are shown, followed by glaciers from the eastern Alps, Scandinavia and Iceland. These seventeen glaciers, which all retreated between 1850 and 1980, were selected because their length records are among the longest of all European glacier records. We calculated reconstructions of the mass balance and ELA with the analytical model and compared the results with climate reconstructions derived from other methods (Section 5.4).

We then extended this investigation to global glacier length changes and derived historic fluctuations in the equilibrium line altitude (ELA) on the basis of nineteen glacier length records from different parts of the world. So far, the climatic interpretation of worldwide glacier length fluctuations has been studied occasionally. More frequently, length fluctuations of single glaciers or glaciers in one specific region were investigated, e.g. the European Alps (Haeberli and Hoelzle, 1995), the Central Italian Alps (Pelfini and Smiraglia, 1997), Scandinavia (Bogen et al., 1989), the Patagonian Ice Fields (Warren and Sugden, 1993; Aniya, 1999), the North Cascade glaciers (Pelto and Hedlund, 2001), Northern Eurasia (Solomina, 2000), Tien Shan (Savoskul, 1997), New Zealand (Chinn, 1996) and the tropics (Kaser, 1999). On the other hand, Oerlemans (1994) compared the retreat of 48 glaciers from different regions of the world. He estimated from this a global linear warming trend of 0.66 K per century, but did not take the response time of the glaciers into account. Hoelzle et al. (2003) also investigated 90 glaciers worldwide. They estimated –from cumulative glacier length changes– a global mean specific mass balance of –0.25 m w.e. a⁻¹ since 1900. The analytical model that we used for the climatic interpretation of length fluctuations is more sophisticated than the models of Oerlemans (1994) and Hoelzle et al. (2003), as we corrected for the length response time and treated the geometry of each glacier more comprehensively.

In Section 5.5, we describe the nineteen glaciers of our global data set. The ELA and mass-balance reconstructions derived from the length records are presented in Section 5.6. The results are also interpreted in terms of changes in air temperature (Section 5.6.4). Section 5.7 contains a discussion of the results and Section 5.8 a summary.
5.2 THE ANALYTICAL MODEL

5.2.1 LINEAR RESPONSE EQUATION

A glacier responds to a change in the mass balance (and the ELA) by changing its length. The size of the length change depends on several factors, such as the geometry, the slope and the mass-balance profile of the glacier. For relatively small length fluctuations compared to the total glacier length, it is assumed that the change in glacier length can be described by a linear response equation:

\[
\frac{dL'(t)}{dt} = -\frac{cE'(t) + L'(t)}{t_{rL}}
\]  

(5.1)

where \(L'(t)\) is the glacier length with regard to a reference length \((L_0)\) (m), \(t\) is time (a), \(c\) is the climate sensitivity (–), \(E'(t)\) is the ELA with regard to a reference altitude \((E_0)\) (m) and \(t_{rL}\) is the length response time of the glacier (a) (see Figure 5.2 for explanation of the symbols). The concept of this analytical model was first put forward by Oerlemans (2001).

![Figure 5.2: Schematic outline of a glacier, showing some of the parameters of the analytical model.](image)

The reference length, \(L_0\), is calculated as the mean of a glacier length record. The climate sensitivity \((c)\) is a factor that relates a change in the
glacier’s steady-state length to a change in the ELA. For large length fluctuations, the analytical model is not valid because the glacier geometry can change significantly when the glacier retreats or advances over a large distance. As a result, the response time and the climate sensitivity cannot be kept constant.

The inverse of the linear response equation can be used to calculate the historic ELAs:

$$E'(t) = -\frac{1}{c} \left( L' + t_{rl} \frac{dL'}{dt} \right)$$ (5.2)

If $E'(t)$ is multiplied by the mass-balance gradient at the equilibrium line altitude ($\beta_L$), this equation can be used to calculate a mass-balance reconstruction ($B'(t)$). Methods to estimate the length response time and the climate sensitivity are given in Section 5.2.2. First, how the time derivative of a length record can be calculated is described.

Generally, glacier length records are not smooth records (see Figure 5.1). Linear interpolation between the observed data points is applied, assuming that the glaciers have not fluctuated substantially during the periods between these points. It is likely that they were not larger during the period in between the data points, otherwise moraines would have been deposited. However, glaciers could have been smaller during these periods. Taking the time derivative from these length records, which are needed for Equation (5.2), is not a straightforward exercise because the derivatives will be discontinuous at the data points. To obtain smooth time derivatives, we tried to fit polynomials to the length records and to calculate the time derivative of these polynomials. However, we found that not every length record can be represented well by a polynomial fit of a certain degree. We then used Fourier functions to describe the length records and calculate the time derivatives. The advantage of Fourier series is that this method allows separation of time scales. Nevertheless, most length records only show a retreat in length and little fluctuations. The results revealed that these length records especially are difficult to describe by Fourier decompositions. The use of cubic spline interpolation leads to smooth length records. However, this method often causes an increase in the maxima of the length records and was therefore rejected.

We concluded that a better solution would be to apply a Gaussian filter to the length records and to calculate the time derivatives with central differences. A Gaussian filter is a weighted average, and the weight for each year ($w_i$) is defined by a Gaussian function. The filtered length record $L'(t)$ can be calculated with:
\[ L'(t) = \frac{\sum_{-N}^{N} w_i' \Lambda'(t + i)}{\sum_{-N}^{N} w_i} \]  

(5.3)

where \( \Lambda'(t) \) is the original length record, as shown in Figure 5.1. The Gaussian function is defined as:

\[ w_i = e^{-i^2/\tau^2} \]  

(5.4)

where \( \tau \) is the timescale. As we concentrated on fluctuations in the ELA occurring on a decadal timescale, the timescale of the Gaussian filter was taken as 10 years. Accordingly, \( N \) was taken as 15 years. Figure 5.3 gives an example of the filtered length record and its time derivative of Hintereisferner. The time derivative does not show any discontinuities at the data points.

