PARAMETERIZATION OF GLACIER INVENTORY DATA FROM JOTUNHEIMEN/ NORWAY IN COMPARISON TO THE EUROPEAN ALPS AND THE SOUTHERN ALPS OF NEW ZEALAND

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With 9 figures, 11 tables, and 2 appendices
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Summary: A parameterization by Haeberli and Hoelzle (1995) was applied to the glacier inventory data of Jotunheimen, Southern Norway, for three distinct time steps: 'Little Ice Age' maximum, 1980s, and 2003. Input data are measured or mapped values of the surface area, the length of the glacier flowline, and the minimum and maximum altitude. Additionally, the mass balance gradient and the glacier geometry had to be determined. Data from three glaciers with direct mass balance measurements available were applied to evaluate the parameterization results. According to the data, the area could be separated in a more maritime western and a more continental eastern part. These results were compared with existing parameterization results for the European Alps (Haeberli and Hoelzle 1995) and the Southern Alps of New Zealand (Hoelzle et al. 2007), and for the time period from the 'Little Ice Age' maximum until the 1970s/80s. The area loss was largest in New Zealand (-49%), moderate in the European Alps (-35%), and lowest in Jotunheimen (-27%). The corresponding volume loss was about 61% for New Zealand, 48% for the European Alps, and 42% for Jotunheimen. Jotunheimen, therefore, is the most continental of these three regions.

Zusammenfassung: Die Parametrisierung von Haeberli and Hoelzle (1995) wurde auf die Inventardaten der Gletscher Jotunheimens, Südnorwegen, für drei Zeitschritte angewendet: Maximum der „Kleinen Eiszeit", 1980er Jahre, 2003. Die Eingabedaten waren gemessene oder kartierte Werte der Gletscherfläche, Länge der Fliesslinie, der Gletscherunter- und -obergrenze. Zusätzlich mussten der Massenbilanzgradient und die Gletschergeometrie definiert werden. Zur Validierung der Ergebnisse der Parametrisierung wurden Daten von drei Gletschern, an denen direkte Massenbilanzmessungen durchgeführt werden, verwendet. Aufgrund der unterschiedlichen Werte der Massenbilanzgradienten dieser Gletscher konnte die Region in einen mehr maritimen Westen und einen eher kontinental geprägten Osten geteilt werden. Diese Ergebnisse wurden mit den Ergebnissen der Parametrisierung der Europäischen Alpen (Haeberli and Hoelzle 1995) und der Southern Alps auf Neuseeland (Hoelzle et al. 2007) seit dem Maximum der „Kleinen Eiszeit" bis zu den 1970er/80er Jahren verglichen. Der Flächenverlust ist in Neuseeland am größten (-49%), geringer in den Europäischen Alpen (-35%) und am geringsten in Jotunheimen (-27%). Der entsprechende Volumenverlust liegt bei 61% in Neuseeland, 48% in den Europäischen Alpen und 42% in Jotunheimen. Jotunheimen repräsentiert das kontinentalste Gebiet in diesem Vergleich.

Keywords: 'Little Ice Age'; glaciers; Jotunheimen; parameterization

1 Introduction

The significance of glaciers as indicators of climate change has been widely acknowledged (e.g. IPCC 2007). However, for any successful application of this information provided by the variations of glaciers, it is necessary to gain a representative regional climate signal from the glacier cover rather than signals from few selected individual glaciers. One individual glacier in most cases hardly could represent a whole mountain system (Hoelzle et al. 2007; WGMS 2008). Therefore, it would give an unreliable and subjective basis for further investigations and related conclusions. In addition, global effects of climate change can be achieved only by comparing long-term behaviour of glaciers within different mountain systems (Hoelzle et al. 2007).

The development and behaviour of glaciers and their response to climate change is important for hydropower production, especially in Norway: 98% of the domestic electricity is produced using hydropower and 15% of the exploited runoff is derived from glacierized river basins (Andreassen et al. 2005).

One attempt to analyze this situation is the parameterization developed by Haeberli and Hoelzle (1995). It has already successfully been applied to the European Alps (Haeberli and Hoelzle 1995) and the Southern Alps of New Zealand (Hoelzle et al. 2007) and was lately used by Paul and Svoboda (2009) for Baffin Island. The aim of this study is to present the application of this parameterization on the glacier inventory of Jotunheimen, South Norway, and to compare these results with the other two regions mentioned above.

2 Study area

2.1 Local settings

All three glacier regions have high-alpine characters with mostly individual glaciers of alpine morphology (Hoel and Werenskiold 1962; Østrem et al. 1988; Chinn 2001; Fitzsimons and Veit 2001; Lawson and Fitzsimons 2001; Andreassen et al. 2008; Zemp et al. 2008) and a climate gradient in continentality (Aune 1993; Førland 1993; Moen 1998; Salinger 2001; Sturman 2001a, b; Sturman and Wanner 2001; Bätzing 2005). The European Alps and the Southern Alps of New Zealand represent whole mountain systems. Jotunheimen, however, represents one mountain region within Norway. Therefore, the areal extent of the study areas and the number of glaciers differ remarkably.

The glaciers of Jotunheimen (Central South Norway; 61.5° N, 8.3° E; Fig. 1) are mostly small individual valley-type and cirque-type glaciers, separated by steep rock-walls, and only a few ice caps exist (Andreassen et al. 2008). The highest peaks of Norway (Galdhøpiggen 2469 m a.s.l., Glittertind 2464 m a.s.l.) are located here. In 2003, the glaciers ranged from 1300 to 2300 m a.s.l (Andreassen et al. 2008). In South Norway, there is a considerable climate gradient of raising continentality from the coast eastwards to the drier interior (Hoel and Werenskiold 1962; Andreassen et al. 2008; Winkler 2009), recorded by long-term mass balance measurements along a West-East profile (e.g. Østrem et al. (1988); Rasmussen et al. (2007)). The glacier area of Jotunheimen constitutes the most continental glaciological regime in Norway (Østrem et al. 1988), including a subregional gradient with a relatively maritime regime in the West and a relatively continental East (Matthews 2005). Since the 'Little Ice Age' (LIA) maximum, the glaciers retreated more or less continuously until the 1980s. Subsequently, the increase in volume of the maritime glaciers in Norway, especially in the 1990s (Matthews and Briffa 2005) and until the end of the 20th century, was also visible in the glaciers of Jotunheimen: a slightly positive net balance in the early 1990s (Kjøllmoen 2009).

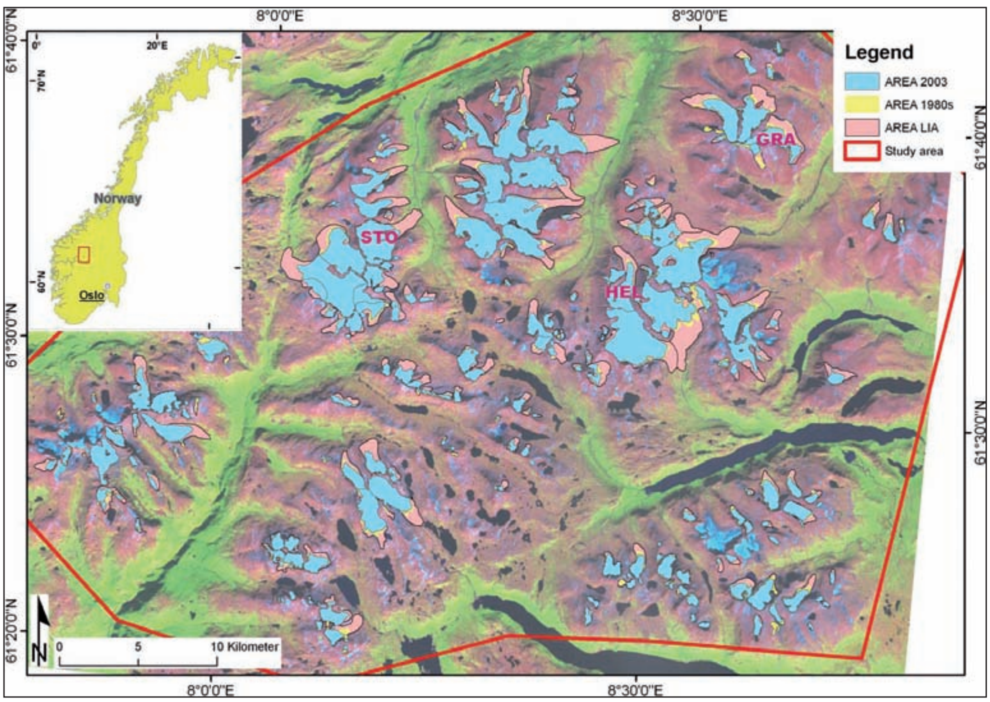

Fig. 1: Location of the glacier area Jotunheimen (inlay) and glacier extent during 'Little Ice Age' maximum, 1980s, and 2003 for glaciers used in the parameterization. Background: Landsat TM 5 August 2003. Letter codes denote: STO = Storbreen, HEL = Hellstugubreen, GRA = Gråsubreen. (Source inlay: ESRI templates; satellite image: Norsk Satellittdataarkivet)

According to CIPRA (2007), the European Alps roughly range between Grenoble (FR) in the West, Vienna (AT) in the East, Kempten (DE) in the North and Lake Garda (IT) in the South and have a curved shape. This vast area comprises several different climatic zones, but is generally located within the mid-latitudinal Westerlies. Precipitation is higher in the outer than in the inner areas because of orographic effects (Rott et al. 1993). This results in a more humid area in the outer northern Alps, dry inner-alpine regions, a maritime western and a continental eastern part and a Mediterranean influenced region in the South (Bätzing 2005). Since the LIA maximum, the glaciers in the European Alps have shown a general retreat with three phases of intermittent advance: in the 1890s, 1920s, and from 1970 through the 1980s (Zemp et al. 2008). The area declined mainly after 1985, and the acceleration of the retreat was more pronounced in 1985–1999 as compared with 1850–1973 (Paul et al. 2004, 2007).

