Abstract

Climate modelling and climate monitoring is based on information of past climatic fluctuations. Climatic fluctuations lead to changes in the mass balance of glaciers. A widely used index for the mass budget of a glacier is the equilibrium line of altitude (ELA). Six different methods and affiliated modifications are reviewed and their potential and weaknesses are discussed. The application of these approaches is exemplarily conducted for the maximum, supposedly high-Wuermian glacial stadium of the Kleinen Arbersee glacier (Bavarian Forest, Germany). The glacial setting of the Kleinen Arbersee glacier is ideal as the erosional and depositional glacial features are well preserved allowing a plausible reconstruction of the former glacier outline. A Digital Elevation Model (DEM) was produced and the former glacier outline as well as the ice-surface was inferred from geomorphological evidence. The necessary spatial analyses for the appliance of the different methods of ELA-determination were computed with a Geographic Information System (GIS)-software. The different methods of ELA-reconstruction were applied to the glacier and evaluated by defining maximum and minimum thresholds. The resulting ELA for the Kleiner Arbersee glacier can be located at an elevation of 1069 m asl (standard deviation σ =12.1 m). The Accumulation-area-ratio (AAR) approach with a ratio of 0.5 seems most reliable. The described proceedings of ELA-determination can readily be transferred to other glacial settings.

Table of Contents

Table of figures

Table of tables

1 Introduction and rationale

Climate change and its regional implications are of crucial importance to the world's population. Considerable effort has been put into research that is linked to climate change and climate modelling. Climatic changes initially influence sensitive ecological regions, in particular mountain ranges. An important indicator of climatic fluctuations is hence the mass balance of glaciers, as they respond sensibly to change in climatic parameters and leave behind evidence of their former extent. The mass balance of a glacier is predominantly governed by summer ablation and winter precipitation. HOINKES (1970) found in empirical studies that for mid-latitude, alpine glaciers the summer ablation is the dominating influence on the annually-integrated mass balance. Glaciers attempt to accommodate to variations in the mass budget, striving for equilibrium conditions. Yet glacier response is not instantaneous, but somewhat retarded allowing to distinguish different glacial stages in the field by geomorphic evidence.

To reconstruct glacier fluctuations over time and interpret them climatologically, a parameterisation of glaciers has to be conducted. Only if the specific configuration of a glacier is reduced to simple geometric features and indexed, a comparison to other glacial settings is feasible. A widely used index that describes the geometry of a glacier as a result of mass balance is the equilibrium line of altitude (ELA). The ELA can either be determined for modern glaciers by direct measurement of mass balance parameters on the glacier's surface or reconstructed for former glaciers by interpretation of the glacier's geometry. The latter is in particular important for the understanding of past glacier dynamics.

Several methods have been theoretically developed or inferred from glacier mass balance studies on modern glaciers for modelling the ELA of former glaciers. They are all based on knowledge of glacier geometry and glacier dynamic principles, and have been tested and evaluated in studying present-day glacier dynamics. The different approaches of constraining the ELA are based on visible morphological glacial features, depositional as well as erosional. It has thus to be borne in mind that these stratigraphical records are the result of a complex interaction between local climate and topographic setting, complicating the inference of their climatological implication (fig. 7).

Methods of determining the ELA have been published as early as in the 19th century; recently most of the methods have been reviewed and evaluated (BENN & LEHMKUHL 2000, CARRIVICK & BREWER 2004).

This study aims to apply the different methods of ELA-reconstruction to a restricted formerly glaciated valley and to develop algorithms that can be used to compute the necessary parameters for the ELA-determination with a GIS-software.

As a showcase catchment the valley of the Kleiner Arbersee glacier in the Bavarian Forest in south-eastern Germany was chosen, which yields an ideal glacial setting. The resulting ELAs for the different approaches are compared and evaluated regarding the local situation. Studying the Pleistocene glaciation in the Bavarian Forest is of special importance and climatological interest as the Bavarian Forest formed a small isolated glacial island in the periglacial environment between the southern fringe of the Scandinavian ice sheet and the northern rim of the Alpine glaciation during the Pleistocene. (see fig. 1)

It is intended to compile a GIS-based methodology that facilitates the application to other study areas more readily.

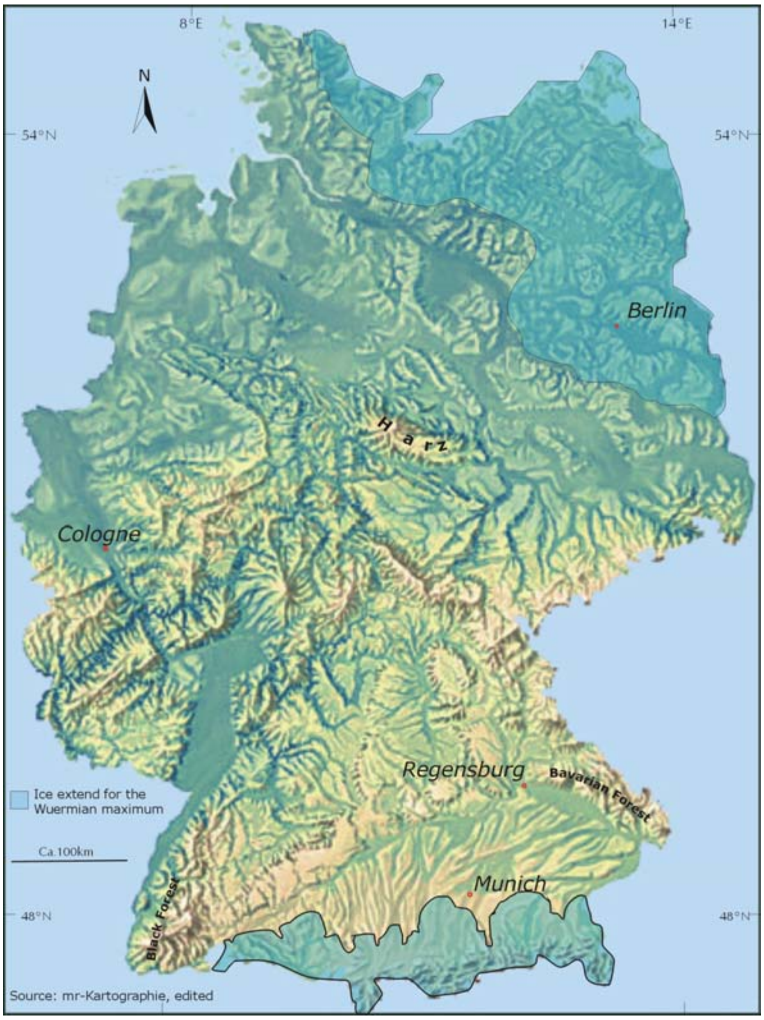


Fig. 1 Location of the ice margins in northern and southern Germany during the maximum of the Wuermian glaciation

This study is organised in eight chapters. Firstly the study area is introduced (chap. 2), then the theoretical background of ELA-determination is reviewed (chap. 3), followed by a chapter on the different methods of ELA-reconstruction with their underlying principles, their potential and weaknesses (chap. 4). In chapter 5 the parameterization of the Kleiner Arbersee glacier is presented and in chapter 6 the different ELA-approaches are applied to the glacial setting. Discussion and interpretation of the results (chap. 7) is followed by the conclusion and perspectives for further research (chap. 8).

2 Study area: Catchment of Kleiner Arbersee

The Kleiner Arbersee glacier catchment was chosen as a suitable valley to compare different methods of reconstructing glacier mass balance with the equilibrium-line of altitude (ELA) approach. The formerly glaciated area of the Kleiner Arbersee glacier is well studied, mapped and surveyed; hence high-resolution topographic data exists, leaving no drawbacks to the study from the viewpoint of data availability. The Kleiner Arbersee glacier was restricted in extent and left behind a staircase of well preserved moraine arcs and distinct features of glacial erosion, allowing to draw an outline of the former glacier with some confidence. The glacier had a simple geometry and shape, which rules out a number of reservations associated with certain approaches of modelling the ELA.

2.1 Geological and geomorphological setting of the study area

The study site Kleiner Arbersee (N49°07', E13°07') is located in the eastern part of the Bavarian Forest (Hoher or Hinterer Bayerischer Wald) and is also referred to as the Bohemian Forest (Böhmer Wald). From a geological point of view the Bavarian Forest is part of the geological complex Bohemian Massif (Böhmische Masse). This Precambrian basement is subdivided into different units by large tectonically fractured zones. The study location is part of the Moldanubicum, which forms the southern and south-western boundary of the Bohemian Massif. (MÜLLER-HOHENSTEIN 1973, TROLL 1967a, b) The Moldanubicum largely consists of crystalline rocks of metamorphic or magmatic origin. During the initial orogeny clay-sandy sediments were transformed into paragneiss by metamorphism and subsequently folded repeatedly and uplifted as a mountain range. The orogeny was associated with granitic intrusions in the gneiss. (TROLL 1967b, ROHRMÜLLER et al. 1996, GÜMBEL 1894)
One of the major fault zones of the Moldanubicum is the Bavarian Pfahl which divides the landscape unit Oberpfälzer-Bayerischer Wald and forms the boundary line between the Hoher and Niederer Bayerischen Wald along the cities of Viechtach, Regen and Grafenau.
Three prominent mountain ranges are located in the Hohen Bayerischen Wald: (1) Hoher Bogen, (2) Arber-Kaitersberg mountain range and (3) the mountain crests of Osser-Falkenstein-Rachel-Lusen, with a general Hercynian trend that runs parallel to the tectonic zone of the Pfahl. The mountain crests are interrupted by small cross

valleys and are separated from each other by wide longitudinal valleys. This tectonic division into NW-SE striking mountain crests is result of the Variscan orogeny. (MÜLLER-HOHENSTEIN 1973, TROLL 1967a, b, ROHRMÜLLER et al. 1996)

Until the late Tertiary the mountain ranges were deeply eroded and large peneplains formed under tropical climatic conditions. The formation of peneplains was interrupted by different upheaval phases. Extensive plains in different altitudes still bear witness to the interplay of erosion and uplift in the past. (MÜLLER-HOHENSTEIN 1973, TROLL 1967b)

The relief of the old-aged basement was superimposed by glacial and periglacial activity in the Pleistocene climate. Nowadays the glacial and periglacial deposits and landforms are ubiquitous as for instance moraines, scalloped cirques, moraines, mountain tarns, erratic boulders, periglacial slope deposits. (e.g. PFAFFL 1997, HAUNER 1980, ERGENZINGER 1967) Especially the glacial lakes are an important indicator of the glacial overprinting of the ancient basement and the peneplain landscape.