![Figure 5.3: The length record of Hintereisferner filtered with the Gaussian filter and the time derivative of the length record (dashed line). The black dots indicate the individual data points.](image)

### 5.2.2 Calculation of the Length Response Time and the Climate Sensitivity

The length response time and the climate sensitivity of a glacier can be determined by a numerical flowline model, as done previously for Hintereisferner (Greuell, 1992), Pasterzenkees (Zuo and Oerlemans, 1997a), Unterer Grindelwaldgletscher (Schmeits and Oerlemans, 1997), Rhonegletscher (Wallinga and Van de Wal, 1998), Nigardsbreen (Oerlemans, 1997), Glacier
d'Argentière (Huybrechts et al., 1989) and Sólheimajökull (Mackintosh, 2000). However, we prefer a simpler method based on a perturbation analysis on the continuity equation (Oerlemans, 2001). After Reynolds decomposition and neglecting higher-order terms of the continuity equation for a glacier volume, a perturbation equation is obtained:

$$\frac{dV'}{dt} = \frac{dA'}{dt} + A_0 \frac{d\bar{H}'}{dt}$$

(5.5)

where $V'$ is the glacier volume with regard to the reference volume ($m^3$), $\bar{H}'$ is the mean glacier thickness with regard to the reference thickness ($\bar{H}_0$) ($m$), and $A'$ is the glacier area with regard to the reference area ($A_0$) ($m^2$). The reference situation of the glacier is thus defined by $L_0$ and $A_0$.

According to Equation (5.5), a change in glacier volume is directly coupled to a change in the area and the thickness of a glacier.

It is assumed that a change in the glacier area is only related to a glacier length fluctuation. The corresponding volume change is then calculated by multiplying the length change by the width and the thickness of the glacier tongue. The width of the glacier tongue is assumed to be constant. If we also assume that the change in glacier thickness is proportional to the change in glacier length, the volume change of a glacier can be written as:

$$\frac{dV'}{dt} = (w_f \bar{H}_f + \eta A_0) \frac{dL'}{dt}$$

(5.6)

where $\eta$ is a constant that relates the mean glacier thickness to the glacier length, $w_f$ is the characteristic width ($m$) and $\bar{H}_f$ the characteristic thickness of the glacier tongue ($m$) (see Figure 5.2).

Changes in the glacier volume are caused by changes in the mass balance, described by:

$$\frac{dV'}{dt} = A_0 B' + A_0 \beta \bar{H}' + w_f B_f L'$$

(5.7)

The first term on the right hand side is the volume change caused by a perturbation of the mass-balance rate, $B'$ ($m$). $B'$ can be expressed in a change in the ELA ($E'$) by dividing $B'$ by the mass-balance gradient at the equilibrium line altitude ($\beta_E$). The second term is a volume change resulting from the feedback between the mass balance and the surface elevation of the glacier. $\beta$ is the average mass-balance gradient over the glacier and $\bar{H}'$ can be described as $\eta L'$. The last term represents a volume change due to a
change in glacier length, where $B_f$ is the melt rate at the glacier terminus (m). Combining and rewriting Equation (5.6) and (5.7) yields:

$$\frac{dL'}{dt} = \frac{\beta E_0}{\eta A_0 + w_f H_f} - \frac{B_f}{\eta A_0 + w_f H_f} L' + \frac{\eta \beta A_0 + w_f B_f}{\eta A_0 + w_f H_f} E'$$  \hspace{1cm} (5.8)

Comparing this equation with Equation (5.1), the length response time and the climate sensitivity can be derived:

$$t_{rL} = -\frac{\eta A_0 + w_f H_f}{\eta \beta A_0 + w_f B_f}$$  \hspace{1cm} (5.9)

$$c = \frac{\beta E_0}{\eta \beta A_0 + w_f B_f}$$  \hspace{1cm} (5.10)

We calculated the climate sensitivity and the length response time of each glacier with these equations. It should be noted that Equation (5.9), the expression for the length response time, also holds for the volume response time because volume and length changes are coupled (Equation (5.6)). Normally, the volume response time is shorter than the length response time because the glacier volume is more directly affected by changes in the mass balance (Oerlemans, 1997). Therefore, length response times calculated with Equation (5.9) are expected to be shorter than the real length response times.

If the mass balance – surface elevation feedback is not taken into account (i.e. $\eta = 0$), the length response time (Equation (5.9)) corresponds to the volume time scale derived by Jóhannesson et al. (1989). The volume time scale of Jóhannesson is always shorter than the length response time calculated with Equation (5.9). Furthermore, the climate sensitivity decreases if the mass balance - surface elevation feedback is discounted. If we also assume that the glacier’s width is constant along the glacier, the climate sensitivity corresponds to the often-used expression reported in Paterson (1994):

$$\frac{dL'}{dB'} = \frac{L_0}{B_f}$$  \hspace{1cm} (5.11)

However, most glaciers do not have a uniform width. Normally, the glacier tongue is narrower than the mean glacier width. In that case, Equation (5.11) underestimates the climate sensitivity.
We obtained most input data to run the model from Haeberli et al. (1998) and from topographical maps. $\beta_E$ and $\beta$ were calculated from mass-balance measurements. However, mass-balance measurements were not available for all glaciers, and if absent, the mean of the mass-balance gradients of nearby glaciers was used. The thickness of the glacier snout ($H_f$) was derived from glacier slope and length. If we assume a constant driving stress, the ice thickness at any point on the glacier can be calculated from the surface slope (Paterson, 1994). The surface slope multiplied by the glacier thickness is then a constant, and the thickness of the glacier snout follows:

$$H_f = \frac{\overline{H} \cdot s}{s_f}$$

(5.12)

where $s_f$ is the surface slope at the glacier snout, estimated from a topographical map, and $s$ is the average surface slope of the glacier. We derived $s$ from the maximum elevation, the minimum elevation and the length of the glacier. $\overline{H}$ is the mean glacier thickness, which we calculated from an expression proposed by Oerlemans (2001):

$$\overline{H} = \left( \frac{\mu L}{1 + \nu s} \right)^{1/2}$$

(5.13)

where $L$ is the glacier length (m) and $m$ and $n$ are constants (~9 m and ~30 respectively) determined by a numerical model (Oerlemans, 2001). Taking the derivative of Equation (5.13) yields an expression for $\eta$:

$$\eta = \frac{d \overline{H}}{dL} = \frac{1}{2} \left( \frac{\mu}{(1 + \nu s)L} \right)^{1/2}$$

(5.14)

A typical value for $\eta$ is 0.006.