The Southern Alps of New Zealand (called 'New Zealand Alps' here) are situated close to the West coast of the South Island of New Zealand (42.0° – 45.9° S, 167.3° – 173.8° E). Many of the largest valley glaciers in New Zealand are debris-covered (Röthlisberger 1986) and exhibit proglacial lakes (Chinn et al. 2005). New Zealand has a (super)humid maritime climate with a strong gradient in precipitation (Chinn et al. 2005). The mean annual precipitation, evenly distributed over the entire year, is 3000 mm along the Western coastal plains, and rises up to 15000 mm in the western ranges of the Alps immediate west of the main divide (Chinn 2000). Towards the eastern ranges, precipitation drops to about 1000 mm/a. After termination of the LIA, glaciers shrank in area and volume until the mid-1970s (Chinn et al. 2008). Comparable with South Norway, an advance started in the early 1980s until about 2000 (Chinn et al. 2005). This advance was recorded at the majority of the index glaciers throughout the New Zealand Alps (Chinn et al. 2005), but not at those large debris-covered valley glaciers with proglacial lakes (see Dykes et al. (this issue); Winkler et al. (this issue)). Since mid-2005, the glacier tongues of Franz Josef and Fox glacier have started to advance again (Winkler 2009).

All three regions are dominated by specific atmospheric circulation patterns.

The climate of Jotunheimen (as well as all of Norway) and the European Alps can be characterized by the circulation indices North Atlantic Oscillation (NAO) and the Arctic Oscillation. They measure the strength of zonal air flow in Northwest Europe quite adequately (Nesje et al. 2000; Winkler and Nesje 2009). In New Zealand, the Southern Oscillation Index (SOI) and Pacific Decadal Oscillation can be applied for the same purpose. The NAO implies an antiphase relationship between the European Alps and South Norway: a positive NAO index results in lower summer temperatures and especially high winter precipitation in southern Norway (Matthews and Briffa 2005), whereas a temperature rise and low precipitation over the European Alps is recorded (Holzhauser et al. 2005). Therefore, the Alpine glacier maxima during the LIA and more recently the readvances during the 20th century did not occur in Southern Norway (Günther and Widlewski 1986; Grove 2004; Matthews and Briffa 2005).

2.2 LIA maximum

In Jotunheimen, glaciers reached their maximum extent since Neoglaciation (i.e. during the Holocene) at the LIA maximum (Grove 1988; Matthews 1991; Matthews et al. 2000; Grove 2004). As they have not been overridden by any subsequent advances, moraines formed during the LIA maximum can, therefore, be applied for the reconstruction of the LIA glacier outlines. These advances were generally caused by higher winter precipitation and lower summer temperatures (Grove 2001; Holzhauser et al. 2005; Nesje et al. 2008b). Timing of the LIA maximum falls roughly between 1750 and 1800 in Jotunheimen (Winkler 2002; Matthews 2005). The distinct regional pattern divides between West and Central Jotunheimen with outermost moraines dating from about 1750 and East Jotunheimen with slightly younger moraines dated to about 1780/1800 (Fig. 3) (Baumann et al. 2009). Terminal moraines with double ridged outermost moraines are restricted to West and Central Jotunheimen and give evidence of a two-phase pattern of the LIA maximum (Winkler 2001). A small number of ice-cored moraines, mainly found in front of small high-lying cirque glaciers, are mostly located in eastern Jotunheimen (Østrem 1964; Østrem et al. 1988).

In the European Alps, glacier advances during the LIA occurred in the decades around 1320, 1600, 1700 and 1810 (Grove 2001; Grove 2004). Three maxima within the LIA, remarkably similar in extent, were identified around 1350, 1650 and 1850 in the Swiss Alps (Matthews and Briffa 2005). Although a certain degree of spatial differentiation of the timing of the LIA maxima has to be mentioned for the European Alps, the general pattern of three roughly similar maxima is a good approximation.

In New Zealand, the range of dated LIA moraines reveals no clear regional pattern yet. The tentative conclusion relating this inconclusive timing to different response times of the glaciers needs to be taken with considerable care, as methodological problems with the applied dating techniques might highly influence any interpretation (cf. Röthlisberger (1986); Winkler (2004); Burrows (2005); Schaefer et al. (2009)). The earliest maximum related to the LIA is assumed to be about 1600 or even earlier. Later maxima or readvances occurred between mid to late 1700s, early to mid 1800s, and around 1900. A maximum advance at several glaciers in Mt Cook National park was dated ~1750 (Winkler 2004), comparable with Jotunheimen. That was followed by several readvances, closely reaching the maximum or even overriding it (e.g. Tasman glacier) (Winkler 2004).

In Jotunheimen (~1750) and the European Alps (~1850), the term 'maximum' seems to more or less correctly describe the former circumstances, because no other glacier extent was larger than this one during the LIA (Erikstad and Sollid 1986; Winkler 2001; Matthews 2005; Grove 2008; Matthews and Dresser 2008; Nesje et al. 2008a; Nesje 2009). In New Zealand, the LIA maximum showed a slightly different pattern, based on the available dating: no clear regional maximum is identifiable yet, just a rather broad time span between 1600 (or earlier) and around 1900 (Burrows 2005). However, during the decades from 1750 until 1900, and partly until 1930, most glaciers seem not to have undergone much variation and remained quite close to their maximum positions (Chinn et al. 2005). Therefore, Hoelzle et al. (2007) set the date of the LIA maximum extent arbitrarily to 1850 in their parameterization to perform the comparison with the European Alps on similar time scales.

3 Material and Methods

For analyzing the inventory data of the three regions, a simple parameterization scheme developed by Haeberli and Hoelzle (1995) is used. To run the parameterization, the necessary input data are surface area, minimum and maximum altitude, and length of the glacier flowline.

3.1 Material

The three regions are very well suited for comparison. The time of compiling the glacier inventories that were used as the main basis for the investigation is very similar, and the inventories show the same high level of accuracy.

- European Alps: early 1970s (Haeberli and Hoelzle 1995)
- New Zealand Alps: 1978 (Hoelzle et al. 2007)
- Jotunheimen: 1980s (Østrem et al. 1988).

In recent studies, the parameterization has already been applied to the inventory data of the European and New Zealand Alps (Haeberli and Hoelzle 1995; Hoelzle et al. 2007). In addition to the inventories from the 1970s and 1978 mentioned above, selected data from the LIA maximum was available for the parameterization. Further information about the data inventories used can be found in Hoelzle (1994), Haeberli and Hoelzle (1995), and Hoelzle et al. (2007). The results of these studies are used in the comparison with the parameterization data of Jotunheimen.

For the parameterization of Jotunheimen, the inventory data of the LIA maximum (Baumann et al. 2009), the 1980s (Østrem et al. 1988), and 2003 (Andreassen et al. 2008) are used. For each of these inventories, digital outlines of all glaciers are available. As the general date of the LIA maximum, despite regional differences, the year 1750 is chosen. Because of the parameterization data from the European and New Zealand Alps, the comparison is only made with the results of the 1980s and selected results of the LIA maximum from Jotunheimen.

3.2 Methods

3.2.1 Parameterization scheme

The parameterization scheme was first developed by Haeberli and Hoelzle (1995). Detailed information about the parameterization can be found there, but an overview pointing out the most important issues is given here. All parameters and calculations of the variables are listed in Appendix A.

The parameterization is based on the concepts of Jòhannesson et al. (1989) and Nye (1960). Aim is the estimation of the glacier behaviour based on (measured) inventory data (area, length, minimum and maximum altitude) caused by a disturbance of air temperature and/or precipitation. The basic assumption is a glacier in steady-state that returns to a new steady-state after adaption to the new conditions by a step change of the equilibrium line altitude (ELA) and the mass balance disturbance (δb). This leads to a change in glacier length (δL) depending on the original length (L_0) and the ablation at the glacier tongue (b_t) (Fig. 2).

$$\delta b = b_t * \delta L / L_0 \quad (1)$$

The time needed is the response time (t_{resp}).

$$t_{resp} = h_{max} / b_t \quad (2)$$

The term h_{max} is the maximum ice thickness (see equation in Appendix A) and b_t the average annual ablation at the glacier tongue. The calculation of h_{max} is based on ice thickness measurements on various glaciers world-wide (Hoelzle et al. 2007) and was adopted from Haeberli and Hoelzle (1995). The ablation at the glacier tongue is calculated as:

$$b_t = (H_{mean} - H_{min}) * db/dH \quad (3)$$

H_{mean} is the mean glacier altitude (see equation in Appendix A), representing the ELA, and db/dH the mass-balance gradient. In a recent study, Braithwaite and Raper (2009) demonstrated that mean glacier altitudes give relatively good estimates of the balanced-budget ELA derived from mass balance data.

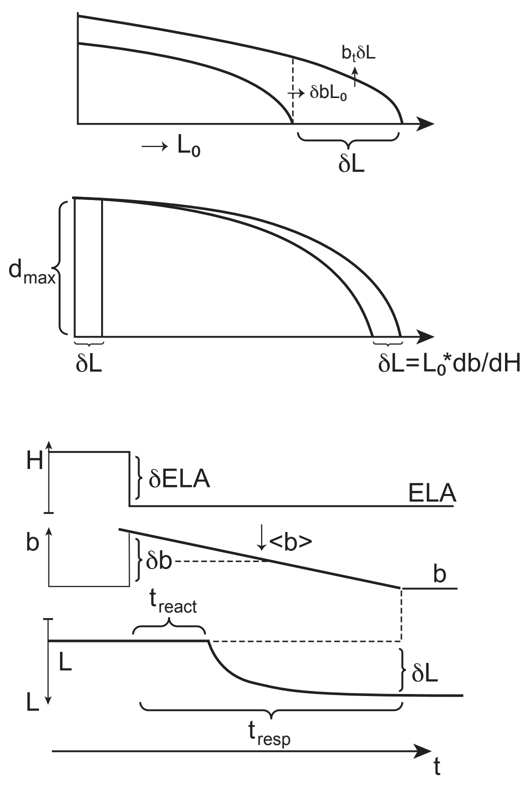

Fig. 2: Glacier reaction after a step change of the equilibrium line altitude (δELA) and a thereafter change in mass balance (δb) (modified after Haeberli (1991)). Terms see text and Appendix A

In the parameterization, only glaciers larger than 0.2 km² are used because they react more distinctly to changes in climate dynamics (Hoelzle et al. 2007). Additionally, large glaciers have a predominant influence on regional total mass changes (Haeberli and Hoelzle 1995), and average size glaciers (some km²) represent the largest part of the glaciated area in the investigated regions (Haeberli 1998).