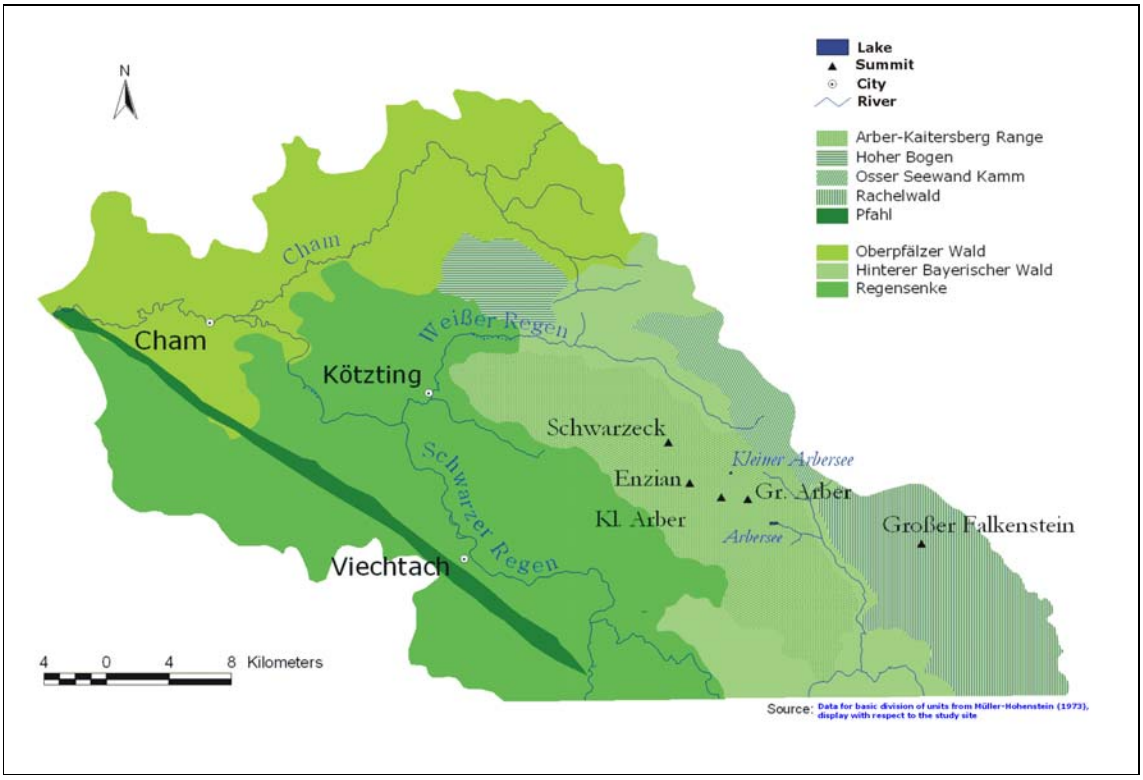


Fig. 2 Map of the Bavarian Forest and the natural setting

The NW-SE striking Arber-Kaitersberg mountain range forms the heart of the Bavarian Forest, with the summit of the Grosse Arber reaching the highest elevation of the mountain range with 1456 m asl. From this highest point, the Arber range slopes down towards NW along the peaks Kleiner Arber (1384 m asl), Enzian (1285 m asl) and Schwarzeck (1238 m asl). The Kaitersberg range adjoins the Arber range in the north.

On the Arber mountain, relicts of former peneplains can be found in different altitudes, for instance Spitzberg-Filzriegel (900 m asl), Wildau (1200 m asl) and the plain at the base of the Arber summit (1250 m asl). (MÜLLER-HOHENSTEIN 1973, ELLING et al. 1987)

The Arber range consists of shaly banded or granular sillimanite-cordierite gneiss and is partly intruded by granite. Granites are not exposed in the study area. The gneiss is a facial homogeneous paragneiss, originating in clay schist and greywacke. The metamorphic bedrock is heavily jointed with the joint planes dominating the erosion, reflected in steep bedrock cliffs on either side of the Arber range. The Arber summit shows the exposed bedrock, block trains can be found on the high slopes. (TROLL 1967b, ROHRMÜLLER et al. 1996)

The glacial lake – Kleiner Arbersee – was artificially dammed in 1885 for the reason of wood drifting, in consequence the former surface of ca. 2.7 ha rose to an extent of 9.4 ha and a depth of ca. 8.7 to 9.7 m. (BLWW 1983/1987, SCHEUERER 1991) The present-day lake level is about 917 m asl. The catchment size is 2.79 km² and drained by the Seebach river. The receiving stream is the Grosser Regen river (mouth at 600 m asl.) and the Danubia river.

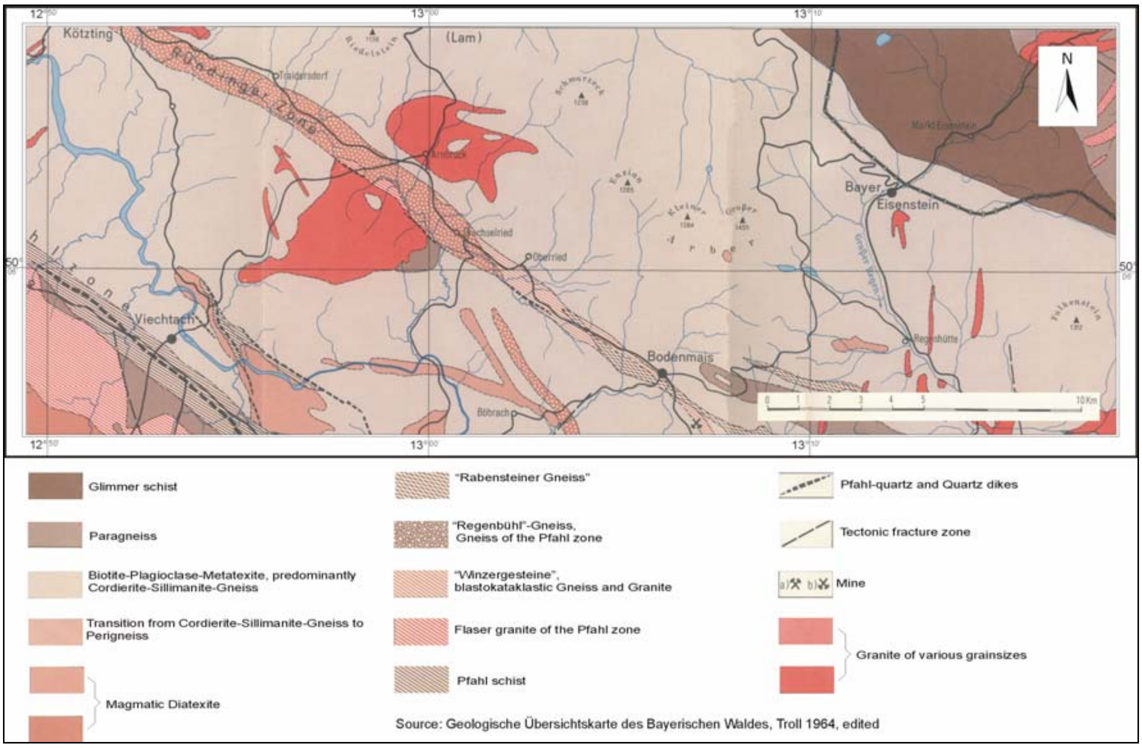


Fig. 3 Geological map of the Arber area, Bavarian Forest

2.2 Climate of the study area

Climate in south-eastern Germany is warm-moderate with precipitation in all seasons and an annual distribution of the climate parameters depending on continentality and elevation according to the climate classification of KÖPPEN & GEIGER (1972). Bavaria is situated in the transition zone from maritime in the west, to continental climate in the east. (BAYKLIMFO 1996) The eastern Bavarian Forest is primarily influenced by the westerly air masses, but with reduced precipitation due to the continental position (compared to the climate of the Black Forest about 500 km further west). (MÜLLER-WESTERMANN 1996) However, in the winter season cold air masses and fall winds (Boehmwind) (ELLING et al. 1987) from the east bring cold air masses. Due to the Hercynian strike (NW-SE) of the mountain ranges in the eastern Bavarian Forest (Hinterer Bayerischer Wald) the slopes experience a precipitation maximum on the luv sides (SE to SW) and a corresponding minimum in the precipitation shadow. (see data in table 1) The mountain crests receive a primary precipitation maximum in the summer and a secondary maximum in the winter months, mostly due to convective precipitation.

The eastern Bavarian Forest receives a mean annual precipitation of 1000 – 1800 mm with a mean annual temperature of 3-7°C. The annual temperature amplitude is with 17-18°K fairly low (lowest in the Bavarian Forest) and another indicator for a maritime influence in this part of the Bavarian Forest. (KNOCH 1952, BAYKLIMFO 1996) The degree of continentality for the Arber range is comparatively low due to the described orogenic position. (SCHEUERER 1997, HAUNER 1980) The calculation of the degree of continentality for the city of Zwiesel (further SE) however yields according to the equation of IVANOV (1959)[1] a value of about 150, with values of more than 100 showing continental climates.

The local mesoclimate is rather different from the general macroclimatic conditions. The Arber summit receives a mean precipitation of 1949 mm (data compilation by SCHEUERER 1997), where about 40% occurs in the summer months (May-August) and 32% in the winter season (November-February). The mean annual temperature is with 3-4°C low enough to maintain a snow cover for almost five months of the year. In January the mean temperatures at the summit is -4 to -5°C. (SCHEUERER 1997)

---

[1] Simplified version of the equation from IVANOV (1959) as presented in www.klimageographie.de.vu (verified for availability 08.07.04), climate data for Zwiesel from BAUMGARTNER (1970).

The Kleiner Arbersee is situated in a sheltered hollow-position, where cold-air pockets form due to nocturnal katabatic winds flowing down the steep slopes that surround the lake. The mean annual temperature is hence only 5°C, despite the altitudinal difference of more than 500 m. This low adiabatic lapse rate of about 0.5°C/100 m was already reported by BAUMGARTNER (1970) for other locations in the Bavarian Forest. The valley of the Kleiner Arbersee receives an annual amount of 1475 mm total precipitation with a mean January temperature of -4 to –3°C. The fraction of solid precipitation is yet rather high (30-40%). At the Arber summit an unbroken snowcover is reported for 150 days/yr (SCHEUERER 1997, HAUNER 1980), MÜLLER-HOHENSTEIN (1979) even reports 170 days/yr.