This analytical model thus takes into account the length response time of the glacier. This implies that the ELA reconstruction shifts backward in time as the length response time increases, which also influences the amplitude of the ELA fluctuations. The geometry of the glacier is included in the model, mainly defined by the width of the glacier tongue and the glacier area. Glaciers with relatively small glacier tongues compared to the total surface area will therefore have larger climate sensitivities and response times. Furthermore, the effect of the mass balance - surface elevation feedback is included. For steep glaciers, $\eta$ is small (the surface elevation changes little when the glacier length changes), implying a weak
mass balance - surface elevation feedback, resulting in shorter length response times and lower climate sensitivities.

5.3 PARAMETER SENSITIVITY

Climate sensitivity and length response time are determined by eight parameters. These parameter values may contain errors because some of them are unknown or difficult to estimate. We therefore investigated the effect of a change in these parameters on length response time and climate sensitivity and, finally, on the ELA reconstruction of a fictitious glacier. Table 5.1 lists the parameter values of this fictitious glacier. The values are supposed to represent an average valley glacier and lead to a length response time of 85 years and a climate sensitivity of 70.

Table 5.1: Parameters of the fictitious glacier.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$L_0$</td>
<td>15 km</td>
</tr>
<tr>
<td>$A_0$</td>
<td>20 km$^2$</td>
</tr>
<tr>
<td>$s$</td>
<td>10$^\circ$</td>
</tr>
<tr>
<td>$s_f$</td>
<td>10$^\circ$</td>
</tr>
<tr>
<td>$w_f$</td>
<td>500 m</td>
</tr>
<tr>
<td>$B_f$</td>
<td>-5 m w.e. a$^{-1}$</td>
</tr>
<tr>
<td>$\beta$</td>
<td>0.005 m w.e. a$^{-1}$ m$^{-1}$</td>
</tr>
<tr>
<td>$\beta_E$</td>
<td>0.007 m w.e. a$^{-1}$ m$^{-1}$</td>
</tr>
</tbody>
</table>

We changed each of these parameters by 30%. The resulting length response times and climate sensitivities are plotted in Figure 5.4. They are most sensitive to changes in the melt rate at the glacier terminus, the width of the glacier tongue and the mass-balance gradient at the equilibrium line altitude. Also, the glacier area seems to influence the value of $tr_L$ and $c$ substantially. Nevertheless, we do not expect the surface area to be an important contributor to the uncertainties in $tr_L$ and $c$, as it is often known within 10%. An increase in the melt rate at the glacier tongue or an increase in its width both lead to shorter response times, as both of them cause a larger mass turnover. However, an increase in the width of the glacier tongue also implies that more ice needs to be transported down the glacier to make up for an equal change in glacier length. This counterbalances the effect of a larger mass turnover on $tr_L$. Therefore, the length response time is more sensitive to melt rate than to width of the glacier tongue. Regarding climate sensitivity, changes in glacier width or melt rate at the glacier terminus lead to the same effect. Generally, $tr_L$ and $c$ change into the same direction when one of the input parameters is varied.
As a second step, we determined the effect of a change in length response time and climate sensitivity on the ELA reconstruction of this fictitious glacier. We derived historic ELAs from a (fictitious) length record, which we defined as a sine function with an amplitude ($L'_{\text{max}}$) of 1.5 km and a period ($P$) of 300 years. Instead of smoothing this record with a Gaussian filter and calculating the derivatives with central differences, we directly inserted this function and its time derivative in Equation (5.2). This leads to:

$$E'(t) = -\lambda \sin \left(2\pi \frac{t + \varphi}{P}\right)$$  \hspace{1cm} (5.15)

where $\lambda$ and $\varphi$ are the amplitude and the phase difference of the ELA reconstruction (see Figure 5.5) and are described by:

$$\lambda = \frac{1}{c} \sqrt{1 + \left(2\pi \frac{tr_L}{P}\right)^2 L'_{\text{max}}}$$  \hspace{1cm} (5.16)
\[ \varphi = \frac{1}{4} P - \frac{P}{2\pi} \arctan \left( \frac{P}{2\pi r_L} \right) \]  

(5.17)

Figure 5.5 shows the length and the reconstructed ELA as function of time. The amplitude of the ELA reconstruction thus depends on the period of the length record, the length response time and the climate sensitivity. The reconstruction experiences a phase difference depending on the length response time. For very large length response times, the phase difference approaches \(0.25P\).

![Figure 5.5: Length record and reconstructed ELAs of the fictitious glacier. The length response time is 85 years and the climate sensitivity 70.](image)

We varied the length response time between 65 and 123 years and the climate sensitivity between 53 and 101. These numbers are the limits set by a change of 25\% (plus or minus) in the most sensitive parameter, \(B_j\) (Figure 5.4). The effects on the amplitude and phase difference of the ELA reconstruction are plotted in Figure 5.6. The amplitude varies roughly from 30 to 59 m and the phase difference from 18 to 30 years. Note that an increase in \(r_L\) leads to larger amplitudes and an increase in \(c\) to lower amplitudes. The total effect on the ELA reconstruction constitutes the sum of both, and is therefore smaller than the individual effects. Figure 5.6c shows that the total effect on the amplitude of the ELA reconstruction is at maximum 10\%, when the melt rate at the glacier tongue is varied by 25\%.

Based on this sensitivity test, we conclude that the length response time and the climate sensitivity are most sensitive to uncertainties in the width of the glacier tongue, the melt rate at the glacier snout and the mass-balance gradient at the equilibrium line altitude. The effect of these
uncertainties on the final ELA reconstruction is, at most, 10% in the amplitude, and a phase difference of 6.

Figure 5.6: Amplitude (solid) and phase (dashed) of the ELA reconstruction of the fictitious glacier, when $\tau_r$ is varied and $c$ kept at 70 (a), when $c$ is varied and $\tau_r$ is kept constant at 85 (b), and when $B_f$ is varied by 25% (c).

