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Reconstructing the Last Glacial Maximum (LGM) ice surface geometry and flowlines in the Central Swiss Alps

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Key words: Trimlines, paleoglaciology, Last Glacial Maximum (LGM), Swiss Alps, GIS, paleoclimate

ABSTRACT

The surface geometry of the accumulation area and flow patterns of the ice at the Last Glacial Maximum (LGM) have been reconstructed in the Central Swiss Alps, where the ice surface formed two individual ice domes. One of them was located in the headwaters of the Rhein river and attained a minimum altitude of 2700 m. The second dome culminated at about 2900 m in the Upper Rhone valley. According to 174 trimline data, the ice surface gradients were steeper towards the south and north than to the southwest and northeast. Flowlines are perpendicular to ice-surface contours and underline the presence of a dome-type glaciation with radial outflow. The ice-flow pattern indicates drainage through the large longitudinal valleys and ice-transfluences over all major passes and lower cols to the north and south. Modelling of the LGM surface configuration by means of a Geographical Information System (GIS) likewise displays a surface geometry consisting of multiple domes with the ice divides located overhead the Rhone- and Rhein valleys, respectively. This configuration was then used to reconstruct the prevalent precipitation pattern and as the main atmospheric paleocirculation for the Alps. The position of both ice domes to the south of the Alpine main weather divide indicates that the atmospheric pressure system during the last glaciation was markedly different from today's, with prevailing southerly instead of westerly circulation. This is correlated with the southward displacement of the atmospheric Polar Front in the North Atlantic to the latitude of 40–50° due to the advancing sea ice cover. This caused the related storm tracks to move to the south of the Alps across the Mediterranean, resulting in increased rainfall in the Southern Alps and in notably reduced precipitation to the north of the Alps where dry katabatic winds promoted permafrost conditions

ZUSAMMENFASSUNG

Die Oberflächengeometrie und die Fliessmuster des Eises für das Akkumulationsgebiet während des letzteiszeitlichen Maximums (LGM) wurden in den zentralen Schweizer Alpen rekonstruiert. Die Eisoberfläche bildete in diesem Gebiet zwei individuelle Eisdome; der eine befand sich in den Oberläufen des Rheintales (Surselva) und erreichte eine minimale Höhe von 2700 m, während der zweite Eisdom auf ungefähr 2900 m im oberen Rhonetal (Obergoms) kulminierte. Die Analyse von 174 kartierten Schliffgrenzen zeigt, dass die Eisoberfläche nach Süden und Norden steiler einfiel als nach Südwesten und Nordosten. Die von Striemungsdaten abgeleiteten Eisfliesslinen stehen senkrecht zu den Höhenlinien der rekonstruierten Eisoberfläche und unterstreichen die Existenz einer domartigen Eiskappe mit radialem Abfluss. Die Rekonstruktion des Eisfliessmusters deutet auf eine Hauptentwässerung des Eises durch die grossen Längstäler hin, sowie auf Transfluenzen nach Norden und Süden. Die Berechnung der Eiszeittopographie anhand eines Geographischen Informationssystems (GIS) liefert zudem das Modell einer, aus mehreren Eisdomen bestehenden Oberfläche. Die Eisscheiden lagen dabei über dem Rhone- bzw. Rheintal. Diese Konfiguration des ehemaligen Akkumulationsgebietes wurde im weiteren dazu verwendet, um das Hauptniederschlagsmuster sowie die vorherrschende atmosphärische Paleozirkulation für die Alpen nachzubilden. Die Lage beider Eisdome südlich der alpinen Wetterscheide deutet auf eine gegenüber heute markanten Änderung des atmosphärischen Drucksystems hin, mit vorherrschender Süd- anstelle von Westwindzirkulation. Diese Umstellung der Hauptwetterlagen wird mit der vorrückenden Meereisdecke im Nordatlantik in Verbindung gebracht. Dies führte zu einer Verlagerung der atmosphärischen Polarfront zur Breite von 40–50°N so, dass die Hauptzugbahn zykonaler Störungen südlich der Alpen zu liegen kam. In der Folge nahmen die Niederschläge südlich der Alpen und in Südeuropa zu und verringerten sich im Gebiet nördlich der Alpen. Dort förderten zudem trockene katabatische Winde die Bildung von Dauerfrostböden.

Introduction

The late Quaternary history of glaciations in the Alps has been debated for almost two centuries. Investigations on vanished glaciers are generally done 'by reversing the order of study usually applied to climate-glacier relationships' (Haeberli & Penz, 1985; p. 352). Ice margins and bed characteristics of former ice bodies can be derived from a range of geomorphological/geological field evidence, including the distribution of terminal and lateral moraines, the scattering of erratic boulders, trimlines, till limits, periglacial features and fluvioglacial landforms. Consequently, detailed mapping of depositional and of erosional features provides the elementary information needed for most paleoglaciological reconstructions, e.g. equilibri-

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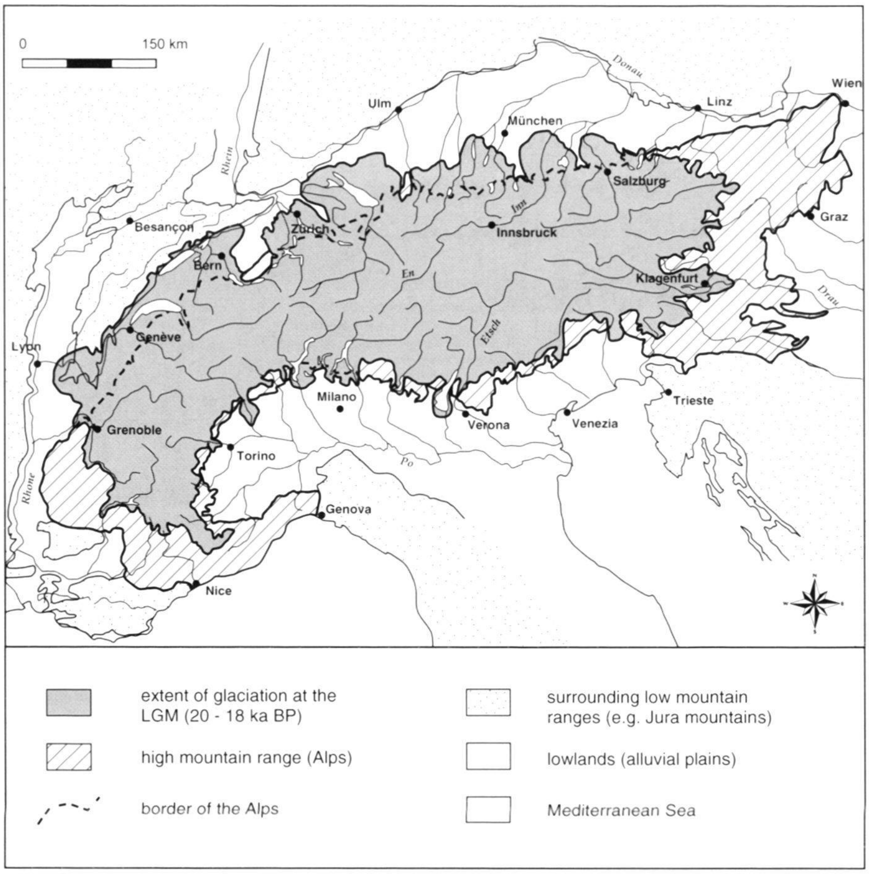


Fig. 1. The maximum extent of the Alpine glaciation in the northern and southern forelands at the LGM, 20'000 to 18'000 years ago. Glaciation of the Jura Mountains, Vosges and Black Forest are not shown. Dashed line gives the morphological boundary of the Alps (adapted from Jäckli 1962/70, van Husen 1987, de Beaulieu et al. 1991).

um line altitudes (ELA). Given the limits of the ice body, further glaciological (e.g. flow velocity, mass balance, shear stress values) and climatological (e.g. balance gradient, precipitation patterns) parameters can be inferred.

In 1884, Favre published the first detailed map of the last glaciation for the whole of Switzerland not only showing the maximum extent of the glaciation, the distribution of erratic boulders, trimlines and the different glacier catchment areas (Rhein, Aare, Rhone etc.), but also distinguishing the former ablation area from the zone of accumulation. Further studies focused on the morphology, sedimentology, stratigraphy and petrography of the Ice Age deposits in the former ablation area. As a result, the maximum extent of the last glaciation outside the boundaries of the Alps was recognised as broad outlet piedmont glaciers spreading out well into the northern and southern Alpine forelands (Fig. 1). Jäckli (1962, 1970) attempted the last detailed reconstruction of the Last Glacial Maximum (LGM) in the Swiss Alps whereas the maps of van Husen (1987) and de Beaulieu et al. (1991) cover the Eastern Alps and the Southwestern Alps. These reconstructions were done by compiling information accumulated in numerous local investigations based on moraine morphology in the northern Foreland and on trimline evidence in the inner Alpine area. Apart from some local errors (c.f. Rhein glacier: Keller & Krayss 1993, Southern Switzerland: Bini 1987; Felber 1993) these survey maps still rank as some of most accurate small-scale reconstructions presently available and are still widely used. This is particularly true with respect to the area of the Alps proper because today, despite more than 150 years of study on the Late Pleistocene glaciations, surface contours of the LGM ice body in the former accumulation area remain uncertain.

