Note to Chapter 4: Modelled climate sensitivity of the mass balance of Morteratschglatscher and its dependence on albedo parameterisation:

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MODELLED CLIMATE SENSITIVITY OF THE
MASS BALANCE OF MORTERATSCHGLETSCHER
AND ITS DEPENDENCE ON ALBEDO
PARAMETERISATION

Abstract – This chapter presents a study of the climate sensitivity of the mass balance of Morteratschgletscher in Switzerland. We ran a two-dimensional mass balance model for the period 1980 to 2002 and compared our mass-balance simulations with measurements carried out on Morteratschgletscher and six other glaciers. We calculated the mass-balance sensitivity by perturbing the air temperature by values ranging from –2 to +2 K and precipitation by values ranging from –20 to +20 percent. The mass-balance sensitivity to temperature and precipitation are –59 cm w.e. a⁻¹ K⁻¹ and at 17 cm w.e. a⁻¹ per 10 percent respectively. The albedo parameterisation that we used for these simulations relates the snow albedo to snow age and snow depth, whereas it considers the ice albedo constant in time and space. As the chosen albedo scheme is often the largest error source in mass-balance models, we investigated the impact of using three other albedo parameterisations on mass-balance sensitivity. One parameterisation uses ice albedo values that vary in space and were acquired from a Landsat image. Another parameterisation calculates the snow albedo from accumulated daily maximum temperatures since snowfall. The third parameterisation is the simplest and uses two constant albedo values, one for snow and one for ice. The differences between the mass-balance sensitivities calculated from the different albedo parameterisations are generally small. However, when the simplest parameterisation is used, the mass-balance sensitivity to temperature decreases to –52 cm w.e. a⁻¹ K⁻¹. This is mainly due to the constant snow albedo, which implies that the albedo feedback is not fully considered. For an accurate estimate of the mass-balance sensitivity, the albedo parameterisation should capture the process of a decreasing snow albedo when a snow pack gets older or thinner.

4.1 INTRODUCTION

As volume and mass balance of glaciers vary with changing climate, many model studies and measurements have been carried out to investigate this link (e.g. Oerlemans and Hoogendoorn, 1989; Oerlemans, 1992; Jóhannesson et al., 1995, Greuell and Böhm, 1998; Vallon et al., 1998; Kayastha et al. 1999; Klok and Oerlemans, 2002 (Chapter 3); De Ruyter de Wildt and Oerlemans, 2003).

Generally, net shortwave radiation is the most important energy source for the melting process of glaciers. Therefore, glacier melt rates depend to a large degree on the surface albedo. Van de Wal et al. (1992) concluded from measurements on Hintereisferner (Austria) that 70% of the variation in ablation is explained by differences in the surface albedo. Nevertheless, within mass-balance models, the albedo scheme is often regarded as the largest source of error (Arnold et al., 1996; Klok and Oerlemans, 2002 (Chapter 3)). Usually, albedo parameterisations cannot account for all spatial and temporal variations in the surface albedo. Several studies demonstrated that the modelled mass balance is very sensitive to small changes in the albedo. Oerlemans and Hoogendoorn (1989) showed that for a perturbation in the surface albedo of 0.1, the change in the mass balance of a fictitious glacier in the Alps varies between 0.5 to 1.5 m w.e. a⁻¹, depending on elevation. The largest sensitivity was found at the glacier terminus. Munro (1991) found a change in the mass balance of 0.66 m w.e. a⁻¹ for Peyto glacier, Canada, when the albedo was increased by 5% and a change of -0.83 m w.e. a⁻¹ when the albedo was decreased by 5%. In the latter two studies, the albedo perturbation was applied to both snow and ice surfaces. Klok and Oerlemans (2002) (Chapter 3) investigated the effect of a change in only the ice albedo for Morteratschgletscher and found a lower sensitivity. The mass balance decreased with 0.15 m w.e. a⁻¹ when the ice albedo was lowered with 0.08.