Station	**Elevation *(m asl)***	**Jan**	**Feb**	**Mar**	**Apr**	**May**	**Jun**	**Jul**	**Aug**	**Sep**	**Oct**	**Nov**	**Dec**	**Annual**
Lam	575	62	51	57	66	85	97	119	110	72	65	54	71	899
Zwiesel	590	91	72	64	68	80	90	112	94	72	70	66	99	978
Bodenmais	627	90	91	84	82	109	142	133	110	81	86	75	110	1208
Lohberg	650	96	69	69	82	92	100	120	107	82	89	76	103	1085
Kl.Arbersee	917	122	93	95	113	126	137	162	146	113	122	105	141	1475
Gr.Arbersee	920	148	110	106	119	133	148	180	157	123	127	112	161	1624
Gr. Arber	1325	179	169	125	127	160	197	244	183	138	146	125	155	1949
Regensburg	366	43	39	37	44	61	79	72	72	51	44	49	48	642

Tab. 1 Precipitation values (in mm precipitation) after SCHEUERER (1997) (climatic period: 1891-1930, Gr. Arber: 1951-1975) and data for Regensburg after MÜLLER-WESTERMANN (1996) (climatic period: 1961-1990)

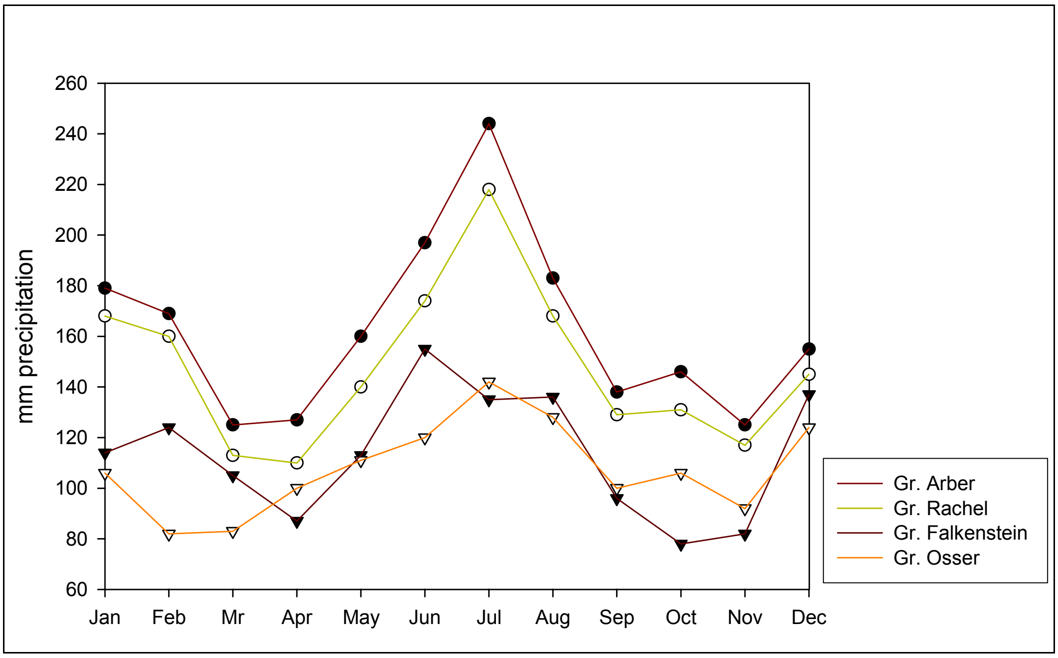


Fig. 4 Precipitation values (in mm precipitation) after SCHEUERER (1997) (climatic period: 1891-1930, Gr. Arber: 1951-1975)

The cirque locations receive a lot of snow even in the precipitation shadow due to snowdrift. In the sheltered cirques the snow melts considerably later than in more exposed locations. SCHEUERER (1997) and BAUMGARTNER (1970) report snowcover in the sheltered position of the Seeloch cirque (Kleiner Arbersee) usually until May or even June.

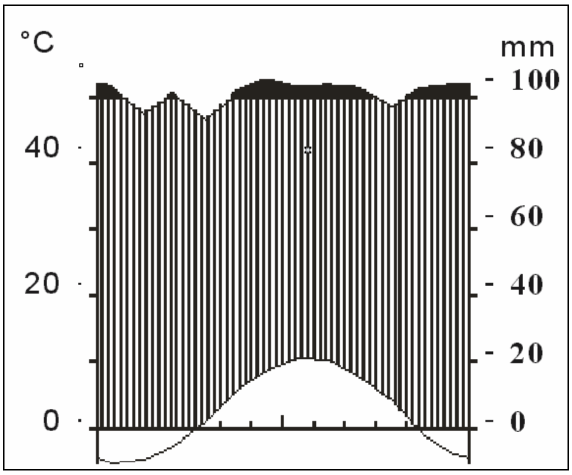


Fig. 5 Climate diagram for Grosser Arber Mountain, climate data from SCHEUERER (1991) and BAUMGARTNER (1970)

2.3 Vegetation and soil

The area of the Kleiner Arbersee is a nature reserve due to its unique coincidence of rare species and vegetation types. The dominating vegetation type is the forest with spruce, birch, fir and sycamore. Human forestry activity impacted the natural structure of the forest and favoured the spruce starting in the 8^{th} century under the rule of Charlemagne.

There is an elevational change in vegetation: between 950 and 1150 m asl. Birch, fir and sycamore are present, at lake level the spruce is ubiquitous as well as on crest positions. The Grosser Arber is the only mountain in the Bavarian Forest that reaches above the timberline. The summit of the Grosser Arber is a subalpine refuge for ice-age plants that cannot spread in present-day climatic conditions. There is a whole list of Pleistocene relict plants that occur only on the mountain summit and that are highly endangered. (SCHEUERER 1991 and 1997)

The substratum is influenced by the ubiquitous saprolites that cover the gneissic bedrock. The saprolites formed under tropical climatic conditions in the Mesozoic and the Tertiary by intensive deep chemical weathering. (VÖLKEL 1999) Bedrock outcrops formed where glacial erosion removed the overlying sediments. The dominating soil type in the study area is cambisol[2] (RAAB 1999), the altitudinal range of 800–1100 m asl so called “Lockerbraunerden” (MAHR 1998) can be found. On the upper slopes the cambisols can be podzolic. Very thin regosols are present in crest and summit locations that are barren of vegetation. In different parts the impermeable subsoil and high precipitation resulted in the formation of swamps and bogs. The dominating soil type on the valley bottoms along the creek is gley. (MÜLLER-HOHENSTEIN 1979, ELLING et al.1987) The non-Wurmian glaciated regions are covered with periglacial slope deposits. (VÖLKEL 1995, RAAB & VÖLKEL 2003)

[2] After soil classification of the **F**ood and **A**griculture **O**rganisation of the UN

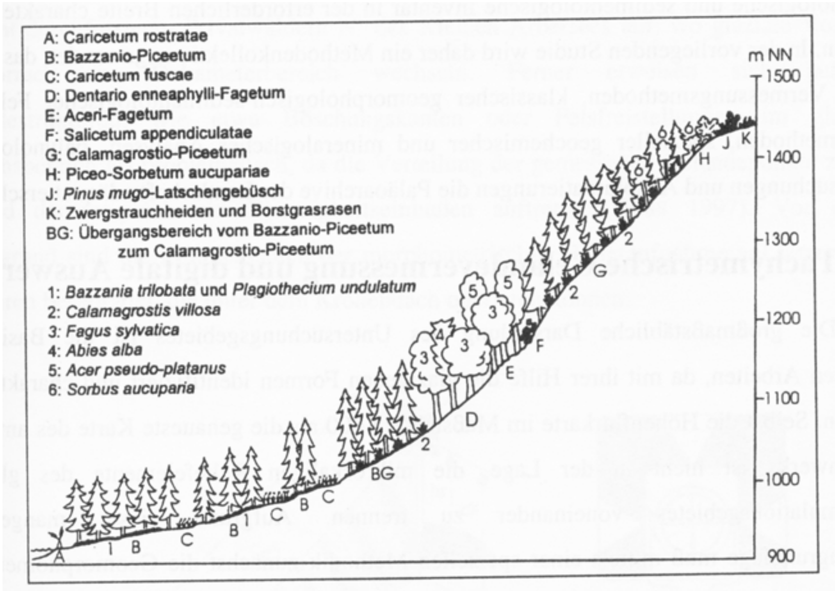


Fig. 6 Vegetation structure according to SCHEUERER (1991)

2.4 Glaciation of the Kleiner Arbersee catchment

The glacial character of the Bavarian forest was already recognized early. PARTSCH (1882) was the first to publish a fairly detailed description of the glacial landscape around the Kleiner Arbersee. More thorough early studies were realized for instance by BAYBERGER (1886), RATHSBURGER (1927), PRIEHÄUSSER (1927). Later investigations were primarily carried out by PRIEHÄUSSER (e.g. 1930), HAUNER (e.g. 1980) and PFAFFL (e.g. 1989); they published a whole list of studies on the glacial landscape around the Arber mountain. The most recent, less descriptive and more geomorphologically orientated studies were carried out by BUCHER (1999) and RAAB (1999).

Despite a controversy on possible Pre-Wuermian glaciations and the extent of the glaciated area, most investigators agree on the fact that the Kleiner Arber lake was formed by glacial erosion and that the most recent glaciation left behind a series of glacial erosive and accumulative landforms. The Wurmian Kleiner Arbersee glacier developed in a staircase of cirques at the valley head of the Seebach which drains the catchment that stretches between the Grosser and the Kleiner Arber mountain. Multiple cirques can be found on these valley slopes in elevations of about 1260 m asl., 1100 m asl., 1040 m asl. and 940 m asl. (ERGENZINGER 1967, MANSKE 1989, BUCHER 1999 and own field observations) The cirque floors are flat to sloping with an inclination of up to 35°. Down the valley of the cirques, a valley glacier formed that left behind a series of distinct lateral and terminal moraines. The moraine ramparts are about 5-10 m high, sometimes even higher, covered with many large erratic boulders (height up to 4 m). The till shows a large fraction of boulders in a loamy-sandy ground matrix. The Kleiner Arbersee is hence a glacial lake (BUCHER 1999), that is dammed by a morainic arc of terminal, possibly recessional moraines. South of the lake there is a longitudinal swampy area that ascents slowly until it hits the steep slopes that run down from the cirques. Glacially eroded bedrock, described first by RATHSBURG (1927) as roche moutonées, striations and glacial polish as well as dead-ice kettles are evidence of the glacial overprinting of the area.

The Kleiner Arbersee glacier extended about 2.5 km from the saddle between the Grosser and the Kleiner Arber northwards, with a width of about 130 to 830 m and a maximum thickness of approximately 800 m. Two more glaciers developed in the Wuermian that excavated the basin of the Grosser Arbersee to the south-east of the summit of the Grosser Arber and the smaller, already drained lake at the Bänkelschwelle further to the south of the summit.

On the other mountain ranges in the Bavarian and Bohemian forest glaciers developed in the Late Pleistocene. The extent of the different glaciers is still subject of discussion but an extensive; ice sheet glaciation as proposed by BAYBERGER (1886) can be ruled out. Even though ERGENZINGER (1967) avoids taking bias in the controversial discussion on the extent of the glaciation in the Pre-Wuermian, the ice extend in his overview map of Wuermian ice extent seems still to be the general consensus. Generally speaking, cirque glaciers formed in the Bavarian forest on north to southeast facing slopes on mountains that rise up to more than 1300 m asl. The extent of a pre-Wuermian glaciation still remains subject to speculation.