The results acquired in the ablation area indicate that the surface geometry and flow directions of the ice were strongly influenced by the underlying topography; it was usually assumed that the last glaciation in the Alps showed a dendritic pattern, similar to the present river network. However, since mapping of glacial-erosion features at high altitudes in the Central Alps indicate up-valley flow of some paleoglaciers and ice transfluence over high altitude passes, the validity of this hypothesis is limited to the ablation area. Such features cannot be explained with existing reconstructions of the ice-surface configuration. Therefore, a realistic geometry of the LGM can only be achieved by detailed investigations in the inner Alps. In doing so, the basic problems are: 1) limited field evidence (overprinting by younger glacier advances, weathering, burial by snow, ice and sediments, etc.) in the former accumulation area and 2) the question of whether mapped field evidence

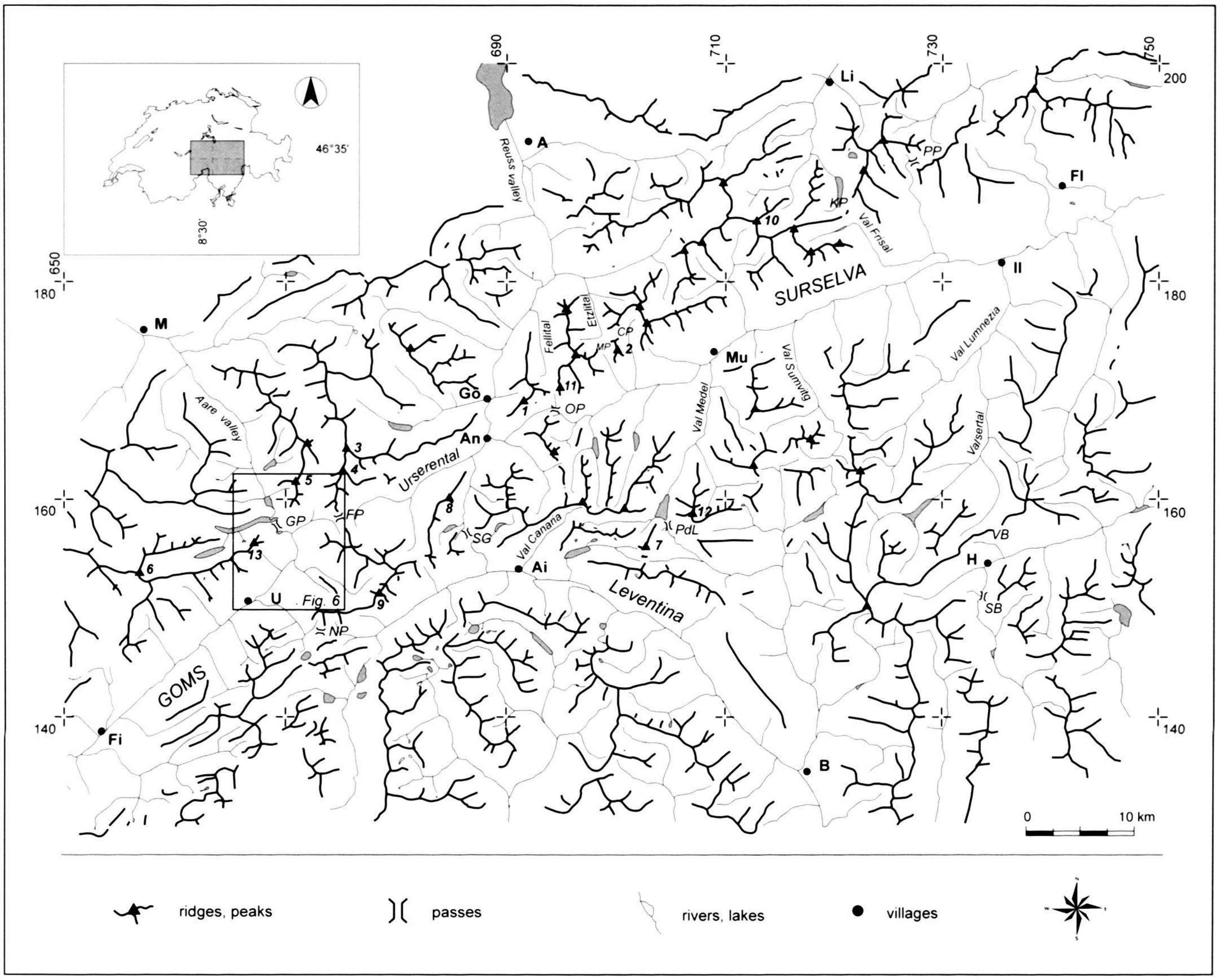


Fig. 2. Overview map of the study area, showing the location of sites mentioned in the text and the position of Fig. 6. Abbreviations: A- Altdorf, Ai – Airolo, An – Andermatt, B – Biasca, Fi – Fiesch, Fl – Flims, Gö – Göschenen, H – Hinterrhein, Il – Ilanz, M – Meiringen, Mu – Mustér, U – Ulrichen, CP – Chrüzlipass, FE – Fellilücke, FP – Furkapsss, GP- Grimselpass, KP – Kistenpass, LP – Lolenpass, MP – Mittelplatten, NP – Nufenenpass, OP – Oberalppass, PdL – Passo del Lucomagno, PP – Panixerpass, SB – Passo del San Bernardino, SG – Passo del San Gottardo, VB – Valserberg; 1 – Chl Schijen, 2 – Chrüzlistock, 3 – Dammastock, 4 – Galenstock, 5 – Gärstenhörner, 6 – Oberaarhorn, 7 – Pizz dell'Uomo, 8 – Pizzo d'Orsiono, 9 – Pizzo Rotondo, 10 – Piz Russein (Tödi), 11 – Piz Tiarms, 12 – Scopi, 13 – Sidelhorn

above (e. g. trimlines) and below (e.g. terminal moraines) the former equilibrium line have been produced by the same event.

In recent years, paleoglaciological modelling experienced considerable progress following a significant development in computer technology. As a consequence, there have been attempts to model the climatological and glaciological characteristics of the last Pleistocene glaciation. Thus, mass balance, balance gradients, shear stress values and flow velocities of LGM glaciers (Körner 1983; Haeberli & Penz 1985; Keller & Krayss 1993) as well as the temperature distribution in Alpine glaciers (Blatter & Haeberli 1984) have been roughly estimated by the use of 2-dimensional steady-state models and simple approaches. The effects of the Late Pleistocene glaciation on geothermal heat flux (Haeberli et al. 1984), isostasy (Gudmundsson 1994) and hydrogeology (Speck 1994) could on the other hand be assessed by means of sensivity studies. However, accurate modelling of realistic conditions for Ice Age glaciers can only be achieved by the application of a 'complex 3-dimensional, thermomechanically coupled system of time dependent ice and permafrost over the whole glacier area' (Haeberli & Schlüchter 1987). But as long as quality and quantity of the

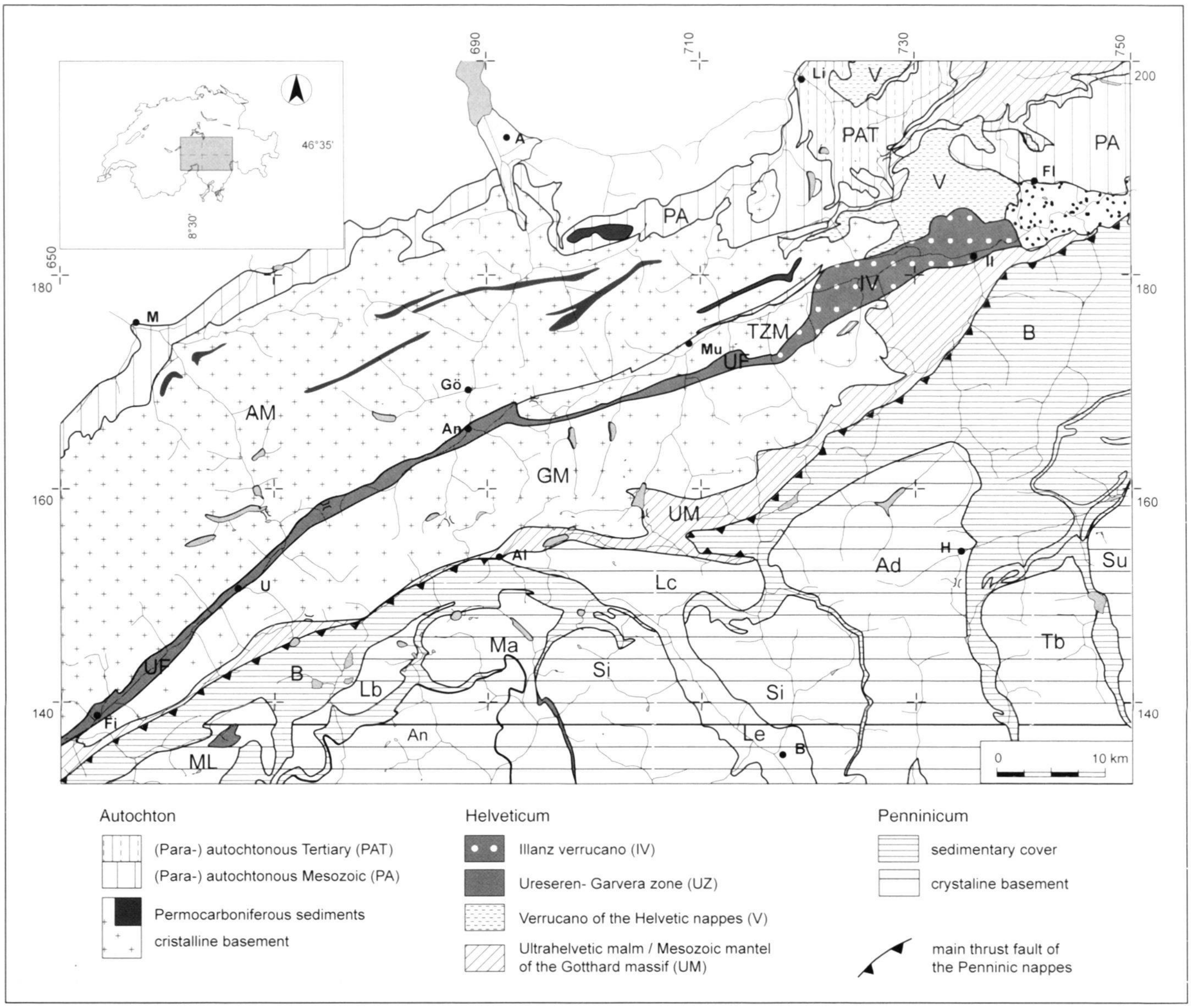


Fig. 3. Highly simplified geologic and tectonic map of the Central Swiss Alps area (modified from Spicher 1980); Abbreviations and areal cover as Fig 2.

input data will remain the same and progress only takes place in one direction, namely in the field of numerical modelling, theoretical glaciology and computer sciences, even the most sophisticated models will not be able to provide any new results at all because every (paleo-) glaciological reconstruction is only as reliable as its geological/geomorphological input data.