Also the mass-balance sensitivity to a change in climate depends to a large extent on the parameterisation of albedo and albedo feedback. The albedo feedback enhances the response of the mass balance to a change in climate in the sense that changes in temperature and snowfall influence the melt rate and the average albedo of the glacier, which in turn influences the melt rate. Several processes are responsible for a change in average glacier albedo when, for instance, the temperature increases. First of all, a temperature increase causes more melting. Therefore, snow disappears earlier in the melt season, the duration of the period of exposed glacier ice increases, and the average albedo becomes lower. Secondly, a temperature increase influences the growth of ice crystals, resulting in larger grain sizes, which have a lower albedo. Thirdly, increased melt rates result in a larger melt water content at the surface, which also causes a decrease in albedo.
Fourthly, increased melt rates result in thinner snow packs and lower albedos because the effect of the lower albedo of the underlying ice or debris is greater when snow depth decreases. Lastly, a temperature increase also leads to a decrease in snowfall. Less fresh snow implies lower albedos and also implies thinner snow packs and a longer duration of the period of exposed ice.

Greuell and Böhm (1998) demonstrated the effect of the albedo feedback on the mass-balance sensitivity of an Austrian glacier (Pasterze). They showed that the sensitivity to a warming of 1 K increased by 91% (from –0.47 to –0.90 m w.e. a⁻¹). De Ruyter de Wildt and Oerlemans (2003) investigated the effect of the albedo parameterisation on the mass-balance sensitivity of Vatnajökull. They found a mass-balance sensitivity of –0.65 m w.e. a⁻¹ for an increase in temperature of 1 K when they modelled the snow albedo as a function of snow age. However, when they used a constant snow albedo of 0.72, the mass-balance sensitivity decreased by 25%.

![Figure 4.1: Morteratschglacier (17 km²). Outlines of the glacier are indicated as well as the locations of the weather stations (M1, M2 and M3).](image)

To address the impact of the albedo parameterisation on the modelled mass balance, this study evaluates in more detail the differences between various albedo parameterisations. We focused on the modelled response of the mass balance to climate change. For this purpose, we used the two-dimensional mass-balance model of Klok and Oerlemans (2002) (Chapter 3). They developed this model to investigate the spatial distribution of the energy and mass-balance fluxes of a glacier. They applied it to Morteratsch-
gletscher located in southeast Switzerland (46°24’N 8°02’E) (Figure 4.1), and simulated the mass balance for two years: 1999 and 2000.

For the present research, we ran the model for a period of 23 years: from 1980 to 2002. We also applied the model to Morteratschgletscher and used meteorological input from synoptic weather stations located near the glacier. We briefly describe the mass-balance model in Section 4.2. The albedo parameterisation that we used for the simulations follows Oerlemans and Knap (1998). We studied the performance of the model and compared the modelled mass balance with accumulation and ablation measurements carried out on the glacier tongue (described in Section 4.3). We also examined differences between the modelled mass balance of Morteratschgletscher and mass-balance measurements carried out on other glaciers in the Alps (Section 4.4). Then, we calculated the annual mass balance of Morteratschgletscher over the period 1980 to 2002 by using three other albedo parameterisations. The different albedo parameterisations are explained in Section 4.5. We estimated the mass-balance sensitivity to changes in air temperature and precipitation for the four albedo parameterisations. We perturbed these variables over the period 1980–2002 and then computed the average change in the mass balance. We present and discuss the results in Sections 4.6 and 4.7.

4.2 THE TWO-DIMENSIONAL MASS-BALANCE MODEL

As we used the two-dimensional mass-balance model of Klok and Oerlemans (2002) (Chapter 3), we only give a brief explanation here and refer to the earlier Chapter 3 for a more extensive discussion.

