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ScienceDirect
Global and Planetary Change 56 (2007) 69–82
GLOBAL AND PLANETARY CHANGE
www.elsevier.com/locate/gloplacha

The application of glacier inventory data for estimating past climate change effects on mountain glaciers: A comparison between the European Alps and the Southern Alps of New Zealand

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Received 18 August 2005; accepted 21 July 2006
Available online 14 September 2006

Abstract

This study uses the database from national glacier inventories in the European Alps and the Southern Alps of New Zealand (hereinafter called the New Zealand Alps), which contain for the time of the mid-1970s a total of 5154 and 3132 perennial surface ice bodies, covering 2909 km^2 and 1139 km^2 respectively, and applies to the mid-1970s. Only 1763 (35%) for the European Alps and 702 (22%) for the New Zealand Alps, of these are ice bodies larger than 0.2 km^2, covering 2533 km^2 (88%) and 979 km^2 (86%) of the total surface area, respectively containing useful information on surface area, total length, and maximum and minimum altitude. A parameterisation scheme using these four variables to estimate specific mean mass balance and glacier volumes in the mid-1970s and in the '1850 extent' applied to the samples with surface areas greater than 0.2 km^2, yielded a total volume of 126 km^3 for the European Alps and 67 km^3 for the Southern Alps of New Zealand. The calculated area change since the '1850 extent' is −49% for the New Zealand Alps and −35% for the European Alps, with a corresponding volume loss of −61% and −48%, respectively. From cumulative measured length change data an average mass balance for the investigated period could be determined at −0.33 m water equivalent (we) per year for the European Alps and −1.25 m we for the 'wet' and −0.54 m we per year for the 'dry' glaciers of the New Zealand Alps. However, there is some uncertainty in several unknown factors, such as the values used in the parameterisation scheme of mass balance gradients, which, in New Zealand vary between 5 and 25 mm m^{-1}.

Keywords: glacier fluctuations; glacier length changes; glacier mass changes; climate change; reconstruction; volume change; area change

1. Introduction

The last report of the Intergovernmental Panel on Climate Change (IPCC, 2001) stated that glaciers are the best natural indicators of climate. Hence, glacier changes are observed world wide within the framework of the GTN-G (Global Terrestrial Network on Glaciers) of the Global Climate Observing Systems (GCOS/GTOS, 1997a; GCOS/GTOS, 2004; Haeberli et al., 2000;

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doi:10.1016/j.gloplacha.2006.07.001

Haeberli and Dedieu, 2004; Haeberli et al., 2002; WMO, 1997). The GTN-G is led by the World Glacier Monitoring Service (WGMS, http://www.wgms.ch), the successor to the 1894 founded international glacier commission (Forel, 1895). WGMS collects data based on the Global Hierarchical Observing Strategy (GHOST), which consists of five tiers (GCOS/GTOS, 1997b; IUGG (CCS)/UNEP/UNESCO, 2005). An extensive amount of data on topographic glacier parameters, based on tier 5 of the GHOST, has been built up in past regional glacier inventories (IAHS(ICSI)/UNEP/UNESCO, 1989), which are currently updated by modern remote sensing methods within the Global Land Ice Measurements from Space (GLIMS, http://www.glims.org) project (Bishop et al., 2004; Kieffer et al., 2000). Repetition of the glacier inventories should be undertaken at periods compatible with the characteristic dynamic response times of mountain glaciers (a few decades). However, current glacier down-wasting as observed in several mountain areas will probably require more updates of inventories in shorter time intervals (Paul et al., 2007-a).

The glacier inventories help with assessing the problem of representivity of continuous measurements within different mountain areas, which can only be carried out on a few selected glaciers. A climate signal extracted from one single glacier is often not very representative for a whole mountain range. The understanding of global effects of climate change can only be achieved by

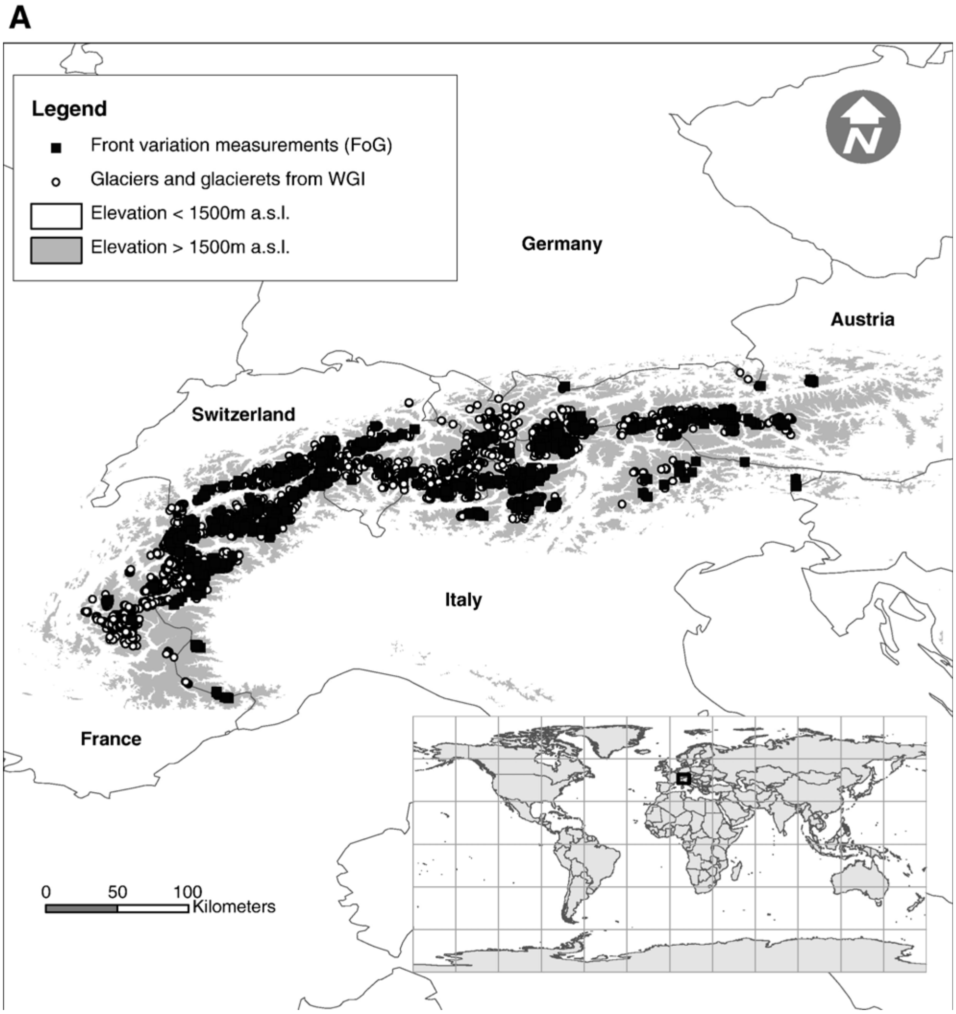

Fig. 1. A)Distribution of inventory glaciers (WGI = World Glacier Inventory) and of glaciers with length change measurements (FoG = Fluctuations of Glaciers) in the European Alps. B)Distribution of inventory glaciers (WGI) and of glaciers with length change measurements (FoG) in New Zealand. Only glaciers on the Southern Island of New Zealand were used in the analysis. The letters a) to g) are used to differentiate between 'wet' and 'dry' glaciers in the Southern Alps of New Zealand (a) = North_dry, (b) Fiord ('wet'), (c) West ('wet'), (d) East_wet and (e) East_dry). The digital elevation model used to draw the elevations is based on http://edc.usgs.gov/products/elevation/gtopo30/hydro/index.html.