3.2.2 Compilation of the Jotunheimen data

The inventory data of the 1970s was the basis for the parameterization of the European (Haeberli and Hoelzle 1995) and New Zealand Alps (Hoelzle et al. 2007). To be as close in time as possible, the inventory data of Jotunheimen from the 1980s was chosen as basis to selecting glaciers usable for the parameterization. First, all glaciers > 0.2 km² are used. Second, the flowline length of these glaciers must be comparable in all three years. The second selection results from the disintegration of the glacier area depending on glacier retreat. From the total amount of 218 glaciers covering 207.8 km² in the 1980s, 125 glaciers (57.3%) remained after the selection with a total area of 182.5 km² (87.8%). Depending on the remaining area, this selection is representative for the region and usable as basis for the parameterization.

Some further assumptions have to be defined. The mass balance data of all glaciers with mass balance measurements in the area were analyzed (data source: Kjøllmoen (2006; 2008; 2009), and WGMS database). These measurements were conducted at three glaciers: Storbreen (STO), Hellstugubreen (HEL) and Gråsubreen (GRA) (see location in Fig. 1). First, all years with an annual net mass balance of 0 ± 0.19 m w.e. (net mass balance averaged over the entire glacier surface; called steady-state years here) were chosen to calculate the mass balance gradient in the ablation area (Hoelzle 1994) and the mean accumulation area ratio (AAR_0; i.e. mean of all available AAR; for detailed information see Baumann 2010). The mass balance gradient is the most critical, but also the most important assumption for the parameterization (Hoelzle et al. 2007). Variables of steady-state years were chosen because they most likely represent glaciers during equilibrium. The resulting values of the mass balance gradient correspond quite well with the results of Rasmussen and Andreassen (2005) (Tab. 1). They vary between 0.2 m w.e./100 m/a for Gråsubreen in the East and around 0.6 m w.e./100 m/a for Stor- and Hellstugubreen in the central part. Gråsubreen is a high-elevation glacier

Tab. 1: Mass balance gradient db/dH of all years with an annual net mass balance of 0 ± 0.19 m w.e. (steady-state years) and vertical profiles of net balance available (not just glacier-average net balance). Mean value is resulting net mass balance gradient. *Year not used for calculation. Sources: (1) Rasmussen and Andreassen (2005); raw data db/dH: WGMS database

Glacier \ Year	1968	1971	1979	1991	1992	1994	1995	1998	2000	2005	2008	Mean	(1)
db/dH [m w.e./100 m/a]													
STO	–	–	–	1.18	0.77	–	–	–	–	0.72	0.71	0.85	0.61
HEL	0.58	0.67	0.70	–	0.52	0.68	0.65	0.61	0.71	–	0.74	0.65	0.57
GRA	0.17	–	–	–	–	0.18	-0.09*	–	–	–	–	0.18	0.20

with an impressive ice-cored moraine. This indicates a special type of glacier behaviour where the snout remains almost at the same position for long time periods with a restricted variability to a certain extent (Østrem 1964). Furthermore, this glacier might be polythermal (Lie et al. 2004) or even entirely cold-based (personal comment N. Haakensen). In another study (Rasmussen and Andreassen 2005), the winter mass balance gradient of Gråsubreen showed unusual results. Therefore, the resulting gradient must be taken with precaution, or even better, should not be used for estimating the mass balance gradient in the parameterizatioNonetheless, these results point towards a difference of the gradients between the western/central and the eastern part. Hence, the area is divided in two sub-regions West and East (Fig. 3), depending on the calculated mean elevation at the LIA maximum (Baumann et al. 2009) and the occurrence of double ridged moraines. In order to select a specific value of the mass balance gradients, additional analyses need to be made.

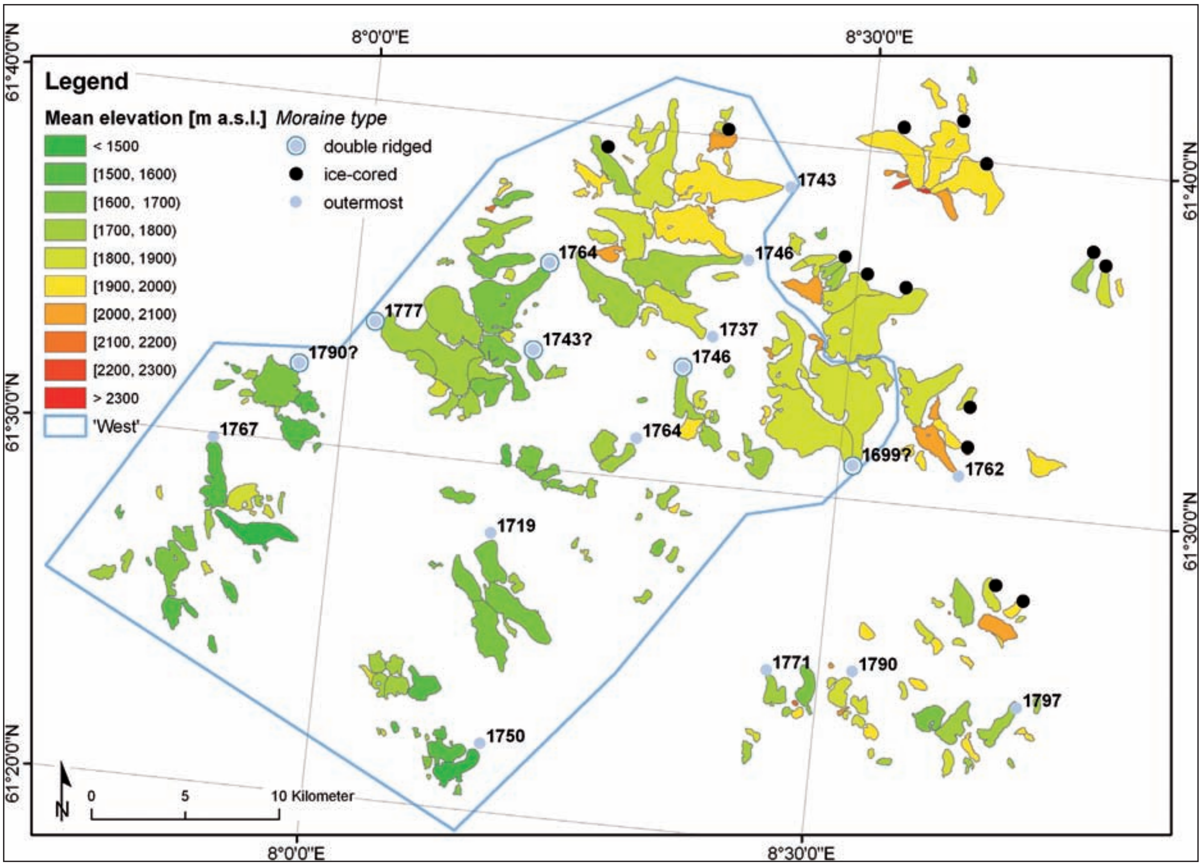

Fig. 3: Separation between the two sub-regions West and East in Jotunheimen. Specification of moraine type and timing of the LIA maximum after different sources. Glacier areas during the 'Little Ice Age' maximum are shown colour-coded based on mean elevation at the LIA maximum. (Source moraine type and timing: Erikstad and Sollid (1986); Winkler (2001); Matthews (2005))

First, the annual net mass balance from the years 1980 to 2003 was used to calculate the mean annual mass balance $\bar{b}n$. The data from these years was selected despite the availability of a data set that went back much farther in time (until 1949 for Storbreen) because of the direct comparability between measured glaciological *in situ* data and calculated values by the parameterization (1980s–2003). Second, the mass balance at the glacier tongue was derived from measurements of steady-state years from each mass balance glacier. Therefore, the net mass balance at the terminus is used. The means of these values are compared with calculations of b_t by the parameterization (see Equ. (3)), calculated with different values of the mass balance gradient to achieve the best fitting correlation (Tab. 2). According to this process, a mass balance gradient of 0.9 m w.e./100 m/a for the West and of 0.3 m w.e./100 m/a for the East fit best. These values are only a little bit higher than the ones calculated by using steady-state years.

The estimation of ELA ~ H_{mean} (setting of the parameterization) was not changed. This means that no uniform AAR_0 is chosen for all glaciers, because this variable depends on a defined area distribution ratio over the entire glacier range. Calculations of the AAR_0 at glaciers with mass balance measurements in Jotunheimen showed an AAR_0 close to 0.5, i.e. AAR_0 of 0.5 for STO, and 0.45 for HEL and GRA, and no significant change during the available periods.

The geometry shape factor f was chosen according to Paterson (1994). The half-width W, the ice thickness and the cross-section profile of the glacier must be known for the assessment of f. The shape of the cross-section of the glaciers in the area is not known and, to our knowledge, has not been measured on any glacier in the area yet. Hoel and Werenskiold (1962) sketched a cross-section of Hellstugubreen in the ablation area, estimating a parabolic or semi-elliptic profile. For the parameterization, a semi-elliptic geometry was chosen. NVE (2006) reported a mean ice thickness of 115 m for Storbreen in 1997. The estimated half-width of Storbreen was ~1065 m in 2003. Using this values results in W = 9.3 and f = 1. Measured thicknesses of Styggedalsbreen at several points were reported by Ahlmann (1928) in 1923/24, but no width was given for this glacier in these specific years. Due to the lack of other values or calculation possibilities, the value of Storbreen was extrapolated to the entire region. In conclusion, there were no valley-glaciers with steep valley slopes in Jotunheimen. An overview of the parameterized values used for all three regions is given in table 3.