Contrary to the large Laurentide and Fennoscandian ice sheets where the ice-surface topography can only be modelled by means of numerical reconstructions and theoretical approaches (Hughes 1981; Boulton et al. 1985; Hindmarsh et al. 1989; Huybrechts & T'siobbel 1995), glacial trimlines in Alpine type mountain ranges provide powerful field evidence for direct and detailed 3-dimensional reconstructions of the surface configuration of the eroding ice masses.

In the following we present new field data concerning LGM elevation and general paleo-ice flow directions in the area of Surselva - Urseren - Goms in the Central Swiss Alps. A reconstruction of the former ice-field geometry across varying bedrock lithologies is given, based on geological and geomorphological field evidence using a Geographical Information System (GIS). The main emphases are: (a) a detailed reconstruction of the LGM surface geometry and (b) modelling of the resulting flowlines with a validation by measured field evidence for flow direction. We show that in the study area a detailed reconstruction of the LGM surface geometry is possible by means

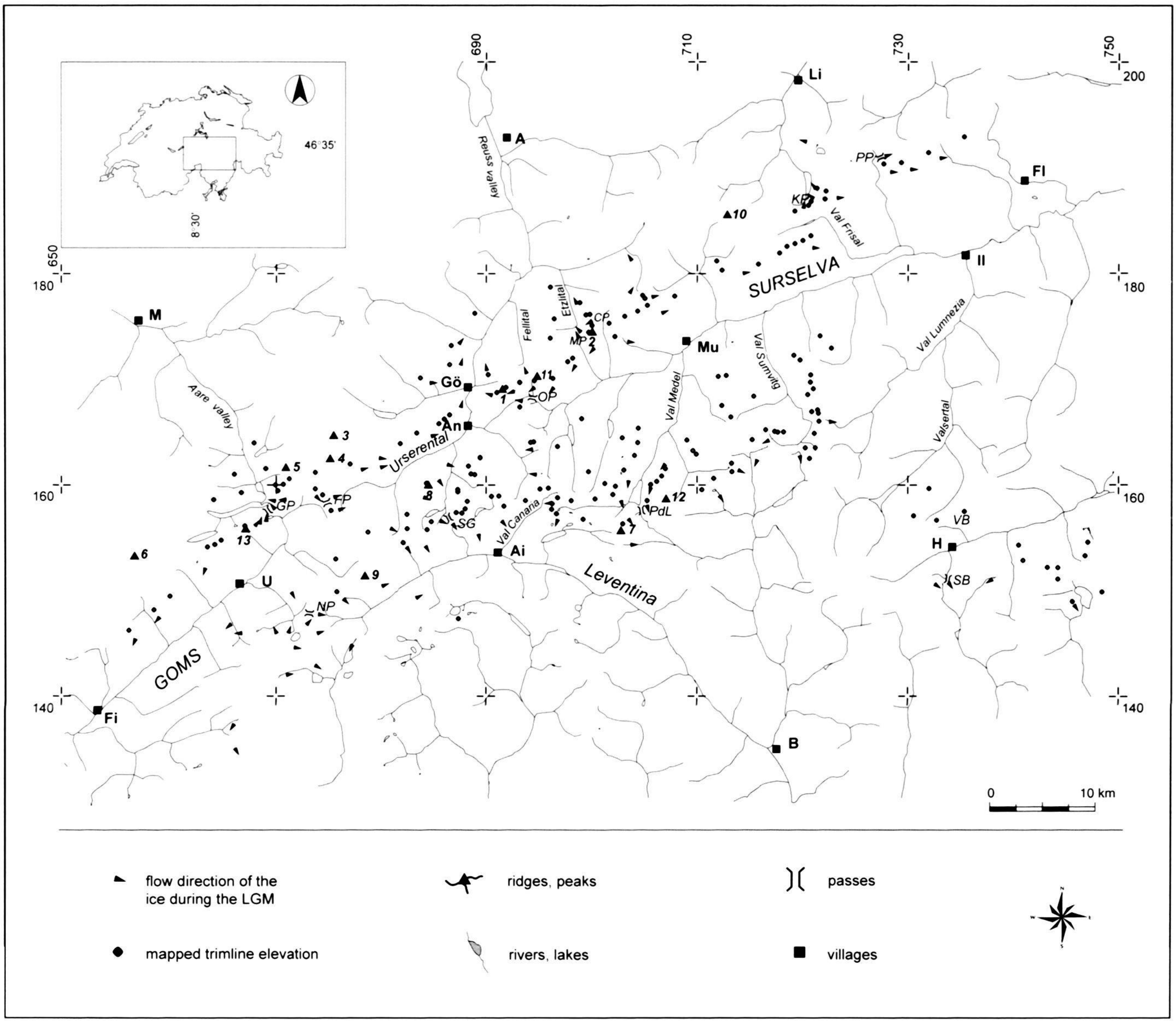


Fig. 4. Index map showing the localities of mapped features for LGM flow directions and trimline positions; Abbreviations and areal cover as Fig 2.

of trimline evidence. Moreover, the flow directions modelled using GIS and the detailed field evidence for flow direction are consistent, both showing radial flow out of two centres located above the longitudinal inner Alpine valleys to the south of the area with the highest modern precipitation rates. We interpret this with the surface geometry of the late Pleistocene ice field consisting of two dispersal centres, which developed during the build up of ice due to increased southerly circulation.

Field area

The study area extends from Flims (Graubünden) in the northeast to Fiesch (Wallis) in the southwest and includes the Central Swiss Alps and a part of adjacent Italy to the south (Fig. 2). Altitudes range between 400 m and 4200 m. The massif is divided by the Rhone and Upper Rhein valleys in the east-west direction and by the Reuss and Ticino valleys from north to south; this produces a north-eastern group of ranges (Glarner, North Bündner and East Urner Alps), the northwestern Alps (West Urner and Bernese Alps), the southwestern Alps (Valais and Ticino Alps) and the southeastern Alps (South Bündner Alps).

Numerous peaks along the northeastern and northwestern Alps exceed 3400m with Oberaarhorn (3637m), Galenstock (3583m), Dammastock (3630m), Piz Russein / Tödi (3614m), etc. The southwestern and southeastern Alps are somewhat lower; the mean altitude measures around 2700 m and the highest peaks reach 3100 to 3300 m. The whole area is characterised by well developed Alpine erosional landforms including cirques, hanging valleys, serrated ridges, arrêtes, steep peaks and spurs with the highest part of the mountains remaining glaciated today. The main glaciated areas coincide with the highest mean annual precipitation and are located to the north (Elm: 1650 mm/yr; Linthal: 1816 mm/yr; Meiringen: 1356 mm/yr) and south (Fusio: 1599 mm/yr; Airolo: 1552 mm/yr; Biasca: 1615mm/yr) of the longitudinal valleys, which show a more continental climate with mean annual precipitation rates of 916 mm/yr for Ilanz and 899 mm/yr for Fiesch (Fliri 1984).

Forming part of the three main continental river basins, the study area is drained by the Rhone to the Mediterranean, by the Ticino to the Adriatic Sea and by the Rhein, Reuss and Aare to the North Sea. The continental divide runs south of the Rhein valley, then crosses the passes of San Gottardo, Furka and Grimsel to continue westwards along the ridge to the north of the Rhone valley.

The longitudinal valleys (Goms, Urseren, Surselva and the upper part of the Leventina valley) which are strongly controlled by tectonic boundaries (root of the Helvetic nappes) display in their upper parts surprisingly low gradients (1–2%) in contrast to the generally much shorter and steeper, N–S striking tributaries.

Outline of the Geology

The glacial history in the accumulation zone of formerly glaciated areas is inferred for the most part from glacial-erosion traces on bedrock. Bedrock lithology is thus crucial for the shape, development and weathering resistance of all glacial-erosion features. Stratigraphy, petrography and tectonics of the area as described by Bearth (1951), Oberholzer (1955), Stalder (1964), Labhart (1977, 1985), Trümpy (1980), Wyss (1986) and Steinmann (1994) are significant to our interpretations. In order to exclude dependency of the trimline altitude to bedrock lithology, the investigated peaks and ridges lie within a roughly 100 km long and 30 km wide corridor aligned parallel to the general strike of the main nappe boundaries and include all major lithologic units (Fig. 3).

The entire sequence of rocks shows effects of low regional Alpine metamorphism that increases from northwest to southeast. Minor structural elements occur in approximately all units; strong foliation is developed especially in the Urseren- and Garverazones as well as in the Tavetsch massif. This is important for physical and chemical weathering and thus for the conservation of glacial erosion features discussed later.

Methods

Glacial-erosion features were recorded on more than 200 mountains and ridges in order to determine maximum LGM elevation and flow directions of the ice. Mapping was done by fieldwork and occasionally by interpretation of oblique aerial photographs. Coverage was good on both sides of Urseren valley, on the passes of Grimsel and Lucomagno and all along the ridge to the north of the Rhein valley; it was generally poor to the south of the Rhein- and in the Rhone valley. The map in Figure 4 shows the main localities of the LGM - trimline positions and the ice-flow directional record.

Glacial erosion

At different scales, form and size as well as the processes and patterns of glacial erosion are controlled by glaciological variables (shear and normal stress) and the thermal regime (cold or temperate ice) at the glacier bed as well as by the physical characteristics (lithology, jointing, layering and preglacial weathering) and morphology of the glacier bed. Generally, the preservation of erosional features is best in fine-grained igneous rocks, while it is poorest in carbonate rocks, slates and gneisses due to rapid weathering by granular disintegration or exfoliation.