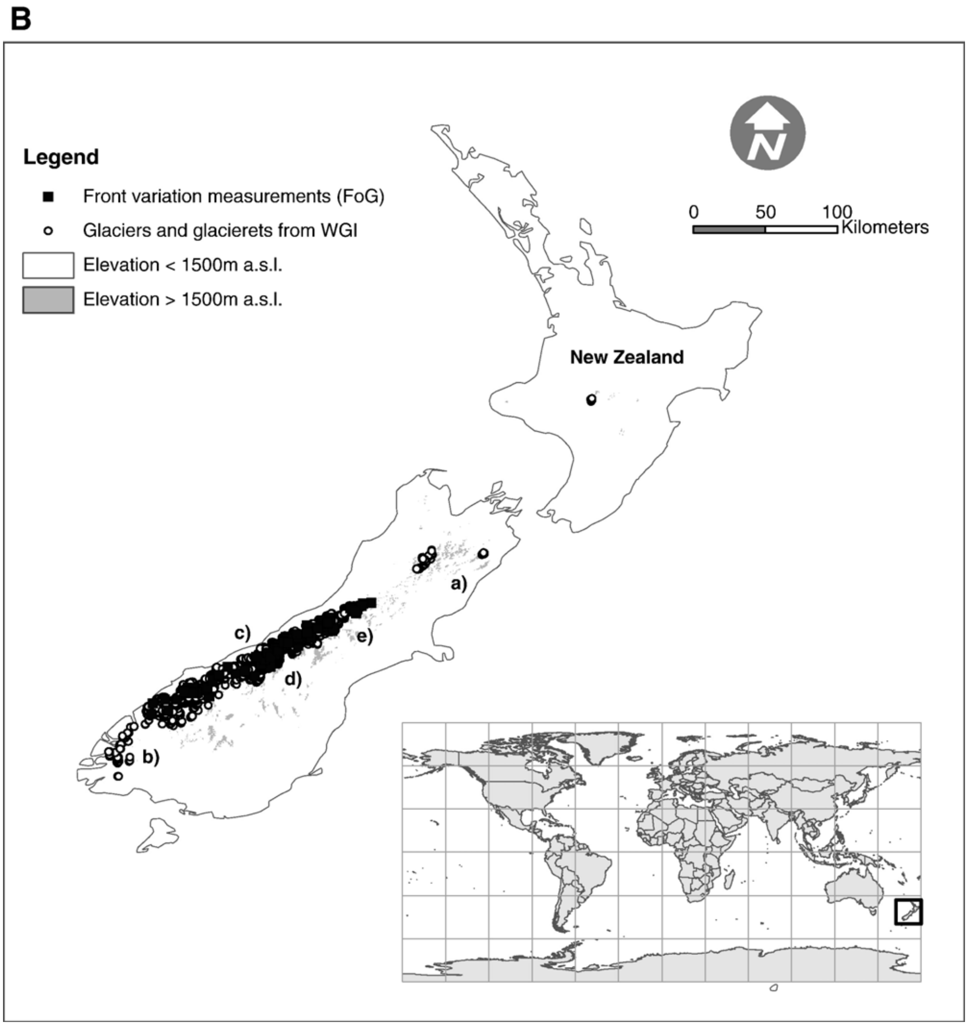

Fig. 1 (*continued*).

comparing the long-term behaviour of glaciers within different mountain ranges. The European Alps were analysed in detail by Haeberli and Hoelzle (1995) and a reanalysis was recently done by Zemp et al. (in press-b). This data is now ready to compare with other mountain ranges. The Southern Alps of New Zealand (hereinafter called the New Zealand Alps) are particularly interesting for such a comparison, because they are situated at a similar latitude in the Southern Hemisphere as the European Alps in the Northern Hemisphere. However, both mountain ranges have quite different climatic conditions. The method used to assess climate change effects on glacier change in both regions are based on simple dynamic considerations and steady state conditions, which could be approximated for 1850 and the mid 1970s. Both inventories show a very high accuracy and are perfectly suited to apply the method described in detail by Haeberli and Hoelzle (1995). Based on this application the New Zealand inventory (Chinn, 1991; 2001) is compared with the European Alps in this study (Fig. 1). Both inventories are included in the WGMS database stored at

Table 1
Parameterised values used in the calculations

Parameters	EU Alps	NZ Alps ‘wet’	NZ Alps ‘dry’
τ [Pa]	$1.3{\cdot}10^5$	$1.8{\cdot}10^5$	$1.2{\cdot}10^5$
$\delta b/\delta h$ [mm m^{-1}]	7.5	15	5
ρ [kg m^{-3}]	900	900	900
g [m s^{-2}]	9.81	9.81	9.81
A [a^{-1} Pa^{-8}]	0.16	0.16	0.16
n	3	3	3

WGMS in Zurich, Switzerland and at the National Snow and Ice Data Center (NSIDC, http://nsidc.org) in Boulder, Colorado, USA (Hoelzle and Trindler, 1998).

In this paper we show: a) an application of an existing glacier parameterisation scheme to two glacier inventories in two different mountain ranges to compare characteristic variables like balance at the tongue, ice thickness, etc., b) a method of mean specific mass balance and volume reconstruction based on observed equilibrium line altitude (ELA) changes, c) a reconstruction of mean mass balance change by using glacier length change measurements and d) a comparison of the different approaches within the different mountain ranges.