The mass-balance model is based on the surface energy balance of a glacier. Its spatial resolution is 25 m and half-hourly time steps are used. The model accounts for the effects of shading, slope, aspect, obstruction of the sky and reflection from the surrounding slopes on the incoming shortwave radiation. The albedo of each grid cell is estimated with the method of Oerlemans and Knap (1998), in which the snow albedo depends on the snow depth and the time since the previous snowfall event. We defined a snowfall event as a day with a snowfall > 3 mm w.e. The parameterisation of the turbulent heat fluxes follows the model of Oerlemans and Grisogono (2002). The turbulent exchange coefficient of this parameterisation was used to tune the model results to glacier melt measured at M1 (Klok and Oerlemans, 2002 (Chapter 3)). The conductive heat flux into the glacier is estimated from a simple 2-layer subsurface model. The temperature of the lowest boundary is fixed at 273.15 K because measurements indicated that the glacier is temperate (Schwerzmann, personal communication).

The model is driven by meteorological input from synoptic weather stations of Meteo Schweiz located in the vicinity of Morteratschgletscher.
We used half-hourly means of air temperature, relative humidity and incoming shortwave radiation from the stations Samedan and Corvatsch. Samedan is located at 1705 m a.s.l. and about 12 km from the Morteratsch-gletscher and Corvatsch at 3315 m a.s.l. and about 8.5 km from the glacier. Shortwave radiation was used to estimate cloud cover, which is needed for the shortwave and longwave radiation parameterisations. We computed the amount of snowfall from daily precipitation data measured at Pontresina and used a precipitation gradient of 0.5 mm a\(^{-1}\) m\(^{-1}\). Pontresina is located at 1774 m a.s.l., at about 8 km from the glacier.

We ran the model for 23 years (1980 to 2002). The first two years were used to initiate the model because we did not have a complete input data set for these years. Temperature measurements at Corvatsch began at 17 October 1980, and at Samedan and Corvatsch, relative humidity recordings started at 13 May 1981. For the period up to these dates, average temperatures or relative humidity were used, based on the available data. We did not incorporate the results for the first two years into this study.

4.3 COMPARISON OF SIMULATIONS WITH MEASUREMENTS

To study the performance of the model, we compared simulated snow depth, surface melt and mass balance with measurements carried out at the locations M1, M2 and M3 (Figure 4.1). The Institute for Marine and Atmospheric Research Utrecht operates automatic weather stations there and measures among others snow depth, ablation and surface height from stakes, sonic rangers and snow density profiles (Klok and Oerlemans, 2002 (Chapter 3); Oerlemans and Klok, 2002).

![Figure 4.2: Simulated snow depth as function of measured snow depth at M1 and M2.](image)

Figure 4.2: Simulated snow depth as function of measured snow depth at M1 and M2.
4.3.1 Snow Depth

We derived snow depth from snow profiles taken at M1 and M2. At M1, snow pits have been dug at an average frequency of three times each winter since 1995. At M2, four snow profiles have been taken since 2000. Figure 4.2 reveals that simulated and measured snow depth are close to the 1:1 line, especially for M1. Figure 4.3 shows simulated and measured snow depth as function of time, for M1. For this location, the mean difference between simulated and measured snow depth is 1 cm w.e. and the standard deviation 7 cm w.e. For M2, the mean difference is −2 cm w.e. and the standard deviation 26 cm w.e. On average, measured snow depth at M1 was overestimated with 4%, and it was underestimated with 3% for M2.

Figure 4.3: Simulated and measured snow depth at M1 as function of time.

Figure 4.4: Simulated surface melt between two stake readings as function of measured surface melt estimated from three stakes at each location.
4.3.2 Surface Melt

Stake readings are carried out at all three locations all the time. At each location, three stakes are drilled into the ice. The distance between the stakes is typically 20 m. Surface melt is estimated as the average of three stake readings. At M1, stake measurements have been made at an average frequency of about four times per year since 1995, at M2 four times per year since 1999, and at M3, three times a year since 1999.