Ice-flow directional record

Linear scratches and arch shaped fracture marks on bedrock surfaces have long been recognised as products of glacial erosion (e.g. Agassiz 1838; Chamberlain 1888; Gilbert 1906) and could also be described by mechanical experiments (e.g. Ficker et al. 1980). These features are striae, crescentic fractures (Parabelrisse), crescentic gouges (Sichelbrüche) and lunate fractures (Sichelwannen). Striae strike parallel to the local ice movement and thus provide the orientation of former ice flow but not its direction. This is also the case for the arched crescentic fractures, which may have their horns pointing up-glacier or down-glacier (Harris 1943; Dreimanis 1953). Crescentic gouges and lunate fractures are similar, except that their horns point in opposite directions; the open or concave side of crescentic gouges faces up-glacier whereas it points down-ice for lunate fractures (Fig. 5a). Both features are bound by two fracture planes, a principal fracture plane (p) dipping down-ice and a vertical one (v), intersecting the principal plane. As the principal fracture plane always dips in the direction of glacier flow - independently of bedrock lithology – crescentic gouges and lunate fractures provide a perfect method for reconstructing former ice flow directions (Fig. 5a) (Harris 1943; Dreimanis 1953; Ficker et al. 1980; Wintges & Heuberger 1980). Their diameters range from a few centimetres to about one meter. Commonly these superficial, small-scale erosional marks are superimposed on erosional bedfoms of intermediate scale, ranging in size from 1m up to 1 km, including whalebacks and roches moutonnées. Whalebacks describe elongate, symmetrical and streamlined bedrock clumps that have been smoothed on all sides by the glacier (e.g. Sugden & John 1976), primarily representing a product of glacial abrasion. Striae and other small-scale features may therefore cover any surface of the whaleback. On the other hand, roches moutonnées show elongate, upstanding and asymmetrical morphologies with a gently sloping abraded stoss face and a steeper, rougher and quarried lee side. Small-scale erosional forms are restricted to the stoss sides, sometimes showing even patches of glacial polish (Fig. 5b). Bedrock lithology and structure (jointing, bedding and foliation) may strongly influence the detailed morphology of roches moutonnées (Rastas & Seppälä 1981; Laitakari & Aro 1985; Sugden et al. 1992). Nevertheless, statistically the orientation of the stoss slopes and lee sides and in particular superimposed crescentic gouges and lunar fractures allow roches moutonnées to serve as highly reliable ice flow indicators. Finally, small- and intermediate-scale erosional features are on the surface of the glacially sculptured large-scale forms such as glacial troughs and U-shaped valleys. The latter are best developed in the crystalline rocks of the Aare and Gotthard massifs.

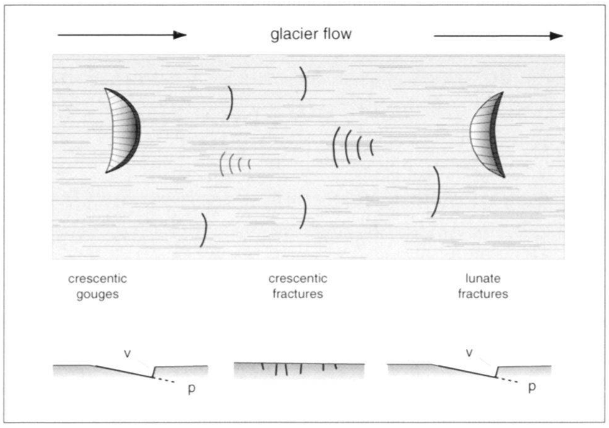
Ice-elevation record

The upper limit of glacially moulded bedrock and erosional features is often discernible in the field as a straight line engraved into topography. This line, also known as trimline, denotes the active ice surface of the last glaciation at its maximum thickness. It is the maximum erosion level of the glacier advance into pre-existing weathered bedrock (Fig. 5c). Thorp (1981, p. 49) defined it as 'a narrow zone separating strongly ice-moulded, from frost-affected bedrock, below and above the trimline respectively'. Bedrock ridges and spurs above the trimline are serrated and lack striae, whereas below the trimline striae, chattermarks and sometimes glacial polish are present. The boundary interval varies from a few meters to 50–100 m. It is clearest on truncated ridges oriented perpendicular to the main ice flow direction, forming obstacles. In addition, the trimline itself describes a distinct break in slope of the U-shaped valley walls, the character of which mainly depends on bedrock lithology. In the field, we generally distinguish two types: A decrease of slope below the trimline is referred to as (a) concave (Schliffkehle), an increase as (b) convex. Concave forms are typical in massive igneous rocks (e.g. granites, diorites, and gabbros), whereas convex forms mainly occur in 'soft' rocks (sediments in general or intensely jointed crystalline rocks). Because crystalline rocks are more resistent to weathering, most conspicuous trimlines show concave morphologies and are best developed where numerous mountain ridges and steep peaks poke up above the former ice surface as nunataks. Moraines are absent at the trimline.

Assumptions

As mentioned above, the trimline describes the maximum erosion level of a former glacier in (weathered) bedrock. The basic problem is whether or not the origin of this maximum erosion level in the Central Alps and the well-preserved terminal moraines in the northern Alpine foreland (e.g. Wangen a. d. Aare, Killwangen, Schaffhausen) pertain to the same event and time period. Deposition of the terminal moraines has usu-

a)



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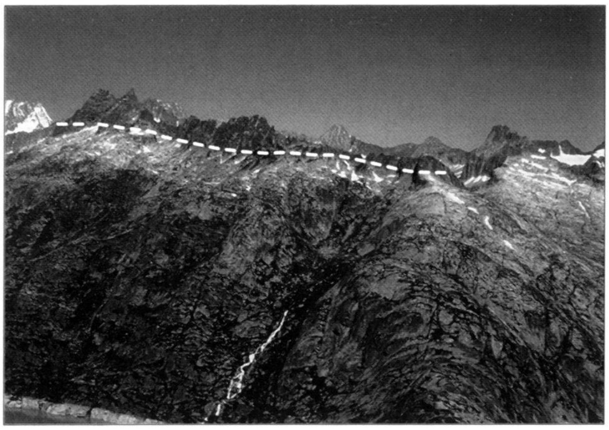

Fig. 5. Characteristic features of glacial-erosion used for the reconstruction of former ice elevation and flow-direction: a) Small-scale forms of glacial erosion (plane and profile); b) Glacially shaped roche moutonnée on Passo di San Gottardo showing a gentle, striated stoss face formed by abrasion and a steep plucked lee-slope. Ice flow was from right to left; c) Glacial trimline on Juchlistock to the northwest of Grimselpass, separating bossed, smoothed, ice-moulded topography from an upper serrated crest of weathered bedrock. The boundary marks the upper limit of glacial erosion, describing the surface of the last glaciation at its maximum thickness.

ally been correlated with the global ice-volume maximum (e.g. Schlüchter 1989); this is portrayed by the marine oxygen-isotope record, which is assumed to largely reflect global ice-volume fluctuations. According to the deep-sea δ18O record (e.g. Chappell & Shackleton 1986; Martinson et al. 1987; Shackleton 1987), the 'marine LGM' occurred at around 20'000–18'000 14C years BP. The correlation of the terrestrial LGM with the marine LGM was confirmed only recently by means of exposure age dating of several erratic boulders (Ivy-Ochs 1996) on the terminal moraine at Wangen a. d. Aare (Nussbaum 1910, 1951). Accordingly, the Rhone glacier reached its maximum extent at 20'000 ± 1800 years ago (Ivy-Ochs 1996) in concert with global ice-volume maximum of isotope stage 2. In contrast to the erratic boulders on the moraines, it is more difficult to determine the age of the trimline. Hence, presumptions of a glacier expansion to the trimline during an earlier stage of the last glaciation (e.g. isotope stage 4) or even during a glaciation prior to the Würm glaciation (Haeberli & Penz 1985; Haeberli & Schlüchter 1987) cannot be excluded. However, when taking weathering into account, assuming the calculated erosion rate of 1–1.5 mm/yr for the last 15'000 years in eastern Switzerland (Müller 1993), there would be more than 65 m of erosion since isotope stage 4 or 200 m since stage 6; even a very small erosion rate of 0.01 mm/yr for a mountainous climate with steep relief (Press & Siever 1986) would produce at least 0.65 m and 2 m of erosion, respectively. As a result, not only glacial polish and small-scale erosional features would have been removed, but even the distinct trimline would not be visible any more. From this point of view, until exposure age datings become available, the timing of the trimline as described here is assumed to have occurred simultaneously with the LGM in terms of greatest global ice extent at around 20'000 years BP (isotope stage 2). A second assumption concerns the accuracy of the use of the trimline to reconstruct the configuration of ice age glaciers, as the trimline only marks a minimum elevation of the ice surface. This is because the process of abrasion and plucking at the glacier bed requires a minimum ice-thickness that could have amounted to a few tens of meters. Because the excess thickness of ice is unknown, we infer former ice surfaces from the trimline directly. Nevertheless, in areas with alpine topography, trimlines allow an accurate reconstruction of the vertical dimensions of vanished valley glaciers, providing important input for glaciological models or paleoclimatic studies (Thorp 1981, 1991; Ballantyne 1989, 1990; Florineth 1998).

Field results

The state of preservation of striae and friction marks is highly dependent on bedrock lithology. Glacial polish was found on only a few roches moutonnées on the passes of San Gottardo and Grimsel where the bedrock is a fine grained granite and ice-flow was constricted. In order to exclude features due to Lateglacial or Holocene glacier advances, the field survey was restricted to peaks and ridges near the trimline and to passes outside the moraines of the Younger Dryas advance.

In the following sections, we describe the quality of mapped field evidence for trimline altitude and flow direction as shown in Figure 4.