Of course, this test provides no evidence that an uncertainty in the input parameters always leads to small changes in the ELA reconstructions because other glaciers may have different input parameters and length records than our fictitious glacier. Therefore, we also carried out a sensitivity test for Nigardsbreen, a glacier for which we have a long length record. We investigated the effect on the ELA reconstruction when the most sensitive parameters ($B_f$, $\omega$, and $\beta_E$) were changed within their uncertainty ranges (Figure 5.7). This test indicates that the uncertainty in the width of the glacier tongue has the largest impact on the ELA reconstruction of Nigardsbreen, leading to a standard deviation of 4 m in $E'(t)$, and a
maximum deviation in the ELA of 9 m for the ELA maximum, which occurred in 1978.

![ELA reconstruction for Nigardsbreen for different values of the input parameters.](image)

Figure 5.7: ELA reconstruction for Nigardsbreen for different values of the input parameters.

5.4 CASE STUDY: EUROPEAN GLACIERS

We tested the analytical model on length records of seventeen European glaciers (Figure 5.1). The length response times and climate sensitivities calculated from Equations (5.9) and (5.10) were compared to values derived from numerical flowline models (Section 5.4.1). The ELA reconstructions, which are presented in Section 5.4.2, were compared to reconstructions derived from numerical models and from temperature and precipitation records (Section 5.4.3).

5.4.1 LENGTH RESPONSE TIME AND CLIMATE-SENSITIVITY RESULTS

Table 5.2 shows the length response times and the climate sensitivities for the seventeen European glaciers calculated with the analytical model. For some glaciers the climate sensitivity, the length response time and the volume response time calculated with numerical models are given. The analytical length response times were expected to be shorter than the numerical length response times because the analytical length response time is in fact a volume response time (see Section 5.2.2). However, the results indicate that the analytical length response times correspond better
to the numerical length response times than to the numerical volume response times. Hintereisferner is an exception because the numerical length response time of Hintereisferner (94 a) is much longer than the analytical length response time (62 a).

The differences between the numerical and analytical values can be associated with topographical effects. For instance, the climate sensitivity increases if the glacier tongue reaches a point where the valley is narrow or the bed slope small. A numerical model takes into account these topographical features, whereas the analytical method assumes a constant bed slope and a constant width of the glacier tongue.

Table 5.2: Climate sensitivities and length response times calculated by the analytical method (ana) and by numerical models (num) and volume response times calculated by numerical models for several European glaciers; the numerical values are taken from literature referenced in this chapter.

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Climate sensitivity</th>
<th>Length response time (a)</th>
<th>Volume response time (num) (a)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ana</td>
<td>num</td>
<td>ana</td>
</tr>
<tr>
<td>Glacier d’Argentière</td>
<td>25</td>
<td>35</td>
<td>33</td>
</tr>
<tr>
<td>Mer de Glace</td>
<td>57</td>
<td></td>
<td>56</td>
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<tr>
<td>Bas Glacier d’Arolla</td>
<td>16</td>
<td></td>
<td>48</td>
</tr>
<tr>
<td>Rhonegletscher</td>
<td>31</td>
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<tr>
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<td>Forni</td>
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<tr>
<td>Svinafellsjökull</td>
<td>20</td>
<td></td>
<td>60</td>
</tr>
</tbody>
</table>

5.4.2 Reconstructions of the Equilibrium Line Altitude

The reconstructed equilibrium line altitudes of the seventeen European glaciers are shown in Figure 5.8. The ELAs of most glaciers show a similar pattern: after 1850 the ELAs increase, indicating a warmer period or less snowfall. At the beginning of the twentieth century, there is a small
decrease in most ELAs. After that, the ELAs increase until around 1950 and then decrease slightly. The ELA reconstructions of Unterer Grindelwaldgletscher and Glacier d’Argentière are amongst the longest records and show very similar fluctuations. However, the amplitudes of the fluctuations differ.

The ELA reconstruction of Rhonegletscher shows a large increase between 1850 and 1870 compared to the other reconstructions. This strong increase is probably an artefact of the analytical model, which uses a constant climate sensitivity and a constant length response time. Between 1850 and 1870, the terminus of Rhonegletscher rested on a small sloping surface, much smaller than the mean slope, implying that the glacier length was actually very sensitive to a change in the ELA. If a larger climate sensitivity was applied to Rhonegletscher, a smaller increase in the ELA

![Diagram of reconstructed ELA records of seventeen European glaciers calculated with the analytical model.](image)

*Figure 5.8: Reconstructed ELA records of seventeen European glaciers calculated with the analytical model.*
would have been calculated.

Hintereisferner reveals two very steep increases in the ELA after 1850, interrupted by significantly lower ELAs at the beginning of the twentieth century. The total increase in the ELA of Hintereisferner is rather large compared to other ELA reconstructions of glaciers in the Alps. The analytical model is probably less valid for Hintereisferner because the relative decrease in length over the total period of Hintereisferner is large: 30%. The other glaciers in the Alps retreated on average 18%.

Sólheimajökull’s high ELAs around 1760 and low ELAs just before 1800 are necessary to explain its enormous retreat before and growth after 1780. Svinafellsjökull, however, does not show low ELAs before 1800. Still, the large ELA fluctuations of Sólheimajökull fit with documented changes in the Icelandic climate and sea ice extent (Ogilvie, 1992): during the 1780s, sea ice remained unusually close to Iceland. Besides, the large ELA fluctuations are confirmed by ELA reconstructions of Mackintosh and Dugmore (2000) that were calculated for Sólheimajökull with a numerical flowline model.