3 Introduction to equilibrium line altitude (ELA) determination

A widely used parameter to characterize Pleistocene and Holocene climate conditions in mountainous areas and to correlate former glacial stages is the equilibrium line of altitude (ELA) of a glacier. The ELA is an important proxy to reproduce glacier mass budget changes. It yields an approximation of the local climatic conditions in a mountain range. Any vertical fluctuation of the ELA indicates changes in the mass budget of a glacier and in consequence indicates changes in the climatic conditions. For studies that determine ELAs on modern glaciers, the necessary climatic parameters can be measured directly, allowing the calculation of the ELA for a defined time period. If ELAs of former glaciers are to be reconstructed, no direct measurement can be achieved; thus different methods have been developed to estimate its elevation on a glacier. In most cases the approximation of ELA for certain glacial stages is conducted by the interpretation of erosional or depositional glacial features and the geometry of the glacier bed.

However, this approach of mass budget determination is useful, but simplified. The relation between climatic conditions and the fluctuations of the ELA is in most cases complex. In different topographic settings the limiting climatic factor for glacier fluctuations might change over time. To successfully reconstruct former climates a thorough understanding of the present-day climate and climate-glacier interactions is needed. However, the climatic interpretation of glacial fluctuations is dependent on the time-scale. Glacier response to changing climate is not instantaneous, but on a time resolution of 10 to 10^2 years. The time resolution that can be inferred from geomorphological evidence is on a larger time scale and thus allows time-integrated statements on glacial stages at which moraines were deposited, and hence can be taken as an (almost) steady-state ELA. SELTZER (1994) and LOCKE (1990) point out that most research questions towards palaeoglacial fluctuations are restricted to periods of 10^2-10^4 years; over this time period several influences on the mass budget of a glacier might cancel each other out.

3.1 Definition of terms

No real consensus exists on the terminology of the altitudinal line on a glacier that reflects the mass balance. Therefore the most widely used terms and their controversial utilization are summarized here in reference to the literature. The definition that seems most appropriate for this study is introduced.

A glacier is an open system with input, storage, transfer and output of mass (fig. 7), hence any change in the mass budget will result in glacier fluctuations. The zoning of a glacier into the accumulation area, the ablation area and a separating line of zero mass balance is typically found as a result of the altitudinal decrease of temperature in the atmosphere (lapse rate) and a resulting increase in solid precipitation. The line of zero mass balance fluctuates in the course of a budget year, while changing from the accumulation to the ablation season. The main influencing parameters are the air temperature and the precipitation, reflecting the local climate. A glacier is in steady-state with the local climatic conditions and its environment, if the annual mass balance of the entire glacier is zero (no melt, no surge). (Andrews 1975, Gross et al. 1977, Flint 1970, Braithwaite & Müller 1978) Benn & Lehmkuhl (2000) point out that the steady-state of a glacier is a theoretical construct, as a particular glacier is in most instances out of equilibrium with the local climate to a greater or lesser degree. Moreover, the mass budget of a glacier is not solely governed by precipitation and altitudinal decrease of air temperature, but also by local topographic and climatological factors such as aspect, deposition of blown snow, debris cover, ice falls or steep steps in the longitudinal profile of the glacier bed; hence the altitudinal position of the line of zero mass balance can deviate from the expected elevation. (e.g. Benn & Lehmkuhl 2000, Hoinkes 1970, Porter 1975a, Gross 1983)

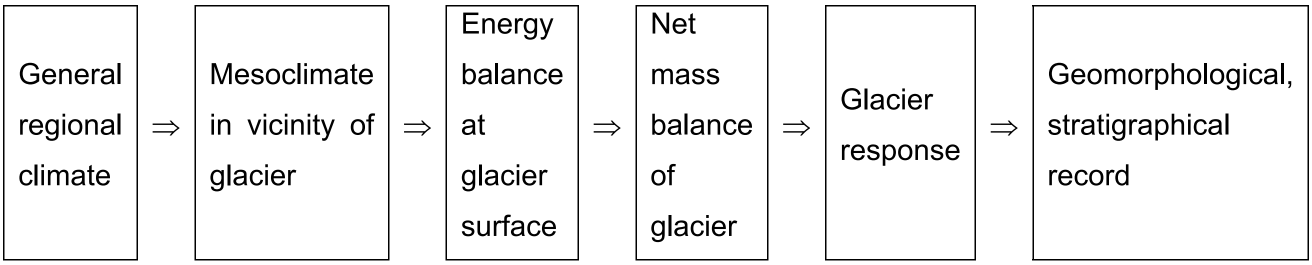


Fig. 7 Interplay between general climatic conditions, the glacial response and the resulting geomorphological record (modified from Meier 1965)

3.1.1 Equilibrium line of altitude

The equilibrium line of altitude (ELA) on a glacier is defined as the line, where net accumulation is balanced by net ablation of the glacier ice. The ELA marks, strictly speaking, the elevation, where the mass budget of the glacier is zero, integrated over

the period of a budget year. (UNESCO/IAHS 1970, ANDREWS 1975, HOINKES 1970, PORTER 1975a, PATERSON 1994, BRAITHWAITE & MÜLLER 1978)
Usually the ELA for a present-day glacier is determined by the measurement of the specific balance quantities at many points of the glacier surface for a budget year, facilitating the drawing of a reliable isoline of zero mass balance. The altitudinal dependency of mass budget of a glacier is a prerequisite for the fact that the ELA can be drawn as a contour line across the whole glacier. Usually the mass balance of a glacier is measured for the longitudinal profile and projected to horizontal areas of equal elevation. It should be noted that a position of the ELA is often very complicated and not a straight line. (GROSS et al. 1977, HOINKES 1970, BRAITHWAITE & MÜLLER 1978)
In the above summarized *senso stricto*, the ELA is the line of zero mass balance for a budget year of a glacier. It should be noted that ELAs in this strict sense can only be determined on modern glaciers, where mass balance measurements were carried out for the budget year in question. However, the term ELA is used with different facets and is often not strictly applied to mass balance measurements. It is frequently used interchangeably with the term snowline, which has little reference to the actual mass balance. (OHMURA et al. 1992, GROSS et al. 1977)

3.1.2 Firn line

The differentiation between firn line and firn edge and the discrimination to other terms is not very intuitive.
The **firn line** is according to UNESCO/IASH (1970) the lower boundary of snow deposits from the past year, which will be carried over as firn into the next balance year. Strictly speaking and in contrast to UNESCO/IASH (1970), the snow that accumulated in the budget year and marks the firn line by the end of the season cannot yet be called firn. HOINKES (1970) introduced the more appropriate German word "Altschnee" (metamorphosed snow) for the snow of the previous year that is not yet firn. Usually the firn line *senso stricto* is covered by the "Altschnee" and cannot be seen.
The **firn edge** or **firn limit** is according to HOINKES (1970) and GROSS et al. (1977) the zone, where the transition between (old) firn snow and glacier ice takes place and equals the elevation, where no further ablation occurs. The firn edge does not reflect the glaciers mass balance of the previous year, but the mean of the last couple of years. This zone can be seen more easily on aerial photographs of a glacier and is often confused for the firn line.

For temperate glaciers with simple geometry the elevation of the firn line is almost identical to the line of zero mass balance, integrated over a couple years, and can be used as a fairly good approximation of this balance line for present-day glaciers, where mass budget measurements cannot be carried out. (MÜLLER et al. 1976, MÜLLER & SCHERLER 1980, UNESCO/IASH 1970)

3.1.3 Snowline

The definition of the snowline is often very imprecise. Many authors define it vaguely as the dividing line between the area of accumulation and ablation or between the snow-free and the snow-covered part of the glacier. (FLINT 1970, LOUIS 1955, UNESCO/IASH 1970, ANDREWS 1975, LICHTENECKER 1938) GROSS et al. (1977) propose the general use of snowline as the mean of zero-mass balance lines over a couple of years. ANONYMOUS (1969) and FLINT (1970) introduce the term of an “annual snowline” for the snowline that can be seen at the end of the summer melting season. In the strict sense suggested by HOINKES (1970), the snowline is called the “transient” or “temporary snowline” at any time of the hydrological year (01.10-30.09) and turns into the firn line (Altschnee-line – see above) by the end of the budget year. This definition seems widely accepted (MÜLLER et al. 1976, MÜLLER & SCHERLER 1980, CHARLESWORTH 1957, GROSS et al. 1977, ANONYMOUS 1969, WISSMANN 1959). LOUIS (1955) even points out the fluctuations of the temporary snowline on a daily basis.

3.1.4 Regional and Climatic Snowline

The concept of the ELA of a glacier is often used to compare present-day and former glaciation and to reconstruct climatic changes. In this sense the terms regional or climatic snowline are often used.

(1) Regional snowline
The mean of all local ELA-values for a larger area is called the regional snowline. All local influences such as exposure or shading against precipitation, wind or radiation for a particular glacier are supposedly eliminated, and anything that can be traced back to the local climate conditions is smoothed out, when calculating a regional snowline. (GROSS et al. 1977, LICHTENECKER 1938, ANDREWS 1975) The resulting altitudinal zone (averaged ELAs) supposedly varies for a whole region solely with temperature and

precipitation and is often referred to as the regional snowline (FLINT 1970). Different authors (e.g. LOUIS 1955, WISSMANN 1959) use the term climatic snowline in the above sense.

(2) Climatic snowline
The climatic snowline is a theoretical concept of an ideal plane that reflects the lower limit of perennial snow on fully exposed flat surfaces. This is, however, very conceptual and does not relate necessarily to the natural environmental conditions. The climatic snowline is thus independent of relief and wind influence and cannot be empirically detected. (FLINT 1970, GROSS et al. 1977, KERSCHNER 1990)

Generally speaking, the climatic snowline as well as the regional snowline is difficult to determine by observations, especially due to the strong influence of topography on accumulation and ablation of the snowcover.

3.1.5 Definition for this study

Owing to the confusing terminology, exclusively the strict meanings of the above defined terms will be used in the following study. To refer to an equilibrium line that is reconstructed through morphological evidence for former glaciations the term *paleo-ELA* (pELA) will be applied.

3.2 Considerations for the determination of the ELA

Basically, the ELA can be estimated for glaciers by two different approaches:
(1) Methods that are based on mass balance measurements (glaciological approach), and (2) methods that take the morphology of the glacial catchment (cirque and moraine morphology), and erosional as well as depositional features into account (geomorphological approach). Over the course of the last 150 years many different approaches have been tested to model indices that represent an ELA for a glacier either for a defined year or a more or less well known period of time.
An introduction to the reconstruction of the present-day as well as the palaeo-ELA, the ELA depression as an important indicator and factors that govern the mass budget of a glacier in general is given in a following section, as these are important factors to bear in mind when modelling the (p)ELA.