Figure 4.4 shows the amount of simulated ice melt between two stake readings as function of measured melt derived from the stakes. The mean difference between simulations and measurements is –2, –2 and –5 cm ice for M1, M2 and M3 respectively. This implies that the model slightly underestimates ice melt at all locations. The respective standard deviations are 16, 24 and 11 cm ice. At M1, total accumulated simulated surface melt was 3% less than measured from stakes. At M2, it was 3% less and at M3, 4% less.

At all locations, also sonic rangers measure surface melt. For M1, a continuous record since 1998 was available. Records for periods in 1999 and 2000 were available for M2 and M3. Table 4.1 presents the results of the comparisons between surface ice melt derived from the sonic rangers and simulated ice melt. At M1, simulated surface melt is overestimated by 2% compared to the measurements, but at the other two locations simulated ice melt is underestimated by ~10%.

<table>
<thead>
<tr>
<th>Location</th>
<th>Period</th>
<th>Measured m ice</th>
<th>Simulated m ice</th>
<th>Difference m ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>15 Jul 1998 — 20 Sep 2002</td>
<td>29.37</td>
<td>30.09</td>
<td>0.72</td>
</tr>
<tr>
<td>M2</td>
<td>29 Jul 1999 — 26 Sep 1999</td>
<td>2.19</td>
<td>1.94</td>
<td>–0.25</td>
</tr>
<tr>
<td>M3</td>
<td>28 Jun 2000 — 20 Sep 2000</td>
<td>2.32</td>
<td>2.09</td>
<td>–0.23</td>
</tr>
<tr>
<td>M3</td>
<td>30 Jul 1999 — 22 Sep 1999</td>
<td>2.32</td>
<td>2.09</td>
<td>–0.23</td>
</tr>
<tr>
<td>M3</td>
<td>28 Jun 2000 — 16 Sep 2000</td>
<td>3.48</td>
<td>3.40</td>
<td>–0.08</td>
</tr>
</tbody>
</table>

Figure 4.5 depicts accumulated simulated surface ice melt at M1 and compares them to data from the sonic ranger and the stake readings. The arrow indicates the start of the record of the sonic ranger. The stake readings record starts one year later. The accumulated simulated surface melt is in between the two measured records. At the end of the ablation period of 2002 (indicated by a dotted line), the simulations overestimated accumulated ice melt by 0.79 m ice compared to data from the sonic ranger and underestimated accumulated ice melt by 1.6 m ice compared to the stake measurements. This implies that the surface mass balance at M1 is very well simulated, but the discrepancy between stakes and sonic ranger is
large. The difference is 8%, which can be regarded as a measure of the accuracy of the measured ice melt.

![Figure 4.5: Surface ice melt at M1 simulated with the mass-balance model and measured with stakes and sonic ranger. The sonic ranger also shows snow depth (m snow) in the winter months. The arrow indicates the start of the sonic data record and the dotted line indicates the end of the ablation period of 2002.](image)

### 4.4 Mass-Balance Calculations for the Period 1980–2002

The net mass balance of a glacier is normally measured over a hydrological year, which runs from 1 October to 30 September. We calculated the net mass balance over this period for Morteratschgletscher from 1982 until 2002. Net mass balance, snowfall and melt are depicted in Figure 4.6. All simulated net mass balances are negative, except for 2001. Compared to the other years, this year received most snow: 197 cm w.e., which is 81% more than the average over 1982–2002. The average net balance is –78 cm w.e. Years with high annual snowfall often coincide with low annual melt rates and vice versa (see years 1997, 1998, 2001). This is presumably due to the albedo feedback because high snowfall amounts signify a shorter duration of the period of bare ice and thus a higher average albedo, which in turn leads to a decrease in the melt rate.

We compared the modelled annual mass balance of Morteratschgletscher to the measured mass balance of six nearby glaciers (Figure 4.7). Overall, the simulated net mass balance of Morteratschgletscher is somewhat lower than the measurements of these six glaciers. This might be caused by regional differences in climate or by the uncertainties in the
model calculations. For instance, snowfall could have been underestimated for the accumulation area. Due to the absence of measurements in the accumulation area, we could not validate the mass balance model for this area.