Trimline

North of the Rhein river a conspicuous trimline occurs on almost all peaks and steep ridges and shows consistent elevations across lithologic and structural breaks. It is best developed and coherent in the igneous granitic rocks of the Aar massif. There, the trimline separates not only the smoothed and striated crests from serrated ridges that lack striations but it also marks a distinct change of colour from light grey below to darker grey above; a phenomenon which is also known from today's glacier retreats. The trimline elevations continuously rise northeastwards from 2510 m at Chli Schijen in the west to 2620 m at Piz Tiarms and 2640–2680 m between Chrüzlistock and Piz Dado; it then decreases to 2540 m eastwards to Panixerpass. Northwards, the trimline slopes from 2650 m above the Rhein valley to 2620 m at Chrüzlistock and then at a steeper gradient along Felli- and Etzlital to 2200 m on the north ridge of Bristenstock. East of Piz Russein (Tödi), physical and chemical weathering of the autochtonous sediments of the Aar massif aggravated the determination of the exact position of the trimline. South of the Rhein valley trimline data are less frequent because large areas are covered by unfavourable bedrock such as slates (Bündnerschiefer), dolomites and gneisses (cf. Fig 3). Trimline elevations often had to be recognised just by means of ambiguous evidence like the break in slope of the valley walls. Nevertheless, they show consistent elevations with those on spurs to the north of the Rhein valley and display surprisingly constant elevations of 2620–2660 m in the area between Val Maighels and Val Lumnezia. Furthermore, trimlines along tributary valleys match trimlines of the Rhein valley at confluences and rise in smaller tributaries. However, south of the Rhein valley trimline elevations generally decrease from north to south. The specific elevation values on crests along the western flank of Val Medel are, from north to south: 2640, 2620, 2630, 2620, 2610 and 2580 m at Pizzo dell'Uomo. A similar trend of the trimline can be seen along Val Sumvitg.

The Urseren valley, in the central part of the study area, is underlain by the Hercynian granitic and granodioritic intrusions of the Aar and Gotthard massifs, showing a well-developed trimline with concave morphology. The northern flank exhibits a coherent trimline that gradually decreases from 2760 m on Furkapass in the west to 2400 m in the east above Andermatt and 2300 m to the northwest of Göschenen. Ridges on the southern flank also show consistent trimline elevations. On the nunatakker of Pizzo d'Orsino for example, the trimline is at 2640 m and regularly slopes towards south to 2590 m at Passo del San Gottardo and to about 2500 m in the Valle Leventina.

Finally, trimline evidence is very scanty and becomes less satisfactory in the western part of the field area. It is limited to the crystalline crest to the north of the Goms and to the granitic body of 'Rotondogranit' to the north of Nufenenpass. Along the northern flank of the Rhone valley, trimlines gently rise from 2640 m in the southwest to about 2800 at Juchlistock, then slope down to about 2700 m at Sidelhorn to the west of Grimselpass and in very distinct form further northwards along the Aare valley to 2450 m at Alplistock. To the northeast of Grimselpass a similarly well-defined trimline at Gärstenhörner marks the former ice surface at 2780 m and gently rises towards the Rhone glacier. Trimline elevations to the north of Nufenenpass are unclear, but lying between 2680–2750 m, they are consistent with those around Grimselpass.

In summary, trimline mapping yields a detailed picture of the LGM ice sheet surface, revealing individual ice domes located in the headwaters of the Rhone- and Rhein rivers drained by ice streams. This implies a northerly shift of the position of ice divide at the LGM with regard to the present position of the water divide. Maximum trimline elevations of the eastern dome attain about 2650 - 2700 m and thus, are somewhat lower than the western dome. From these two high elevation zones, trimline altitudes gradually descend in all directions. The smooth decrease to the east and west, parallel to the longitudinal valleys is thereby in contrast to a much steeper decline across the ridges to the north and to the south.

Directional record

The Rhein valley

The area of Oberalppass (2044 m) between Rhein and Urseren valley shows evidence of overriding by ice. The pass area in cross section has a typical U-shape and on both sides numerous striated surfaces occur to an altitude of 2600 m. Stoss and lee features as well as crescentic gouges indicate a westerly transfluence of Rhein to Reuss ice not only at Oberalppass but also further south at the glacially-moulded surfaces of Lolenpass (2399 m) and Pass Maighels (2420 m) (Eckhardt 1957; Hantke 1980). North of the Rhein valley glacially smoothed granitic outcrops of the Aar massif are common, with well-preserved small and intermediate erosional features, but without glacial polish. Taken together, striae, chattermarks and the orientation of roches moutonnées give evidence of the main ice-flow to the north - northeast, with northerly ice-transfluences over Fellilücke (2478 m), Mittelplatten (2472 m) and Chrüzlipass (2347 m). As described before, the glacially smoothed surfaces of the jointed sedimentary rocks of the Aar massif east of Piz Russein suffered exfoliation and granular disintegration that removed most of the small scale erosional features. Nevertheless, occasional relict stoss-and-lee features as well as the large-scale morphology of the east-west oriented crest between Val Frisal and Limmernsee indicate that ice spilled to the north at both Kistenpass (2625 m) and Panixerpass (2407 m).

We found only few striations in the upper parts of Valsertal and Hinterrhein. Here, the gneisses of the Penninic Adula nappe show overall smoothing on a large scale but exhibit moderate weathering (exfoliation) on a medium scale and thus mainly lack small scale features. Nevertheless, flow direction is supposed to have been towards south across Valserberg (2504 m) and Passo del San Bernardino (2065 m).

South of Mustér (Disentis) the northern ridge of Piz Scopi shows evidence of overriding by ice, with the altitude of glacially smoothed surfaces ranging from 1900 to 2640 m. Former ice flow was towards the south and southwest (!) as can be inferred from stoss-and-lee features. The coarse-grained granitic bedrock displays a well-preserved trimline, but surface granular disintegration removed most of the shallow striae and glacial polish below. The orientation of rock faces with striae, crescentic gouges and numerous roches moutonnées in the area of Passo del Lucomagno (1908 m) document likewise a consistent flow direction to the south (e.g. Penck & Brückner 1909; Bleuler 1998). One exception are west-east trending erosional features at the confluence of Val Medel and Val Cadlimo. Here, ice flowing down-valley from Val Cadlimo merged with Rhein ice, which resulted in a deflection of its natural flow direction of more than 90°. On the other hand, at the western end of Val Cadlimo the ice was drained to the west across Botta di Cadlimo and along Val Canaria to the Ticino glacier.

The Urseren valley

Smoothed-and-bossed granite outcrops and a typical U-shape in cross section characterise the whole area of Passo del San Gottardo (2109 m). Again, the trend of striae and the orientations of abundant crescentic gouges, lunar fractures and roches moutonnées all show southward flow (Penck & Brückner 1909; Lautensach 1912; Hantke 1980). Moreover, perfectly preserved patches of highly polished granitic surfaces on the stoss side of roches moutonnées around the hospice area are indicative of strongly erosive basal ice. This can be explained by the fact that ice flow across the pass area was constricted and thus, in order to preserve ice discharge with constant thickness, surface velocity had to be increased. From mapped trimline elevations, the minimum ice thickness is calculated to at least 480–500 m for this southerly transfluence to the Ticino glacier. Above that, the Ticino glacier was additionally supplied with ice from the catchment area of the Rhone river. At Nufenenpass (2478 m) crystalline outcrops with stoss-and-lee features and occasional preserved striated patches around the hospice and to the north of it, show former ice flow to the east - southeast. However, sedimentary rocks (Bündnerschiefer) to the south of the col show intense physical weathering and no longer provide evidence of glacial erosion.

The Rhone valley

The Rhone glacier 'lost' ice not only by the southeasterly transfluence across Nufenenpass but also across Furkapass (2431 m) to the Reuss glacier (Hantke 1980) in the northeast and towards the north over the Grimselpass (2165 m) to the Aare glacier (Agassiz 1838; Penck & Brückner 1909) (Fig. 4).

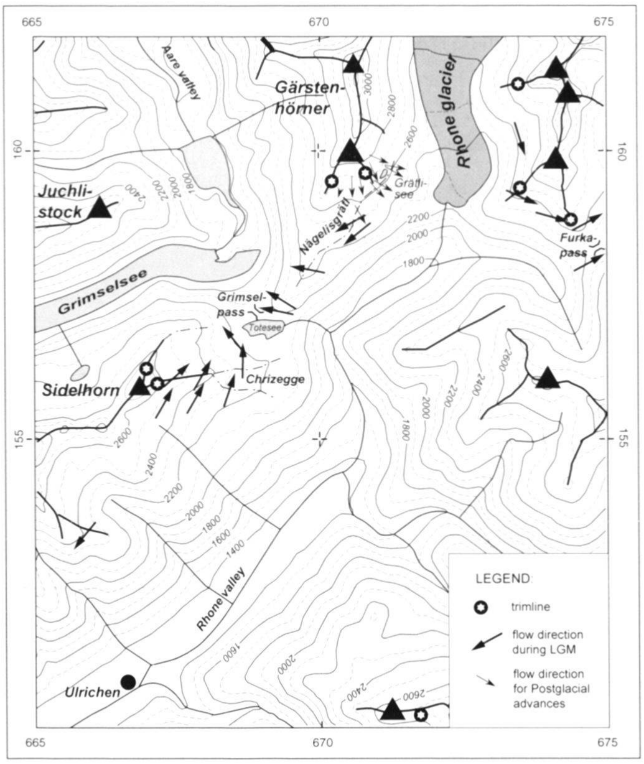

Fig. 6. Detailed index map of the Grimselpass area. The arrows picture measured flow directions associated with the LGM ice extent and Postglacial advances and the stars give the positions for observed trimline altitudes. Note the rotation of flow directions to the east and west of the pass. For general position see Fig 2.

