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Last Glacial Maximum climatic conditions in the Polish part of the High Tatra Mountains (Western Carpathians)

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Palaeoclimatic conditions in the Polish part of the High Tatra Mountains during the Last Glacial Maximum (LGM) were reconstructed based on the former glacial topography with the use of two independent glacier-climate models. The extent of the palaeoglaciers was determined using the glacial-geomorphological record of terminal and lateral moraines as well as trimlines. Two north-faced prominent glaciers were reconstructed (Sucha Woda/Pańszczyca and Biała Woda) with their surface areas as 15.2 and 40.3 km^2, respectively. The equilibrium line altitudes (ELA) of these glaciers were determined at the level of 1460 and 1480 m a.s.l. Modelled palaeoclimatic parameters show a mean summer temperature of about 0.3°C and mean annual precipitation of around 580 mm at the equilibrium line altitude of the former glaciers. This means that the summer temperature was lower by 10°C and precipitation was lower by about 60% in relation to the modern conditions. The mean annual temperature was lower by at least 12°C. On the basis of palaeoclimatic data the modern snowline altitude was established at the level of 2450–2550 m a.s.l. This indicates an ELA depression of 1000–1100 metres. Reconstructed climatic parameters at the ELA and ablation gradients indicate that both glaciers were similar to the modern glaciers in Canadian Arctic. Such climatic conditions in Central Europe indicate a cold and dry climate characteristic of the subarctic zone.

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INTRODUCTION

The Last Glacial Maximum (LGM) is defined as the latest time interval in Earth history, when glaciers reached their maximum volume. This is widely inferred based on sea level minima record across the globe (Lambeck *et al*., 2002). The first pulse of the most recent sea level low-stand is marked at around 30 000 years ago and its maximum occurred between 26 500 and 19 000 years ago (Lambeck *et al*., 2002; Clark *et al*., 2009). The onset of this maximum corresponds well with the time period when almost all ice sheets attained their maximum extent. At around 24 000 years ago the lowest temperatures occurred during Heinrich Event 2 (Stuiver and Grootes, 2000). This was interpreted based on Greenland ice-core analysis and this episode is defined as Greenland Stadial 3 (Lowe *et al*., 2008).

The Tatra Mountains are located in the Western Carpathians, in the transition zone between the European Alps and the Balkan orogens (Fig. 1). They are the highest massif of the Carpathians and rise about 2000 metres above surrounding depressions, reaching up to 2655 m a.s.l. (Gerlach). The Tatra Mountains are the northernmost Carpathic mountain range glaciated during the Pleistocene. Therefore, they are a significant area for the reconstruction of palaeoclimatic conditions in Central Europe between the area formerly covered by the Fennoscandian Ice Sheet and the mountain ranges of Southern Europe.

Mountain glaciers are sensitive indicators of climate changes. Their mass balance is strictly determined by climatic conditions. Former glacier extents marked by moraines and trimlines can reflect temporal fluctuations of accumulation and ablation. Formation of prominent terminal moraines or glacial trimlines requires glacier activity on average in a steady – state condition for at least a few decades. Hence, it is possible to obtain the annual amount of precipitation and the summer temperature at the equilibrium line altitude of stable glaciers.

In this study we focus on the glacial forms which give the basis for reconstruction of the glaciers' geometry. For this purpose the distribution of the most extended glacial trimlines, lateral moraines and terminal moraines was used.

The aim of this paper is to reconstruct climatic conditions in the High Tatra Mountains during the LGM, using a simple gla-

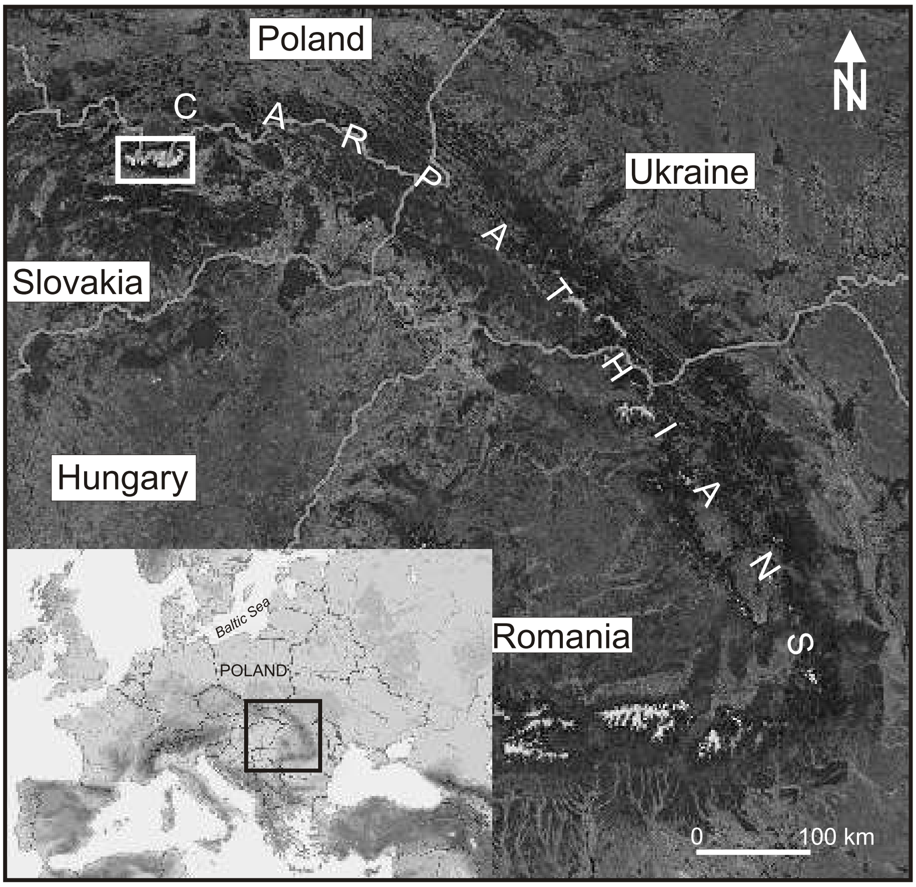

Fig. 1. Location of the Tatra Mountains

cier – climate model (Kull and Grosjean, 2000; Ivy-Ochs *et al*., 2006) and a one-dimensional flow model (Sarikaya *et al*., 2008, 2009).

STUDY AREA

The Polish part of the High Tatra Mountains contains four large north-faced valleys (Sucha Woda, Pańszczyca, Pięć Stawów Polskich and Rybi Potok). They built two large drainage basins: the Biała Woda drainage basin, which in part is located on Slovak territory and the Sucha Woda/Pańszczyca drainage basin (Fig. 2). In general, the northern exposure of the study area permitted the accumulation of large ice masses. Thus, glacial cirques and troughs on the northern and northeastern slopes of the High Tatra Mountains are the deepest incised into the entire massif. A predominant part of the study area is composed of resistant granodiorite. In the north, the granitic intrusion is covered with Mesozoic nappes which built the northernmost, lower part of drainage basins (Bac-Moszaszwili *et al*., 1979).

The very prominent glacial relief of the High Tatra Mountains is an effect of glacial activity during eight glaciations (Lindner *et al*., 2003). However, the last glacial cycle (Marine Isotope Stage 2 – MIS 2) had a decisive influence on formation and development of glacial forms. The glacial erosional features located in the upper parts of glacial valleys were likely formed during the LGM and the Late Glacial (Makos, 2008, 2010). The lowermost, well-preserved moraines located far downvalley were formed during the LGM also, as inferred from exposure dating (^{36}Cl; Dzierżek, 2009) and luminescence dating (OSL; Baumgart-Kotarba and Kotarba, 2002). The timing of the LGM in the High Tatra Mountains is limited to between 25 000 and 21 000 years ago. Moraine deposits north of the LGM moraines are considered to be the remnants of older glaciations (Klimaszewski, 1988; Baumgart-Kotarba and Kotarba, 1997) or older stadials of the last glacial period (Lindner, 1994; Lindner *et al*., 2003; Dzierżek, 2009). Currently, the High Tatra Mountains are free of glacial ice and only small perennial snow patches or small glacierets supplied by avalanches occupy north-faced niches or glacial cirques below steep and high rock walls (Gądek, 2008).

The modern snowline altitude on the northern slope of the Tatra Mountains has been estimated at between 2200 m a.s.l. (Hess, 1996) and 2500–2600 m a.s.l. (Zasadni and Kłapyta, 2009). This discrepancy is an effect of different criteria used in estimations. According to Hess (1965), even the lower location of the snowline at an altitude of 2200 m a.s.l., well below the highest summits, does not allow the production of glaciers, due to topographic conditions. There is no place above the elevation of 2200 m a.s.l. suitable for longer accumulation of snow and ice.