![Graph showing mass balance, snowfall, and melt over time.](image)

**Figure 4.6:** Simulated net mass balance by using albedo parameterisation (I), annual snowfall, and melt averaged over Morteratschgletscher.

![Graph showing net mass balance of Morteratsch and other glaciers.](image)

**Figure 4.7:** Simulated net mass balance of Morteratsch glacier and measured net mass balance of six other glaciers in the Alps (see Table 4.2).
Table 4.2 lists the coordinates of the glaciers, the distance to the Morteratsch Glacier and the correlation coefficient with the modelled mass balance of Morteratsch Glacier. The mass-balance measurements of Silvretta correlate best with the modelled mass balance (R = 0.91). This glacier is also closest to Morteratsch Glacier. All mass-balance records show a dramatic low net mass balance in 1998. Besides, all measured net balances have a maximum at 1995, except for the simulated mass balance of Morteratsch Glacier. This could be explained by large regional differences in snowfall. Laternser and Schneebeli (2003) showed that in the area of Morteratsch Glacier, snow depths were 80% less for 1995 compared to the long-term mean, while in some other parts of the Alps snow depths were 80% more than normal.

### 4.5 Albedo Parameterisations

The foregoing results were calculated by using the albedo parameterisation of Oerlemans and Knap (1998), which we call parameterisation (I). In addition, we applied three other common-used parameterisations to simulate the mass balance and its sensitivity to climate change and to investigate any resulting differences. We explain each parameterisation in this section. Table 4.3 at the end of this section gives a summary.

#### 4.5.1 Parameterisation I

The albedo parameterisation developed by Oerlemans and Knap (1998) defines glacier albedo, $\alpha^0$ as a function of snow depth:
\[ \alpha^{(i)} = \alpha_{\text{snow}}^{(i)} + (\alpha_{\text{ice}} - \alpha_{\text{snow}}^{(i)}) \exp \left( \frac{-d}{d^*} \right) \] (4.1)

where \( \alpha_{\text{snow}} \) is the albedo of snow, \( \alpha_{\text{ice}} \) the albedo of ice, \( d \) is the snow depth (mm w.e.) and \( d^* \) a characteristic snow depth scale (11 mm w.e.). By relating snow albedo to snow depth, one assumes that the underlying ice or debris influences the surface albedo of snow. The snow albedo also depends on the time since the previous snowfall event. The time since the previous snowfall event actually acts as a surrogate variable for the increase in grain size, melt water, and impurity content when snow gets older. The albedo of snow is calculated from:

\[ \alpha_{\text{snow}}^{(i)} = \alpha_{\text{firn}} + (\alpha_{\text{frsnow}} - \alpha_{\text{firn}}) \exp \left( \frac{s - i}{t^*} \right) \] (4.2)

where \( \alpha_{\text{firn}} \) is the albedo of firn (0.53), \( \alpha_{\text{frsnow}} \) is the fresh snow albedo (0.9), \( s \) is the time since the previous snowfall event (in days), \( i \) is the actual time (in days) and \( t^* \) is a time scale (21.9 days). When the snow depth is zero, the albedo equals the ice albedo, which is constant in time and space (0.34). The numerical values of the parameters mentioned above were found by Oerlemans and Knap (1998) using a data set from the weather station at M1 of one year (1995/1996) (Figure 4.1). The values of \( \alpha_{\text{frsnow}} \) and \( d^* \) were chosen to fit the measurements at M1 (Klok and Oerlemans, 2002 (Chapter 3)).