The ELA reconstructions of glaciers located in the same area exhibit resemblances. Therefore, means of the ELA reconstructions of the Icelandic and the Scandinavian glaciers and the glaciers of the western and eastern Alps were calculated (Figure 5.9). The eastern part of the Alps was separated from the western part by the 9˚E meridian. Means were calculated over the period 1872–1974 because ELA reconstructions of all glaciers were performed for this period. The mean ELAs show an increase

![Figure 5.9: Means of the reconstructed ELA records of Scandinavia, Iceland, the eastern and the western Alps.](image-url)
after 1915. This is 45 m for the Icelandic glaciers, 59 m for the Scandinavian glaciers, 47 m for the western Alps and 64 m for the eastern Alps. Apparently, air temperatures and snowfall have changed more in the eastern Alps than in the western Alps. The ELAs in northern Europe reached a maximum before 1950 and then decreased, unlike the ELAs of the Alps, which were at maximum after 1950 and show a smaller decrease after that. Furthermore, it is striking that the ELAs in the western Alps were also at a minimum before 1900, unlike the other ELA reconstructions.

5.4.3 COMPARISON OF THE RESULTS WITH OTHER RECORDS

It would be ideal to compare the ELA reconstructions with long records of ELA or mass-balance observations. However, these long observation records do not exist. Therefore, the results of the analytical model were compared with different types of data: results from a simple method and from a numerical model and from historical temperature and precipitation records.

Haeberli and Hoelzle (1995) calculated the average change in the mass balance over 1850–1970 from 13 length records of glaciers in the Alps. They assumed that the glaciers were stationary at the beginning and at the end of this period, and between 1890 and 1925. Subsequently, they supposed that a full dynamic response to a step change in the mass balance would explain the glaciers’ retreat over these period. They then calculated an average mass loss of –0.33±0.09 m w.e. a⁻¹ during 1850–1970. Mass-balance changes over the same period were also calculated for the glaciers in the Alps with the analytical model. The average of the mass-balance changes resulted in –0.21 ± 0.16 m w.e a⁻¹, which is smaller compared to the step change in mass balance calculated by Haeberli and Hoelzle. The difference between the results of the two methods could be explained by the different approaches. First, the analytical model takes into account the response time and does not assume steady state situations before 1850 and after 1970. Second, Haeberli and Hoelzle use Equation (5.11) to determine the climate sensitivity of a glacier, which underestimates the climate sensitivity.

Figure 5.10 shows how the analytical ELA reconstruction of Nigardsbreen compares to the ELA reconstruction calculated from a numerical model (Oerlemans, 1997). The variation in the ELA reconstructions is of the same size, but there is a phase difference between the two reconstructions. Although the time at which the ELAs start to decrease or increase is similar, the locations of the maxima and minima differ. Firstly, this difference could be due to the numerical model, which uses a succession of step functions. Nine step functions were used to obtain the numerical ELA reconstruction as shown in Figure 5.10. Increasing the number of step functions will certainly influence the ELA reconstruction and probably lead to a reconstruction that is closer to the analytical
reconstruction. Secondly, the analytical model is based on the assumption that the glacier length directly responds to a change in the ELA. However, from numerical studies it is clear that glacier volume does indeed respond immediately to a change in the ELA, but glacier length starts reacting somewhat later. Therefore, in the analytical model, a change in the ELA is closer followed by a length change. In the numerical model, it takes more time for the glacier length to respond to a change in the ELA. The analytical ELA reconstruction is accordingly shifted forward in time compared to the numerical reconstruction. Thirdly, numerical models may respond too slowly to a mass-balance change due to the grid point spacing, which was 100 m for Nigardsbreen.

Comparing a mass-balance reconstruction with climate records is difficult because changes in the mass balance can be due to fluctuations in temperature, precipitation, sunshine duration or solar radiation. A method to derive a mass-balance history from historical meteorological data is using seasonal sensitivity characteristics (Oerlemans and Reichert, 2000). A seasonal sensitivity characteristic is the dependence of the mass balance on monthly anomalies in temperature and precipitation. Figure 5.11 shows the seasonal sensitivity characteristics of Griesgletscher and Rhonegletscher calculated by Oerlemans and Reichert (2000) from a mass-balance model of Oerlemans (1992). The figure illustrates that the mass balance of Rhonegletscher is more sensitive to changes in precipitation than the mass balance of Griesgletscher. Furthermore, a temperature increase in winter does not change the mass balance of Griesgletscher and hardly influences the mass balance of Rhonegletscher.
Figure 5.11: Seasonal sensitivity characteristics of Griesgletscher (a) and Rhonegletscher (b) calculated from a mass-balance model (Oerlemans and Reichert, 2000).

Figure 5.12: Mean mass-balance reconstruction of glaciers in the western Alps (dashed) and mass-balance reconstructions calculated with seasonal sensitivity characteristics of Griesgletscher (dotted) and Rhonegletscher (solid). A Gaussian filter is applied to the mass-balance reconstructions.

Mass-balance reconstructions of these two glaciers in the Alps were calculated from the seasonal sensitivity characteristics and compared with the reconstructions calculated with the analytical model. Therefore, long records of monthly precipitation and temperature anomalies were needed. The monthly precipitation of Beatenberg (Switzerland) and the monthly homogenised high-elevation (above 1500 m a.s.l.) temperature record of 46°
N, 8° E taken from Böhm et al. (2001) were used for both glaciers. The mass-
balance reconstructions calculated from the seasonal sensitivity
characteristics were then filtered with a Gaussian filter.

The mass-balance reconstructions of Griesgletscher and Rhone-
gletscher and the mean mass-balance history of the glaciers of the western
part of the Alps calculated with the analytical model are shown in Figure
5.12. The mass-balance reconstructions show similar fluctuations with
amplitudes of similar magnitude. However, there is again a phase
difference between the reconstructions. The analytical mass-balance
reconstruction is shifted about 10 years forward in time, which is less than
the phase difference between the analytical and the numerical model
results.

5.5.4 Conclusions of the case-study

Application of the analytical model to seventeen European glaciers
shows that the model is useful to derive a climate signal from a glacier
length record. The calculated mass-balance fluctuations are in agreement
with mass-balance reconstructions derived from numerical models and
from temperature and precipitation records using seasonal sensitivity
characteristics. The length response times and climate sensitivities
calculated with the analytical model are in agreement with values
calculated from numerical models.