2. Methods

The parameterisation scheme developed by Haeberli and Hoelzle (1995) provides the possibility of analyzing

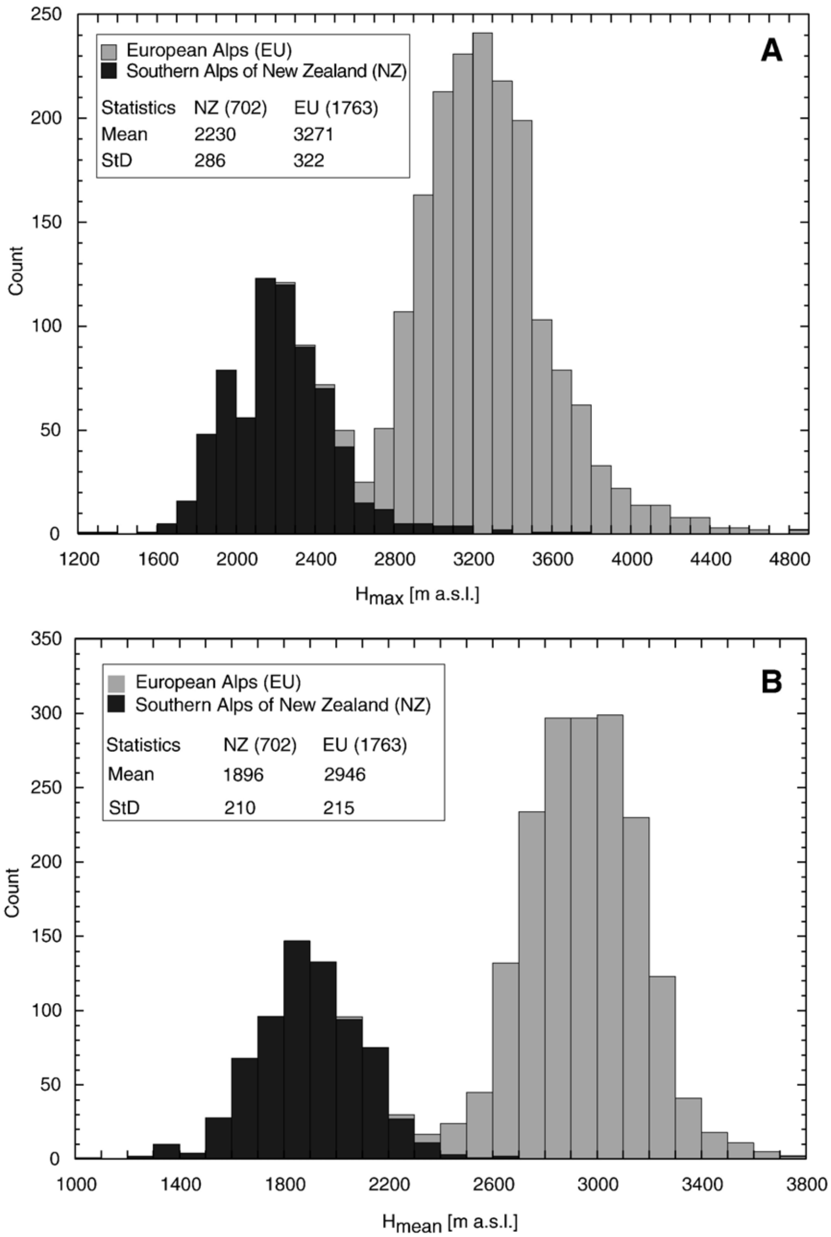

Fig. 2. Distribution of A) maximum (H_{max}), B) mean (H_{mean}), C) minimum (H_{min}) altitude and D) the slope (α) of inventory glaciers of the European Alps and the Southern Alps of New Zealand.

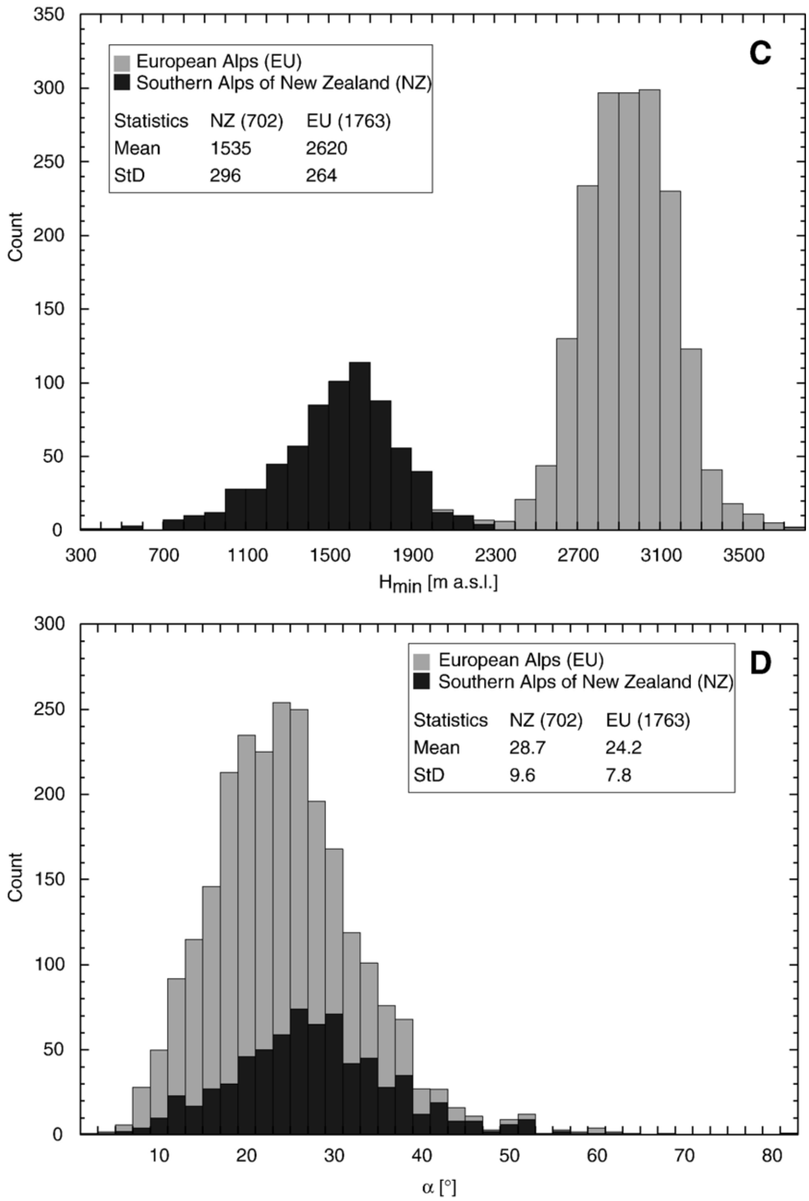

Fig. 2 (*continued*).

large amounts of topographic glacier inventory data stored in the current databases at WGMS and NSIDC and will probably be archived in future databases at the NSIDC within the framework of the GLIMS project (Raup et al., 2003; 2007). Detailed information about the full parameterisation scheme can be found in Haeberli and Hoelzle (1995). Therefore, we discuss only the most important parameterisations applied in this paper. Based on theoretical concepts of Jòhannesson et al. (1989) and Nye (1960). We use only four measured input variables from the inventories, namely maximum altitude (H_{max}), minimum altitude (H_{min}), length (L_0) and total surface area (F). All other variables are calculated or taken from measurements (see Table 1). Where specific equations are not given, the corresponding references are cited.