### 4.5.2 Parameterisation II

A disadvantage of the preceding parameterisation is that the ice albedo is constant in space, while in reality the ice albedo varies across the glacier due to, for instance, variations in debris concentration. Therefore, we used Landsat TM and ETM+ images to estimate the ice albedo for each grid cell of the mass-balance model. We calculated an average ice albedo from two Landsat scenes (Figure 4.8). We chose 21 August 2000 and 15 September 2000 (Klok et al., 2003 (Chapter 2)) because these images were acquired rather late in the ablation season when a large part of the glacier is free of snow. Grid cells with an albedo greater than 0.5 were assumed to be snow. We assigned a value of 0.4 for the ice albedo to grid cells that were covered by snow at both images or not calculated at all due to shading or too steep slopes. It turned out that the resulting mean ice albedo of the ablation area is about 0.23, which is much lower than 0.34 used in parameterisation (I). This is mainly due to debris. The glacier tongue is characterised by bands of low albedo at the sides and high albedo at the middle of the glacier tongue, relating to ice with higher and lower debris concentrations (Klok et al., 2003).
In short, parameterisation (II) equals parameterisation (I), but uses an ice albedo that varies in space and is constant in time. Note that we only used the Landsat images to include the spatial variation in ice albedo. We did not account for the temporal variation in ice albedo, for which a large time series of Landsat images would be required. Meanwhile, the temporal resolution of useful Landsat images is low, due to the short return period of the satellite and the restriction of cloudless days and days with small solar zenith angles (Klok et al., 2003 (Chapter 2)).

Figure 4.8: Landsat-derived surface ice albedo for Morteratschgletscher.

4.5.3 Parameterisation III

In parameterisations (I) and (II), the snow albedo is a function of time since the previous snowfall event. Instead as a function of time, the snow albedo is sometimes presented as a function of the temperature sum since the previous snowfall event (e.g. Winther, 1993; Henneman and Stefan, 1999; Brock et al., 2000). Comparable to time since snowfall as in parameterisations (I) and (II), also the temperature since the previous snowfall event is a proxy for the increase in grain size and impurity content, which leads to lower albedos. Therefore, the third parameterisation uses a temperature sum. It is a combination of the method of Brock et al. (2000), which is based on measurements during an ablation season at Haut Glacier d’Arolla in Switzerland and the parameterisation of Winther (1993), which is based on
albedo measurements during a winter period at Helligsdaghaugden, a research site in Norway. Brock et al. (2000) calculated albedos for deep snow (> 0.5 cm w.e.) and shallow snow (< 0.5 cm w.e.), where the surface albedo depends on the accumulated maximum temperature since snowfall. For deep snow, they used a logarithmic function of the accumulated daily maximum temperature > 0 °C. In contrast, Winther (1993) used a linear model and estimated the snow albedo as function of the accumulated maximum temperature in °F (!) and also solar radiation. We found that for deep snow, the albedo was better estimated by the linear model of Winther (1993) than with the method of Brock et al. (2000), as the snow albedo also decays when temperatures are below melting point. However, for our calculations, we did not use the temperature in °F but accumulated daily maximum temperature relative to −20 °C \(T_{\text{sum}}\). The albedo for shallow snow and the albedos for the transition between deep and shallow snow were calculated according to Brock et al. (2000). The final equations read:

\[
\alpha^{(i)}_{\text{ds}} = 0.9 - 0.000485 T_{\text{sum}} \quad (4.3)
\]

\[
\alpha^{(i)}_{\text{ss}} = \alpha_{\text{ice}} + 0.304 \exp(-0.00735 T_{\text{sum}}) \quad (4.4)
\]

\[
\alpha^{(i)} = (1 - \exp\left(-\frac{d}{24}\right))\alpha^{(i)}_{\text{ds}} + \exp\left(-\frac{d}{24}\right)\alpha^{(i)}_{\text{ss}} \quad (4.5)
\]

where \(\alpha_{\text{ds}}\) is the albedo of deep snow and \(\alpha_{\text{ss}}\) the albedo of shallow snow. For large values of \(T_{\text{sum}}\), the albedo of deep snow may decrease to below the albedo of shallow snow. Therefore, \(\alpha_{\text{ds}}\) is constrained by a lower limit that equals \(\alpha_{\text{ss}}\). The coefficients of Equations (4.3) and (4.4) were found by using data sets of M1, similar to the calibration procedure of parameterisation (I). For the ice albedo, we used the Landsat ice albedo, as with parameterisation (II).