The present-day climatic conditions in the Tatra Mountains are determined by the prevalent northwestern circulation and transport of humid, polar air masses from the North Atlantic. This gives the highest amounts of annual precipitation on the northwestern slope of the massif while on the southern side the amount of total precipitation is clearly lower (Niedźwiedź, 1992). This spatial distribution of precipitation is also characteristic for other, E–W oriented mountainous areas in Central Europe e.g., the Alps (Florineth and Schlüchter, 2000; Kuhlemann *et al*., 2009). The mean annual temperature changes along the vertical profile from 6°C at the foot of the massif to –4°C at the highest peaks. The precipitation gradient is more complicated and its course along the vertical profile is non-linear with a decreasing trend above an altitude of 2000 m a.s.l. during the summer season (Hess, 1996). In winter, precipitation rises with elevation. In general, water-vapour condensa-

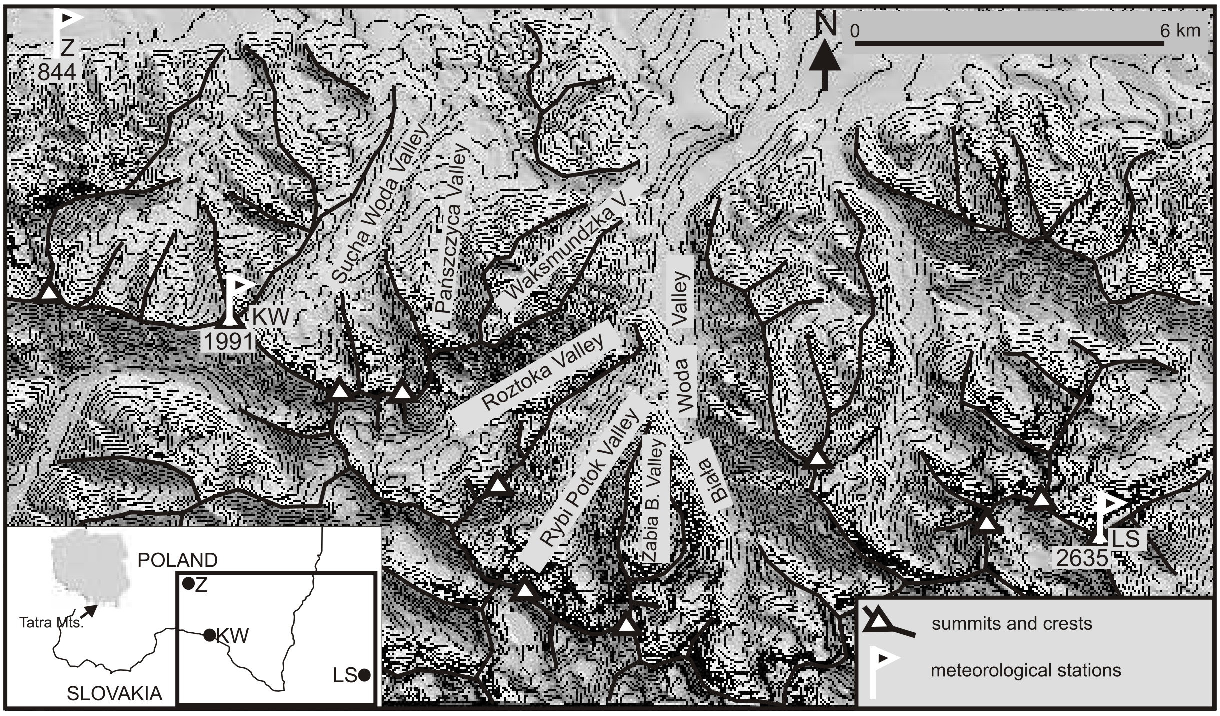

Fig. 2. Location of the study area

KW – Kasprowy Wierch, LS – Lomnicky Štit, Z – Zakopane; contour line interval is 50 metres

tion totals also rise with elevation (Hess, 1996). The natural timberline lies at 1550 m a.s.l. (Obidowicz, 1996) on the northern slope of the Tatra Mountains on the opposite side this level is placed 100 metres higher (Plesník, 1971).

MATERIALS AND METHODS

RECONSTRUCTION OF THE GLACIER GEOMETRY

Mapping of glacial forms was applied to reconstruct the geometry of two glaciers which occupied the drainage basins of Biała Woda and Sucha Woda/Pańszczyca. Fieldwork and analysis of a digital elevation model (DEM) of the study area allowed the distinction of many glacial features that determine the limits of glacier thickness and extent. The data obtained were supported by a wealth of archival materials, especially in the case of the location of terminal and lateral moraines (e.g., Kliamszewski, 1988; Baumgart-Kotarba and Kotarba, 1997, 2001). These forms are major indicators of glacier shape in an ablation area. In order to reconstruct geometry of the glacier above the ELA in the accumulation area, the location of glacial trimlines was used. These data were previously in part presented by Makos (2008) and Makos and Nowacki (2009). The term “trimline” relates to a boundary on the slope of glacial trough or glacial cirque, below which there is evidence for glacial erosion such as polish, and above which the bedrock is frost-weathered and jagged (Thorp, 1981; Ballantyne, 1997; Kelly *et al*., 2004). It is assumed that glacial trimlines mark the elevation of the former surface of active glacier and hence, it gives us information about ice thickness. The uppermost trimlines (generation I) are thought to have formed during the last maximum extent of glaciers in the Tatra Mountains. Based on dating of the northernmost terminal moraines (Dzierżek *et al*., 1999; Dzierżek, 2009) in the Sucha Woda Valley it was inferred that glaciers reached their last maximum extent at least about 21 000 years ago. The latest exposure dating of abraded bedrock below glacial trimlines suggested that the generation I surface had been exposed between 21 500 and 16 500 years ago (Makos, 2010). Therefore, the onset of ice accumulation in the Tatra Mountains during the last cold stage occurred no later than 22 000 years ago. The exposure age of the most extended moraines and trimlines shows that both features were formed during the same glacial episode, that of Marine Isotope Stage 2 (MIS 2). Thus the geometry of LGM glaciers has been reconstructed based on the location of maximum terminal and lateral moraines and trimlines of generation I. This ice surface geometry was controlled by basal shear stress (τ_b) values along the centre line of the glacier. For most valley glaciers, the critical value of basal shear stress lies between 50 and 150 kPa (Paterson, 1994).

EQUILIBRIUM LINE ALTITUDE (ELA)

The equilibrium line altitude (ELA) is a boundary between the accumulation zone and the ablation zone. This line connects points where the mass balance of the glacier equals zero. The surface of the glacier in the vicinity of the ELA is usually flat. Above this boundary the glacier surface is concave while below it is convex due to the different trajectories of ice movement. In the mid-latitude mountain ranges the ELA often is located at the altitude of the climatic snowline (Jania, 1997). Thus it is possible to obtain the altitude of the climatic snowline from calculation of the average ELA of neighbouring glaciers in one mountain range (Gross *et al*., 1977). The ELA is one of the most important factors for palaeoclimatic reconstructions.

The ELA of palaeoglaciers is determined using different methods. The most popular and widely used is the AAR (accumulation area ratio) method. The AAR determines the relation between the accumulation area and the total surface of the glacier. Analyses of the ELA of modern glaciers show that this relation is approximately 2:1. An AAR value of 0.65 has been used for ELA reconstruction in formerly glaciated areas (e.g., Porter, 1975). Gross *et al.* (1977) inferred that an AAR value of 0.67 comprised the best fit to the modern and Late Glacial glaciers in the European Alps. The most recent research on the AAR of modern glaciers allowed determination of the dependence of the AAR on the total area of the glacier (Kern and László, 2010). The authors calculated that an AAR value of 0.44 ±0.04 is best applied to glaciers with an area in the range between 0.1–1 km^2, 0.54 ±0.07 for glaciers 1–4 km^2 in size and 0.64 ±0.04 for glaciers larger than 4 km^2. These data are based on calculation of the AAR on 46 modern glaciers. According to Kern and László (2010), the relation between accumulation area and the surface of the whole glacier is optimally described by a logarithmic function:

$$\mathrm{ssAAR} = 0.0648 \ln S + 0.483 \qquad [1]$$

where: ssAAR – size specific steady state AAR and S – surface of the glacier.

We used this equation to determine the AAR of the glaciers reconstructed in the Tatra Mountains.

PALAEOCLIMATIC RECONSTRUCTION

If the geometry of the palaeoglacier is reconstructed it is then possible to make a palaeoclimatic reconstruction with a simple glacier flow model and a glacier-climate model. The required parameters of the glacier (glacier width, ice thickness, surface slope) were taken from the a the GIS-based model of the glacier. Based on the topography of the glacier, the sensitive glaciological parameters (ELA, basal shear stress) were obtained. Then, the ice velocities in selected cross-sections were estimated. The ice velocity is a sum of the basal sliding velocity and the deformation velocity. Both were calculated with the use of parameters for basal sliding and ice deformation after Budd *et al.* (1979). For calculation of the mass balance of the glacier we used a simple glacier flow model, which calculates the net ablation along the tongue of the glacier below the ELA. Details can be found in Ivy-Ochs *et al.* (2006) and Kerschner and Ivy-Ochs (2008). First, the model calculates ice-flux through given cross-sections between the ELA and the end moraine. Then, the net ablation between two neighbouring cross-sections can be estimated. This is the relation between the ice-flux difference and the surface area of the glacier between these cross-sections. All necessary equations and variables are presented in Table 1.