This approach considers the step changes after full dynamic response and new equilibrium of the glacier has been achieved, when mass balance disturbance Δb leads to a corresponding glacier length change ΔL that depends on the original length L_o and the average annual mass balance (ablation) at the glacier terminus

b_t. The term b_t is calculated as $b_t = \delta b/\delta h\ (H_{mean} - H_{min})$, where H_{mean} is determined by $(H_{max} + H_{min})/2$:

$$\Delta b = b_t \cdot \Delta L / L_0 \tag{1}$$

Glacier thickness (h) is determined according to Eq. (2) (Paterson, 1994) where α is the slope, τ the basal shear stress, ρ the density of ice and g the acceleration due to gravity (see Table 1 for used values of τ, ρ and g).

$$h = \tau / \rho \cdot g \sin\alpha \tag{2}$$

The dynamic response time t_{resp} is calculated after Jòhannesson et al. (1989), where h_{max} is a characteristic ice thickness, usually taken at the equilibrium line where ice depths are near maximum. h_{max} is calculated as 2.5 h, as estimated from known ice thickness measurements on various alpine glaciers world-wide (Bauder et al., 2003; March, 2000).

$$t_{resp} = h_{max} / b_t \tag{3}$$

Assuming a linear change of the mass balance from b to zero during the dynamic response, the average mass balance $<b>$ can be calculated according to Eq. (4). $<b>$ values are annual ice thickness change (meters of water equivalent (we) per year) averaged over the entire glacier surface, which can be directly compared with values measured in the field. Although the method is quite simple, the results compare very well with long-term observations (Hoelzle et al., 2003). The factor n_{resp} denotes the count of possible response times for each glacier within the considered time period.

$$<b> = \Delta b / 2 \cdot n_{resp} \tag{4}$$

The reaction time t_{react} is calculated, based on the kinematic wave velocity (Nye, 1965; Paterson, 1994) between the onset of the mass balance change and the first reaction at the glacier terminus from:

$$t_{react} = L_a / c \tag{5}$$

where L_a is the length of the ablation area and c is taken as $4 \cdot u_{s,a}$ (surface velocity in the ablation area) which corresponds quite well to observations (Müller, 1988).

3. Results

3.1. Glacierization in the 1970s

The inventories of the New Zealand Alps and the European Alps are well suited for a comparative study, because both have a high level of accuracy and both were compiled in the 1970s. The data bases from the national inventories used in this study contain a total of 5154 perennial surface ice bodies in the European Alps and 3132 for the Mountains of New Zealand. They were compiled for the early 1970s for the European Alps and 1978 for the Southern Alps of New Zealand. Only 1763 (35%) for the European Alps and 702 (22%) for the Southern Alps of New Zealand of these total numbers are ice bodies larger than 0.2 km^2 having useful information available about surface area, total length, maximum and minimum altitude. All calculations presented here were performed with these two sub-samples. This is justified by the fact that by far the bulk of the ice is contained within the large glaciers, such as Aletsch glacier, which stores around 11% of the total ice mass within the European Alps and in New Zealand, nearly 50% is contained in the 5 largest glaciers. In addition, the applied parameterisation is better suited to larger ice bodies, because of their more distinct dynamical behaviour (Leysinger Vieli and Gudmundsson, 2004).

The measured surface area of the 1763 glaciers in the European Alps and 702 glaciers in the Southern Alps of New Zealand >0.2 km^2 are 2533 km^2 and 979 km^2 corresponding to 88% and 86% of the total surface area, respectively. The calculated total volume of these glaciers is based on the ice thickness (Eq. (2)) multiplied with the measured area in the 1970s and equals 126 km^3 for the European Alps and 67 km^3 for the New Zealand Alps. These volumes correspond to a sea-level rise equivalent to 0.35 mm for the European Alps and about 0.18 mm for the New Zealand Alps. These small values point to the limited significance for sea-level rise, but mainly to the vulnerability to climate effects of glaciers in mountain areas with predominantly small glaciers (Barnett et al., 2005). Mountain ranges with such small glaciers could now be deglaciated within a few decades, whereas larger glaciers with high elevation ranges will persist somewhat longer before complete deglaciation. Small glaciers, especially in densely populated areas like the European Alps, have a strong impact on natural hazards, energy production, irrigations and/or tourism (Haeberli and Burn, 2002; Kääb et al., 2005a, b). The volume calculated for the Southern Alps of New Zealand (67 km^3) is somewhat higher than the value determined by Chinn (2001) and Heydenrych et al. (2002) of 53.3 km^3 and closer to a value determined earlier (63 km^3) by Anderton (1973). The mean maximum glacier elevation in the Southern Alps of New Zealand is 2236 m a.s.l.±289 m, and for the European Alps a value of 3271 m a.s.l.±322 m was found (Fig. 2A). Mean glacier elevations, which roughly equate to the equilibrium line altitude (ELA, here assumed as

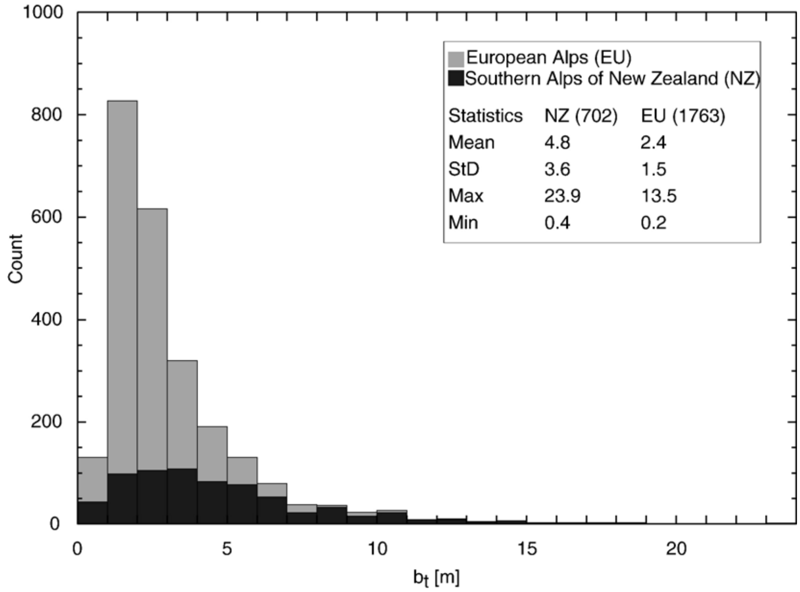

Fig. 3. Calculated values of mass balance at the tongue (b_t) the inventory glaciers of the European Alps and the Southern Alps of New Zealand.

the steady-state ELA) is, for the European Alps, 2945 m a.s.l. ± 214 m and for New Zealand Alps 1904 m a.s.l. ± 220 m (Fig. 2B). The minimum elevation of the glaciers in the European Alps is 2620 m a.s.l. ± 264 m and for the New Zealand Alps a value of 1545 ± 308 m was found (Fig. 2C). The overall slopes of the glaciers are steeper in New Zealand (28.7°) than in Europe (24.2°) (Fig. 2D).