4 Overview input data 1970s/80s

The basic input data of all three regions consist of area, length, and minimum and maximum altitude of the 1970s/80s. The total glacier area varied quite a lot depending on the number of glaciers (Tab. 4). The largest glacier in Jotunheimen is only about a tenth of the largest glacier in the European and New Zealand Alps. However, the mean glacier area is very similar, and in all regions more than 90% of the glaciers are < 5.0 km² (Tab. 5). The western and eastern parts of Jotunheimen differ e.g. regarding mean and maximum value (Tab. 4). The distribution pattern of the glacier length is comparable: The mean is nearly the same in all regions, but the maximum in Jotunheimen is only a fifth of the maximum in the European and New Zealand Alps (Tab. 6). The frequency distributions of the maximum elevation are very similar for the European and the New Zealand Alps, but the values for mean, maximum and minimum are about 1000 m higher in Europe (Fig. 4). The mean maximum elevation is also about 1000 m higher in the European Alps than in Jotunheimen. The range of the maximum altitude is much smaller in Jotunheimen compared with the other two regions. The mean maximum elevation is lower in the western (2025 ± 155 m a.s.l.) than in the eastern part (2096 ± 122 m a.s.l.) of Jotunheimen. The mean value of the minimum elevation is pretty similar in the New Zealand Alps and Jotunheimen, but is again about 1000 m higher in the European Alps (Fig. 5). The lowest value of the minimum elevation is 305 m a.s.l. in the New Zealand Alps (Fox Glacier) and ~1200 m a.s.l. in the European Alps (Bossons Glacier) and Jotunheimen (Riingsbreen). Both the minimum (difference East – West = 265 m) and the mean value (difference East – West = 147 m) of the minimum elevation are remarkably higher in the eastern compared with the western part of Jotunheimen.

5 Results

Differences between the published values by Hoelzle et al. (2007) and the values presented in sections 5.1 and 5.2 are due to rounding errors.

5.1 Comparison of 1970s/80s parameterization data

The mean elevation is $(H_{max} + H_{min}) / 2$, and its value is taken as an estimation of the ELA. The mean value for the European Alps is ~1000 m high-

Tab. 2: Ablation at the glacier tongue in the ablation area, measured from steady-state years in different time-periods and calculated by the parameterization with different values of mass balance gradient. Explanation of calculation in the text. b_t [m/a], db/dH [m w.e./100 m/a]. (Raw data measured b_t: WGMS database)

measured				calculated					
Period [a]	*1991-2008*	*1991/92*		**Period [a]**	*1980-2003*				
b_t STO	-2.9	-3.4			-2.9	-6.3	-2.4	-1.6	-0.6
Period [a]	*1968-2008*	*1979-2008*	*1979-2000*	**db/dH (STO, HEL)**	**0.9**	**2**	**0.75**	**0.5**	**0.2**
b_t HEL	-2.5	-2.6	-2.5		-3.1	-7.0	-2.6	-1.7	-0.7
Period [a]	*1968-1995*	*1994/95*		**db/dH (GRA)**	**0.3**	**2**	**0.75**	**0.5**	**0.2**
b_t GRA	-0.3	-0.3			-1.0	-6.8	-2.5	-1.7	-0.7

Tab. 3: Overview of the parameterized values used in Jotunheimen, the European and New Zealand Alps, and sub-regions. *Data from Hoelzle et al. (2007)

Region / Parameter	European Alps*	New Zealand Alps* 'wet'	New Zealand Alps* 'dry'	Jotunheimen West	Jotunheimen East
A	0.16				
n	3				
ϱ	900			917	
g	9.81				
f	0.8			1	
db/dH	0.75	1.5	0.5	0.9	0.3
τ [bar]	$1.3*10^5$	$1.8*10^5$	$1.2*10^5$	calculated	

Tab. 4: Statistics of the glacier area in 1970s/80s for Jotunheimen, the European and New Zealand Alps, and sub-regions

Region / S [km²]	European Alps	New Zealand Alps	Jotunheimen	Jotunheimen West	Jotunheimen East
Mean	1.44	1.40	1.46	1.68	1.09
Max	86.76	98.34	8.90	8.90	7.62
Min	0.21	0.20	0.21	0.22	0.21
Std	3.69	4.81	1.75	1.89	1.39
Sum	2544.38	982.04	182.50	131.38	51.13
Number [n]	1763	702	125	78	47

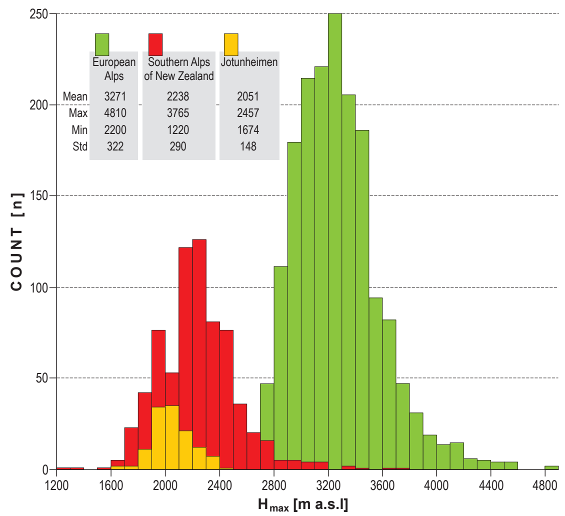

Fig. 4: Distribution of glacier maximum altitude (H_{max}) during 1970s/80s in Jotunheimen, and the European and New Zealand Alps

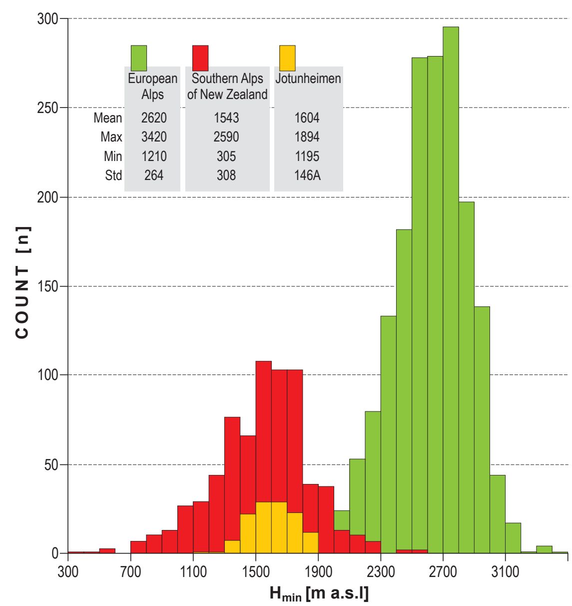

Fig. 5: Distribution of glacier minimum altitude (H_{min}) during 1970s/80s in Jotunheimen, and the European and New Zealand Alps

Tab. 5: Classification of the glacier area in 1970s/80s into size intervals for Jotunheimen, and the European and New Zealand Alps

Region / Area interval [km²]	European Alps Number [n]	[%]	New Zealand Alps Number [n]	[%]	Jotunheimen Number [n]	[%]
≤ 1	1241	70.4	544	77.5	70	56.0
[1, 2)	250	14.2	77	11.0	30	24.0
[2, 3)	103	5.8	25	3.6	4	3.2
[3, 4)	49	2.8	19	2.7	9	7.2
[4, 5)	21	1.2	7	1.0	4	3.2
[5, 10)	66	3.7	14	2.0	8	6.4
[10, 15)	14	0.8	7	1.0	–	–
[15, 20)	12	0.7	4	0.6	–	–
[20, 25)	2	0.1	–	–	–	–
[25, 30)	1	0.1	1	0.1	–	–
[30, 35)	2	0.1	2	0.3	–	–
[35, 40)	–	–	1	0.1	–	–
[40, 45)	–	–	–	–	–	–
[45, 50)	–	–	–	–	–	–
[50, 55)	–	–	–	–	–	–
[55, 60)	–	–	–	–	–	–
[60, 65)	–	–	–	–	–	–
[65, 70)	1	0.1	–	–	–	–
[70, 75)	–	–	–	–	–	–
[75, 80)	–	–	–	–	–	–
[80, 85)	–	–	–	–	–	–
[85, 90)	1	0.1	–	–	–	–
≥ 90	–	–	1	0.1	–	–
Total	1763	100	702	100	125	100

Tab. 6: Statistics of the length of the glacier flowline in 1970s/80s for Jotunheimen, the European and New Zealand Alps, and sub-regions

Region / L_0 [km]	European Alps	New Zealand Alps	Jotunheimen	Jotunheimen West	Jotunheimen East
Mean	1.63	1.58	1.64	1.73	1.50
Max	24.70	28.50	5.12	4.56	5.12
Min	0.30	0.05	0.33	0.33	0.53
Std	1.53	1.99	1.09	1.13	0.98
Number [n]	1763	702	125	78	47

er compared with the value of the New Zealand Alps and Jotunheimen (Fig. 6). The mean value differs ~100 m between the eastern (1896 ± 94 m a.s.l.) and the western part (1787 ± 113 m a.s.l.) of Jotunheimen. The range of values in Jotunheimen is much smaller than in the other two regions. The calculated total volume, based on an inverse ice-flow-law, is 126 km³ for the European, 60 km³ for the New Zealand Alps, and 6 km³ for Jotunheimen. These volumes correspond to a potential sea-level rise of 0.40 mm for the European, 0.18 mm for the New Zealand Alps, and 0.02 mm for Jotunheimen (calculated after IPCC (2007)). The mean slopes in Jotunheimen are much more gentle (18.3°) than in the European (24.3°) and the New Zealand Alps (28.6°), as well as the maximum values (Fig. 7). A difference is recognisable between East and West Jotunheimen: the eastern part is slightly more gentle (17.2°) than the western part (18.9°). Calculated ablation at the tongue, b_t, (Tab. 7) shows the lowest mean value in Jotunheimen (1.6 m/a), higher values in the European Alps (2.4 m/a) and twice as high ablation rates in New Zealand. The maximum value of b_t is highest in the New Zealand