3.3 Present-day ELA

The reconstruction of the present-day ELA can be most precisely conducted by mass balance measurements on glaciers. The altitude of zero mass balance fluctuates in the course of a year around a mean position. If this mean position is constant over several years, the glacier is in steady-state and the mean altitude represents the steady-state ELA. If the mass budget measurements were carried out in a period of negative or positive glacial regime the altitude of the EL does not reflect the steady-state ELA and might be too low or too high, respectively.
Besides direct mass budget measurements, the ELAs for present-day glaciers can also be derived by indirect methods (see chapter 4). Namely the Hess-contour line method can only be applied to existing glaciers, as well as the balance-ration method that require information on the mass budget; the AAR-method is more readily applied to present-day glaciers, because the outline of the glacier is well known.
Knowledge of the mass budget of a glacier can be used to evaluate the quality of palaeo-ELA determination by different methods as shown by TORSNES et al. (1993) for glaciers in western Norway.

3.4 Palaeo ELA

Information on the mass budget of former glaciers is not available and measurements cannot be conducted. There are, however, a series of methods to approximate the mass budget (and the pELA as its indicator) by hypsographic and geomorphological evidence. In most cases, it is necessary to reconstruct the catchment and the outer limit of the glacier. The field evidence, that is found to delimit the glacier, is usually based on moraines and erosional features in the glacier bed or along the valley walls (e.g. trimlines). All of these features require a certain time period to form and for simplification a steady-state for the period of the deposition and erosion, respectively, is assumed. In consequence the reconstructed pELAs are referred to as steady-state altitudes. Most published ELA measurement series show great variability from year to year, with differences of several hundred metres between maximum and minimum ELA-values (e.g. SLUPETZKY 1974, GROSS et al. 1977). At the same time it can be speculated that secular variations of glacier ELAs are only of the order of tens of metres per century. If the state of glacierization of a region is to be described with respect to the secular time scale, it would be wrong to include ELA-values for only a single year, unless that can be shown to be “typical”. (BRAITHWAITE & MÜLLER 1978)

Even though it is a fairly rough estimate, the reconstruction of the pELAs for glacial stages provides proxy data on the palaeoclimatic conditions, with the altitudinal line being an indicator for the climatic conditions.

3.5 ELA Depression

EL altitude values are a local phenomenon, depending on the local climatic conditions, the topography, the altitude of the mountain range and many other influences. For any palaeoclimatic conclusion a better comparable proxy is needed. The depression of the (p)ELA represents such a parameter. ELA depressions are often used to correlate former glacial stages, as glacial features that indicate the same pELA are supposedly of the same age.

To calculate the ELA depression a reference altitude, the reference level, has to be determined. This can either be the modern ELA of a glacier or the pELA for a well known glacial stage, such as the Little Ice Age (LIA). The latter is more favourable, because the glacier is assumed to have been in near steady-state at the time of the deposition of the LIA moraines. (KERSCHNER 1990) In contrast, the modern ELA is in most cases not an indicator for a steady-state mass budget. According to MEIERDING (1982) the theoretic modern ELA can also be approximated from the summer freezing level, if a mountain range is not high enough for glaciers to form. To calculate the ELA depression the altitude of the reference level and the pELA for the glacial stage in question have to be modelled. The approximation of the pELA for the glacial stages can be inferred using any of the geometric approaches (see chapter 4). The difference between the level of reference and any glacial stage yields the respective ELA depression for the glacier at the time of a particular glacial stage. ELA depressions are difficult to derive if the former glacier consisted of multiple catchments and hence the reference level is hard to determine. KERSCHNER (1990) proposes for this case to subdivide a glacier into different tongues and to assign a LIA ELA to each part. The level of reference is then the weighted mean of the different ELAs. (see fig. 8) No ELA depressions can be calculated for areas where neither LIA nor modern ELAs can be derived. Only if absolute ages of moraine deposits (e.g. cosmogenic nuclide dating, radiocarbon dating) exist, it is possible to calculate ELA depressions for the respective stages. ELA depressions should only be calculated, if the glacier geometry was comparable for both glacial stages and the same method of ELA determination can be applied.

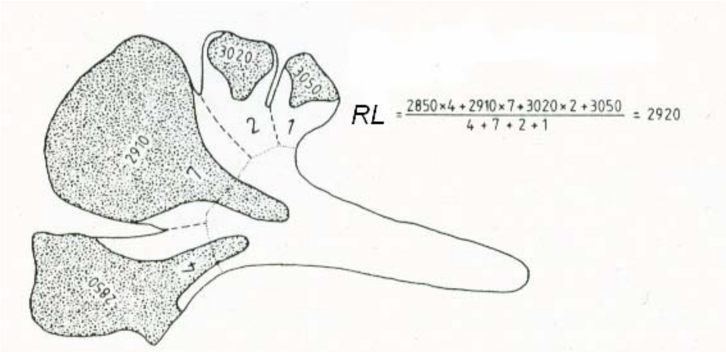


Fig. 8 Determination of the reference level (RL) for a glacier with multiple tongues. The accumulation area is divided in different tongues; to each of them a corresponding LIA-ELA is assigned and weighted by the area of each tongue. (from KERSCHNER 1990, modified)

The relative values of the ELA depression can thus be compared to depressions in the neighbouring valleys or other mountain ranges. From this difference palaeoclimatic implications on a spatial scale can be inferred. MAISCH (1987), however, warns that differences in the ELA depression might not reflect regional variations in precipitation, but might only be due to topographic variations or deviating methodological approaches. Strictly speaking, the ELA depressions can just be compared between glaciers that existed in climatological homogeneous areas.

It was proposed to produce maps from all over the world with ELA depressions for the mountain ranges, to determine a global pattern of climate change, for instance for the Last Glacial Maximum (LGM). (e.g. PORTER 1975a,b) Reservations are however expressed towards this large-scale approach (e.g. KERSCHNER 1990, MÜLLER & SCHERLER 1980), as there are too many uncertainties inherited to the methodology; for instance isostatic or tectonic uplift might influence the absolute EL altitudes and subsequently the ELA depressions considerably.

ELA depressions are valuable proxies to support palaeoclimatic interpretations, but have to be used with caution.

3.6 Factors influencing glacier mass budget with respect to the ELA

Glacier fluctuations are mainly governed by the climate parameters **precipitation** and **temperature**. Greatest influence on the glacier mass budget has the summer temperature, which governs the ablation and the amount of solid precipitation that is key to accumulation. To model the exact mass balance of a glacier some other climatic parameters have to be taken into account, such as the **radiation balance**, the distribution of **cloud cover**, warm **advection** on the glacier surface, the ratio of rain to snow, accumulation due to **wind drifted** snow as described in more detail by SELTZER (1994), MEIERDING (1982), GROSS (1983), KUHN (1980 & 1983 & 1984), TORSNES et al. (1993) and HOINKES (1970).

Besides the climatic influences the **topography of the glacier bed**, glacier **hypsometry** and the **aspect** of the glacier is crucial for the spatial distribution of accumulation and ablation zones and hence the (p)ELA. TORSNES et al. (1993) found that different angles of the subglacial bed influence the glacial response to changing climatic conditions. They found that glaciers with a gently sloping bed usually have a longer response time. WILLIAMS (1983), FURBISH & ANDREWS (1984), GROSS (1983) and HOINKES (1970) for instance even found that glaciers, which advance over steep icefalls in the altitudinal range of the ELA, do not show a simple response to mass budget change in the terminal position of the glacier. Consequently, to glaciers with no simple geometry many reservations apply towards the reconstruction of the (p)ELA.

When using the catchment geometry for the determination of the pELA (e.g. AAR-method), attention has to be paid to **avalanche nourishment** affecting the accumulation area and **debris cover** affecting the ablation area. Both show considerable interaction between topography and climatic conditions and might blur the effective catchment size. Avalanche nourishment increases the accumulation independent of climatic conditions. As pointed out by BENN & LEHMKUHL (2000) and SISSONS & SUTHERLAND (1976) the amount of snow accumulation from avalanches is controlled by a series of factors, including the catchment area and glacial hypsometry, the dominant wind direction, the altitude of the accumulation area and the characteristics of possible intermediate storage sites.

Debris-cover on a glacier changes the ablation or might even prevent ablation, as shown by rock glaciers or ice-cored moraines. (CLARK et al. 1994, LEHMKUHL & BENN 2002, KULKARNI 1992) CLARK et al. (1994) found for glaciers in the Sierra Nevada that debris cover of more than 1m leads to a significant reduction of ablation. BENN & LEHMKUHL (2000) point out that a thin debris cover might in contrast increase the ablation due to higher absorption of radiation from the debris. As it is almost impossible to account for the debris cover they call for caution, when calculating pELAs for debris-covered glaciers.

The increased ablation by **calving ice** from the glacier tongue can, according to KERSCHNER (1980), be neglected. In very steep terrain, however, caution should be taken when reconstructing ELA.

4 Methods to determine the (paleo)-ELA

In the following chapter a summary of the most widely-used techniques of (paleo)-ELA determination will be presented. Many of the techniques require detailed field work or extensive image interpretation, hence, not every approach is suitable for each glacial setting.

4.1 Accumulation–area ratio (AAR) method

4.1.1 Accumulation-area ratio

The accumulation-area ratio (AAR) method reflects general glaciological knowledge on the mass balance of glaciers. Each glacier that is in an equilibrium state over the course of a certain time period has a constant ratio of the accumulation area to the ablation area. The AA-ratio is usually expressed as area of accumulation of a glacier to its total extent. (ANONYMOUS 1969) The AAR can hence be used to locate the pELA of a glacier. Changes in the mass balance of a glacier lead to glacier fluctuations, and hence to variations in the steady-state ratio.

$$AAR = \frac{Sc}{S} \qquad \text{[Eq. 1]}$$

where Sc is the areal extent of the accumulation zone and S is the total glaciated area.

The assumption of a fixed ratio between the accumulation and the ablation area for steady-state conditions was verified by mass balance measurements on modern glaciers over different time periods. A relatively stable ratio of 0.67 for alpine areas was found (e.g. GROSS et al. 1977, HAWKINS 1985, KERSCHNER 1990). In more humid areas there is a tendency towards proportionally slightly smaller accumulation areas (AAR: 0.63), whereas in drier areas the accumulation area is slightly larger (AAR: 0.7) (GROSS et al. 1977, KERSCHNER 1990). Hence glaciers in regions with a maritime climate have a smaller accumulation area for a steady-state mass balance, whereas glaciers in more continental climates have a larger accumulation area, respectively. The data from present-day mass balance measurements and subsequent AAR determinations show the influence of topography and climate on the ratio. Deviations from the widely used AA-ratio can emerge in areas with comparable climate, but different general topographical set-up. The above given AAR values are valid for alpine glaciers, but values for ice caps, piedmont glaciers and plateau glaciers differ significantly (HAWKINS 1985, KUHLE 1986). Therefore, the AAR-approach is only applied to alpine valley or cirque glaciers. The determination of AA-ratios on modern glaciers allows extrapolation of the ratio to former glacier geometry only to a certain extent. On the one hand the present climate is certainly not representative for the general circulation during ice ages and on the other hand most glaciers are presently not in equilibrium, and hence cannot reflect a mass balance for a steady-state AAR. KERSCHNER (1990), GROSS et al. (1977) and TORSNES et al. (1993) thus use the fairly stable conditions of the Little Ice Age for the calibration of the method and again found the AAR of 0.67 to be valid.