The calculated average mass balance at the tongue (b_t) is, in the New Zealand Alps, around twice the ablation found in the European Alps. A maximum value was calculated as 23.9 m a^{-1} (Fox glacier) for the New Zealand Alps and 13.5 m a^{-1} (Bossons glacier) for the European Alps (Fig. 3). Calculated response times (the time taken to reach equilibrium after a 'step' climate change) after Eq.

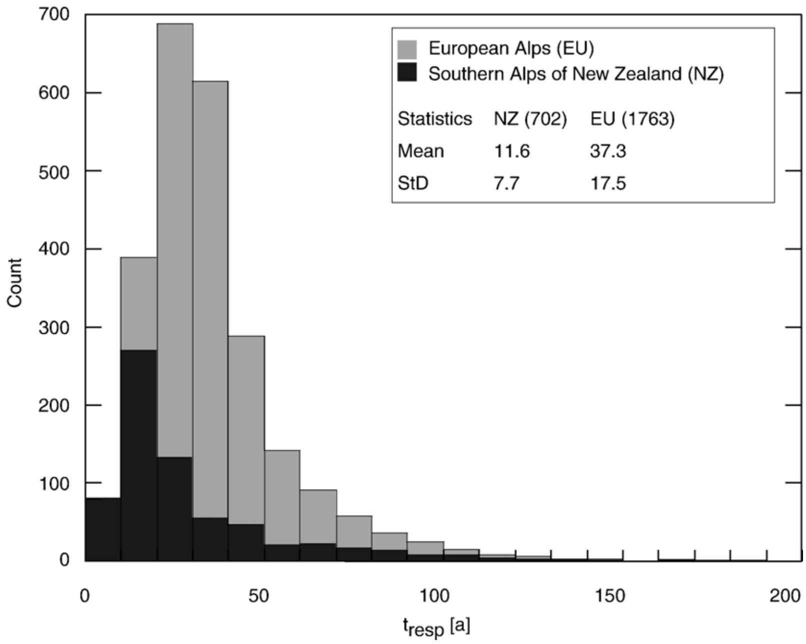

Fig. 4. Calculated values of response times (t_{resp}) for the inventory glaciers of the European and Southern Alps of New Zealand.

(3) have a mean value of 11.6 years in the New Zealand Alps and 37.4 years in the European Alps (Fig. 4). As an example, the calculated response time for Franz Joseph glacier is around 20 years, which corresponds well to values found with a numerical model by Oerlemans (1997). The reaction time of Franz Joseph glacier is around 7 years as estimated in this study (Table 3b). This value corresponds very well with the values found in other studies (Chinn, 1996; Hooker, 1995; Hooker and Fitzharris, 1999; Woo and Fitzharris, 1992). In contrast, the response time of Aletsch Glacier in the European Alps is around 80 years (Table 3b).

Even though the sub-samples were selected for larger glaciers, these glaciers mainly have surface areas smaller than 10 km^2, lengths shorter than 5 km and overall slopes steeper than 10°. This means that the sample of presently existing alpine glaciers is dominated by small and steep mountain glaciers with average thicknesses of a few tens of meters (see Fig. 5).

3.2. Reconstruction of the mean specific mass balances

The reconstruction of the mean specific mass balances since the '1850 extent' is valuable because it is directly comparable to present day measurements and therefore to current trends. The precise year of the maximum glacier extension in the Little Ice Age (LIA) period is often difficult to determine (Grove, 1988). In the European Alps, the maximum glacier extension differs considerably from glacier to glacier (e.g. Aletsch 1859–60, Gorner 1859–65 and Unt. Grindelwald 1820–22, see Holzhauser et al., 2005) and for New Zealand, recent datings (Winkler, 2004) and studies at Franz Joseph glacier (Anderson, 2003) have shown that maximum LIA glacier extension was at the end of the 18th century, but that there was only minor glacier retreat to the end of the 19th century. Therefore, here the time of LIA maximum is arbitrarily set at 1850 AD although there are numerous cases where different dates for the maximum extents have been found.

Kuhn (1989; 1993) calculated that a temperature change of +1 °C would increase the ELA by 170 m with an accuracy of ±50 m per 1 °C. Chinn (1996) reported a possible shift in ELA of up to 200 m for the Southern Alps of New Zealand, which corresponds to an air temperature increase of about 1.2 °C, which is in good agreement with a measured temperature change suggested by Salinger (1979) of around 1 °C for New Zealand. If the ELA-change is known, the corresponding change in mass (Δb) can be calculated together using the mass balance gradient ($\delta b/\delta h$) after Eq. (6).

$$\Delta b = \delta b/\delta h \cdot \delta \mathrm{ELA}/\delta T_{\mathrm{air}} \cdot \Delta T_{\mathrm{air}} \tag{6}$$

where $\delta\mathrm{ELA}/\delta T_{\mathrm{air}}$ describes the vertical shift of ELA per 1 °C, and integrates the change of all climate parameters, i.e. radiation, humidity, accumulation and air temperature, as well as feedback effects for any temperate glaciers (Kuhn, 1993).