The next step is calculation of the net ablation gradient ($\partial a/\partial z$) which is an input data to calculate annual precipitation (P). This is then used to calculate summer temperature (T_S).

Palaeoclimatological interpretation is based on the statistical models which relate annual precipitation (accumulation parameter) and summer temperature (ablation parameter) at the ELA of the former glacier. Based on analysis of the relation of

Table 1

Equations for the calculation of ice flow, mass flux and balance gradients
(*cf*. Kull and Grosjean, 2000; Ivy-Ochs *et al*., 2006)

Parameter	Equations and variables
Basal shear stress – τ_b [kPa]	$\tau_b = \rho g h F \sin\alpha$ ρ – density of ice (900 kgm^{-3}) g – acceleration due to gravity (9.81 ms^{-2}) h – ice thickness α – surface slope F – shape factor for a channel with a parabolic cross-section after Nye (1952), optional
Average ice velocity in cross-section – U [ma^{-1}]	$U = U_d + U_b = f_d \tau_b^3 h + f_b \tau_b^3/h\, c$ f_d – parameter for ice deformation (1.9 × 10^{-24}Pa^{-3}s^{-1}) f_b – parameter for basal sliping (5.7 × 10^{-20}Pa^{-3}m^{2}s^{-1}) both parameters after Budd *et al*. (1979) c – correction factor for sliding velocity, optional
Ice-flux through cross-section – Q [m^{3}a^{-1}]	$Q = UA$ A – area of the cross-section
Net ablation between neighbouring cross-sections – a [ma^{-1}]	$a = \Delta Q/S$ S – surface area between neighbouring cross-sections

precipitation and summer temperature, Ohmura *et al.* (1992) inferred that:

$$P = 9T_S^2 + 296T_S + 645 \qquad [2]$$

where: P – sum of winter accumulation and summer precipitation and T_S – summer (June–August) temperature.

Ivy-Ochs *et al.* (2006) re-analysed the data set of Ohmura *et al.* (1992) and presented the equation with T_S as the dependent variable:

$$T_S = 0.1815\ P^{0.5}–4.1 \qquad [3]$$

To obtain the value of precipitation, being an item of input data to equation [4], Ivy-Ochs *et al.* (2006) used:

$$P = –14\ (\partial a/\partial z)^2 – 381\partial a/\partial z – 321 \qquad [4]$$

This equation is an effect of analysis of ablation gradients of modern glaciers (Kuhn, 1984) and values of annual precipitation at their ELA's (Ohmura *et al*., 1992).

Additionally, we used the other, one-dimensional ice flow model which creates a valley glacier along the flow line based on equations of ice flow (Paterson, 1994; Haeberli, 1996). The modelling procedure is described in detail in Sarikaya *et al.* (2008). The mass balance of the glacier, which determines its growth, is calculated by using the accumulation predicted by snowfall which is assumed as precipitation occurring below 0°C and ablation of snow and ice with use of degree day factors: 3 mm (water equivalent)/day/°C for snow and 9 mm (w.e.)/day/°C for ice (Braithwaite and Zhang, 2000). The model assumes no basal sliding and ice is assumed to be isothermal. The ice deformation velocity is calculated without a valley shape factor. The model allows us to recreate a valley glacier along the flow line as a function of prescribed surface air temperature and precipitation. Starting from the present-day valley topography, the model creates a glacier until steady state conditions are attained. The elevation of the point of zero mass balance is the ELA of the glacier. As input data model requires the surface topography of the valley, the spatial distribution of monthly mean temperatures along the valley and the annual precipitation rates along the valley. Surface topography was obtained from a digital elevation model of the Tatra Mountains. The meteorological data for the time interval between 1960 and 2010 were downloaded from the Global Historical Climatology Network. To calculate the temperature gradient along the vertical profile of the Tatra Mountains, the data from meteorological stations in Zakopane (844 m a.s.l.), Kasprowy Wierch (1991 m a.s.l.) and Lomnicky Štit (2635 m a.s.l.) were used. Additionally, the precipitation values from Konček and Orlicz (1974) and Trepińska (2010) were used to supplement several gaps in the precipitation record from the *GHCN* server. Finally, for the first model, the summer (JJA – June, July, August) temperature gradient is 0.6°C/100 m. For the second model this gradient was calculated for each month separately and its value is in the range between 0.33 and 0.67°C/100 m. Estimation of the precipitation gradient is more problematic due to its non-linear character. Generally, to calculate the temperature and precipitation change we used a precipitation gradient of 64 mm/100 m based on data from Zakopane and Kasprowy Wierch. The data set of Trepińska (2010) shows that the mean precipitation gradient along the vertical profile between 900 and 2600 m a.s.l. is 20 mm/100 m. On the other hand, considering values of water-vapour condensation totals presented by Hess (1996), the precipitation gradient between Kasprowy Wierch and Lomnicky Štit equals 64 mm/100 m. Inversion of the precipitation gradient above 1900 m a.s.l. is an effect of the data set from the Lomnicky Štit which is located on the southern side of the massif in the "precipitation shadow" (Niedźwiedź, 1992). On the northern slopes of the Tatra Mountains this inversion is absent, likely due to orographically-induced precipitation.

RESULTS AND DISCUSSION

RECONSTRUCTION OF THE GLACIER TOPOGRAPHY

In this study two large glacial systems were reconstructed based on the spatial distribution of moraine ridges and glacial trimlines (Fig. 3). The first glacier fills the drainage basin of the Sucha Woda/Pańszczyca, and the second one occupies the drainage basin of the Biała Woda. The former contains two glaciers (the Sucha Woda Glacier and the Pańszczyca Glacier). In the uppermost cirques, the glacial ice reached an elevation of 2150 m a.s.l. The glacier terminated at an elevation of 1100 m a.s.l. where a very prominent terminal moraine was produced. The shape of the whole glacier in the ablation zone is precisely delineated by terminal and lateral moraines. The accumulation area is well constrained by the upper limit of glacial trimlines. The surface area of the Sucha Woda/Pańszczyca Glacier was 15.2 km^2.

The glacier filling the drainage basin of Biała Woda contains the main Biała Woda Glacier and four tributaries (Żabi Białczański Glacier, Rybi Potok Glacier, Roztoka Glacier and Waksmundzki Glacier). In the accumulation areas, the upper limit of glacial ice is constrained by glacial trimlines reaching up to 2250 m a.s.l. The glacier tongue is delimited by lateral moraine ridges and the prominent terminal moraine on the Łysa Polana at an elevation of 950 m a.s.l. The surface area of the whole glacier is 40.3 km^2.

The topography of both glaciers was controlled by values of the basal shear stress along the centre line. As the value of basal shear stress is very sensitive to changes of the ice slope and ice thickness, the topography of the glacier has to be consistent with topography of the valley floor. This means that if the glacier thickness rises, its slope should decrease and *vice versa*. Such conditions are needed for typical values of the basal shear stress. These values are in the range between 40 and 85 kPa along the tongues of the reconstructed glaciers. The stress rises up to 100–120 kPa inside the glacial cirques. The highest values of basal shear stress were obtained below high thresholds due to rising ice discharge below tributaries. This process was described in detail by e.g., Anderson *et al.* (2006), MacGregor