However, the ELA reconstructions calculated with the analytical
model are shifted forward in time by a decade compared to the numerical
mass-balance reconstruction and the mass-balance reconstructions
calculated from temperature and precipitations records. This is due to the
model assumption that glacier length immediately responds to a change in
the mass balance. Another limitation of the model is that topography is
represented in a schematic way. The topography is especially important
when the slope of the bedrock and the valley width change along the
glacier. Then, the climate sensitivity and the length response time depend
considerably on the position of the terminus, which in turn will influence
the mass-balance reconstruction. Therefore, the analytical model is not
suitable for glaciers with strong variations in the slope and width of the
glacier valley.

5.5 Global glaciers and their length records

Building a data set of useful length records of worldwide glaciers was
difficult because most non-European glaciers have only been measured
since the beginning of the twentieth century. We aimed to find records that
go back to before 1900. A second demand of our data set was that the length
fluctuations needed to be small ($< \sim 20\%$) compared to the total glacier length, to justify the assumption of a quasi-linear response. Thirdly, the glacier type was bound to be valley or outlet. Lastly, we needed information about the geometry and the mass balance to calculate the response time and the climate sensitivity. This led to a collection of fourteen glaciers from countries other than European, to which we added five European records to obtain a global picture (Figure 5.13).

The glaciers of our data set are located in six regions: Canada, U.S.A., South America, Europe, Asia and New Zealand. All of them have retreating over the previous century, but 60% of them also slightly advanced.

Figure 5.13: Glacier length fluctuations of the nineteen glaciers of our data set.
Table 5.3: Model parameters of the nineteen glaciers, the calculated length response time ($t_rL$), climate sensitivity ($c$), and the mean annual precipitation (Zuo and Oerlemans, 1997b). The first five letters of their names abbreviate the glacier names. (L₀: reference glacier length, A₀: reference glacier area, s: mean surface slope, s_f: surface slope of the glacier snout, w_f: width of glacier tongue, B_f: melt rate at the glacier terminus, β_E: mass-balance gradient at the ELA, β: mass-balance gradient of the glacier)

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after 1950 (Figure 5.13). Retreat rates, total retreat and periods of retreat vary widely between the glaciers. Table 5.3 gives information about the geometry and the mass-balance regime of the glaciers. These parameter values were used as model input. The length records and the input data were either collected or estimated from data of the World Glacier Monitoring Service (Haeberli et al., 1998), the World Glacier Inventory, from Dyurgerov (2002) or from other references mentioned in the text.

### 5.5.1 Canada

The three Canadian glaciers of the data set are Athabasca (Rockies), Clendenning and Havoc (Coast Mountains). Athabasca is an outlet glacier.
of the Columbia Ice Field, an ice cap of roughly 325 km². The glacier flows over three icefalls into an alpine valley (Reynolds and Young, 1997). It has retreated rapidly since 1910 until 1960, after which a slowdown in the retreat rate occurred. Clendenning and Havoc, two valley type glaciers, are located closer to the ocean and their glacier termini reach further down (±1000 m a.s.l.) than Athabasca (±1950 m a.s.l.). The length records of Clendenning and Havoc do not show similar patterns, although both are located in the Clendenning valley. This could be attributed to differences in glacier geometry because the glacier tongue of Havoc is much steeper and narrower.

5.5.2 U.S.A.

White and Blue Glaciers are located in the Olympic Mountains, which is ±55 km from the Pacific Ocean. The climate of the Olympic mountains is strongly maritime and involves the greatest precipitation of any area in the U.S.A., excluding Alaska and Hawaii (Armstrong, 1989). Blue Glacier has a steeper and narrower glacier tongue than White Glacier and has retreated less. South Cascade lies in the North Cascade Range, 250 km from the ocean. This glacier receives less snow (±3 m w.e.) than White and Blue Glaciers (±4 m w.e.) (Rasmussen and Conway, 2001). Mass-balance measurements have been carried out on Blue Glacier (Conway et al., 1999) and South Cascade (Krimmel, 2000). Armstrong (1989) showed that Blue Glacier had a slightly positive mass balance between 1956 and 1986 (0.3 m) and Krimmel (1989) found a mean negative balance for South Cascade (−0.22 m) for 1959 to 1985. Blue Glacier has not retreated significantly since 1950, while the length record of South Cascade shows an ongoing retreat. Rasmussen and Conway (2001) concluded that South Cascade has been more out of balance than White Glacier, which they ascribed to differences in geometry between the two glaciers, rather than in climate.

5.5.3 South America

Frías Glacier (Argentina) is an outlet glacier situated on the east side of Mount Tronador. This ice cap is the northernmost ice body of Argentina. Measurements from a weather station in the Rio Frías Valley of Frías Glacier indicated that the annual precipitation amounts to 4300 mm a⁻¹ (Perez Moreau, 1945). The glacier fluctuations have been recorded since 1976 and Villalba et al. (1990) dated oscillations of Frías Glacier by using tree-ring analysis. Between 1850 and 1900, the retreat rate was 7 m a⁻¹ and increased to 10 m a⁻¹ between 1910 and 1940.

San Quintín is further south than Frías Glacier and is the largest glacier of our data set. It is a piedmont outlet glacier of the North Patagonian Ice Field in Chile. San Quintín flows to the west, onto an
outwash plain. It receives between 3700 and 6700 mm precipitation per year. Winchester and Harrison (1996) investigated the ice front retreats and advances and concluded that precipitation is the main factor controlling the fluctuations. The length record of San Quintín shows a peculiar glacier retreat just before 1935 from which it recovers after 1935, which does not show up in the other length records.

5.5.4 Europe

Sørbreen is a glacier on Jan Mayen, the northernmost island on the Mid-Atlantic Ridge (71˚N, 8˚W). This glacier flows southwards to the ocean, from the 2277-m high Beerenberg volcano. The climate is cool oceanic with an annual mean air temperature of –1.2 ºC at sea level. The island is surrounded with pack ice during winter and spring (Anda et al., 1985). Nigardsbreen is an outlet glacier in Norway, flowing from the largest ice cap of continental Europe (Jostedalsbreen) in a southeasterly direction. It is a maritime glacier, located close to the Atlantic ocean. It advanced very rapidly between 1710 and 1748 and after 1988, it slightly advanced again. Pohjola and Rogers (1997) claimed that this advance is due to enhanced westerly maritime flow after 1980 leading to high winter accumulation and low summer ablation.