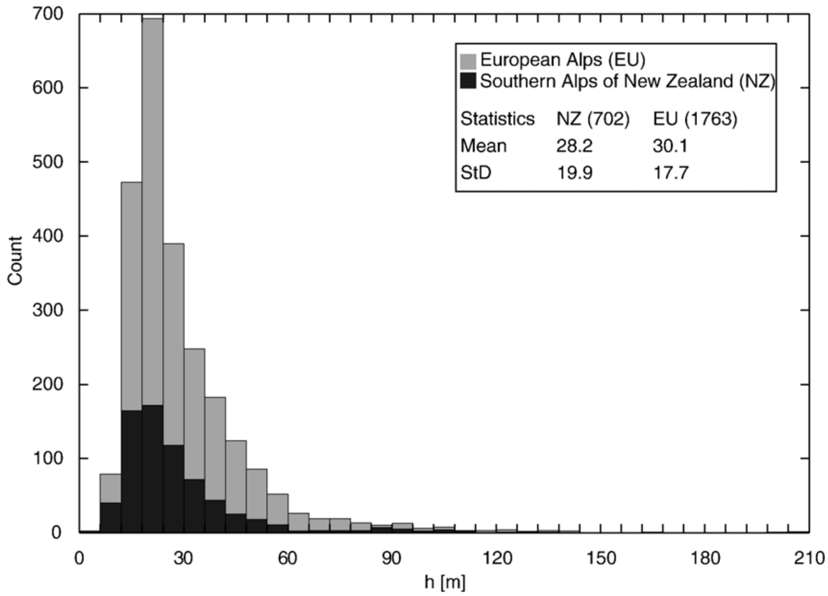

Fig. 5. Calculated values for mean ice thickness (h) along the flowlines for the inventory glaciers of the European Alps and the Southern Alps of New Zealand.

Mass balance gradients play a very important role in the parameterisation scheme. It is therefore fundamental to select realistic gradients for each area. For the European Alps a mass balance gradient of 7.5 mm m^{-1} was chosen for the whole sample. In contrast, the New Zealand Alps are characterized by much stronger precipitation gradients in comparison to the European Alps (Chinn et al., 2005a; Fitzharris et al., 1997), with the associated large range of mass balance gradients. Although, several studies in the New Zealand Alps have shown that there are no large differences in response between glaciers and climate of the glaciers west and east of the Main Divide (Chinn, 1996, 1999; Chinn et al., 2005a), mass balance gradients are completely different between glaciers west and east of the Main Divide (Chinn et al., 2005a). Therefore, the New Zealand sample was divided into two sub-samples, 'wet' to the west with a mass balance gradient of 15 mm m^{-1} and 'dry' to the east with a mass balance gradient of 5 mm m^{-1}. The 'wet' mass balance gradients were applied to three regions designated 'west', 'fiord' and 'east_wet' and the 'dry' mass balance gradients applied to 'east_dry' and 'north_dry' (Chinn et al., 2005a) (Fig. 1B). In the New Zealand Alps, an ELA shift of 200 m, will induce a mass balance change of 3.0 m for the 'wet' glaciers and 1.0 m for the 'dry' glaciers. The calculation of the <*b*> value was done by taking into account each individual response time and multiples thereof (Eq. (4)). The results of these calculations are presented in Fig. 6. The values range between −0.67 m we (water equivalent) in the 'west' region and −0.54 m we in the 'east_dry' region. No differentiation for the mass balance gradient was made for the European Alps, because there are no clear regional differences although, locally mass balance gradients can vary strongly.

3.3. Reconstruction of the 1850 area/volume and subsequent losses

In determining the volume change since the '1850 extent' and the original volume at 1850, the volume was calculated according to the parameterisation method described in Haeberli and Hoelzle (1995). Using the calculated mean specific mass balance for each area, the total mass loss could be estimated from the mean area of the 1970s and the 1850 area (F_{1850}), given in the following equation:

$$F_{1850} = 0.002 + .0285 L_{1850} + 0.219 \quad (L_{1850})^2 - 0.004 \quad (L_{1850})^3 \qquad (7)$$

where L_{1850} is the estimated length at the end of the Little Ice Age, calculated from Eq. (1). The third-degree

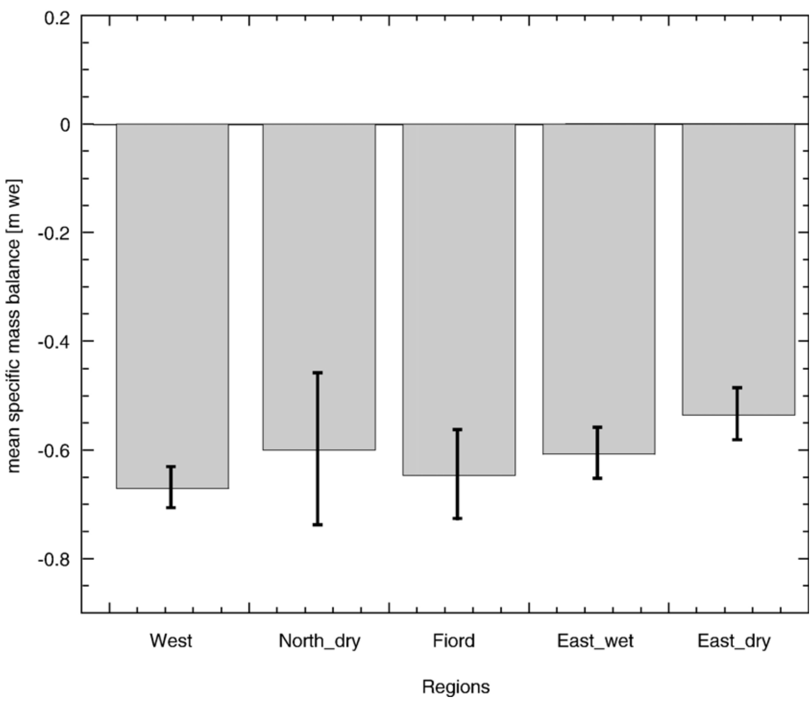

Fig. 6. Calculated mean specific mass balance (<*b*>) for the time period between the mid 1970s and the '1850 extent' for the different regions a) to e) for the Southern Alps of New Zealand. The error bars are only a statistical value for the standard error.

Table 2
Data used and calculated for all glaciers >0.2 km^2

Regions	Area 1970s (km^2)	Area LIA (km^2)	Count of glaciers	<b>used (m a^{-1})	Volume 1970s (km^3)	Volume LIA (km^3)	Volume loss (km^3)	Volume loss %
North_dry	0.69	3.81	2	−0.6	0.0079	0.16	−0.15	
East_wet	350.26	640.43	204	−0.61	32.37	66.40	−34.03	
East_dry	122.80	257.39	129	−0.53	5.367	16.71	−11.35	
West	464.20	951.67	301	−0.67	27.56	80.98	−53.43	
Fiord	40.80	78.36	63	−0.65	1.467	5.82	−4.36	
NZ Alps	978.75	1931.66	702		66.77	170.10	−103.33	−61
EU Alps	2544.38	3914.61	1763	−0.33	126	241.35	−115.35	−48

polynomial fit to the data is chosen to avoid negative area values for the smallest glaciers and to optimally reproduce the length/area-relation for large valley glaciers (Haeberli and Hoelzle, 1995). The so-calculated F_{1850} is around twice the 1978 area for the New Zealand Alps, and around 1.5 times the 1970s area for the European Alps. However, this result not only neglects the area changes for the ice bodies <0.2 km^2 in the current inventories, but also excludes all ice bodies, which had completely disappeared before the mid 1970s. The reconstructed area of Zemp et al. (in press-b), included both the small and vanished glaciers, and is therefore a value of 4470 km^2. Their corresponding value of area loss from 1850 to 1970 was −35% and is equal to the calculated loss of this study. Calculated ice volumes are 126 km^3 for the 1970s in the European Alps and 67 km^3 for 1978 for the New Zealand Alps (see details in Table 2). The calculation is based on the thickness along the flowline (Eq. (2)). Volume loss is determined from the calculated mean specific mass balances for each region and the mean area for the 1970s and 1850. Around 61% of the original volume has been lost in the New Zealand Alps and around 48% in the European Alps (Table 2 and Figs. 7 and 8).