The mass balance of a glacier is not just a function of precipitation and ablation. As described above, the additional snow accumulation by avalanching from the surrounding cirque and valley walls has to be accounted for. These additional nourishing areas have to be added to the accumulation area in locations with intensive avalanche activity. The influence of a possible debris cover has to be incorporated in the modelling of the glacier outline. When calculating the AAR for a particular glacier the regional influence of topographic shielding, leeside deposition of snow as well as the glaciers altitudinal geometry (such as a topographically forced narrow tongue and wide accumulation area) or subglacial topography has also be taken into account. (FURBISH & ANDREWS 1984, MAISCH 1987, GROSS 1983, HOINKES 1970, TORSNES et al. 1993) As pointed out by TORSNES et al. (1993) the volume of the glacier is an important parameter that is not included in the AAR calculations. TORSNES et al. (1993) showed for Norwegian glaciers that glaciers that advance over gentle sloping surfaces may grow more in thickness than in the lateral extent and hence the reconstructed AAR is different from that of a glacier advancing over steeply sloping terrain, even though their ice volume could have been the same.

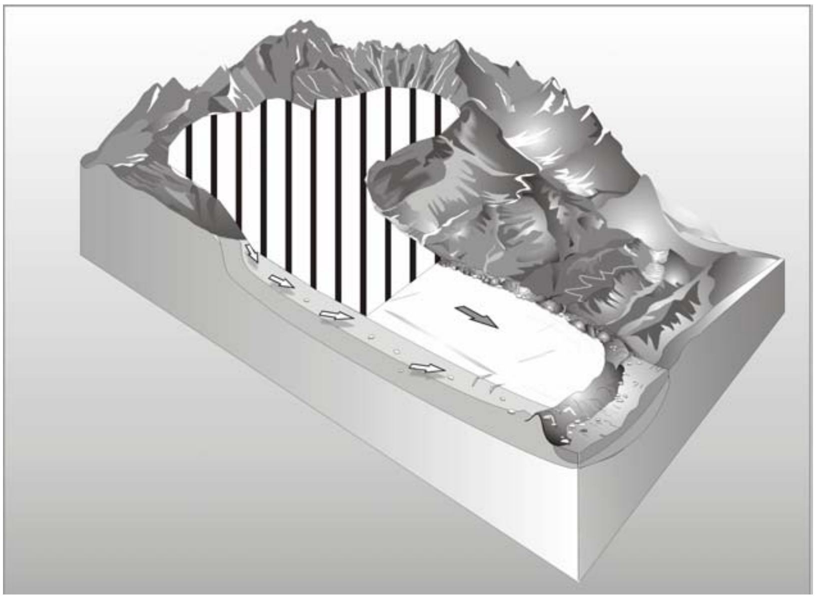


Fig. 9 Sketch of ELA-determination by using the AAR-method (for a hypothetical AAR value)

The calculation of the palaeo-ELA by the AAR method for older glacier-stages is usually accomplished by using reconstructed glacier outlines and the former glacier surface. Well developed and distinct lateral and terminal moraines mark the areal extent of the former ablation area; trimlines, the cirque and other erosional landforms mark the extent of the accumulation area. If any of these landforms are not present or cannot be unambiguously associated with a certain glacial stadium, assumptions have to be made, rendering the reconstruction subjective and spongy. KUHLE (1986) points out that for extra-alpine areas with intensive postglacial and present-day morphodynamics, young moraines are often not well preserved, adding uncertainty to the reconstruction. In cases where moraines are not present in sufficient proportions or the moraine remnants are difficult to correlate, KERSCHNER (1978) proposed to calculate a longitudinal profile of the glacier tongue as shown by Nye for the correlation of moraines and for the reconstruction of the glacier outline. He shows the applicability of this method for the Kaunertal-glacier in Austria.

Best results from the AAR-approach can be achieved if the following parameters or preconditions are considered:
(1) The surface geometry of the glacier should be of a fairly simple valley glacier, especially steep slopes lead to deviations from the typical AAR. (2) The outline of the glacier should be well defined to be rather certain of the areal extent of the glacier at the particular glacial stage. (3) The glacier that is investigated should have reached near equilibrium at the time of the reconstruction (can be assumed with prominent

moraines marking the highest point). (4) KERSCHNER (1990) describes that an area-elevation distribution with the largest surface area in the vicinity the ELA is most favourable.

4.1.2 Kurowski-method (AAR 50)

The Kurowski-method for the determination of the pELA of a glacier is the same as applying a AAR of 0.5 to the glacier. Sometimes the method is referred to as the median-elevation method (MÜLLER & SCHERLER 1980), but this term is not used here to avoid confusion with the Höfer-approach.

KUROWSKI (1891) assumed linear gradients of glacial accumulation and glacial ablation with altitude, and hence postulated the pELA to separate equal areas of accumulation and ablation. A restriction that Kurowski formulated is that the glacier, to which the approach is applied to, has to have a uniform slope in the longitudinal profile and a consistent aspect. Only very few small glaciers meet these preconditions, thus MÜLLER (1978) proposes to solely use this method for a rough ELA assessment for large-scale studies.

The assumption that the mass balance parameters of a glacier are a linear function of altitude has proven to be wrong in most cases; the decrease of ablation with altitude is faster than the increase in accumulation with altitude. (GROSS et al. 1977, SLUPETZKY 1974) Thus the pELA values that are inferred are generally too high and larger ratios than 0.5 have to be applied to reconstruct steady-state conditions. (CHARLESWORTH 1957, GROSS et al. 1977) Despite these drawbacks, the method is widely used (e.g. MÜLLER & SCHERLER 1980, CHARLESWORTH 1957, BRAITHWAITE 1984) and even proposed for inclusion in the world glacier inventory (BRAITHWAITE & MÜLLER 1980). OSMASTON (1989) however proposed the introduction of a weighing factor when determining the pELA with the Kurowski-method to adjust for the wrong assumption about linearity. FINSTERWALDER (1952) adjusted the method in assuming a parabolic change in the snow accumulation and ablation.

4.1.3 Brückner-method (AAR 0.75)

BRÜCKNER (1886) was the first to suggest that the breakdown of the areal extent of a glacier allows inference on the glacial mass balance. He assumed the glacier tongue (ablation area) to take up one quarter of the total area and hence used an AAR of 0.75. RICHTER (1888) developed this method further by stressing that any snow or ice free

cliffs should not be included in the glacier area estimate. Most reliably it can be applied to valley glaciers. This approach was empirically verified by BRÜCKNER (1886) and RICHTER (1888) for glaciers in the eastern Alps but the ratio of 0.75 is rarely used nowadays.
A list of different AA – ratios from the literature can be found in Appendix A1.

4.1.4 Balance ration (BR) method

FURBISH & ANDREWS (1984) introduced the BR-method as another method to determine the ELA. The BR-method takes account of both glacier hypsometry and an approximation of the altitudinal mass-balance distribution. Basis for this approach is the fact that for zero mass balance the annual accumulation is exactly balanced by the net-ablation of a glacier. They express this assumption as the ratio that divides the accumulation area by the ablation area, each of them multiplied by the mean value of accumulation, respectively ablation.
The BR-method requires detailed information on the mass balance. If applied to former glaciers the present-day conditions have to be extrapolated. They found for glaciers in Alaska a balance ratio of around 1.8 and propose the use of a BR of 2 as representative for maritime mid-latitude glaciers.

This relationship suggests linearity in the accumulation and ablation with altitude (BENN & GEMMELL 1997), an assumption that was the key point for criticism of the Kurowski-method. However, the BR-method can only be applied to glaciers with a linear mass-balance gradient, which are not debris covered, not substantially nourished by avalanches and have no calving tongues.

4.2 Median bedrock elevation method

The Median-Bedrock-Elevation (MBE) method accounts for the area-altitude relation - the hypsometry of the glacier. The approach relies on quantification of the aerial extent of the relict glacier bed, rather than its surface expression as done for the AAR-method. The median elevation of the former glacier bed is taken as the MBE and found to be an approximation of the ELA of a glacier. (MANLEY et al. 1997)
The MBE approach is best suited for formerly glaciated valleys. The MBE-pELA can be derived satisfactorily from digital elevation data or by planimetry of topographic maps,

but the generated pELAs are limited by the resolution of available DEMs or the spacing of contour lines on maps. (MANLEY et al. 1997)

4.3 Höfer method

The Höfer-method is historically the one most frequently applied approach when calculating the pELA. The “snowline” is determined according to HÖFER (1879) by the arithmetical mean of the vertical distance between the lowermost altitude of the glacier tongue and the average height of the crest above the firnline.

This original Höfer-method has two major shortcomings: (1) to determine “snowline” with the Höfer-method, the firnline has to be known, and (2) the determination of the “snowline” for a certain glacial stage is always based on the present-day firnline, which already renders this approach of the ELA reconstruction for former glacial stages inaccurate. (GROSS et al. 1977, CHARLESWORTH 1957) The circular argument (determination of snowline by firnline) makes this approach impossible to apply as proposed by HÖFER (1879), hence a number of modifications have been introduced.

4.3.1 Modification of the Höfer-method to apply the approach in the sense of HÖFER (1879)

To apply the Höfer-method without stumbling over the circular argument, the upper endpoint of the vertical distance was taken as the median elevation of the ridge that surmounts the glacier accumulation zone (HAWKINS 1985, MÜLLER 1978, KUHLE 1986). GROSS et al. (1977) proposed to calculate the firnline with the help of the AAR-method (ratio of 0.67).