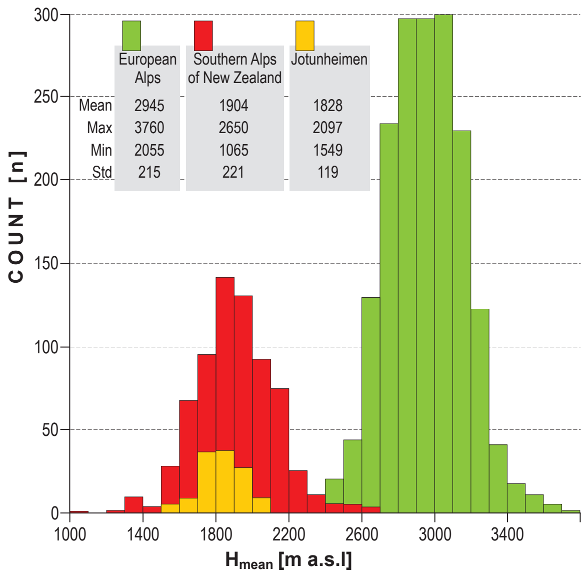

Fig. 6: Distribution of glacier mean altitude (H_{mean}) during 1970s/80s in Jotunheimen, and the European and New Zealand Alps

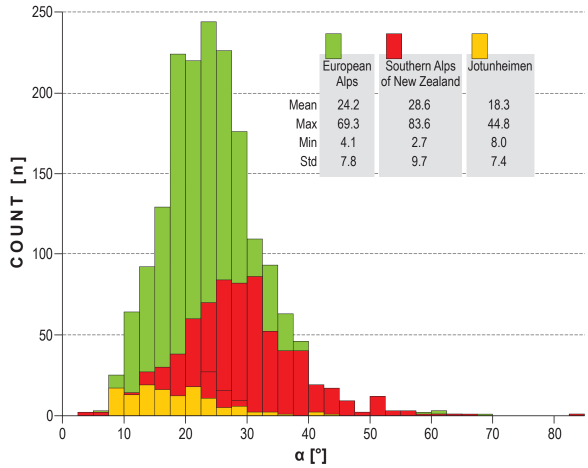

Fig. 7: Distribution of glacier surface slope (α) during 1970s/80s in Jotunheimen, and the European and New Zealand Alps

Alps (23.9 m/a; Fox Glacier), less in the European Alps (13.5 m/a; Bossons Glacier) and lowest in Jotunheimen (4.5 m/a; Styggedalsbreen). The difference between West and East Jotunheimen is considerable (Table 7). The response time is calculated by Equation (2) and this theoretical mean time needed to reach equilibrium is calculated to 37.4 years in the European, 35.7 years in the New Zealand Alps, and 67.0 years in Jotunheimen (Fig. 8). Glaciers in the western part of Jotunheimen reach equilibrium faster than those to the East. The depth-averaged mean flow velocity along the central flowline in the ablation area is taken as the assumption of the mean surface flow velocity along the central flowline in the ablation area. It varies between 15.7 m/a in the European Alps, 18.1 m/a in Jotunheimen, and 36.9 m/a in the New Zealand Alps. The mean surface velocity is about four times higher in the western as compared with the eastern part of Jotunheimen.

The velocity ratio of sliding and surface flow velocity in the ablation area is a measure for the glacier dynamics and gives information about the proportion between sliding and total velocity (Hoelzle 1994). The mean value ranges between 0.89 in the European, 0.95 in the New Zealand Alps, and 0.96 in Jotunheimen. However, because of uncertainties involved within the flow-law parameters, the calculated values of the velocity ratios are uncertain within a wide range (Haeberli and Hoelzle 1995).

5.2 Comparison of LIA reconstruction

The major advantage for Jotunheimen and as compared to both other regions is the availability of a glacier inventory for the LIA maximum in Jotunheimen. Therefore, the parameterization could be applied in Jotunheimen in the same way as for the 1980s. For the European and the New Zealand Alps, no comparable inventory data set is available; only selected parts of the European Alps have LIA inventories (e.g. Austria (Gross 1987)). Surface area,

Tab. 7: Statistics for the parameterization results of the ablation at the glacier tongue in the ablation area in 1970s/80s for Jotunheimen, the European and New Zealand Alps, and sub-regions

Region / b_t [m/a]	European Alps	New Zealand Alps	Jotunheimen	Jotunheimen West	Jotunheimen East
Mean	2.4	4.8	1.6	2.1	0.6
Max	13.5	23.9	4.5	4.5	1.2
Min	0.2	0.2	0.2	0.7	0.2
Std	1.5	3.6	1.0	0.8	0.2
Number [n]	1763	702	125	78	47

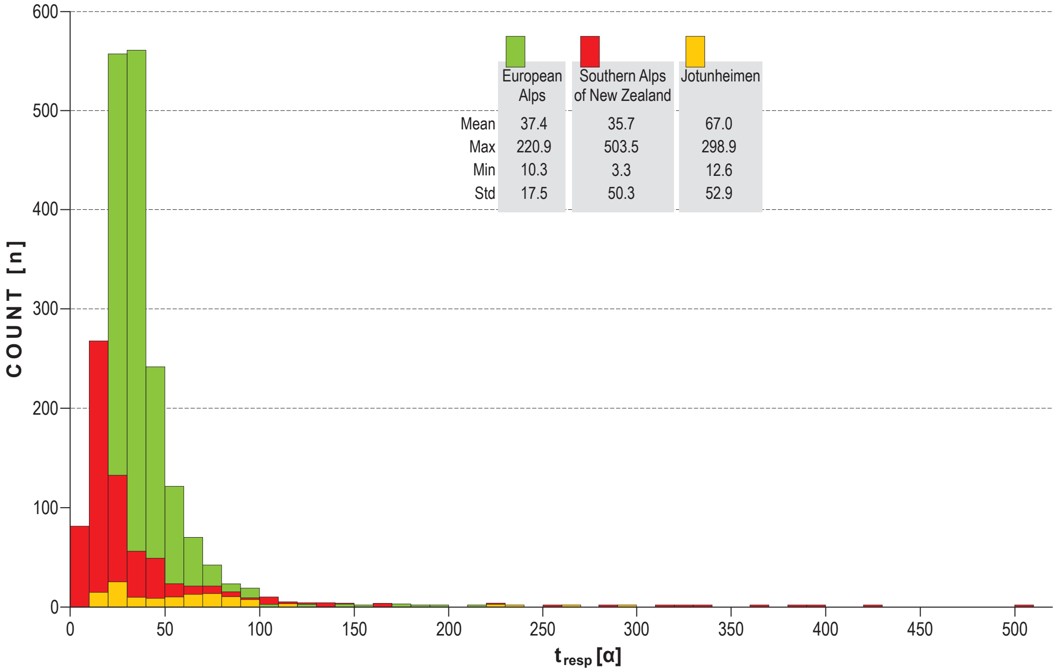

Fig. 8: Distribution of glacier response time (t_{resp}) during 1970s/80s in Jotunheimen, and the European and New Zealand Alps

length, volume, and mean specific mass balance were reconstructed for the European Alps by Haeberli and Hoelzle (1995) and for the New Zealand Alps by Hoelzle et al. (2007), and were calculated slightly differently than shown in Appendix A (for further information see references mentioned above). Hence, the comparison is performed with only these four variables.

An overview of the relative development of the glacier surface area between the LIA maximum and 1970s/80s is given in table 8 and figure 9. The relative area loss between the LIA maximum and the 1970s/80s is highest for the New Zealand Alps and lowest for Jotunheimen. The more maritime and more continental sub-regions of the New Zealand Alps and Jotunheimen do not show large differences. All 'dry' glacier areas in New Zealand (North dry, East dry) had an extent of 261 km² at the LIA maximum and had reduced to 123 km² (- 53%) in 1978. The 'wet' glacier areas (East wet, West, Fjord) decreased by about 49% during the same time. In Jotunheimen, the reduction of area in the more maritime West was about the same as in the East.

The pattern of the relative volume change between the LIA maximum and the 1970s/80s is a little bit different than the area development (Tab. 9 and Fig. 9). Most of the relative volume is lost in the New Zealand Alps. The loss in the European Alps and Jotunheimen is quite similar. In the sub-regions, a difference is visible in the New Zealand Alps. The volume declined from 153 km³ during the LIA maximum to 61 km³ (- 60%) in 1978 in the 'wet' glacier areas, and from 17 km³ to 5 km³ (- 68%) in the 'dry' areas. The difference between the eastern and the western area in Jotunheimen is smaller compared with the sub-regions of New Zealand.

During the LIA maximum, most glacier lengths in all regions were in the interval [1.0; 5.0) km (Tab. 10a). In the 1970s/80s, most of the lengths of the flowlines were still found in this interval, but relatively less than before. The lengths decreased between these two points of time, and relatively more are found in the interval [0.5; 1.0) km. The maximum length decreased in the European Alps from 27.2 to 24.7 km (both Aletsch Glacier), in the New Zealand Alps from 29.8 to 28.5 km (both Tasman Glacier), and in Jotunheimen from 7.3 (Østre Memurubreen) to 5.1 km (Søndre Veobreen). An overview of all length intervals in the sub-regions is given in table 10b.