3.4. Reconstruction of the mean specific mass balance based on length change measurements

The next task was to simulate the 1850 glacial conditions and to check how realistic the proposed scheme is. Present length change measurements were taken from monitored glaciers in the European Alps and the New Zealand Alps (Chinn, 1996). From these length changes, the mean mass balance since 1850 was determined by using Eq. (1) in the reversed manner as used in the previous two subsections (Hoelzle et al., 2003). Only those glaciers were selected, which are not calving in lakes or are heavily debris covered, as these

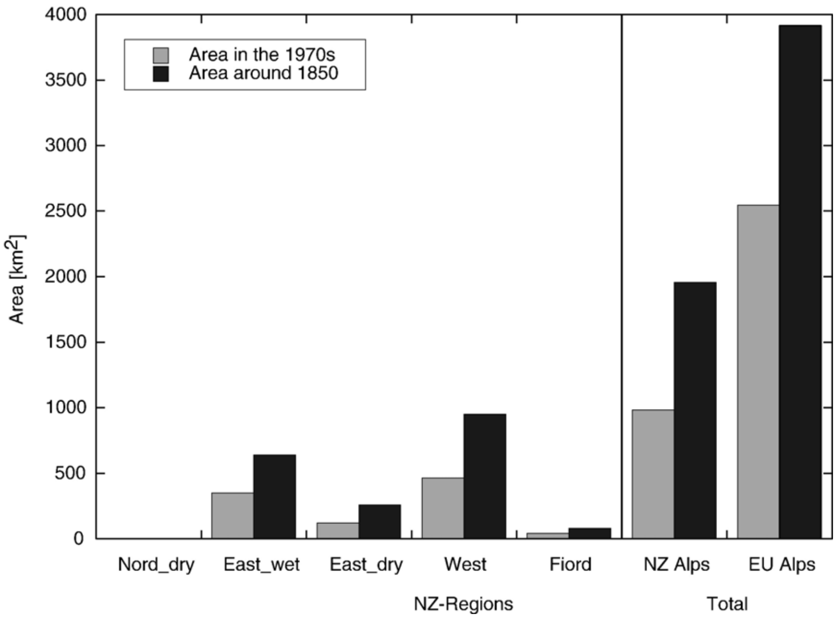

Fig. 7. Calculated areas for the time of the mid 1970s and the ‘1850 extent’ in the European Alps and the Soutern Alps of New Zealand.

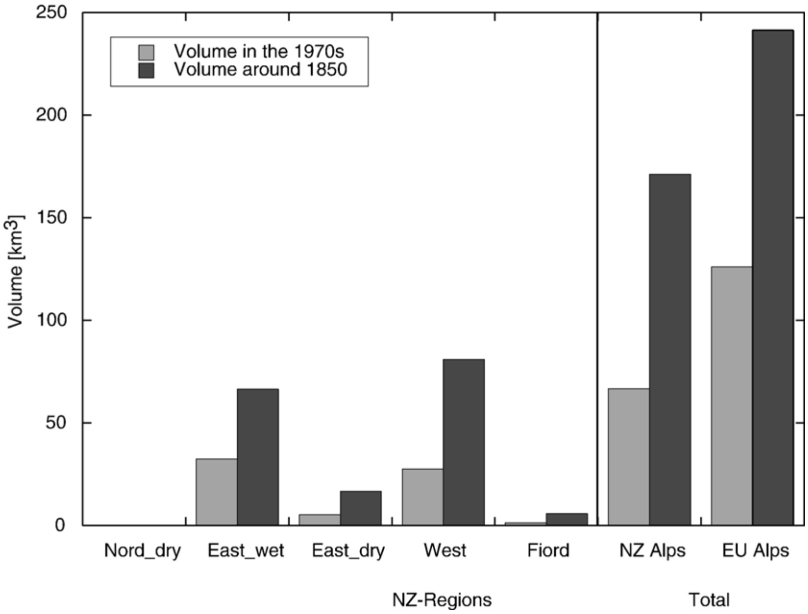

Fig. 8. Calculated volumes for the time of the mid 1970s and the '1850 extent' in the European and Southern Alps of New Zealand.

types of glaciers are frequently divorced from climate forcing by accelerated retreat (lakes) and thermal insulation (debris covered). Many of the large glaciers in New Zealand are heavy debris covered and/or calving into proglacial lakes.

The time interval considered is 125 years for the European Alps and 128 years for the Southern Alps of New Zealand. A single step positive mass balance change of 1 m we per year was assumed for the entire time interval for the European Alps and the 'dry' glaciers for the New Zealand Alps. For the 'wet' glaciers in New Zealand a value of 3 m was chosen. A comparison between calculated and geomorphologically reconstructed length changes for selected glaciers of the New Zealand 'dry' and 'wet' glacier samples show that the different sensitivities of long-term glacier length change as a response to a uniform mass balance forcing can be quite well reproduced (Table 3) and that the chosen mass balance forcing appears to slightly underestimate the real evolution, especially for glaciers in the 'wet' sample. Differences between measured and calculated overall length changes for individual glaciers

Table 3
Three selected glaciers from the parameterisation scheme: Fox, Franz Joseph and Aletsch. a) shows the input parameters for the parametrisation scheme. b) shows some selected calculated output parameters (u_d = deformation velocity, u_b = sliding velocity, u_s surface velocity are calculated after Haeberli and Hoelzle, 1995)

a)

Glaciers	H_{max} (m a.s.l.)	H_{mean} (m a.s.l.)	H_{min} (m a.s.l.)	Area 1970s (km^2)	Length 1970s (km)	Length change Measured (ΔL_m) (km)
Fox	3500	1900	306	34.69	13.20	2.50
Franz Joseph	2955	1690	425	32.59	10.25	2.95
Aletsch	4140	2830	1520	86.76	24.70	2.50

b)