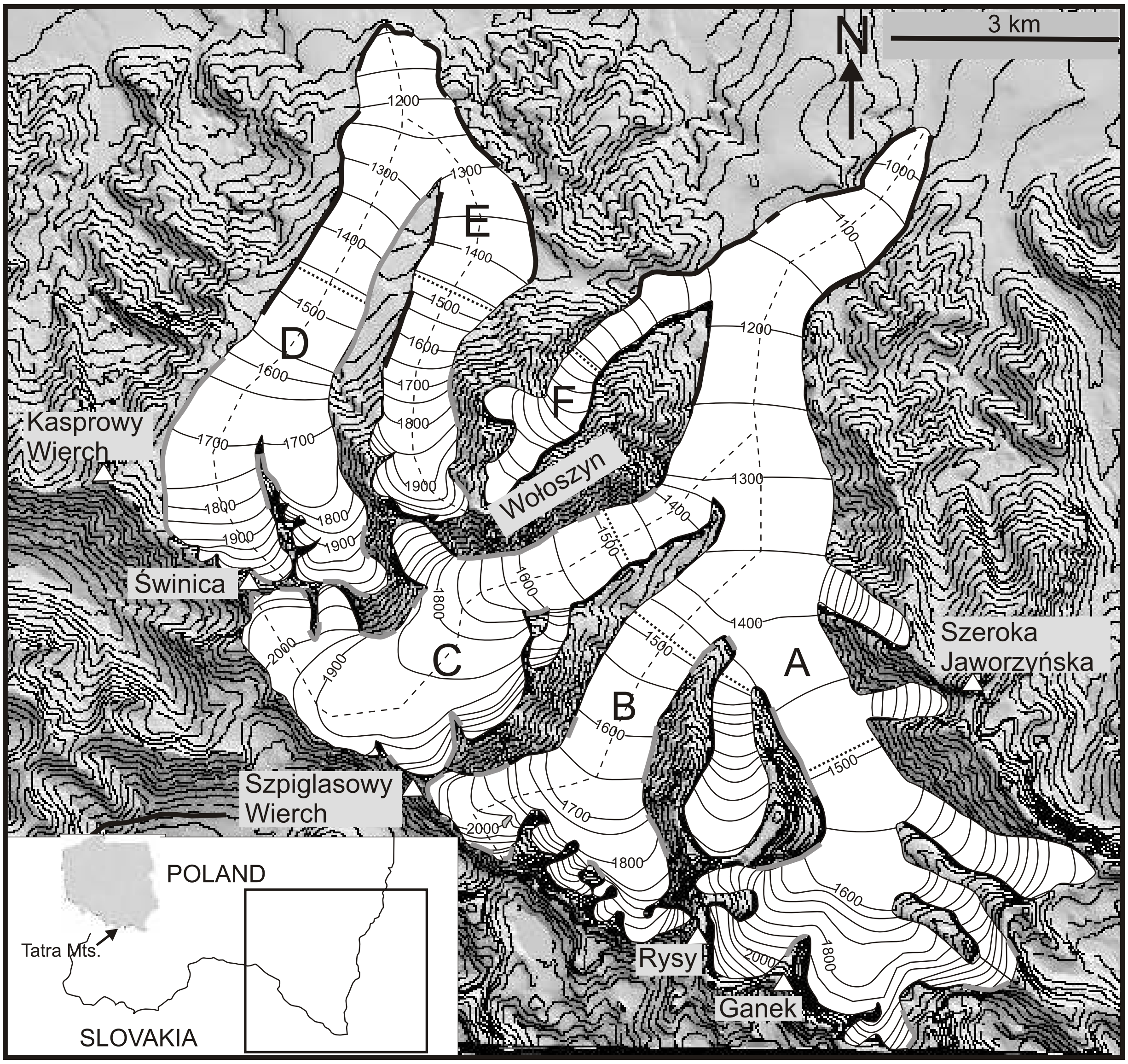

Fig. 3. Map of the LGM Glacier in the study area

A – Biała Woda Glacier, B – Rybi Potok Glacier, C – Roztoka Glacier, D – Sucha Woda Glacier, E – Pańszczyca Glacier, F – Waksmundzka Glacier; thick black lines mark position of terminal and lateral moraines (Klimaszewski, 1988; Baumgart-Kotarba and Kotarba, 1997, 2001); thick grey lines mark position of trimlines of generation I (Makos, 2008, 2010; Makos nad Nowacki, 2009); thin, dashed black lines mark the course of flow line of the reconstructed glaciers; dotted, black lines mark the position of the ELA

et al. (2000, 2009). Increasing ice discharge causes more intensive erosion of the valley bottom and thus overdeepening. The overdeepening has been measured in the Biała Woda Valley with the use of seismic reflection profiles (Baumgart-Kotarba *et al*., 2008). There is from 30 to 140 metres of sediment infill in the lower part of the valley between the tributaries of Roztoka Valley and Polana pod Wołoszynem. These findings are in accordance with our reconstruction of the topography of the parabolic cross-section of the Biała Woda Glacier (Fig. 4). Only cross-section 3 shows almost 60 metres shallower overdeepening than that obtained by seismic reflection analysis. In fact, we do not know on what surface the Biała Woda Glacier advanced during the LGM. The presence of bedrock 140 metres below the present-day valley floor indicates that the Biała Woda Glacier was over 300 m-thick below the tributary Waksmundzka Glacier, the thickness of which was only 50 metres in the zone of confluence. A few hundred metres down the Biała Woda Glacier its thickness was only 155 metres and overdeepening in relation to the present-day topography is 25 metres (Fig. 4). Below the large tributary Roztoka Glacier the sediment infill reaches 50 metres thickness at most. According to simulations of glacial valley formation (MacGregor *et al*., 2000, 2009), the strongest erosion occurs below large tributaries or many tributaries where ice discharge rises dramatically. So, there are steep and high rocky steps forming trough heads, occurring usually below the tributaries. Therefore, within the Biała Woda trough, the deepest overdeepenings should occur below the Roztoka and Rybi Potok valleys. The overdeepening, however, is not only the depth of the sediment infill in the Biała Woda Valley but also a “hanging valley re-

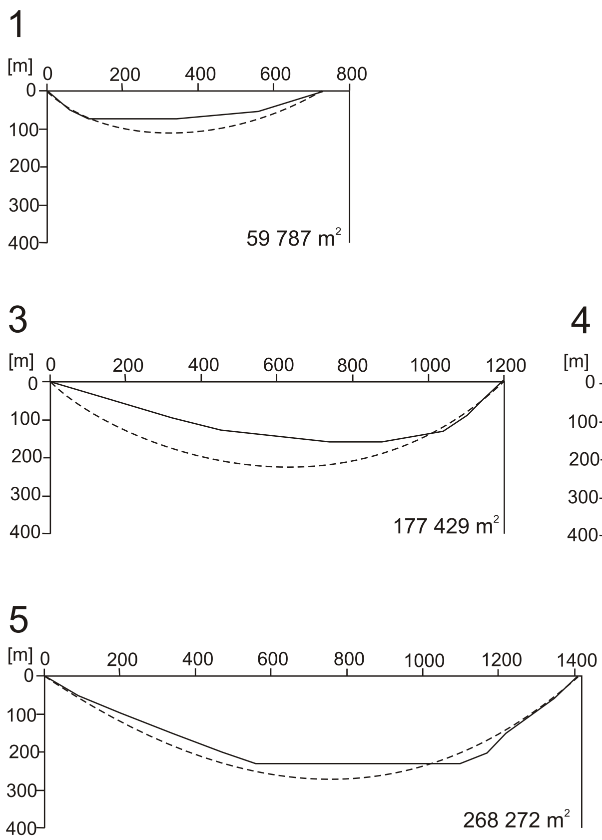

Fig. 4. Cross-section of the lower part of the Biała Woda Glacier

Solid lines represent the present-day topography of the valley; dashed lines are the parabolic cross-sections; numbers 1–5 refer to the position of the cross profiles in Figure 5; surface area of the cross-section are presented in the lower right

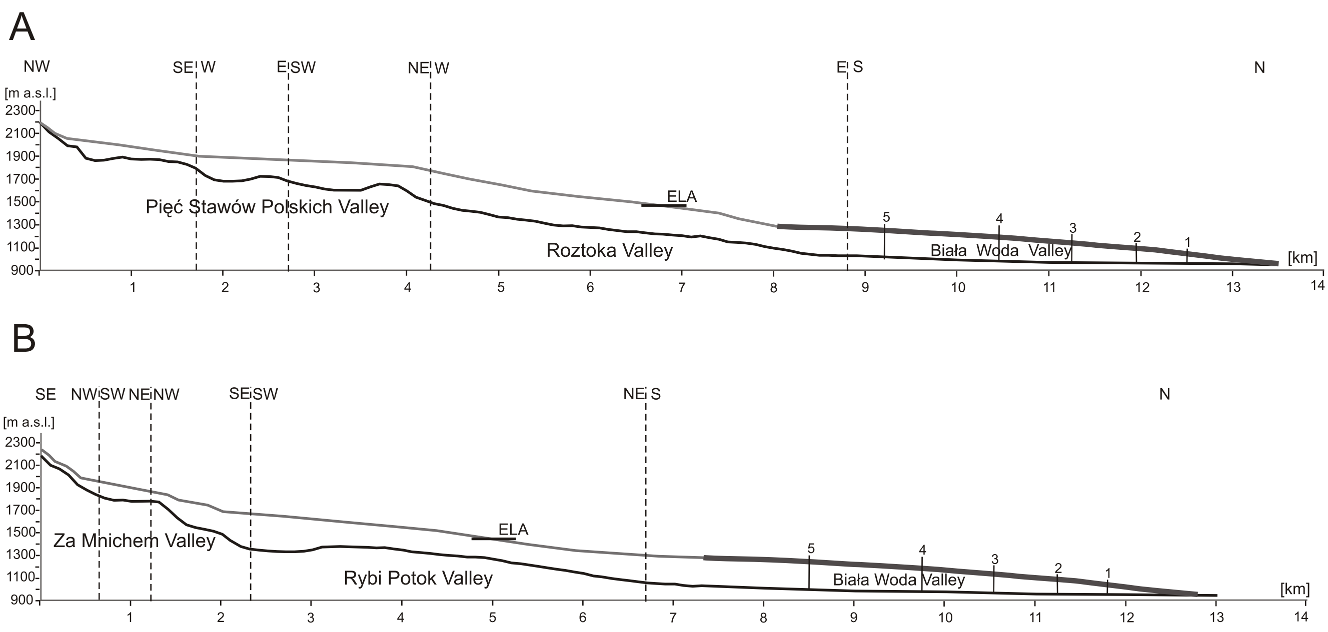

Fig. 5. Longitudinal profiles of the Biała Woda Glacier along: A – Roztoka Valley and B – Rybi Potok Valley