Glacier d’Argentière (France), Morteratschgletscher (Switzerland) and Pasterzenkees (Austria) are valley glaciers in the European Alps, where the climate is more continental. Glacier d’Argentière has been documented well, regarding the numerous fluctuations registered before 1850. Its length record is the longest of our data set and has been studied by Huybrechts et al. (1989). Glacier d’Argentière advanced after 1970, in contrast to Morteratschgletscher and Pasterzenkees. Pasterzenkees is the longest glacier of the Austrian Alps. It retreated roughly during two periods: between 1870 and 1910 and after 1930 (Zuo and Oerlemans, 1997a).

5.5.5 Asia

Marukskiy, Bezengi and Gergeti are valley glaciers in the Caucasus Mountains. Their length records contain only a few measurements, which explains the straight lines in Figure 5.13. The mass-balance gradient of these glaciers is relatively small and the melt rate at the glacier tongue is low (Table 5.3). The maximum elevation of Bezengi and Gergeti is over 5000 m a.s.l. Sofiyskiy glacier is located in the Altai Mountains of central Asia, a border region between Russia and Mongolia. It is a so-called continental, summer-accumulation type glacier. Sofiyskiy glacier is a valley type glacier consisting of three basins. The most rapid retreat of Sofiyskiy glacier occurred between 1900 and 1940 (Pattyn et al., 2003; De Smedt and Pattyn, 2003).
5.5.6 NEW ZEALAND

Fox and Franz Josef are valley glaciers in the maritime climate of the Southern Alps of New Zealand. They receive between 5 and 15 m precipitation per year. Their glacier tongues lie at very low elevations, 425 and 305 m a.s.l. for Franz Josef and Fox Glaciers respectively (Hooker and Fitzharris, 1999). The length records show similar patterns: a rapid retreat after 1940 and a major advance after 1982. This advance phase is characterised by higher precipitation and is also related to a higher frequency of El Niño events (Hooker and Fitzharris, 1999).

5.6 RESULTS FOR WORLDWIDE GLACIERS

5.6.1 LENGTH RESPONSE TIMES AND CLIMATE SENSITIVITIES

We calculated the length response time and climate sensitivity with Equations (5.9.) and (5.10) for all nineteen glaciers and plotted the values in Figure 5.14 (see also Table 5.3 for the precise values of $t_{rL}$ and $c$). Maritime glaciers often have high climate sensitivities (San Quintín, Fox, Nigardsbreen and Franz Josef), since they have large mass-balance gradients. However, the glaciers of the U.S.A. (South Cascade, White and Blue), which are also located in maritime areas, have low climate sensitivities. This is due to their small surface areas and relative wide glacier tongues. Athabasca Glacier has the longest length response time, which is partly caused by its low melt rate at the glacier terminus. Blue

![Figure 5.14](image-url)

*Figure 5.14: Climate sensitivity and length response time for all glaciers. The first five letters of their names abbreviate the names of the glaciers.*
Glacier has the shortest length response time and the lowest climate sensitivity. Since this glacier is situated on a steep slope, the contribution of the mass balance - surface height feedback to \( \tau_r \) and \( c \) is small.

5.6.2 ELA reconstructions

Figure 5.15 shows the reconstructed ELAs. Athabasca and Havoc show similar fluctuations in the ELA although the ELA of Havoc started to increase some years before 1900 and that of Athabasca some years after 1900. The ELA reconstruction of Clendenning shows neither any resemblances with the other Canadian ELAs nor with the reconstructions for the U.S.A. because the ELA does not stabilise around 1950. The reconstructed ELAs of White and Blue Glaciers are similar, but the fluctuation in White Glacier’s ELA is larger.

The ELAs of Frías and San Quintín show hardly any correspondence. The ELA of San Quintín decreases exceptionally after 1920. As San Quintín is a piedmont glacier and not a valley glacier, it could be argued whether its ELA reconstruction is reliable. However, the rapid decrease could be explained by an increase in precipitation between 1919 and 1935 (Winchester and Harrison, 1996). Winchester and Harrison (1996) also found that precipitation is a more dominant factor than air temperature in controlling the fluctuations of San Quintín.

The ELA of the European glaciers increased between 1920 and 1950. Sørbreen reached a maximum ELA before 1950 and the other glaciers after 1950. Morteratschgletscher’s increase in the ELA is large compared to the other ELA reconstructions in the European Alps. Bogen et al. (1989) found that the increase in the ELA of Nigardsbreen that occurred between 1925 and 1955 is related to a series of excessive warm summers.

Marukskij, Bezengi and Gergeti all show slowly increasing ELAs until 1950–1970 and a decrease afterwards. The ELA of Sofiyskiy increases rapidly around 1900 and starts decreasing already after 1930.

The reconstructed ELAs of Fox and Franz Josef show very similar patterns, with a rapid increase in the ELA after 1920 until 1960. The high ELAs around 1960 are strongly linked with changes in the circulation patterns over the southwest Pacific region (Fitzharris et al., 1992). At that time, summer pressures over New Zealand were higher than normal due to a pole-ward shift of the subtropical high, favoring clearer skies and more ablation. Moreover, snow accumulation was unusually low in the 1960s and 1970s.

5.6.3 Worldwide pattern in ELA fluctuations

For all glaciers, ELA reconstructions were calculated for the period between 1910 and 1959. We therefore estimated an average worldwide
Figure 5.15: ELA reconstructions for all glaciers plotted separately for six regions. The first five letters of their names abbreviate the names of the glaciers.
change in the ELA for this period (Figure 5.16a). The average ELA increased by 33 m between 1910 and 1959, with a standard error of 8 m. According to Figure 5.15, this increase was most pronounced in the U.S.A. and New Zealand and less in Asia. Generally, the reconstructed ELAs increased strongly during the first fifty years of the twentieth century. Except for San Quintín, they all decreased again in the second half of the twentieth century until 1980, when most of our ELA reconstructions end. This implies that the first half of the twentieth century was warmer or drier than the period before 1900. After 1960, the climate returned to a situation that favours lower ELAs.