Especially the massive granites (Zentraler Aaregranit) on Grimselpass and to the north of it exhibit one of the most spectacular glacially sculptured landscapes in Switzerland including a wide range of small to large-scale erosional features with excellent preservation. The highest parts of this area remain glacierized today, and a distinct trimline delimits serrated ridges and sharp peaks from bedrock ridges with striae, showing overall smoothing on a large scale (Fig. 5c). Stoss-and-lee features and striated rock surfaces generally show a north- to northeastward ice-flow direction whereas trimline altitudes indicate a thickness of at least 600 m. A detailed survey of the directional features along the ridges between Sidelhorn and Gärstenhörner, however revealed a more complicated and differentiated flow pattern (Fig. 6). On Nägelisgrätli, striae between the moraines of 1850 and the altitude of approximately 2550 m consistently trend south to southeast with 112° to 165° around Grätlisee and with generally decreasing values (112° to ~ 100°) towards the east. Outcrops between 2500 and 2600 m show two sets of crosscutting erosional features. The first set consists of stoss-and-lee features, indicating ice flow towards the west (230°–250°), whereas striated patches overlying their smoothed surfaces were cut by ice flowing from northwest to southeast. The latter probably relate to a Postglacial readvance of smaller local cirque glaciers to the south of Gärstenhörner. This advance however, did not descend below 2500 m, as lower elevation surfaces show only one set of striae that run parallel to the westerly flow direction derived from numerous stoss-and-lee features. We noticed further a rotation of the strike values between Nägelisgrätli and Grimselpass from 230° at 2539m to 300° at 2395m and 340° to the north of Totsee (2160m). In contrast to the high-elevation sites, the bedrock in the pass area exhibits distinct crescentic fractures and gouges, striae and patches of glacial polish with excellent preservation and with hardly any signs of weathering on the stoss faces of abundant roches moutonnées. The clockwise rotation of the strike values is continued also to the west of the pass road. Strike values of both small- and medium-scale erosional features along the eastern ridge of Sidelhorn show that former ice-flow changed from northwestwards (340°) in the east to northwards (360°) at Chrizegge and to northeastwards (48°) to the east of Sidelhorn: this involves an extensive glaciation in the Upper Rhone valley with the ice divide located above Gletsch - Ulrichen rather than just a larger Rhone glacier.

Finally, the rounded U-shaped pass profile, the break in slope on both sides of the col and smoothed outcrops to the north of the hospice point to an ice body no more than 300 m thick which overflowed Furkapass to the east.

In summary, the reconstruction of flowlines by glacial-erosion features in the area of the Central Swiss Alps defines a star-like pattern of former ice-flow directions, similar to reconstructed flow patterns of the late Pleistocene ice sheets in North America, Scandinavia or Greenland. We may distinguish two such dispersal centres from where the ice was able to flow radially in all directions; one is in the Surselva (upper Rhein valley) between Oberalppass and Disentis and a second in the Goms (upper Rhone valley) upstream of Ulrichen. The ice was drained predominately downstream along the Rhein and Rhone valleys but it also spilled over the ridges to the northwest and to the southeast of these longitudinal valleys to merge with the ice streams of the Aare, Reuss and Ticino glaciers. This is consistent with the conclusions drawn from the trimline elevations described above. Moreover, flowlines deduced from striation trends are perpendicular to former ice-surface contours as determined by trimline elevations and likewise confirm the hypothesis of two individual ice-domes. Further, the interpretation of mapped erosional features occurring up to the high altitude trimline as well as the existence of this very trimline shows that during the last glacial the ice body was warm-based and at pressure-melting point throughout when it reached maximum LGM thickness. Had the glaciers been cold-based and frozen to the ground, glacial erosion and especially abrasion would be absent or at least much less obvious.

Modelling of the LGM by GIS

All the results from mapping (trimline evidence, distribution altitudes of erratic boulders and contour lines of older recon-

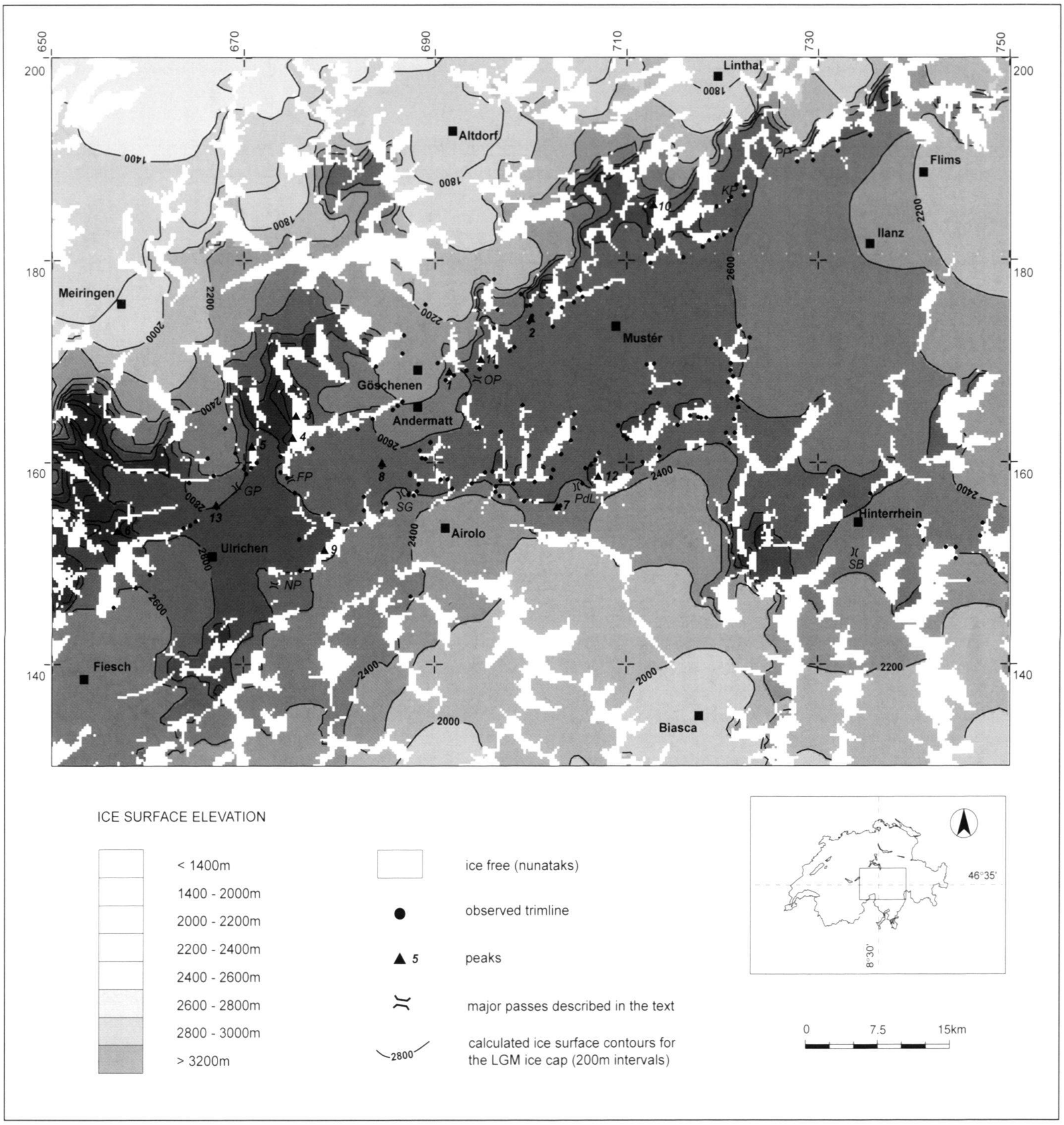

Fig. 7. Three-dimensional reconstruction of the LGM accumulation area in the Swiss Alps with contours at 200m intervals (for the time when ice stood at the trimline). Nunataks are kept in white whereas the ice surface is represented in contoured elevation zones at intervals of 200 m. Darker shading indicates higher altitudes.
The calculations for this GIS based model are principally based on mapped trimline evidence. The area enclosed by the contour line of 2600 m roughly marks the extension of the ice domes. Abbreviations as for Fig. 2.