4.3.2 Modification of the Höfer-method: Toe-to-Headwall altitude ratio (THAR) (Bergschrund-method)

A modification of the Höfer-method is the THAR-method, which assumes that the pELA can be expressed as a constant proportion of the vertical distance between the highest and the lowermost glacier limit. The THA-ratio is hence expressed as a fraction of the distance between the base of the headwall (highest glacier limit) or in the case of a glacierized cirque, the elevation of the bergschrund and the toe of the glacier (lowermost glacier limit). (e.g. HAWKINS 1985, MÜLLER 1978)

A considerable difficulty is the determination of the upper limits of the former glaciers. (HAWKINS 1985, MEIERDING 1982) It is especially difficult to locate precisely on steep cirque headwalls. Consensus seems to be that the upper glacier limit can be located where the headwall steepens to over 60° (the greatest slope attained by ice at modern glacial headwalls). (MEIERDING 1982, MANLEY et al. 1997, BENN & LEHMKUHL 2000) The difficulty to precisely locate the glacial limits shows how sensitive the THAR-method is to subjectivity in the fieldwork and how easily errors can be introduced.
Avalanche nourishment is difficult to incorporate in the reconstruction of the upper limit. (BENN & LEHMKUHL 2000, BURBANK & CHENG 1991)

4.3.3 Modification of the Höfer-method to apply the Toe-to-Summit altitude ratio (TSAR) (Louis-method)

Another modification of the Höfer-method is the TSAR-method, which was described by LOUIS (1955). The line of vertical distance is drawn between the highest elevation of the boundary of the catchment, which equals the highest point of the surrounding ridges and the toe of glacier. The elevation of half that distance is where the snowline can be estimated. By applying his modification of the Höfer-method the accumulation by avalanches from the surrounding cliffs is incorporated in the determination of the mass balance (BURBANK & CHENG 1991, LOUIS 1955)
Even though LEHMKUHL (1995) and BENN & LEHMKUHL (2000) produce pELA estimates with the Louis-method for Asian mountain ranges that are in good agreement with the field observations, they point out that in cases where the highest summit is not representative of the total catchment and may contribute little to the glacier accumulations, incorrect snowline estimates are inferred.

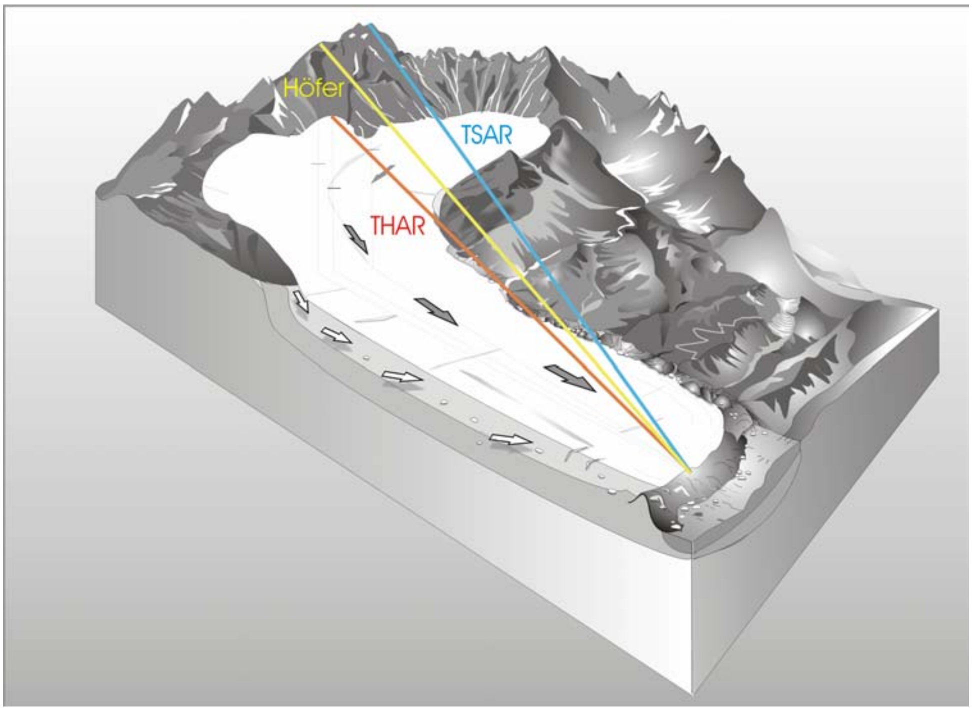


Fig. 10 Sketch of ELA-determination by using the different Höfer approaches (ticks on the vertical distance lines mark the elevation of the ELA for each approach)

4.3.4 Conclusion

The modified Höfer-methods are simply a crude but rapid and easy method to approximate pELAs, but none of them exactly reflects the line of zero mass balance of a glacier. (GROSS et al. 1977, PORTER 1975a,b, HAWKINS 1985, MÜLLER 1978, BENN & LEHMKUHL 2000, HEUBERGER 1966) The fact that the Höfer-methods are only based on two elevations renders them fairly inaccurate. HAWKINS (1985) and TORSNES et al. (1993) criticise that the Höfer-methods are not based on sensible glaciological knowledge. None of the modifications takes valley morphology, surface topography, glacier hypsometry or mass balance into account; coherently the best results are derived for small, geometrically simple glaciers with uniform area-to-altitude distribution. (PORTER 1981 & 175, TORSNES et al. 1993, BENN & LEHMKUHL 2000, OSMASTON 1989) Avalanche nourishment can be accounted for, when using the TSAR-method. Debris-cover influence, however, is neglected in any of the approaches. Empirical studies show that the pELA values that are produced by the Höfer-methods have considerable scatter. (GROSS et al. 1977)

The original Höfer-method used a constant proportion of 0.5 (half the vertical distance) to estimate the ELA. This constant value is probably not applicable for all climates and glacier types. Different authors have found ratios that range from 0.3 to 0.7 for different study areas. GROSS et al. (1977) showed that correlations of any of the Höfer-methods with the AAR-method yield pELA values that are generally too high. Correlation coefficients improved if instead of TSAR-method, the THAR-method was used. HAWKINS (1985) also found the ELAs calculated by the Höfer-method (0.5) produce ELAs that were too high if compared to empirical evidence from modern glaciers. GROSS et al. (1977) and MEIERDING (1982) found the ELAs with a ratio of 0.5 to be at the order of 100 to 150 m to be too high. MEIERDING (1982) showed that a ratio of 0.4 could improve the ELA estimates significantly. A more recent study of TORSNES et al. (1993) shows in contrast that the Höfer-methods underestimate the ELA by up to 340 m for modern glaciers in Norway. To obtain ELAs comparable to the AAR-method a THAR value of 0.7 has to be used for their study area. A list of different ratios from the literature can be found in Appendix A2.

The advantage is that the modified Höfer-methods do not require detailed topographic maps or aerial photographs and knowledge of the mass balance for the estimation of the pELA. (MEIERDING 1982, BENN & LEHMKUHL 2000) The elevation of the upper limit and the toe of the glacier can in most cases be read off a map. Reconstructing the base of the headwall however is more challenging. If distinct terminal moraines are missing, the valley morphology allows an estimation of the maximum glacial extent. The glacial limit can be set to the point where the transition from glacially eroded (U-shaped valley) to stream-eroded (V-shaped valley) in the valley cross section occurs. (e.g. MEIERDING 1982)

4.4 Maximum elevation of lateral moraines (MELM) approach

The snowline can be estimated by the maximum elevation of the lateral moraines of a glacier as suggested by LICHTENECKER (1938). The basic principle is that lateral moraines can only form along the margin of the ablation zone. Glacierdynamically, the deposition of lateral moraines is only possible, where the flow lines of the glacier ice-body point outwards, conveying the englacially transported till to the surface. (ANDREWS 1975, CHARLESWORTH 1957, FINSTERWALDER 1952) These outward pointing flow lines appear only in the ablation zone, and hence deposition of till occurs solely below the ELA. The highest elevation of lateral moraines therefore marks the minimum elevation of the ELA for the glacial stage the moraines belong to.

The upper altitude limit of the preserved moraines generally underestimates the ELA due to fact that the englacial till will not reach the surface exactly at the pELA;

consequently even pELAs that were reconstructed from well preserved lateral moraines might be too low. Postdepositional moraine degradation or non-deposition of material especially for steep valley walls may also lead to erroneously low ELA elevations. (TORSNES et al. 1993, GROSS et al. 1977, BENN & LEHMKUHL 2000)

One precondition is that the ELA was stable at a certain stage for an extended period of time in order to deposit distinct lateral moraines. The MELM-method is based upon the assumption that the maximum elevation of the lateral moraine reflects the steady-state pELA of a glacier, which might not always be the case. (HAWKINS 1985) During a slow melt-down of the glacier, with periods of stagnation or short readvances, the glacier deposits additional moraine arcs. Those younger moraines might blur the distinct morphology of the older moraines of interest and be misleading for the reconstruction. (HAWKINS 1985, BENN & LEHMKUHL 2000) Large talus slopes, debris flows or other mass movements from the valley sides may also obscure the outline of the moraines. Even after detailed field work, it is difficult to determine whether or not the lateral moraine is entirely preserved in the upper parts. This difficulty increases with the age of the landform. (HAWKINS 1985)
However, the MELM-method can be used to approximate modern and palaeo-ELAs. The upper limit of the lateral moraines for different glacial stages can be determined either by traditional field studies or by applying remote sensing techniques. Even though the moraines might still be preserved, it can be somewhat challenging to identify the moraine crest on aerial photographs due to possible tree cover or by confusion of other features such as slope deposits or even bedrock ridges. Tributaries might have cut through the moraine crest, breaking the formerly continuous landform into small ridges that complicates the detection of the continuation. (MEIERDING 1982) However, as pointed out by BENN & LEHMKUHL (2000), the MELM-method is very suitable where a detailed reconstruction of the glaciated area is not possible and mass balance information is not available, and can hence be used as a rapid approximation of the lowest possible elevation of the pELA.
This method is used extensively and often employed to calibrate other approaches by giving a glacierdynamically solid minimum estimate of the pELA. (BURBANK & CHENG 1991, HAWKINS 1985, KERSCHNER 1990, TORSNES et al. 1993, GROSS et al. 1977, MEIERDING 1982, AMMANN et al. 2001, KUHLE 1986, ANDREWS 1975, WILLIAMS 1983)