The mean specific net mass balance for the time period between the LIA maximum and the 1980s in Jotunheimen was linearly calculated in all three areas using several response times if needed (see formula in Appendix A). Mean values of -0.05 m w.e./a for the western part and -0.02 m w.e./a for the eastern

Tab. 8: Development of the glacier surface area between the 'Little Ice Age' maximum and 1970s/80s in Jotunheimen, the European and New Zealand Alps, and sub-regions. ΔS = Difference of surface area between two points of time. *Data from HOELZLE et al. (2007)

Region / S [km²]	European Alps*	New Zealand Alps*	Jotun-heim-en	New Zealand Alps* North dry	East dry	East wet	West	Fiord	Jotunheimen West	East
LIA	3914.61	1931.66	249.83	3.81	257.39	640.43	951.67	78.36	181.56	68.27
1970s/80s	2544.38	978.75	182.50	0.69	122.80	350.26	464.20	40.80	131.38	51.13
ΔS [km²]	1370.23	952.91	67.33	3.12	134.59	290.17	487.47	37.56	50.18	17.15
ΔS [%]	35.0	49.3	26.9	81.9	52.3	45.3	51.2	47.9	27.6	25.1

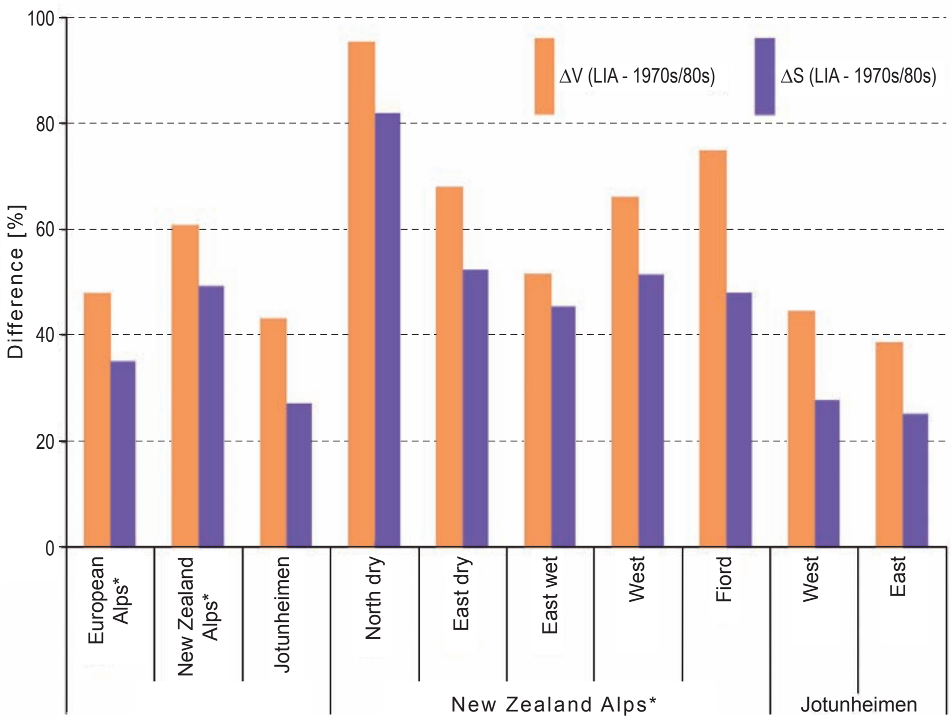

Fig. 9: Development of the glacier surface area and volume between the 'Little Ice Age' maximum and 1970s/80s in Jotunheimen, the European and New Zealand Alps, and sub-regions. Difference between two points of time in [%]. *Data from HOELZLE et al. (2007)

Tab. 9: Development of the glacier volume between the 'Little Ice Age' maximum and 1970s/80s in Jotunheimen, the European and New Zealand Alps, and sub-regions. ΔV = Difference of glacier volume between two points of time. *Data from HOELZLE et al. (2007)

Region / V [km³]	European Alps*	New Zealand Alps*	Jotun-heimen	New Zealand Alps* North dry	East dry	East wet	West	Fiord	Jotunheimen West	East
LIA	241.35	170.10	10.67	0.16	16.71	66.4	80.98	5.82	8.00	2.67
1970s/80s	126.00	66.77	6.08	0.01	5.38	32.37	27.56	1.48	4.44	1.64
ΔV [km³]	115.35	103.33	4.59	0.15	11.34	34.03	53.42	4.35	3.56	1.03
ΔV [%]	47.8	60.7	43.0	95.1	67.9	51.3	66.0	74.8	44.5	38.6

part were calculated. For the European Alps, a value of -0.33 m w.e./a was used. In New Zealand, the values varied between -0.67 m w.e./a in the 'West' area and -0.53 in the 'East dry' area. The values of Jotunheimen are much less negative than in the other two regions.

Tab. 10a: Classification of the length of the glacier flowline during the LIA maximum and in 1970s/80s into size intervals for Jotunheimen, and the European and New Zealand Alps. Numbers in [%]

Region / Length interval [km]	European Alps		New Zealand Alps		Jotunheimen	
	LIA	1970s	LIA	1978	LIA	1980s
Number [%]						
< 0.5	0.1	1.5	1.7	7.1	0.8	3.2
[0.5, 1.0)	3.6	32.7	21.8	38.3	16.0	35.2
[1.0, 5.0)	91.3	62.0	71.7	49.9	76.8	60.8
[5.0, 10.0)	4.1	3.5	3.6	3.6	6.4	0.8
≥ 10.0	0.9	0.3	1.3	1.1	–	–

Tab. 10b: Classification of the length of the glacier flowline during the LIA maximum and in 1978/80s into size intervals for sub-regions of Jotunheimen and the New Zealand Alps. Numbers in [n]

Region / Length interval [km]	New Zealand Alps								Jotunheimen			
	North dry		East dry		West		Fjord		West		East	
	LIA	1978	LIA	1978	LIA	1978	LIA	1978	LIA	1980s	LIA	1980s
Number [n]												
< 0.5	–	–	–	5	8	24	4	14	1	4	–	–
[0.5, 1.0)	–	–	6	55	62	88	30	30	8	23	12	21
[1.0, 5.0)	2	2	122	68	210	169	29	19	62	51	34	25
[5.0, 10.0)	–	–	–	–	17	17	–	–	7	–	1	1
≥ 10.0	–	–	1	1	5	4	–	–	–	–	–	–

6 Discussion

6.1 Selection of variables for Jotunheimen

The chosen db/dH values of 0.9 m w.e./100 m/a for the western and 0.3 m w.e./100 m/a for the eastern part of Jotunheimen are only slightly higher than the calculated ones using the steady-state years (Tab. 1). Reasons for this selection are based on the calculations of the parameterization, especially by the values of the ablation at the glacier tongue (see Equ. (3)). An overview of b_t-values calculated with different values of db/dH is given in table 2. The value for the eastern part was chosen on the basis of the ratio between calculated values from the three mass balance glaciers (Tab. 1). db/dH of Gråsubreen is about one third of db/dH of Stor- and Hellstugubreen, respectively. This ratio was transferred although the gradient of Gråsubreen itself does not seem reliable enough for the parameterization (see below).

There are no other possibilities for adjusting the mass balance gradient inherent in the parameterization. The mass balance gradient is included in the calculation of the mass balance disturbance (see Appendix A), but no measurements are available for this variable. db/dH is included in the calculations of the response time, the surface and the sliding velocities as well. However, there are no measurements available for these variables either. In contrast, measurements are available for the mean net mass balance (Tab. 11). Yet in the calculation, the mass balance gradient is included twice as calculation factor in the numerator and in the denominator and is, hence, reduced.

A consideration that concerns the mass balance gradient is the importance or validity of this variable. It is calculated by using vertical profiles of the net mass balance (Baumann 2010). Rasmussen and Andreassen (2005) found a weak correlation between the net mass balance gradient and the net mass balance on ten Norwegian glaciers (including STO, HEL, and GRA). This resulted from a generally positive correlation between the winter net mass balance and its corresponding gradient, and a negative one between the summer net balance and its gradient. The net mass balance gradient changes only little from year to year in Norway (Rasmussen and Andreassen 2005), and, therefore, the validity of the mass balance gradient concerning the mass balance is probably overestimated (cf. Winkler et al. (2009)).

The values of the mean net balance in Jotunheimen were considerably less negative from the LIA maximum until the 1980s as compared with the European and New Zealand Alps. $\bar{b}n$ in Jotunheimen had its most negative value between the 1980s and 2003, but was still not as negative as in the other two regions. Nonetheless, the calculated values for Stor- and Hellstugubreen fit well to the

Tab. 11: Comparison of the measured and calculated mean annual net mass balance at glaciers to mass balance measurements in Jotunheimen. Measured values from years 1980–2003, calculated values by the parameterization. Letter codes denote: STO = Storbreen, HEL = Hellstugubreen, GRA = Gråsubreen. (Raw data measured $\bar{b}$n: Andreassen et al. 2009)

Glacier / $\bar{b}_n$ [m/a]	STO	HEL	GRA
measured	-0.24	-0.35	-0.29
calculated	-0.26	-0.28	-0.04
difference	0.02	-0.07	-0.25

measured values between the 1980s and 2003, but not for Gråsubreen (Tab. 11). The bad adjustment of the net mass balance for Gråsubreen by the parameterization is due to the restrictions set by the ice-cored moraine and the possible cold-based glacier type. The mean annual air temperature at the ELA, as an expression of continental or maritime climatic conditions, is related to englacial temperatures, mass balance gradients and glacier/permafrost relationships (Zemp et al. 2008). These relationships may also be responsible for the low values of the mean net mass balance in Jotunheimen (see above).

The response time was calculated by using the basal shear stress (see equations in Appendix A), using a relationship between the basal shear stress and glacier mass turnover as defined by the mass balance gradient times the elevation range. A relationship between average basal shear stress and elevation range was empirically calibrated with data from reconstructed lateglacial glaciers in the Grison Alps of Switzerland (Hoelzle 1994; Haeberli and Hoelzle 1995) and varied in accordance with continental or maritime climatic conditions. The transferability of this formula was not tested for Jotunheimen and, hence, can only serve as the best assumption available. The formula could underestimate the basal shear stress for maritime and overestimate it for continental glaciers if these values should be more pronounced in Jotunheimen than in the Grison Alps. This uncertainty also influences the net mass balance.