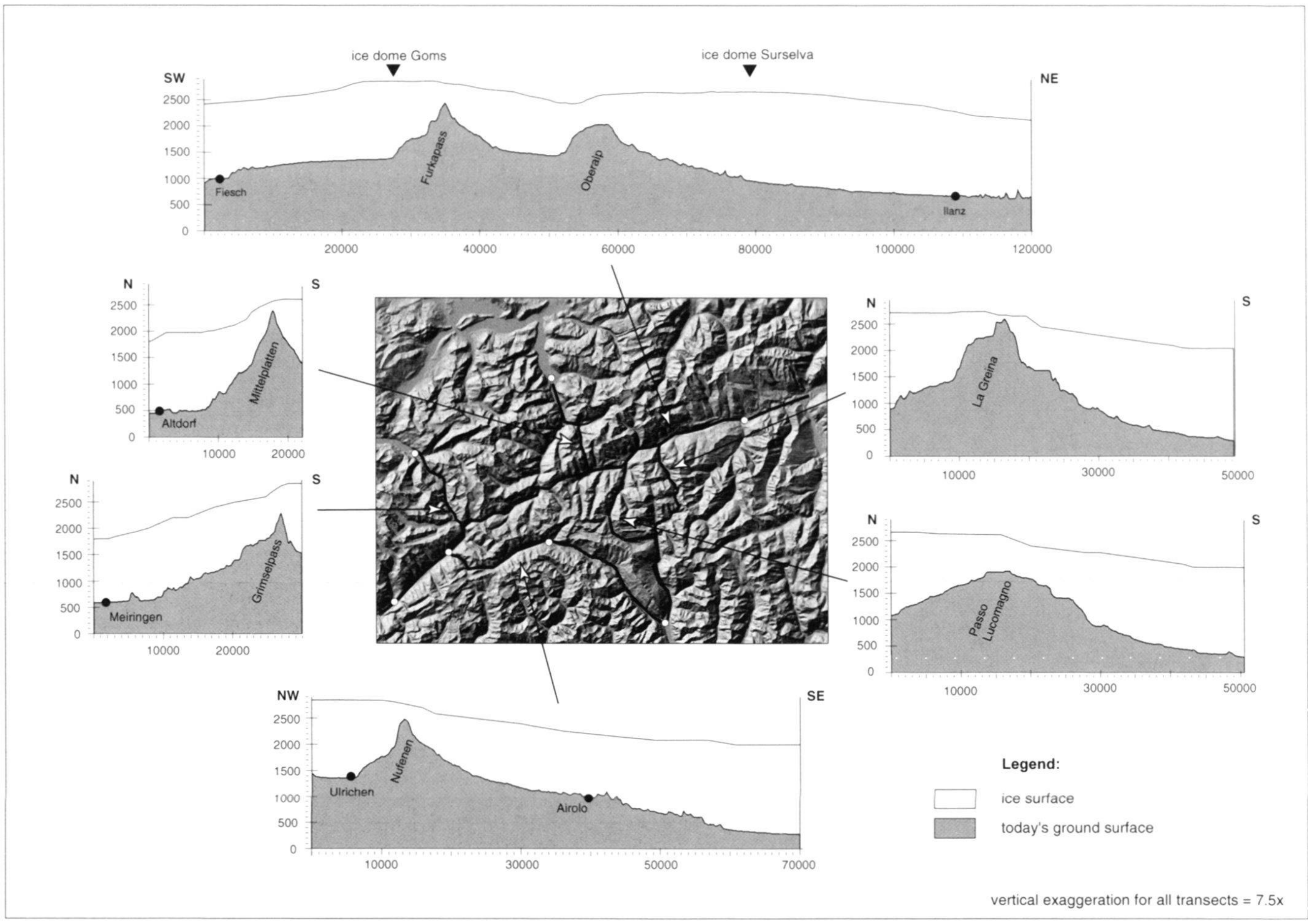

Fig. 8. Surface configuration of the modelled LGM ice surface on different transects aligned along the reconstructed flowlines running parallel and perpendicular to the longitudinal valleys. Note that the profiles display a minimum value for ice thickness because today's ground surface does not always represent the corresponding glacier bed to the reconstructed ice surface.

structions) were used to compute the three-dimensional geometry of the LGM ice surface in the accumulation area with the help of a geographical information system (GIS). The GIS provides sophisticated interpolation methods (algorithms) for terrain modelling and powerful tools for surface analysis (Weibel & Heller 1991), which also proved to be satisfactory in assisting the reconstruction of a former ice surface just by means of trimline evidence (Florineth 1998).

Any digital representation of the continuous variation of a surface is commonly known as digital elevation model (DEM) or digital terrain model (DTM). The shape of a smooth, even surface given by a set of point data can be modelled by applying various types of three-dimensional mathematical functions. However, for a surface of random variation such as the earth's surface, data networks have turned out to be most efficient. These are the gridded network and the triangulated irregular network (TIN) described by Peuker et. al (1978). Both data structures are numerical abstractions or approximations of a surface in the real world. The TIN is a terrain model that uses a sheet of continuous, adjacent triangles, computed from irregularly spaced observation points with geographical location (x, y) and corresponding elevation (z) (Burrough 1987). Among the several known methods of triangulation, the most common one is based on Delaunay triangulation. It consists of a two-dimensional network of non-overlapping triangles where no points in the network are enclosed by the circumscribing circles of any triangle (Delaunay 1934). A triangular plane is an unequivocal description of a plane in space and therefore each triangle edge includes not only elevation but also slope and aspect attributes. Unlike gridded models, the points of a TIN model are stored in an irregular pattern, allowing the density of points to be greater in areas of complex relief and less in those areas where relief is comparatively simple, thus avoiding data redundancy.

Modelling the ice surface

The basis of our surface model consists of a set of irregularly distributed points in three dimensions (trimline elevations) and line data representing digitised contour lines of the maps published by Van Husen (1987) and Jäckli (1970). Digitising of the latter was already carried out in the context of a diploma thesis (Hegner 1995). Contour lines inside the mapped area have subsequently been partly removed or adapted to our field results. The most common formats of representing DTMs are the lattice and surface grid, and consequently, the created TIN surface model was converted likewise to a lattice of regularly spaced sample points for further analysis. We used the linear interpolating method (ESRI 1996) to calculate mesh point z values of the newly generated lattice. The last step of the modelling was to intersect the low resolution DTM of today's ground surface with the calculated lattice representing a continuous glacier surface when ice stood at the trimline. This way it was possible to clip peaks and ridges protruding the LGM ice surface as nunataks so as to get the 'true' area and volume of the ice body on one hand, and to reconstruct a more realistic Ice Age topography on the other hand.

Figure 7 shows the GIS based reconstruction of the LGM in the study area. In general, glacier flow was constrained to flow valleys so that the geometry of the ice body for the most part displays a strong relationship to the present drainage system (cf. Fig. 2). However, the modelled ice surface shows also regions with a drainage pattern that is independent of today's ground surface, especially the areas of elevated ice surface above the headwaters of Rhein and Rhone. Plotting both surfaces simultaneously on different cross-cutting projection planes running parallel as well as perpendicular to the longitudinal Rhone - Urseren - Rhein valleys best illustrates the deviations of the reconstructed ice surface from today's ground surface (Fig. 8). The NW - SE transects show a progressive decline in ice surface altitude to the south- (Fig. 8 a/b) and northeast (Fig. 8 c/d/e), dipping away from the maximum altitude of about 2650 m and 2800 m respectively, located right above the centre of the Rhein and Rhone valleys. Relatively steeply declining ice surfaces can be seen along other NW - SE transects, too. The longitudinal SW - NE transect (Fig. 8 f) shows a dome shaped culmination to the west of Furkapass and a second to the east of Oberalppass with a depression in the area of Andermatt. Surface elevations decrease to the SW and NE from each of the two vertices, but with a smoother gradient. In order to get a clearer picture of today's ground surface, the profiles were selected so as to run parallel to the riverbeds. Furthermore, today's ground surface does not automatically represent the corresponding glacier bed of the modelled ice surface. Instead, it portrays its maximum elevation and thus, a minimum ice thickness. This is true especially for the parts of Aare, Leventina, Reuss, Rhein and Rhone valleys displaying low gradients, where the postglacial sedimentary filling may amount to several hundreds of meters!

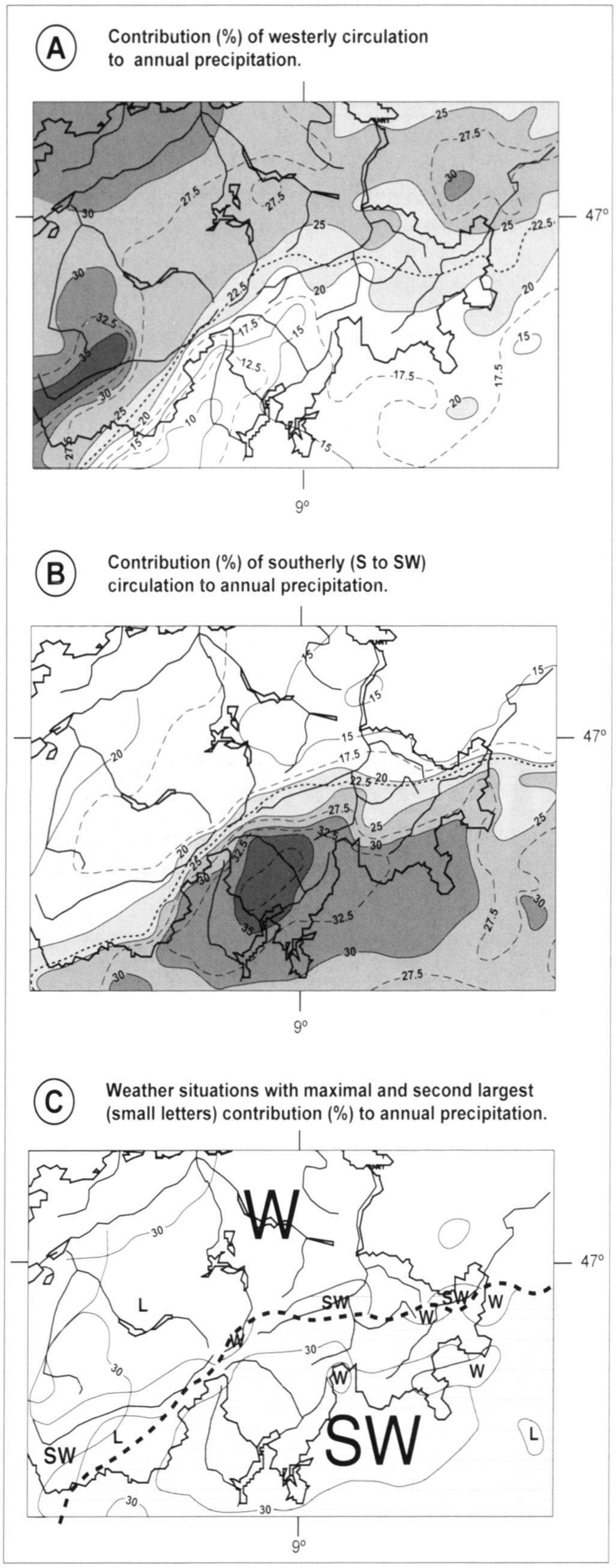

Fig. 9. Characteristics of the Alpine precipitation patterns and related atmospheric circulations (redrawn from Fliri, 1980).

Paleoclimatic implications and discussion

Located on the border between Central and Southern Europe, the weather in the Alps is influenced by all major air currents and they occupy an important place with regard to global climate. The northern Alps, the Central Plateau and the Jura Mountains exhibit a humid and temperate climate that is strongly influenced by the warm surface water of the North Atlantic, whereas climate at the southern side of the Alps is influenced by the Mediterranean with dry summers and precipitation maxima during spring and autumn. Finally, the sudden change from plateau to mountain additionally modifies the precipitation pattern due to uplift of incoming air masses, causing increased precipitation on both sides of the Alps and reduced precipitation in the inner Alpine valleys.