The thick grey line marks the part of the Biała Woda Glacier below the confluence of the Roztoka Glacier; cross-sections (used in model 1) are marked with numbers 1 to 5 lief" (difference between the hanging valley floor elevation and the main trough floor elevation) according to Brocklehurst *et al*. (2008). Even in such a case the overdeepenings below the Roztoka and Rybi Potok valleys are both shallower than that below the Waksumndzka Valley according to Baumgart-Kotarba *et al.* (2008). They inferred glacial trough over 300 metres deep at the Palenica Białczańska which is located nearly below the outlet of the small and thin LGM Waksmundzka Glacier. This means that the Biała Woda Glacier thickness was rising towards the snout. Therefore, we expect that at least a part of 140 metres-thick sediment infill reflects the trough depth from before the LGM. In our reconstruction we assumed 80 metres of sediment infill below the Waksmundzka Valley.

EQUILIBRIUM LINE ALTITUDE

As the geometry of the glacier is well-known, we can apply an AAR method for calculation of the ELA. The accumulation area ratio was determined separately for both glaciers using a new method outlined by Kern and László (2010). As the exact surface of glaciers is calculated, the size-specific steady-state AAR's of Sucha Woda/Pańszczyca Glacier and Biała Woda Glacier are 0.65 and 0.72 respectively. The latter value, however, is slightly higher than that recommended by Kern and László (2010) for glaciers larger than 4 km^2 (0.64 ±0.04). Their equation [1] is based on the analysis of much smaller glaciers than the Biała Woda Glacier. The analysed set of Kern and László (2010) consists of only one glacier with an AAR larger than 0.7 and this is the White Glacier (35.8 km^2) from Canadian Arctic. Hence, we obtained a relative large value of AAR of the Biała Woda Glacier of which the surface area is around 40 km^2. In the palaeoclimatic reconstructions the ELA of adjacent glaciers should be estimated using the same AAR value. Thus, we decided to reconstruct the equilibrium lines of both glaciers using a more typical and widely used value of 0.65 (e.g., Porter, 1975). The ELA of the Sucha Woda/Pańszczyca Glacier is then at 1460 m a.s.l. and the ELA of the Biała Woda Glacier is at 1480 m a.s.l. This consistency of obtained results may provide evidence that both glaciers existed in similar climatic conditions. It has to be considered, however, that complicated glacial systems which consist of several tributary glaciers have different topographic/climatic conditions in each valley. In addition the influence of avalanching, and debris cover on accumulation and ablation should be considered. However, due to lack of knowledge concerning the influence of avalanching and debris cover on palaeoglaciers, it is assumed that reconstructed glaciers were snow-fed and clean. Thus, the ELA's obtained have to be treated as minimum values. The ELA's obtained are 100–150 metres lower than historical data obtained by Lucerna (1908), Partsch (1923) and Halicki (1930). This discrepancy is an effect of the different methodologies used. Former methods (e.g., "cirque-floor", "MELM", "Höffer") are highly imprecise and dependent on surrounding topography. They usually give much higher results than statistical estimations e.g., the AAR method (see Gądek, 1998, table 12). The geomorphological evidence in the Sucha Woda Valley and the Pańszczyca Valley indicates a higher location of the ELA based on the MELM (maximum elevation lateral moraines) method. There are lateral moraines at elevations of 1550 and 1650 m a.s.l. in the Sucha Woda Valley and Pańszczyca Valley respectively. This is much higher than the AAR ELA (1460 m a.s.l.). The MELM method, however, can give false results for many reasons. One is that, when glacier melting is slow, continuous supply of debris can result in the incremental deposition of lateral moraines in an upslope direction, therefore overprinting earlier landforms with moraines associated with higher ELA (Benn and Evans, 1998; Benn and Lehmkuhl, 2000). Baumgart-Kotarba and Kotarba (2001), however, suggest the ELA of the Sucha Woda/Pańszczyca Glacier at 1400 m a.s.l. based on location of the lateral moraine (right hand side) at the Polana pod Wołoszynem. The discrepancy between the elevation of lateral moraines on both sides of the Sucha Woda/Pańszczyca Glacier reaches about 250 metres. There can be a number of reasons for this (see Benn and Lehmkuhl, 2000). There are many uncertainties associated with the MELM method and so, we decided to use ELA values obtained using the AAR method.

CLIMATE RECONSTRUCTION

The morphology and topography of the Biała Woda Glacier are well-known (Table 2), and the topography of the Biała Woda Valley is not complicated especially in its lowermost part, below the confluence with the Roztoka Glacier. Additionally, an almost

Table 2

Parameters of the lower part of the Biała Woda Glacier

Cross-section	Width [m]	Thickness [m]	Surface slope [deg]	Shape factor	Basal shear stress [kPa]
1	730	105	5.2	0.78	65
2	1050	155	3.4	0.77	63
3	1200	230	3.3	0.68	80
4	1170	250	3.15	0.66	80
5	1440	270	3	0.69	85

Table 3

Palaeoclimatic characteristics of the Last Glacial Maximum in the Polish part of the High Tatra Mountains

Percent sliding used	Scenario 1 ($U_b + U_d$)	Scenario 2 50%	Scenario 3 60%	Scenario 4 70%
Mean ice velocity across cross-section 5 [m/a]	15	20	26	34
Ablation gradient (between cross-section 5 and glacier terminus) [kg/m^2/m]	–2.6	–3.9	–4.9	–6.6
P at ELA (1480 m a.s.l.) [mm/a]	580	970	1230	1590
Δ*P* at ELA (1480 m a.s.l.) [%]	–60%	–34%	–16%	+8%
T JJA at ELA (1480 m a.s.l.) [°C]	0.3	1.5	2.3	3.1
Δ*T* JJA at ELA (1480 m a.s.l.) [°C]	–10.0	–8.8	–8.0	–7.2

perfect parabolic cross-section of the valley along the distance described makes the Biała Woda Glacier useful for palaeoclimatic reconstruction by using the glacier-climate model. The ice thickness was measured in parabolic cross-sections which are a few tens of metres deeper than the present-day valley due to sediment infill (Fig. 4). The overdeepening of the valley along the glacier tongue is from 20 to 80 m. The balance gradient was calculated for the lowest 300 metres of the glacier tongue. The length of this section is 4300 metres (Fig. 5A, B). The uppermost cross-section for ice-flux calculation is located at an elevation of 1250 m a.s.l. All other cross-sections were situated below, every 50 metres down the glacier.

The ice-flux and thus, the net ablation values depend greatly on ice velocity. Therefore, some scenarios for different contributions of sliding velocity were assumed. The first scenario assumes the basal sliding velocity and ice deformation velocity both calculated with the shape factor of the valley cross-section (F). Remaining scenarios assume a shape factor (F) for ice deformation and contribution of 50, 60 and 70% of sliding in the total velocity. The results obtained for all scenarios are shown in Table 3 and Figure 6.

The first scenario 1 is characterized by the lowest limit of the results obtained. Ice velocities were calculated with a shape factor which finally resulted in relatively slow ice movement (15 m/a). The sliding velocity is about 30% of the total velocity in this case. This, in consequence, caused the lowest value of the balance gradient within the whole data set (–2.6 kg/m^2/m). Scenarios 2, 3 and 4 gave slightly higher values of the total velocities (20, 26 and 34 m/a respectively), due to the sliding contribution assumed (50–70%). The balance gradients of these three scenarios are in the range between –3.9 and –6.6 kg/m^2/m. The balance gradient in scenario 1 indicates lower ablation and thus lower accumulation in relation to remaining scenarios. The lower accumulation means a lower precipitation sum and hence lower temperature at the ELA. Scenario 1 is therefore much dryer and colder than the remaining three. Its balance gradient is very similar to the White Glacier (Canadian Arctic) and the Gschnitz Glacier (Oldest Dryas alpine glacier – 70% sliding scenario; Ivy-Ochs *et al.*, 2006).