Glaciers	t_{react} (a)	t_{resp} (a)	α (°)	Volume (10^6 m^3)	u_d (m a^{-1})	u_b (m a^{-1})	u_s (m a^{-1})	b_t (m)	h_{max} (m)	Length change calculated (ΔL_c) (km)	ΔL_m/ ΔL_c	b* (m)
Fox	6.0	16.7	13.6	2505.3	50.5	359.4	410	23.9	400.5	1.66	1.51	−1.13
Franz Joseph	7.5	20.7	13.9	2310.9	49.6	207.6	257.2	19.0	392.4	1.62	1.82	−1.37
Aletsch	27.3	76.6	6.1	13719.1	54.4	115.1	169.5	9.8	752.7	2.50	1.00	−0.25

can be considerable and are explained by the uncertainties in the parameterisation scheme applied, and by variable climate/mass balance conditions at each glacier. The mass balance forcing for each glacier can be corrected according to Eq. (8) to fit the measured length changes.

$$b* = \Delta L_m / \Delta L_c \cdot \langle b \rangle_i / 1 \quad (8)$$

where $b*$ is the corrected average mass balance, ΔL_m is the measured length change, ΔL_c is the calculated length change and $\langle b \rangle_i$ is the mean mass balance for the individual regions. These corrections from Eq. (8) to the mass balance forcing for each glacier to fit the measured length change gives an average annual mass balance ($b*$) of -0.33 ± 0.09 m we per year for glaciers in the European Alps. In New Zealand, the average annual mass balance for the 'wet' glaciers is -1.25 ± 0.9 m we per year and for the 'dry' glaciers -0.54 ± 0.5 m we per year. The calculated mass balance change Δb for the 'wet' glaciers in the New Zealand Alps is around 5 m and for the 'dry' glaciers around 1 m. The latter value corresponds quite well with the value used in the parameterisation scheme, whereas the value for the 'wet' glaciers is, much higher than the value of 3 m used in the parameterisation. This suggests that the glaciers have reacted even more sensitively than assumed in our parameterisation and that it is possible that our calculated mass loss is at the lower boundary of the uncertainty range.

4. Discussion and conclusions

The calculations and estimations presented in this study are built on four very simple geometric parameters contained in detailed glacier inventories. This justifies the simplicity of the applied algorithms but also means that the uncertainties involved with the proposed procedure are considerable. Indeed, the large scatter in derived parameters such as mass balance at the tongue, response times etc. points to the fact that the applied parameterisation scheme is more useful for relatively large glaciers than for small ice bodies. The large glaciers dominate the overall mass changes and hence, make the estimates of corresponding changes probably quite realistic. This is clearly indicated by the parameterisation results of the large glaciers such as Aletsch or Franz Joseph, which show quite realistic computation results (Table 3b) in comparison to detailed numerical studies (Oerlemans, 1997). The model is not tuned in any kind to the European Alps, where the model has been applied for the first time. The only factor changed in the model is the mass balance gradient and the basal shear stress (see Table 1). The values used for these parameters in the study are all based on measurements. In any case, the striking sensitivity of glacierization in mountain areas to atmospheric warming trends clearly appears in both mountain regions, although some marked differences do exist. The calculations of the mean specific mass balance change for the investigated period show no large differences between maritime and the 'dry' glaciers within New Zealand (Fig. 6). However, the mean specific mass balances calculated for New Zealand are close to twice than the values determined for the European Alps and indeed for other mountain areas (Hoelzle et al., 2003). In addition, the calculations show that the relative mass loss seems to be considerably larger in the New Zealand Alps than in the European Alps and the comparison of the measured length changes suggests an even more pronounced difference. Therefore, the calculated mass loss given here for the Southern Alps of New Zealand is probably at the lower limit of the uncertainty range.

The ongoing glacier behaviour after the mid-1970s was sometimes quite different between the European Alps and the New Zealand Alps. After a period of glacier advance in the 1970s and early 1980s in the European Alps, the glaciers experienced a strong retreat during the following 20 years with an even more pronounced mass loss at the beginning of the 21st century (Paul et al., 2004). Today the glacier surface area in the European Alps is around 2270 km^2 (Zemp et al., in press-b). In contrast, all glaciers in New Zealand have overall experienced a positive mass balance and some have advanced strongly during the 1990s. This is especially true for the 'wet' glaciers in New Zealand; the sample is clearly dominated by maritime, highly sensitive glaciers with corresponding high precipitation and therefore strong mass turn over. This period of advances was coming to an end at the beginning of the new century (Chinn et al., 2005b). The glaciers in the New Zealand Alps will probably react more sensitively to a future temperature increase than those in the European Alps. Not only because of their greater sensitivity, as expressed by the mass balance gradient, but also because of the generally low altitude of the ELA in the New Zealand Alps promoting a higher percentage of rain rather than snowfall in the future.

Due to increasing uncertainties and pronounced non-linearities such as changing response times with changing glacier size etc., calculations for scenarios with a trend to continuously accelerate climate and mass balance forcing beyond the early decades in the 21st century can be order-of-magnitude-estimates only. Annual mass losses of 2 to 3 m per year such as observed in the year 2003 in the European Alps (Frauenfelder et al., 2005; IUGG(CCS)/

UNEP/UNESCO/WMO, 2005; Zemp et al., 2005) and which must be expected to continue into the future, IPCC scenarios of temperature increase will certainly reduce the surface area and volume of alpine glaciers to a few percent of the values estimated for the '1850 extent' within the 21st century. With such a trend, only the largest and highest-reaching alpine glaciers could persist into the 22nd century. These glaciers will be affected by drastic changes in geometry as well and down-wasting rather than active retreat will be the dominant process of glacier reaction (Paul et al., 2004, 2007-a). This implies that parameterisation schemes like the one presented here would no longer be applicable. Therefore, modern technologies like remote sensing have to be used to measure the accelerating glacier change and alpine wide mass balance (Machguth et al., in press; Paul et al., in press-b) and ELA models (Zemp et al., 2007-a) need to be applied for the extrapolation of glacier down-wasting into the future.

This comparison between two different Alpine regions demonstrates the potential and limitations of the parameterisations of existing inventories. Their use lies especially in quantitatively inferring past average decadal to secular mean specific mass balances for unmeasured glaciers by analyzing cumulative length change from moraine mapping, satellite imagery, aerial photography and long-term observations (Haeberli and Holzhauser, 2003).

Acknowledgments

We would like to thank Ivan Woodhatch for his help in improving the English in the first version of this paper. The present study was carried out within the data analysis work for the World Glacier Monitoring Service with special grants from UNEP, a personal research grant from the Swiss National Science Foundation (PIOI2-111008/1) and as part of the EU Research Programme ALP-IMP (EU: EVK2-CT-2002-00148, BBW: 01.0498-2). The authors are very grateful to the constructive comments of Stephan Winkler, Per Holmlund and Christoph Schneider, which helped to improve the paper.

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