4.5.4 PARAMETERISATION IV

Oerlemans and Knap (1998) also investigated the performance of simpler albedo models than parameterisation (I), but concluded that parameterisation (I) performs best. They found a correlation coefficient of 0.93 for modelled and measured surface albedo at M1. However, simulations were not dramatically worse (\(R=0.88\)) for the simplest model, when effects of snow depth and ageing were excluded. Therefore, the fourth parameterisation only uses two values: a constant value for the snow albedo (0.75) and a constant value for the ice albedo (0.34). The ice albedo is
used when the snow depth is zero, else the surface albedo equals the snow albedo.

Table 4.3: Summary of the albedo parameterisations and change in mass balance (B) due to a perturbation in temperature (T) in cm w.e. a⁻¹ K⁻¹ and precipitation (P) in cm w.e. a⁻¹ 10%⁻¹, calculated for each parameterisation from data plotted in Figure 4.10.

<table>
<thead>
<tr>
<th>Parameterisation</th>
<th>Variables</th>
<th>Ice albedo</th>
<th>dB/dT</th>
<th>dB/dP</th>
</tr>
</thead>
<tbody>
<tr>
<td>(I)</td>
<td>snow depth, snow age</td>
<td>0.34</td>
<td>−59</td>
<td>17</td>
</tr>
<tr>
<td>(II)</td>
<td>snow depth, snow age derived</td>
<td></td>
<td>−62</td>
<td>17</td>
</tr>
<tr>
<td></td>
<td>from Landsat image</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(III)</td>
<td>snow depth, temperature sum</td>
<td>derived</td>
<td>−62</td>
<td>17</td>
</tr>
<tr>
<td></td>
<td>from Landsat image</td>
<td>from</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(IV)</td>
<td>–</td>
<td>0.34</td>
<td>−52</td>
<td>14</td>
</tr>
</tbody>
</table>

4.6 RESULTS FROM DIFFERENT ALBEDO PARAMETERISATIONS

4.6.1 MASS BALANCE

Figure 4.9 shows the modelled net mass balance between 1982 and 2002 for the different albedo parameterisations. The mean net mass balance over 1982–2002 is −76, −94, −92 and −95 cm w.e. for parameterisation I, II, III and IV respectively. When a matrix of correlation coefficients is calculated, the simplest parameterisation (IV) appears to deviate most. Results of parameterisation (I), (II) and (III) correlate well due to a variable snow albedo, although the use of the Landsat ice albedo leads to more ablation and a lower net mass balance. This is caused by the low mean ice albedo of the ablation area (0.23) derived from the Landsat images compared to 0.34 used in parameterisation (I). It is, however, questionable if the simulated melt rates at the sides of the glacier tongue are realistic when the Landsat ice albedo is used. The very low albedos at these locations are a result of high debris concentrations, which also cause an insulating effect and this impedes the melt-enhancing effect of a low albedo. However, the mass-balance model does not take this insulating effect into account.

4.6.2 MASS-BALANCE SENSITIVITY

Even if the mean ablation is similar for different albedo parameterisations, one cannot be sure that the sensitivity is similar as well. For this reason, we also tested the mass-balance sensitivity to a change in climate for the four albedo parameterisations.
We defined mass-balance sensitivity as the change in the mean net mass balance over 1982–2002 due to a perturbation in the climate. We only investigated the mass-balance sensitivity to air temperature and precipitation (Figure 4.10 and Table 4.3). Generally, the results do not differ much for the four albedo parameterisations. The mass-balance sensitivity to a change in temperature is greatest for parameterisations (II) and (III), which use ice albedos derived from Landsat images. Parameterisation (IV) shows the lowest sensitivity because it does not account for the decrease in snow albedo when snow gets older or the snow pack gets thinner, meaning that the albedo feedback regarding snow is not fully covered. Besides, the
ice albedo of parameterisation (IV) is greater than the average Landsat-derived ice albedo. This implies a smaller difference between the albedos of snow and ice, causing a smaller albedo feedback when the period of exposed ice increases and thus a lower sensitivity. This also explains the difference between parameterisation (I) and (II). Hence, the average lower albedo rather than the spatial distribution of the Landsat-derived ice albedo is the reason for the larger sensitivity of parameterisation (II) compared to (I).