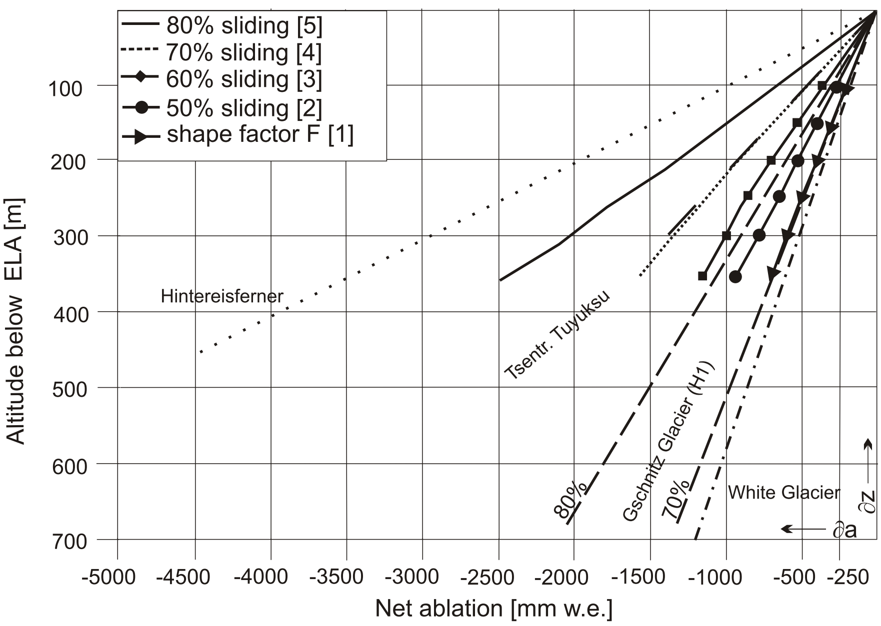

Fig. 6. Net ablation gradients of the Biała Woda Glacier

Numbers in brackets refer to scenarios discussed in the text; net ablation gradients for modern glaciers from Kuhn (1984); net ablation gradients for Gschnitz Glacier from Ivy-Ochs *et al.* (2006)

It shows that scenario 1 indicates a rather cold and dry climate with summer temperature at the ELA of around 0.3°C and precipitation at a level of 580 mm. Present-day climatic conditions at the 1480 m a.s.l. (LGM ELA) were estimated with the use of a temperature lapse rate of 0.6°C/100 m and a precipitation lapse rate of 64 mm/100 m. This means that current mean summer temperature at the LGM ELA level is 10.3°C and the annual precipitation sum is 1470 mm. Hence, scenario 1 indicates summer temperature depression of 10°C and precipitation lowering of 60% in relation to modern conditions. Scenarios 2, 3 and 4 are clearly warmer and wetter. Summer temperatures at the ELA oscillate between 1.5–3.1°C and precipitation totals at the ELA are between 970 and 1580 mm. Scenario 2 has a balance gradient comparable with the Gschnitz Glacier (80% sliding; Ivy-Ochs *et al*., 2006). The summer temperature decrease reached 8.8°C and precipitation was 66% of present-day values. The balance gradient of scenario 3 is similar to that of the Tsentralnyy Tuyuksu Glacier (Tian Shan). It gives a summer temperature depression of 8°C and annual precipitation lowering of 16% in relation to modern conditions. The wetter scenario 4 shows that the balance gradient of the Biała Woda Glacier was higher than that of the Tsentralnyy Tuyuksu Glacier but still lower than the current balance gradient of Hintereisferner (Central Alps). The precipitation sum at the ELA was 8% higher than present-day values and the summer temperature decrease reached 7.2°C.

The one-dimensional flow model (Sarikaya *et al*., 2008) requires a simple topography of the whole valley and especially a non-complicated morphology of the accumulation area. The Biała Woda Glacier was supplied from several sources, each of which was a separate valley glacier with an expanded headwaters area. Thus, it was difficult to apply the model of Sarikaya *et al*. (2008) in the case of such a complex glacial system. The Sucha Woda/Pańszczyca Glacier, however, gave the opportunity to use this model due to the relative simplicity of the valley morphology, particularly above the ELA. The Sucha Woda Glacier and the Pańszczyca Glacier both connect to each other at an elevation of 1300 m a.s.l., about 200 metres above the terminal moraine. This confluence likely caused over 1.5 km further extent of the glacier tongue (Fig. 3). Therefore, both glaciers were modelled separately to obtain climatic conditions in which both glaciers reach the confluence zone. The length of the Sucha Woda Glacier flow line between the headwall and the zone of confluence is 7100 metres. The corresponding distance of the Pańszczyca Glacier is 6000 metres long. The model produces the glacier along the flow line based on simulation of temperature decrease and precipitation changes in relation to modern values (Fig. 7). The modelled Sucha Woda Glacier reached the confluence zone by a temperature decrease of between 7 and 14°C (Fig. 8A). The precipitation sum was then 80% higher and 80% lower respectively. The modelled Pańszczyca Glacier reached the confluence zone with exactly the same conditions as the Sucha Woda Glacier (Fig. 8).

The modelled climatic conditions show a significant discrepancy. Thus, it is necessary to compare the results obtained with proxy data, especially these based on pollen analyses. There are no such data available in the Tatra Mountains. However, we can use information about LGM climatic conditions in Central and Southern Poland (highlands). According to Manikowska (1995) the apogee of coldness during the entire Vistulian Glaciation took place between 20–14 ka BP. In Southern Poland the average annual temperature was lower than –7°C and the mean temperature of July was a few degrees lower than 10°C (Jersak *et al*., 1992). So, if the temperature in July at an elevation of 300 m a.s.l. was about 6–7°C than the temperature at the ELA in the Tatra Mountains would have been about 0°C using the temperature lapse rate of 0.6°C/100 m. These data are in agreement with those presented by Wu *et al*. (2007). They indicated that in Europe the annual winter temperature decrease was between 10 and 17°C. Summer temperatures were 6–12°C lower than today. The mean annual temperature depression, however, was at a level of 10–14°C. Kuhlemann *et al*. (2008) estimated the LGM summer temperature depression in the Mediterranean Basin. In the Pyrenees and Alps this depression reached 10–12°C.

These data suggest rather cold and dry scenarios. According to the climatic data presented by Jersak *et al.* (1992), the expected summer temperature at the ELA_{LGM} should be about 0°C. Thus, we inferred scenario 1 from model 1 as the most realistic. In this case, the summer temperature in the Tatra Mountains would be some 10°C lower than today and the annual precipitation sum was about 40% of present-day values. These data suggest that the Biała Woda Glacier existed in a subarctic climate regime with low precipitation values and slightly positive summer temperatures at the ELA. Scenario 1 assumes both the sliding velocity and ice deformation velocity calculated with use of the valley shape factor. There is no contribution of sliding assumed. The relation of the sliding velocity to the total velocity in the scenario 1 is about 30%. This is in accordance with data presented by Gądek (1998) which indicate that for glaciers some 200 m-thick this relation is less than 0.5. The mean thickness of the lowest 300 m of the Biała Woda Glacier is 200 m.

Considering the temperature data presented by Wu *et al.* (2007) we can state the results from model 2 more precisely. The LGM mean annual temperature decrease of 10–14°C allows us to infer that precipitation was then lowered by between 30 and 80% respectively. However, according to the palaeoclimatic data from the highlands of Southern Poland (Jersak *et al*., 1992) we can infer an annual temperature decrease of more than 10°C. If the precipitation data obtained in the model 1 are correct then the most realistic scenario in model 2 will be the one with a 12°C annual temperature decrease and with 60% precipitation lowering. This scenario gave also the zero mass balance at the altitude of 1455 m a.s.l. which is almost identical with the AAR ELA of Sucha Woda/Pańszczyca Glacier.

Additionally, we tested if our manual reconstruction of the SuchaWoda/Pańszczyca Glacier was concordant with the glacier profile obtained in model 2. We compared glacier profiles along the centrelines (Fig. 7). For the Sucha Woda Glacier we obtained almost identical profiles of the glacier surface in both reconstructions. The Pańszczyca Glacier, however, has shown almost 100 metres discrepancy of ice thickness in the ablation area, near to and below the ELA (Fig. 7B). In our opinion this is an effect of the topography of the Pańszczyca Valley which is shallow and very wide. Furthermore, it is necessary to emphasize that model 2 assumes a present-day topography of the valley. Thus, the model does not take into account sediment infill in the glacial trough which is composed of ground moraine material in the lower part of the valley especially (Klimaszewski, 1988). The thickness of the valley infill, however, is not known. Therefore, the manual reconstruction based on geomorphological evidence can give different results (a lower thickness) than that computed by model 2. Additionally, model 2 calculates the ice velocity without basal sliding and basal shear stress and thus, ice deformation velocity are calculated without a shape factor for the valley cross-section. However, considering the modelled ice thickness in the accumulation area which is almost consistent with that reconstructed using the geomorphological evidence (Fig. 7B), the discrepancy of the topography of both reconstructions is an effect of the sediment infill in the lower part of the glacial trough. The lack of difference between both topographies above ELA can be evidence that there is no significant sediment infill and that absence of ice sliding in estimations is in some degree compensated by the value of the basal shear stress calculated without the valley shape factor.