This large-scale impact of the Alps is supposed to be true also for the climate in the recent geological past. Accordingly, a detailed reconstruction of the LGM ice surface in the Alpine accumulation area permits to investigate past climatic conditions and the link between oceanic circulation in the North Atlantic and the prevailing atmospheric circulation pattern during the time of ice build-up from 28'000–23'000 BP (Schlüchter 1991; Schlüchter & Röthlisberger 1995).

At present, Switzerland's weather is affected by four main European air currents – from the Atlantic, the Mediterranean, the eastern continent and the northern subpolar region. The continental air masses from the north and east rarely cause precipitation, but rather dry and chilly winds in winter and warm winds in summer. Because of its position just inside the zone of the Northern Hemisphere westerlies, northwesterly to westerly winds bring moisture from the North Atlantic across Central France to the Alps, with two distinct peaks in summer and winter [winter 30%, spring 21%, summer 29%, autumn 20% (Fliri 1984)]. The focal point of precipitation lies in the northern part of the Alps and diminishes at a large scale towards the south and east (Fig. 9a). A second group describes the warm and moist southerly oceanic winds (foehn weather situation), with air flowing across Southern France and northern Italy. This situation is particularly important during spring and autumn [winter 25%, spring 29%, summer 18%, autumn 28% (Fliri 1984)]. It brings the highest precipitations in the central and southern Alps, generally decreasing from south to north and from west to east (Fig. 9b). It is thereby possible to characterise the single areas where the different weather situations make the largest and second largest contribution to annual precipitation (Fig. 9c). It is evident that precipitation in the Alps is mainly influenced by westerly and southwesterly airflow. The boundary between the areas of influence of these two prevalent weather situations on annual precipitation, here called 'weather divide', lies north of the continental water. It roughly runs along the Rhone valley in the west across Furkapass and along the ridges to the north of the Urseren and Rhein- valleys to continue eastwards to the south of the Rhein river to the east of Mustér. At present, (north-) westerly circulation is about 40% more frequent than southerly and southwesterly circulation, with precipitation related to southerly and southwesterly circulation to amount at most to 16% of the total annual precipitation, and with more than 22% for (north) west wind weather situations (Schüepp 1979; Fliri 1984). Therefore, total annual precipitation today must be considerably higher in the Central Plateau and the northern Alps than in the inner Alpine valleys and adjacent southern part of the Alps.

However, was the frequency and annual contribution of these 'modern' circulation- and precipitation patterns in the central part of the Alps the same in the past? Paleoclimate archives (e.g. peat bogs, tree rings) generally provide detailed information about former local temperatures and vegetation, but these studies are not yet extensive enough to describe airshed scale trends in past precipitation such as arid, dry, moist or wet. Today we know from numerous pollen analyses of sediments and bogs that both temperatures and precipitation in Central Europe were drastically reduced during glacial times (e.g. Guiot et al. 1989; Frenzel 1990). However, these data have not yet been extrapolated to reconstruct more detailed patterns of past precipitation or former atmospheric circulations. Yet, the surface configuration of a vanished ice body could serve as an important tool for reconstructing the dominant atmospheric circulation. It may provide insight into past precipitation patterns and thus outline main precipitation areas. Despite the large uncertainties due to deformation by ice flow, melting or wind drift, surface contours of the LGM ice body roughly represent the principal inner Alpine precipitation areas during the time of ice accumulation.

From this analysis we recognise that the contours of the modelled ice surface and isolines of precipitation do correspond to a substantial extent to both weather situations. The western ice dome (Goms), situated approximately on the weather divide, can be affected to a similar degree by precipitation related to west and southwest airflow. On the other hand, the eastern dome (Surselva) is located to the south of the weather divide. It therefore seems to show a stronger influence by southerly circulation types. Considering the fact that all passes to the north of the longitudinal Rhone -, Urseren - and Rhein valleys display transfluences towards the north, the hypothesis of an increased contribution of southerly airflow for the whole study area gains further support. In addition, a study in the eastern Swiss Alps (Florineth 1998) revealed likewise increased southerly atmospheric circulation during the last glacial. We assume, therefore, that at least during the last phase of the Würm glaciation, the build-up of the ice in the central part of the Alps was in principal related to precipitation by southerly winds, similar to today's foehn. The quantitative influence of the westerlies to precipitation cannot be assessed here, but in connection with the results from the eastern Swiss Alps; it appears that the maritime influence from the North Atlantic on western Europe was reduced. The stronger influence of southerly circulation compared to westerly circulation for the LGM can be explained by a southward migration of the North Atlantic atmospheric Polar Front in response to the advancing lid of sea ice to as far south as 40–45° (CLIMAP 1976). A corollary of this interpretation is that the production area and the prevailing tracks of the precipitation-bearing mid latitude cyclones were shifted likewise towards south, to flow from west to east across the Mediterranean, promoting increased southerly circulation to the Alps. As a result, the Mediterranean area and the southern fringe of the Alps received increased precipitation relative to the area to the north of the Alps. The reconstruction of the LGM climate on the basis of pollen data supports the hypothesis of relatively wetter conditions to the south of the Alps - Pyrénées line than to the north of it (Peyron et al. 1998). Additionally, the lake records from southern Europe show higher lake levels at 18 ka BP than at present, suggesting moister-than-present conditions during the LGM (Harrison et al. 1996). On the other hand, the decline of precipitation away from the Southern Alps promoted drier conditions to the north of the Alps (Frenzel 1990, 1991, 1992) and, due to the chilling effect of the Alpine ice sheet, severe southwesterly (southerly) katabatic winds assisted the development of arid permafrost conditions in Central Europe as outlined by Washburn (1979).

The conclusions on paleocirculation show good agreement with the results of other paleoclimatological and paleoglaciological research. A study of the glaciological characteristics of the Ice Age glaciers in the Swiss Alps (Haeberli & Penz, 1985) indicated rather inactive glaciers to the north of the Alps showing low values of basal shear stresses and of surface velocities, whereas glaciers on the southern side of the Alps appear to have been much more active. Haeberli & Penz (1985) concluded that the climate was of a cold and accentuated continental type to the north, whereas to the south of the Alps precipitation was more abundant and temperatures higher. These conclusions are supported by a study on the distribution of permafrost, being continuous to the north but not to the south of the Alps (Kaiser 1960). Paleowind indicators on extensive loess sediments (Meyer & Kottmeier 1989) and the orientation of dunes (Poser 1948) likewise indicate severe periglacial conditions in Central Europe and anticyclonic circulation along the southern margin of the Eurasian ice-sheet.

During recent years, global ice age climate has been simulated more or less successfully for the Northern Hemisphere by using general circulation models, (GCM) (Kutzbach & Wright 1985; Broccoli & Manabe 1987; Kutzbach et al. 1991; Ganopolski et al. 1998). All these numerical models show that global climate at 20'000BP was dominated by the large ice sheets that developed over North America, Greenland, Scandinavia and the British Isles, presenting a major topographic barrier to the upper tropospheric westerly jet stream. Each ice sheet was associated with a permanent cell of high pressure due to the cooling of the air above the ice. One consequence of these permanent anticyclones was the split of the jet stream into two branches that flowed along both, the northern and southern margins of the Laurentide and Eurasian ice sheets. According to the results of the GCMs, the position of the continental atmospheric Polar Front to the south of the great ice sheets led to changes in oceanic circulation too, resulting in a markedly increased extent of sea ice in the North Atlantic and a displacement of the oceanic and atmospheric Polar Fronts towards the equator. Finally, most of the climatological models display extremely low precipitation in Central Europe due to reduced evaporation in the North Atlantic and a shift of the prevailing track of the mid latitude cyclones to the south, flowing eastwards across the Mediterranean and causing elevated rainfall to the south and in the southern parts of the Alps.

Conclusions

Detailed mapping of paleoglaciological field evidence in the Central Swiss Alps allowed us to reconstruct the surface configuration of the last glacial accumulation area when ice stood at the trimline. This ice surface is characterised by two individual ice domes located in the headwaters of the Rhone- and Rhine valleys. The analysis of trimline data shows that the western dome reached a minimum altitude of close to 2900 m, whereas the eastern dome culminated at about 2700 m. From these central highs, trimlines gradually descend in all directions with steeper gradients towards the south and north than along the large longitudinal valleys. Flowlines deduced from striation trends below the trimline are perpendicular to ice-surface contours determined from trimline evidence and support the presence of a dome-type glaciation with radial outflow. Ice was drained principally by ice streams to the east and west along the longitudinal valleys but it also overflowed Grimselpass, Fellilücke, Mittelplatten, Chrüzli- and Panixerpass to the north and spilled over La Greina, Passo del Lucomagno, San Gottardo and Nufenenpass to the south. Modelling the topography of the ice surface using a Geographical Information System (GIS) is consistent with these results.

Furthermore, we used the reconstructed geometry of the ice surface to draw some general conclusions about the precipitation patterns responsible for the build-up of the last glacial ice body. The position of both of the ice domes to the south of the main Alpine weather divide suggests increased precipitation due to a generally more southerly atmospheric circulation pattern (foehn weather situation) but minor influence of today's prevailing westerly circulation. This interpretation is consistent with a southward displacement of the atmospheric Polar Front and the associated mid latitude cyclones by 10–15°; this, in turn, was caused by the advancing southern margin of sea ice reaching as far south as 40–50° which caused the related storm tracks to move eastwards across the Mediterranean.

These results agree with other paleoclimatological data, including paleo-glaciologial calculations for Alpine Ice Age glaciers, the mapped southern limit of permafrost, reported evaluations of paleowind indicators as well as with the results of climatological general circulation models for the period of the last glacial maximum (18'000 BP).

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