Adjustment of the parameterization by the use of the three mass balance glaciers seems to be an appropriate method. As ablation at the glacier tongue is critical for the parameterization (Hoelzle 1994), goodness of the parameterization can only be improved by tuning using measured values. Hence, the difference between the periods 'LIA maximum – 1980s' and '1980s–2003' is probably not large enough for measured values to be transferred to this earlier period without adjustment. Nonetheless, it has to be mentioned that values for ablation at the glacier tongue derived from conventional mass balance data averaged over the entire glacier surface might not be reliable in every case (cf. Winkler et al. 2009; Winkler and Nesje 2009).

6.2 Comparison of the three regions

The three selected regions differ considerably regarding area and number of glaciers. The statistical population (n) was 125 glaciers in Jotunheimen, 702 in the New Zealand Alps, and 1763 glaciers in the European Alps. This difference has an influence on standard deviations (σ) because the higher n is, the lower σ is. Therefore, errors were included in comparing the standard deviations. A t-test was performed for the mean of the parameterization results of all comparable variables from the three regions (see Baumann 2010 and Appendix B). Half of the basic input variables were not significant at the 10% level. If a variable is denoted as not significant, n of this variable does not differ significantly between the two chosen data sets. Therefore, the mean of the variables showing no significance could be judged as values spreading around the 'real' mean of the theoretical 'entire' statistic population. This would result in obtaining very similar data sets in the comparison, but would not affect their reliability.

The lowest ablation at the glacier tongue is calculated for Jotunheimen implicating a low mass turnover as typical for continental regimes (Winkler 2009). The considerable difference between West and East Jotunheimen (see section 5.1) depends mainly on the different mass balance gradients (see Equ. (3)). However, ablation at the tongue is still higher in the western part if calculated with db/dH = 0.3 m w.e./100 m/a. The estimated response times of all three regions are highest in Jotunheimen. Response times mainly depend on the glacier slope, i.e. smoother glaciers exhibit longer response times. This relationship is well visible in the calculated data. West Jotunheimen exhibits a much shorter response time, but the mean value is nearly similar to the East if calculated with the same mass balance gradient. The highest mean surface velocity of all three regions is calculated for the New Zealand Alps, which is a sign of maritime influence. Jotunheimen and the European Alps show about the same value which is half of the one for New Zealand. The western part of Jotunheimen shows a faster flow than the eastern part, also if calculated with db/dH =

0.3 m w.e./100m/a. The gentle slopes and the very low mean net balance in Jotunheimen also imply a more continental regime (Winkler and Nesje 2009). Following independent studies on the recent glacier behaviour in Norway (e.g. Andreassen et al. 2005; Winkler et al. 2009), the differentiation in a more maritime western and a more continental eastern part seems appropriately verified and not only dependent on the different mass balance gradients. Following the parameterization results, there is a clear order: The New Zealand Alps are the most maritime region; the European Alps show a transient regime; and Jotunheimen exhibits the most continental one in this comparison.

7 Conclusion

The parameterization of Jotunheimen and the comparison with data from the European and the New Zealand Alps was performed successfully. The main findings are:

The parameterization showed a consistent pattern of the glacier behaviour and glacier regime in Jotunheimen, although the design of the calculations was originally developed for larger glaciers sizes (Haeberli and Hoelzle 1995; Hoelzle et al. 2007).

The results obtained show that the mean annual glacier net balance for the period between LIA maximum and the 1980s in Jotunheimen (-0.05 m w.e.a^{-1} for the western and -0.02 m w.e.$^{-1}$ for the eastern part) were much less negative than for the European (-0.33 m w.e.$^{-1}$) and New Zealand Alps (-0.53 – -0.67 m w.e.$^{-1}$), respectively. The ablation at the glacier tongue, b_t, had its lowest mean value in Jotunheimen (1.6 m/a), slightly higher in the European Alps (2.4 m/a) and twice as high in the New Zealand Alps (4.8 m/a). Nearly the same relationship in reversed order was estimated for the mean response time: 36 a in the New Zealand, 37 a in the European Alps, and 67 a in Jotunheimen. The relative area as well as the relative volume loss was lowest in Jotunheimen and highest in the New Zealand Alps.

It is concluded that the New Zealand Alps have to be characterized as the most maritime, and Jotunheimen as the most continental region of the regions compared in this study. The European Alps have a transient regime.

Analysing the relationship between climate and glacier behaviour using the parameterization results seems possible. For the time steps chosen, the climate situation is mirrored in the parameterization results.

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Appendix A: Overview of parameterization parameters and variables with term, formula, and units for the example Jotunheimen

Name	Term	Calculation	Unit
Surface area	S	–	km^2
Length	L_0	–	km
Minimum altitude	H_{min}	–	m a.s.l.
Maximum altitude	H_{max}	–	m a.s.l.
Length change	δL	$L_{0,old}-L_{0,new}$	m
Mean altitude	H_{mean}	$(H_{max}+H_{min})/2$	m a.s.l.
Equilibrium line altitude	ELA	$\sim H_{mean}$	
Range	ΔH	$H_{max}-H_{min}$	m
Length of the central flowline in ablation area	L_a	$0.5*L_0$ für $L_0 \leq 2$ km; $0.75*L_0$ für $L_0 > 2$ km	km
Average surface slope	α	$\arctan(\Delta H/L_0)$	rad
Average surface slope in ablation area	α_a	$\arctan[(H_{mean}-H_{min})/L_a]$	rad
Mean basal shear-stress	τ	$0.005+1.598*\Delta H-0.435*(\Delta H)^2$ für $\Delta H \leq 1.6$; 1.5 für $\Delta H > 1.6$	bar
Average ice thickness at central flowline	h_f	$\tau/(f*\varrho*g*\sin\alpha)$	m
Average ice thickness at central flowline in ablation area	$h_{f,a}$	$\tau/(f*\varrho*g*\sin\alpha_a)$	m
Average ice thickness over whole glacier	h_F	$(\pi/4)*h_f$	m
Total glacier volume	V	$F*h_F$	km^3
Maximum ice thickness	h_{max}	$2.5*h_{f,a}$	m
Depth-averaged mean flow velocity along central flowline in ablation area	$u_{m,a}$	$[(3*b_t/4)*(L_a/2)]/h_{f,a}$	m/a
Velocity of ice deformation	$u_{d,a}$	$2*A*\tau^n*h_f/(n+1)$	m/a
Sliding velocity in ablation area	$u_{b,a}$	$u_{s,a}-u_{d,a}$	m/a
Mean surface flow velocity along central flowline in ablation area	$u_{s,a}$	$\sim u_{m,a}$	m/a
Velocity ratio	VR	$u_{b,a}/u_{s,a}$	–
Response time	t_{resp}	h_{max}/b_t	a
Reaction time	t_{react}	L_a/c	a
Relaxation time	t_{relax}	$t_{resp}-t_{react}$	a
Kinematic wave velocity	c	$4*u_{s,a}$	m/a
Annual ablation at glacier tongue	b_t	$(db/dH)*(H_{mean}-H_{min})$	m w.e./a
Change in ELA	δELA	$(H_{min,old}-H_{min,new})/2$	m
Mass balance disturbance	δb	$\delta ELA*(db/dH)$	m w.e./a
Mean mass balance	$\bar{b}_n$	$\delta b/(2*n_{resp})$	m w.e./a
Gravitational acceleration	g	9.81	$m*s^{-2}$
Density of ice	ϱ	917	$kg*m^{-3}$
Flow parameter	A	0.16	$a^{-1}*bar^{-3}$
Flow parameter	n	3	–
Shape factor	f	1	–
Mass balance gradient	db/dH	–	m w.e./100m*a^{-1}

Appendix B: Results of a t-test between Jotunheimen – European Alps – New Zealand Alps to compare two means from the respective data sets and to analyze the significant independence of the variables

The t-test was performed to compare to means from individual probes and to analyse the significant independence of the probes. The test shows, whether the difference of the two means is significantly different from zero (null hypothesis). The equations are taken from Bortz (2005).

The standard deviation of the difference of two means is:

$$\hat{\sigma}_{\bar{x}_1-\bar{x}_2} = \left(\sqrt{\frac{\hat{\sigma}_1^2}{n_1}+\frac{\hat{\sigma}_2^2}{n_2}}\right)$$

$\hat{\sigma}_{\bar{x}_1-\bar{x}_2}$ = assumed standard deviation of the difference of the means,
$\hat{\sigma}_1^2$ = assumed variance of statistic population 1,
$\hat{\sigma}_2^2$ = assumed variance of statistic population 2,
n_1 = statistic population of probe 1, n_2 = statistic population of probe 2.

The assumed standard deviation of the difference of the means is needed to calculate

$$\hat{t} = \frac{\bar{x}_1-\bar{x}_2}{\hat{\sigma}_{\bar{x}_1-\bar{x}_2}}$$

Taken a two-tailed significance level α with a calculated degree of freedom (Φ),

$\Phi = n_1+n_2-2$,
the significance is tested by $\hat{t} > t_{\alpha;\Phi}$.

This method was performed with all means from the parameterization in the comparison:
- Jotunheimen – European Alps;
- Jotunheimen – New Zealand Alps;
- European Alps – New Zealand Alps.

Jotunheimen (1) – European Alps (2) – Parameterization data

$n_1 = 125$, $n_2 = 1763$, $\Phi = 125 + 1763 - 2 = 1886$;
$t_{5\%,\,1886} = 1.96$, $t_{1\%,\,1886} = 2.576$, $t_{10\%,\,1886} = 1.645$;
Results: 0 = no, 1 = yes.