AN ATTEMPT TO ASSESS THE ELA DEPRESSION DURING THE LGM

Considering the fact that in mid-latitude mountains the ELA of glaciers occurs at the snowline (Jania, 1997), it can be assumed that the LGM snowline of reconstructed glaciers was located at an elevation of 1460–1480 m a.s.l. The modern climatic snowline can be calculated based on modelled climatic parameters from the LGM (Fig. 9). If the LGM summer temperature and annual precipitation lowering in relation to the modern values are known, then it is possible to obtain the modern snowline altitude using this value as a variable in the following formula:

$$\Delta T_S = (\mathrm{mSL} - \mathrm{ELA}_{\mathrm{LGM}})\ \Delta T/\Delta H + \\ + [(\mathrm{mSL} - h_{\mathrm{ms}})\ \Delta P/\Delta H + (P_{\mathrm{ms}} - P_{\mathrm{LGM}}]/350 \quad [5]$$

where: ΔT_S – LGM summer temperature decrease in relation to the modern value (10°C); mSL – modern altitude of snowline; $\mathrm{ELA}_{\mathrm{LGM}}$ – LGM equilibrium line altitude (1480 m a.s.l.); $\Delta T/\Delta H$ – temperature gradient (0.6°C/100 m); h_{ms} – elevation of meteorological station (Kasprowy Wierch – 1991 m a.s.l.); $\Delta P/\Delta H$ – precipitation gradient (1–20 mm/100 m, 2–64 mm/100 m depending on the meteorological data set used); P_{ms} – mean annual precipitation at the level of the meteorological station; P_{LGM} – precipitation at the ELA during the LGM (580 mm)

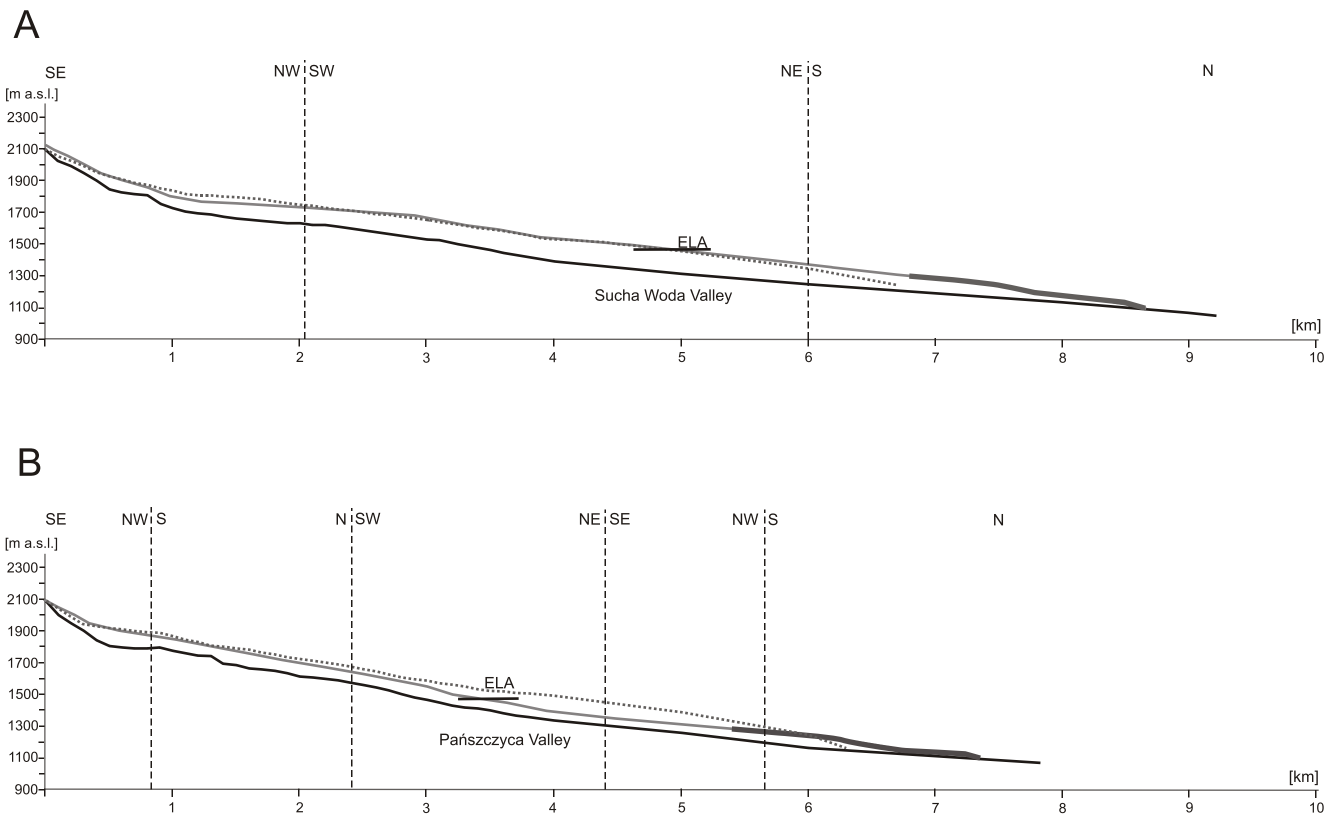

Fig. 7. Longitudinal profiles of the Sucha Woda Glacier (A) and Pańszczyca Glacier (B)

Thick grey line marks the common part of the Sucha Woda and Pańszczyca glaciers; grey dotted line is the profile of the Sucha Woda Glacier produced with the use of model 2

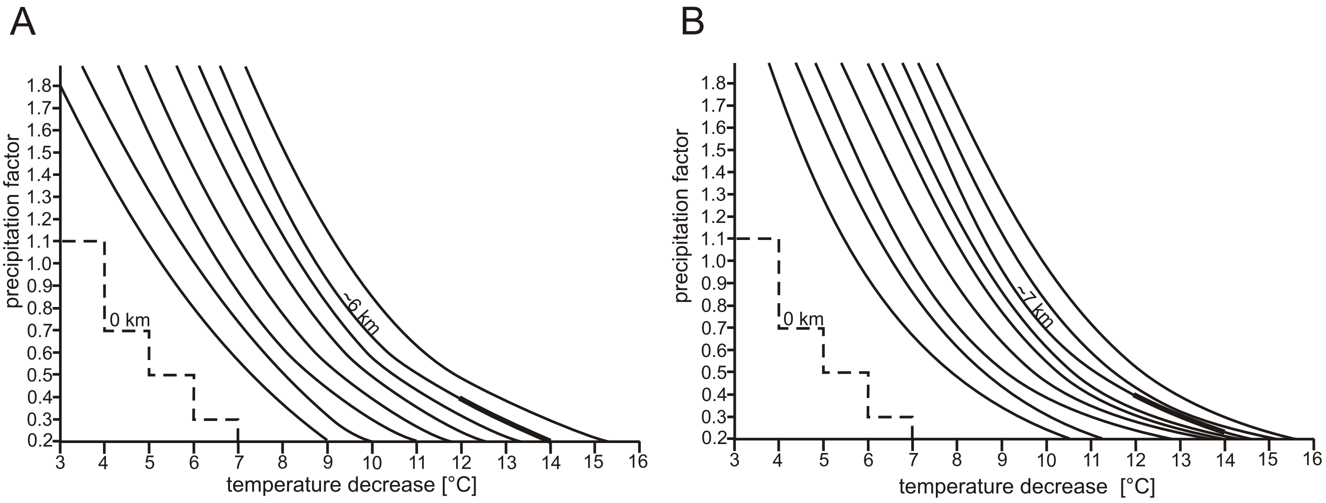

Fig. 8. Modelled length of the Sucha Woda Glacier (A) and Pańszczyca Glacier (B) during the LGM as a function of temperature and precipitation changes from those of today

The thick line shows the possible range of reconstructed climatic conditions best fitted to precipitation changes obtained in model 1; the dashed line (0 km) represents glacier inception

The calculation is additionally based on the assumption of Ohmura *et al*. (1992) that changes of the ELA are strictly related to the fluctuations of temperature and precipitation at the glacier and 1°C of temperature increase is fully compensated by 350 mm higher annual precipitation. When scenario 1 is considered (ΔT_S – 10°C and P_{LGM} – 580 mm), the modern snowline is located at an elevation of 2550 m a.s.l. (precipitation gradient 20 mm/100 m) or 2450 m a.s.l. (precipitation gradient 64 mm/100 m). The values obtained are comparable with data previously presented by Zasadni and Kłapyta (2009). Thus, the ELA depression in the Tatra Mountains during the LGM reached about 1000–1100 metres. Similar values of the ELA (mSL – ELA_{LGM}) are observed across many mountains ranges in Europe (Kuhlemann *et al*., 2009). Higher values of ELA depression (1500 m) occurred mainly in areas which were affected by polar air masses and then by cyclones following the tracks reconstructed by Kuhlemann *et al.* (2009) in the Mediterranean Basin.