Figure 4.11 shows how the mass-balance sensitivity varies with elevation. The sensitivity is lower in the accumulation area because the albedo feedback is smaller. Namely, in the accumulation area, a change in temperature or precipitation only influences the snow albedo and does not result in a change in the duration of the period of exposed ice. Parameterisation (IV) leads to sensitivities that, at all elevations, are smaller than the results of the other parameterisations.

![Figure 4.11: Climate sensitivity with regard to perturbations in air temperature and precipitation as function of elevation, as calculated from four different albedo parameterisations.](image)

4.7 DISCUSSION

The mass-balance sensitivities to perturbations in air temperature and precipitation as calculated from different albedo parameterisations confirm that when ageing of snow is taken into account, the mass-balance sensitivity
increases (Table 4.3). Using a Landsat-derived ice albedo for each grid cell, instead of using a constant value for the entire ablation area, increases the mass-balance sensitivity to temperature. This is not due to the spatial variation of the surface albedo, but is associated with the fact that the mean ice albedo derived from the Landsat images is smaller (~0.23) than the constant ice albedo that is used (0.34).

However, the differences between the results of the different albedo parameterisations are not large. The mass-balance sensitivity to temperature as derived from the most comprehensive parameterisations (II and III) is only 19% larger than the simplest parameterisation (IV). This is in agreement with the results of De Ruyter de Wildt and Oerlemans (2003) mentioned in the introduction. Using Landsat-derived ice albedos instead of a constant ice albedo causes an increase in the mass-balance sensitivity of 5%. Differences between parameterisation (II) and (III) are negligible. The results indicate that for an accurate estimate of the mass-balance sensitivity, the albedo parameterisation should capture the snow albedo decrease when the snow pack gets older or thinner. The snow albedo can be either modelled as function of snow age or as function of the accumulated daily maximum temperatures since the previous snowfall event.

Although this research indicates that the sensitivity of the mass-balance to a change in climate is influenced little by the type of albedo parameterisation, we underline that small changes in surface albedo could still lead to large changes in calculated melt rates. These changes can be of similar magnitude as a warming of 1 °C would cause (Oerlemans and Hoogendoorn, 1989; and Oerlemans, 1992). Therefore, for an accurate calculation of surface melt rates it is important to use albedo parameterisations that capture the spatial and temporal albedo variations well.

A type of albedo parameterisation that we did not include in this research is one in which the ice albedo increases with elevation. Oerlemans (1991) related the ice albedo to the distance from the equilibrium line altitude (ELA) because, so he argued, on many glaciers the amount of accumulated dust, soot, and liquid water increases when going down-glacier. This parameterisation thus calculates the lowest albedos for the glacier terminus. It also captures the process that retreating glaciers become often more heavily debris-covered, resulting in lower ice albedos because glacier retreat is related to an increase in the ELA. However, Landsat images revealed that the ice albedo of Morteratschgletscher is not clearly a function of elevation (Klok et al., 2003 (Chapter 2)). For this reason, we did not use this parameterisation to Morteratschgletscher, but it might be applicable for other glaciers. The mass-balance sensitivity calculated from this parameterisation would likely be larger than the sensitivities calculated from parameterisations I to IV because of the dependence of the ice albedo on the ELA. It causes an albedo feedback that results in a greater mass-balance sensitivity.
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References