1980s/70s

Var.	$\bar{x}_1$	$\bar{x}_2$	$\hat{\sigma}_2^2$	$\hat{\sigma}_2^2$	$\hat{\sigma}$	$\hat{t}$ abs.	$\hat{t} > t_{5,\Phi}$	$\hat{t} > t_{1,\Phi}$	$\hat{t} > t_{10,\Phi}$
S	1.46	1.44	3.04	13.59	0.18	0.09	**0**	**0**	**0**
$\mathbf{H}_{max}$	2051	3271	21803	103736	15.27	79.83	1	1	1
$\mathbf{H}_{min}$	1607	2620	21402	69922	14.52	69.79	1	1	1
$\mathbf{L}_0$	1.65	1.63	1.17	2.34	0.10	0.17	**0**	**0**	**0**
$\mathbf{L}_a$	1.04	1.01	0.80	1.47	0.09	0.39	**0**	**0**	**0**
$\mathbf{H}_m$	1829	2945	13922	46022	11.73	95.22	1	1	1
α	0.31	24.25	0.02	60.90	0.19	128.52	1	1	1
τ	0.62	0.79	0.04	0.10	0.02	8.55	1	1	1
$\mathbf{h}_f$	26.41	30.08	231.96	313.07	1.43	2.57	1	**0**	1
$\mathbf{h}_{f,a}$	32.81	23.62	650.01	193.06	2.30	3.99	1	1	1
V_E	0.05	0.07	0.01	0.17	0.01	1.83	**0**	**0**	1
$\mathbf{h}_{max}$	82	89	4063	5195	5.95	1.16	**0**	**0**	**0**
$\mathbf{b}_t$	1.56	2.44	0.98	2.29	0.10	9.22	1	1	1
$\mathbf{u}_{m,a}$	18.12	15.67	164.89	455.70	1.26	1.95	**0**	**0**	1
$\mathbf{u}_{d,a}$	0.91	2.58	1.56	21.45	0.16	10.62	1	1	1
$\mathbf{u}_{b,a}$	17.20	13.09	143.02	311.30	1.15	3.58	1	1	1
VR	0.96	0.89	0.00	0.01	0.01	11.83	1	1	1
$\mathbf{t}_{resp}$	68.94	18.02	3229.48	50.54	5.09	10.01	1	1	1
$\mathbf{t}_{react}$	0.02	37.37	0.00	306.47	0.42	89.58	1	1	1
$\mathbf{t}_{relax}$	68.93	19.35	3227.76	130.94	5.09	9.74	1	1	1

LIA maximum

Var.	$\bar{x}_1$	$\bar{x}_2$	$\hat{\sigma}_2^2$	$\hat{\sigma}_2^2$	$\hat{\sigma}$	$\hat{t}$ abs.	$\hat{t} > t_{5,\Phi}$	$\hat{t} > t_{1,\Phi}$	$\hat{t} > t_{10,\Phi}$
S	2.00	2.22	5.28	15.91	0.23	0.98	**0**	**0**	**0**
$\mathbf{L}_0$	2.23	2.30	2.21	2.86	0.14	0.47	**0**	**0**	**0**
V	0.09	0.11	0.02	0.05	0.01	1.65	**0**	**0**	**0**

Jotunheimen (1) – New Zealand Alps (2) – Parameterization data

$n_1 = 125$, $n_2 = 702$, $\Phi = 125 + 702 - 2 = 825$;
$t_{5\%, 825} = 1.96$, $t_{1\%, 825} = 2.576$, $t_{10\%, 825} = 1.645$;
Results: 0 = no, 1 = yes.

1980s/70s

Var.	$\bar{x}_1$	$\bar{x}_2$	$\hat{\sigma}_2^2$	$\hat{\sigma}_2^2$	$\hat{\sigma}$	$\hat{t}$ **abs.**	$\hat{t} > t_{5,\Phi}$	$\hat{t} > t_{1,\Phi}$	$\hat{t} > t_{10,\Phi}$
S	1.46	1.40	3.04	23.10	0.24	0.26	**0**	**0**	**0**
$\mathbf{H_{max}}$	2051	2238	21803	83859	17.14	10.88	1	1	1
$\mathbf{H_{min}}$	1607	1543	21402	95120	17.51	3.67	1	1	1
$\mathbf{L_0}$	1.65	1.58	1.17	3.97	0.12	0.56	**0**	**0**	**0**
$\mathbf{L_a}$	1.04	1.19	0.80	2.23	0.10	1.48	**0**	**0**	**0**
$\mathbf{H_m}$	1829	1904	13922	48816	13.45	5.55	1	1	1
α	0.31	0.50	0.02	1.63	0.05	3.73	1	1	1
τ	0.62	0.83	0.04	0.10	0.02	9.61	1	1	1
$\mathbf{h_f}$	26.41	28.37	231.96	393.03	1.55	1.26	**0**	**0**	**0**
V_E	0.05	0.09	0.01	0.43	0.03	1.41	**0**	**0**	**0**
$\mathbf{h_{max}}$	82	105	4063	7665	6.59	3.48	1	1	1
$\mathbf{b_t}$	1.56	4.79	0.98	13.06	0.16	19.89	1	1	1
$\mathbf{u_{m,a}}$	18.12	36.86	164.89	2029.63	2.05	9.13	1	1	1
$\mathbf{u_{d,a}}$	0.91	2.87	1.56	29.30	0.23	8.38	1	1	1
$\mathbf{u_{b,a}}$	17.20	33.99	143.02	1661.16	1.87	8.96	1	1	1
VR	0.93	0.95	0.00	0.01	0.01	1.76	**0**	**0**	1
$\mathbf{t_{resp}}$	68.94	35.68	3229.48	2526.41	5.43	6.13	1	1	1
$\mathbf{t_{react}}$	0.02	11.57	0.00	135.11	0.44	26.33	1	1	1
$\mathbf{t_{relax}}$	68.93	19.35	3227.76	130.94	5.09	9.74	1	1	1
$\mathbf{h_F}$**_E**	26.48	22.28	228.96	242.41	1.48	2.84	1	1	1
$\mathbf{h_F}$**_T**	22.47	18.91	164.97	174.71	1.25	2.84	1	1	1
V_T	0.07	0.07	0.02	0.31	0.02	0.01	**0**	**0**	**0**
$\bar{\mathbf{b}}$**n**	-0.04	-0.41	0.00	0.12	0.01	26.89	1	1	1

LIA maximum

Var.	$\bar{x}_1$	$\bar{x}_2$	$\hat{\sigma}_2^2$	$\hat{\sigma}_2^2$	$\hat{\sigma}$	$\hat{t}$ **abs.**	$\hat{t} > t_{5,\Phi}$	$\hat{t} > t_{1,\Phi}$	$\hat{t} > t_{10,\Phi}$
S	2.00	1.73	5.28	489.40	0.86	0.32	**0**	**0**	**0**
$\mathbf{L_0}$	2.23	2.00	2.21	4.55	0.16	1.48	**0**	**0**	**0**

European Alps (1) – New Zealand Alps (2) – Parameterization data

$n_1 = 1763$, $n_2 = 702$, $\Phi = 1763 + 702 - 2 = 2463$;
$t_{5\%, 2463} = 1.96$, $t_{1\%, 2463} = 2.576$, $t_{10\%, 2463} = 1.645$;
Results: 0 = no, 1 = yes.

1980s/70s

Var.	$\bar{x}_1$	$\bar{x}_2$	$\hat{\sigma}_2^2$	$\hat{\sigma}_2^2$	$\hat{\sigma}$	$\hat{t}$ abs.	$\hat{t} > t_{5,\Phi}$	$\hat{t} > t_{1,\Phi}$	$\hat{t} > t_{10,\Phi}$
S	1.44	1.40	13.59	23.10	0.20	0.22	**0**	**0**	**0**
H_{max}	3271	2238	103736	83859	13.35	77.35	1	1	1
H_{min}	2620	1543	69922	95120	13.23	81.43	1	1	1
L_0	1.63	1.58	2.34	3.97	0.08	0.60	**0**	**0**	**0**
L_a	1.01	1.19	1.47	2.23	0.06	2.82	**0**	**0**	**0**
H_m	2945	1904	46022	48816	9.78	106.53	1	1	1
α	24.25	0.50	60.90	1.63	0.19	123.68	1	1	1
τ	0.79	0.83	0.10	0.10	0.01	2.94	1	1	1
h_f	30.08	28.37	313.07	393.03	0.86	1.99	**0**	**0**	**0**
V_E	0.07	0.09	0.17	0.43	0.03	0.51	**0**	**0**	**0**
h_{max}	88.90	104.95	5195.01	7665.19	3.72	4.31	1	1	1
b_t	2.44	4.79	2.29	13.06	0.14	16.67	1	1	1
$u_{m,a}$	15.67	36.86	455.70	2029.63	1.77	11.94	1	1	1
$u_{d,a}$	2.58	2.87	21.45	29.30	0.23	1.22	1	1	1
$u_{b,a}$	13.09	33.99	311.30	1661.16	1.59	13.11	1	1	1
VR	0.89	0.95	0.01	0.01	0.00	14.01	1	1	1
t_{resp}	18.02	35.68	50.54	2526.41	1.90	9.27	1	1	1
t_{react}	37.37	11.57	306.47	135.11	0.61	42.63	1	1	1

LIA maximum

Var.	$\bar{x}_1$	$\bar{x}_2$	$\hat{\sigma}_2^2$	$\hat{\sigma}_2^2$	$\hat{\sigma}$	$\hat{t}$ abs.	$\hat{t} > t_{5,\Phi}$	$\hat{t} > t_{1,\Phi}$	$\hat{t} > t_{10,\Phi}$
S	2.22	1.73	15.91	489.40	0.84	0.59	**0**	**0**	**0**
L_0	2.30	2.00	2.86	4.55	0.09	3.28	1	1	1

Caption of all tables in Appendix B that have not been explained in the text or Appendix A:

abs. = absolute
h_F_E = Ice thickness over whole glacier with elliptic bed geometry
h_F_T = Ice thickness over whole glacier with triangular bed geometry
Var. = Variable
V_E = Volume with elliptic bed geometry
V_T = Volume with triangular bed geometry
$\hat{\sigma} = \hat{\sigma}_{\bar{x}_1 - \bar{x}_2}$