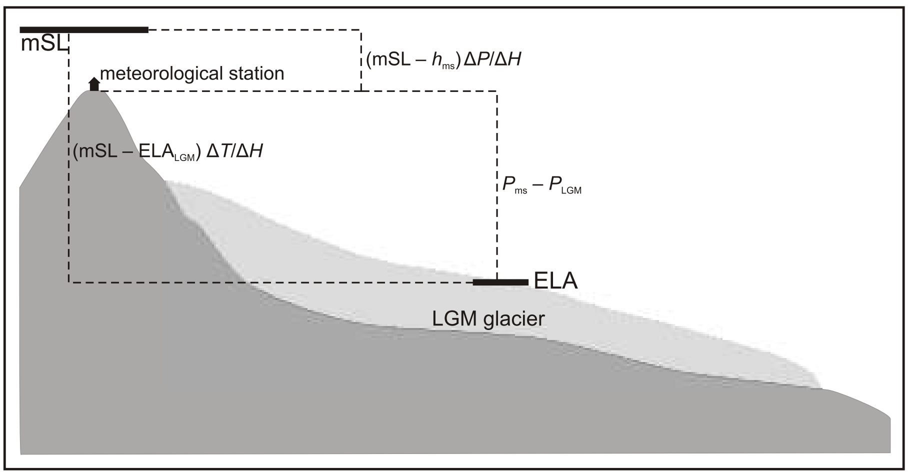

Fig. 9. Graphic presentation of the modern snowline formula 5

The summer temperature decrease is variable depending on the modern snowline altitude. The formula first assumes the ELA depression and calculates the temperature depression using an assumed temperature lapse rate. Then the precipitation change between the modern snowline and the ELA_{LGM} is calculated using the reconstructed precipitation sum at the ELA_{LGM} and an assumed precipitation gradient. The precipitation change is then recalculated into the temperature change using the assumption of Ohmura *et al*. (1992) that a temperature increase of 1°C at the ELA of the glacier is fully compensated by 350 mm higher precipitation

LGM CLIMATIC CONDITIONS IN THE HIGH TATRA MOUNTAINS COMPARED TO OTHER MOUNTAINS IN EUROPE

Pleistocene glaciations affected many mountain ranges across all of Europe. The geomorphological imprint of glacial activity is well-developed in the European Alps (e.g., Ivy-Ochs *et al*., 2008, 2009), the Southern Carpathians (Reuther *et al*., 2007), the Pindus Mountains – Greece (Hughes *et al*., 2006), the Kaçkar Mountains – Eastern Turkey (Akçar *et al*., 2008), Mount Sandiras (Sarikaya *et al*., 2008) and Uludag Mt. (Zahno *et al*., 2010) in Western Turkey.

Alpine glaciers reached their maximum extent during the global Last Glacial Maximum (MIS 2; Ivy-Ochs *et al*., 2008). Glaciers on the southern side of the massif produced the most extended moraines during two advances: an older one between 26.5–23 ka ago and a younger one between 24 and 21 ka ago (Monegato, 2007). As inferred from exposure ages of boulders on the northern Alpine foreland, the Piedmont lobe of the Rhone Glacier reached a maximum between 21 and 19.1 ka ago (Ivy-Ochs *et al*., 2004). The ELA depression in relation to modern conditions reached 1400–1800 metres (Reuther, 2007; Ivy-Ochs *et al*., 2008). Florineth and Schlüchter (2000) suggested a predominance of southerly circulation in the Alps during the LGM. This was an effect of migration of the polar fronts to the latitude of Southern France, which led to domination of moisture transport south of the Alps. Such circulation likely caused a dry and cold climate in Central Europe and much wetter conditions in the Mediterranean area. This is supported by palaeoclimatic data from Turkey and Greece. In the Pindus Mountains the maximum extent of glaciers occurred several thousand years earlier than in the Alps. Dating of alluvial units gave ages of 28.2–24.3 ka (Hughes and Woodward, 2008). Hughes *et al*. (2003, 2006) suggested that the Late Pleistocene glacial stage was cold and humid with annual temperatures 8–9°C lower than today and a higher precipitation by about 10%. The LGM advances in the mountains of Anatolia occurred in general between 23 000 and 19 000 years ago (Sarikaya *et al*., 2008, 2009), correlating well with the global LGM. Palaeoclimatic interpretations, however, show an annual temperature lowering of about 8–11°C and an annual precipitation total up to 90% higher than today on Mount Sandiras (Sarikaya *et al*., 2008), and similar to modern values (20% higher or 25% lower) on Mount Erciyes (Sarikaya *et al*., 2009). In northeastern Turkey, glaciers advanced on to Kaçkar Mountain around 26 000 years ago (Akçar *et al*., 2007, 2008) while in northwestern Turkey, the LGM advance occurred no later than 20.3 ±1.5 ka ago (Zahno *et al*., 2010). The ELA_{LGM} depression in Central Turkey was about 900 metres (Sarikaya *et al*., 2009). Reuther *et al*. (2007) focused on the glacial chronology of the Retezat Mountains (Southern Carpathians). These authors inferred that the Late Würmian advance was asynchronous to the global climate record and the maximum advance (M1) of glaciers had occurred well before global LGM. The LGM moraines were probably overridden by a further advance. The ELA depression in the Retezat Mountains was established between 1175 metres during the maximum advance and 1130 metres during the Late Glacial. The authors inferred more glacier-friendly conditions during the Early Würm (MIS 4) than during MIS 2. These findings are in general agreement with those from Pindus Mountains (Hughes *et al*., 2006) but they indicate a significant discrepancy with the palaeoclimatic reconstructions from Anatolia – in particular, with the timing of glacial episodes.

Luminescence dating (Baumgart-Kotarba and Kotarba, 2002) and ^{36}Cl exposure dating (Dzierżek, 2009) of moraine deposits in the Sucha Woda and Biała Woda valleys as well as ^{36}Cl exposure dating of erosional features in the headwaters of the Pięć Stawów Polskich Valley and the Rybi Potok Valley (Makos, 2010) suggest that glaciers in the Polish part of the High Tatra Mountains reached their last maximum extent between 25 000 and 21 000 years ago. This timing is in good agreement with the global LGM (26 500–19 000 years ago – Clark *et al*., 2009) and with the maximum extent of glaciers in the European Alps and in Anatolia. Palaeoclimatic reconstruction using two independent models reveals a summer temperature decrease of about 10°C and a precipitation decrease of 60% during the LGM. Such a reduction of annual precipitation totals means that annual temperature was lowered by at least 12°C in relation to modern conditions. Reconstructed climatic parameters indicate that ELA_{LGM} depression was at the level of 1000–1100 metres in relation to the modern snowline altitude. That values are very similar to those obtained in the Southern Carpathians for the pre-LGM advance and for the Oldest Dryas (Reuther *et al*., 2007), but they are clearly lower than the ELA depression in the Northern Alps, Pyrenees and Dinarides (Kuhlemann *et al*., 2009). As was suggested, the southerly atmospheric circulation over Europe during the LGM produced wetter conditions in the maritime Alps and Mediterranean and increased aridity in the central part of the continent (Florineth and Schlüchter, 2000; Kuhlemann *et al.*, 2008, 2009).

CONCLUSIONS

Glaciers in the High Tatra Mountains of Southern Poland formed during the LGM under a cold and dry climate. The results obtained indicate a decrease of summer temperature of 10°C and a precipitation lowering of 60% in relation to modern conditions. Mean annual temperature was than depressed by at least 12°C. Reconstructed climatic parameters for the LGM in the Tatra Mountains resemble those occurring currently at the glaciers of the subarctic climate zone. These data find confirmation in pollen-based temperature reconstruction in Central and Eastern Europe presented by e.g., Kageyama *et al*. (2006) and Wu *et al*. (2007). The ELA depression in the Tatra Mountains at the LGM reached a value of 1000–1100 metres which is a comparable result to those from other mountain ranges in Europe (Kuhlemann *et al*., 2008, 2009). The vicinity of the southern margin of Fennoscandian Ice Sheet which occupied almost half of the territory of Poland suggests cold and dry, arctic climatic conditions in the Tatra Mountains. Relatively low values of the annual precipitation in the Tatra Mountains at the LGM (580 mm) in comparison with almost doubled precipitation amounts reconstructed in western Turkey, as well as lower values of ELA depression eastwards from the Alps may reflect increasing continentality of the climate north-east of the Alps.

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