Figure 5.16: (a) All ELA reconstructions (dashed) and the average ELA reconstruction (solid). The ELA reconstruction of each glacier is shifted compared to Figure 5.15 in order to obtain a mean $E'(t)$ of zero over the period 1910 to 1959. (b) Same as Figure 5.16a, but for temperature reconstructions.

5.6.4 RECONSTRUCTIONS OF TEMPERATURE

If we assume that the changes in the ELA were solely due to temperature variations, we can derive historic trends in air temperature from the ELA reconstructions. We then only need to know the sensitivity of the glacier mass balance to the temperature. A change in the ELA ($E'(t)$) can be easily translated into a mass-balance change ($B'(t)$) by multiplying it with the mass-balance gradient. We estimated for each glacier the sensitivity of the mass balance to a change in temperature ($T'(t)$) by using a parameterisation of Oerlemans (2001):

$$\frac{B'(t)}{T'(t)} = -0.271(P_{ann}^{0.597})$$

(5.18)
Oerlemans (2001) based this parameterisation on calculations with a mass-balance model applied to a set of thirteen glaciers. The sensitivity is a function of mean annual precipitation ($P_{an}$) and increases for wetter climates. We chose, for each glacier, a value for the annual precipitation, based on the compiled data of Zuo and Oerlemans (1997b) (Table 5.3). Figure 8b shows the calculated temperature reconstructions for each glacier and an average temperature trend. Between 1910 and 1959, the average increase in air temperature was 0.8 K and the standard error in the mean 0.2 K. The calculated linear trend in temperature was 0.19 K per decade.

5.7 DISCUSSION

Our results agree with those of the IPCC (2001) on an increase in the global surface air temperature between 1910 and 1945 if we assume that the ELA fluctuations were solely explained by changes in temperature. In addition, cooling during the period 1946 to 1975 in the Northern Hemisphere (IPCC, 2001) is consistent with the observed lowering in our reconstructed ELAs. However, the warming rate reported by the IPCC for the period 1910–1945 is less (0.14 K per decade) than estimated from the ELA reconstructions for 1910 to 1959 (0.19 K per decade). This may be associated with the fact that we excluded precipitation and cloudiness as possible causes of the ELA fluctuations. Besides, we estimated this global temperature increase from only a few glacier records that are also not evenly distributed over the globe.

The IPCC also concluded that the Southern Hemisphere has been warming more uniformly during the twentieth century compared with the Northern Hemisphere and shows warming between 1946 and 1975 too. This is in contrast with the decreasing ELAs of the glaciers in New Zealand, unless precipitation has increased simultaneously. Regional differences from the hemispherical mean could also account for this discrepancy.

Our results do not give rise to the conclusion that there was a second period (1976 to 2000) of global warming, as claimed by the IPCC (2001). This is because most of our reconstructed ELAs do not extend beyond 1980, due to the time span of the length records and the Gaussian filter that we applied.

As we related changes in the ELA to changes in the specific mass balance, we could as well determine the mean specific mass balance for each glacier between 1910 and 1959. Averaged over all glaciers, this results in a mean specific mass balance of $-0.18$ m w.e. a$^{-1}$ with a standard error of $0.04$ m w.e. a$^{-1}$. If we assume this value to be representative for all glaciers in the world, the mean rise in sea level due to the retreating glaciers can be roughly estimated. Assuming an area of 527 900 km$^2$ for all of the world’s
small ice caps and glaciers excluding Greenland and Antarctica (Zuo and Oerlemans, 1997b), the estimated total rise in sea level is 1.3 cm over the period 1910–1959 and the mean rate is 0.3 mm a⁻¹. This rate is in accordance with the estimated contribution of glaciers to sea level rise for the period 1910–1990 reported by the IPCC (2001). However, we should realise that this estimate is not very accurate, as we did not include any information from glaciers in Alaska and north-east Canada, which are highly glacierised areas. Neither did we use a weighted average to account for the size of each glacierised region.

5.8 SUMMARY

The purpose of this study was to extract information on the past climate from glacier length fluctuations by means of deriving historical ELAs or mass-balance reconstructions. The model that we developed for extracting ELAs from glacier length fluctuations is a simple analytical one, based on the assumption that the change in length can be described by a linear response equation. Its advantage is that it takes the response time and the geometry of the glacier as well as the mass balance - surface height feedback into account. Furthermore, it requires less information about a glacier than is needed for a numerical model.

We carried out a sensitivity test for a fictitious glacier and found that the uncertainty in the width of the glacier tongue, the melt rate at the glacier terminus and the mass-balance gradient have the largest impact on the ELA reconstruction. More specifically, for Nigardsbreen, the standard deviation in the ELA reconstruction due to uncertainties in the input parameters was estimated to be 4 m.

We performed a case-study, in which we tested the analytical model on length records of seventeen European glaciers. The results indicate that the model is useful to derive a climate signal from a glacier length record. The calculated mass-balance fluctuations are in agreement with mass-balance reconstructions derived from numerical models and from temperature and precipitation records using seasonal sensitivity characteristics. However, the ELA reconstructions calculated with the analytical model are shifted forward in time by a decade compared to the numerical mass-balance reconstruction and the mass-balance reconstructions calculated from temperature and precipitations records. The analytical model is not suitable for glaciers with strong variations in the slope and width of the glacier valley and only valid for relatively small glacier length fluctuations.

We then extracted information on the past climate from glacier length fluctuations for nineteen glaciers from different parts of the world. The results show that all glaciers of our data set experienced an increase in the
ELA between 1900 and 1960. The average increase between 1910 and 1959 is 33 m. After around 1960 until 1980, the ELAs decreased to lower elevations. This implies that during the first half of the twentieth century, climate was warmer or drier than before. The results support the evidence that an average global temperature increase took place between 1910 and 1945 and a sea level increase of 0.3 mm a\(^{-1}\) (IPCC, 2001). The ELA reconstructions also reveal regional differences. Changes in the ELA were most pronounced for North America, Europe and New Zealand. After 1960 up to 1980, the climate reverted to conditions supporting lower ELAs.

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REFERENCES