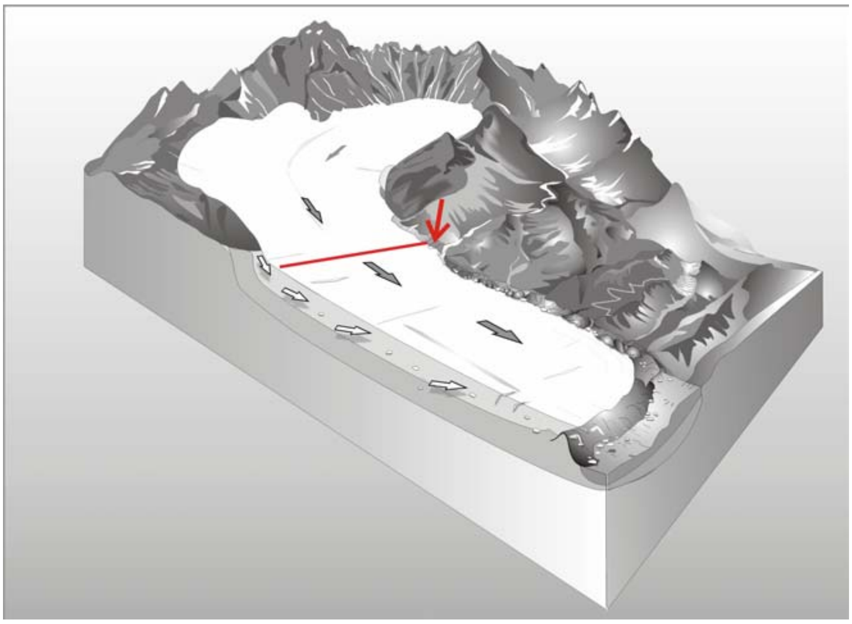


Fig. 11 Sketch of ELA-determination by using the MELM method

4.5 Cirque-floor altitude approach

A fairly simple method of determining the pELA is the elevation of the cirque-floor. For former cirque-glaciers the altitude at the base of the headwall of the cirque (cirque floor) was taken as an indicator for the location of the pELA. (FLINT 1970, MÜLLER 1978, HASTENRATH 1971, ANDREWS et al. 1970, LOCKE 1990, ANDREWS 1975) This method is based on the assumption that the summit and the bergschrund is located above the pELA and that all moraines are deposited below the pELA. Because the mass turnover is greatest at the ELA, cirque glaciers have the highest velocity and thus the greatest erosional potential at the elevation of the ELA. (e.g. LOCKE 1990, CHARLESWORTH 1957, FINSTERWALDER 1952) The erosional potential is additionally increased at or just below the ELA because of the increase of freeze-thaw cycles close to the 0°C summer isotherm. (FLINT 1970)

The determination of the elevation of the cirque-floor is fairly easy when cirques developed their typical arm-chair shape with a steep backwall and a sharp break in the slope.

However, there is no consensus on the cirque that is representative for the pELA of an area if a staircase of cirques can be found for a glacier. CHARLESWORTH (1957) proposes to use the lowest cirque altitude to determine the ELA of the area, whereas MÜLLER (1978) suggests considering a mean of all cirques in an area. MEIERDING (1982) and LOCKE (1990) in disagreement sampled only north-to-eastfacing cirques because they find them to be representative of the regional snowline than cirques.

MEIERDING (1982) also draws attention to the fact that there is generally a lower (climatic) limit in the possible cirque floor altitude but not an upper (climatic) limit. Hence, the more cirques are included in the calculation, the higher the derived pELA. If all cirque floors of a mountain range are sampled and an average is taken, this might more likely reflect different topographical influences than the actual pELA. MEIERDING (1982) therefore suggests sampling solely the lowest north-to-east facing cirques in a region. MEIERDING (1982) and BENN & LEHMKUHL (2000) stress the consideration that the lowest cirque altitude can only reflect a composite pELA for all cirque glaciations that have occurred in an area and hence not be assigned to any particular glacial stage.

MEIERDING (1982) outlines problems compounded when determining cirque-floor altitudes for mid-latitude and fairly low elevation glaciers. The identification of cirques at low elevations is difficult because of the poor morphological formation of the cirques that were not as well developed as in higher elevations where the ice is more erosive and persists longer. The poorly formed cirques have sloping rather than level floors, which hamper the determination of the floor altitudes. MEIERDING (1982) determined consequently cirque-floor altitudes at the location with the lowest down-valley gradient within the cirque.

BENN & LEHMKUHL (2000) summarize studies that showed that cirque floors are strongly controlled by the pre-existing relief, such as Tertiary erosion surfaces and that the cirque basis might have been eroded not primarily during the cirque glaciation, but possibly during more extensive valley glaciations. The median cirque floor elevation might however represent the integrated erosional history of a region across many glaciations. (LOCKE 1990)

The cirque-floor altitude method to determine the ELA is as the MELM-method only a morphological estimation, which does not include glacier mass balance values. The cirque-floors can readily be reconstructed from topographic maps in areas with well developed cirques and might be useful for a quick estimation of regional trends in the ELA. It can also be used to check the soundness of other methods to determine the pELA as shown by LEHMKUHL (1995).

4.6 Glaciation limit approach (Partsch "summit –method")

A method to reconstruct the regional snowline for an area, by averaging the elevation of glaciated and non-glaciated mountains, is the glaciation limit method or glaciation threshold method. It makes use of the fact that glaciers will more likely occur on higher mountains in a particular area than on lower ones. The glaciation limit is defined to be the mean altitude between the summit of the lowest glaciated mountain and the highest neighbouring mountain in a region that is not glaciated. (PARTSCH 1882, ØSTREM 1966, PORTER 1975b & 1977, MEIERDING 1982, CHARLESWORTH 1957, HUMLUM 1986)

The more observations available for a region, the more significant is the resulting regional snowline estimate. The application of this method to Pleistocene glaciation is limited due to the difficulty in reconstructing the extent of the glaciation of each mountain for a particular glacial stage and in assigning the glaciers to surrounding mountains. (HAWKINS 1985, CHARLESWORTH 1957, MÜLLER 1978, MEIERDING 1982) As ØSTREM (1966) points out, it is possible that a mountain summit does not carry a glacier, but exceeds the glaciation limit, because it is unable to hold back the snow due to its relief. This has to be considered when calculating the glaciation limit.

The glaciation-limit method generally exaggerates the height of the regional snowline as the mountain summits are used for the reconstruction. (ØSTREM 1966, CHARLESWORTH 1957) The method is best adapted for highlands that carried a large number of small, local glaciers, but were never completely covered with glacier ice. (FLINT 1970)

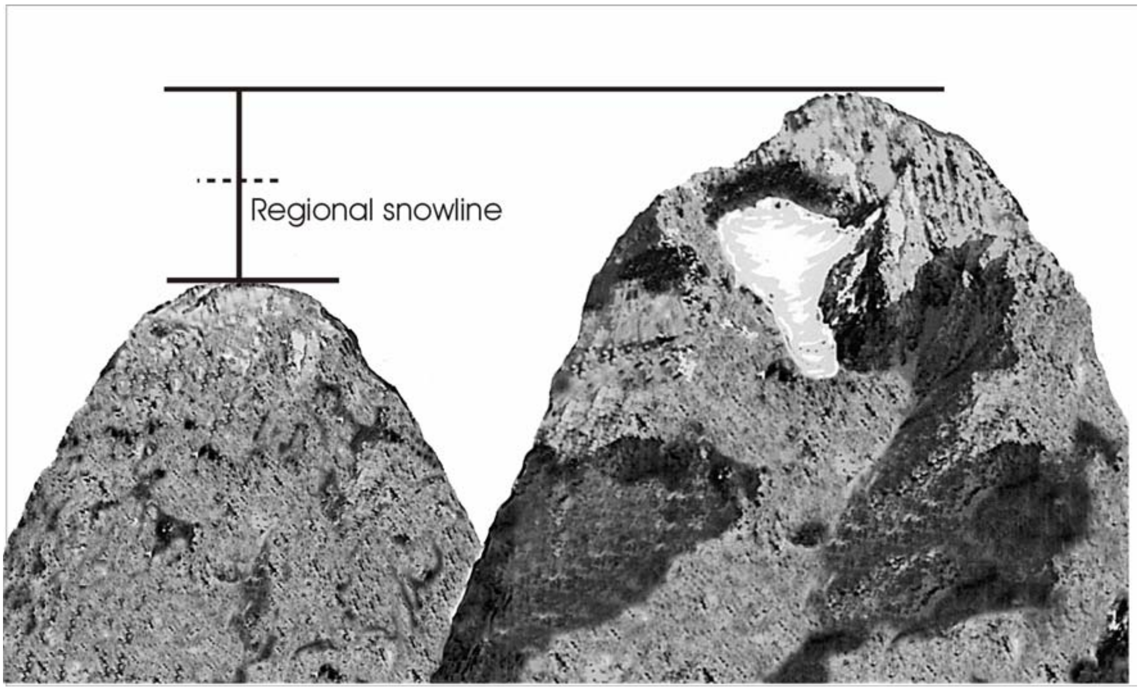


Fig. 12 Principle of the glaciation limit approach

4.7 Contour-line method (Hess-method)

This method is based on the glaciological observation that the accumulation zone of a glacier has a concave surface and the contours in the ablation zone show a convex pattern. This approach was first used by HESS (1904) and later by ØSTREM (1966). In the accumulation zone the flow lines generally emerge, in the ablation zone they immerge. The contours in the accumulation zone bend up the surrounding cliffwalls, whereas the contours in the convex glacier tongue bend sharply when they meet the valley sides. (e.g. HESS 1094, CHARLESWORTH 1957, UNESCO/IASH 1970) The ELA is assumed to be located at the transition from the concave to the convex part of the glacier. Complex topography of the glacier and disturbance of the ideal shape of the glacier surface limit the possible field of application for the method. (MÜLLER 1978, ANDREWS 1975) It can only be applied to glaciers that do not show steep parts around the ELA as pointed out by WISSMANN (1959)

The Hess-contour line method cannot be employed to reconstruct pELA of former glaciers, but it can serve to establish present-day ELAs for calibration. (MÜLLER 1978, BURBANK & CHENG 1991, CHARLESWORTH 1957, WILLIAMS 1983)

5 Parameterisation of the Kleiner Arbersee glacier

In order to apply any of the above described methods of ELA-modelling to the Kleinen Arbersee glacier, its glacial limits have to be reconstructed. To successfully model the pELAs, the outline of the glacier, the headwall or bergschrund, the cirque floor, the maximum elevation of lateral moraines and the surrounding mountain ridges as well as the toe of the glacier have to be parameterized.

As this study aims to show how the reconstruction of pELAs can be conducted using a GIS-Software, only one glacial stage was studied. The glacial stadium that was chosen for this study is the M2a glacial stage after BUCHER (1999) (see Fig. 21). This is the maximum position with the best developed moraine morphology for lateral as well as for the terminal moraines. The blurry stadium that can be found only a little further down-valley is not well developed or conserved, which might be an indicator that it was a short surge from the maximum position (comparable to the “supermaximum” for the alpine Wuermian glaciation after JERZ 1983). Taking the maximum position for the reconstruction allows to interpret the highest elevation of glacial erosion to belong to this stadium. Any pre-Wuermian glaciation can be ruled as the origin of these features as the glacial polish is well preserved and not significantly weathered.

5.1 Data basis

The study area is well mapped and tachymetrically surveyed. For this study the 1:5,000 map (Flurkarte) and the map published by RAAB (1999) of the tachymetric survey (see fig. 13) was used. The maps were scanned and later georeferenced by writing a tfw-file (see fig. 14), before being transferred into a GIS-project. For this study the ESRI GIS ArcView© 3.2 was utilized as well as the extensions Spatial Analyst© and 3D-Analyst©.

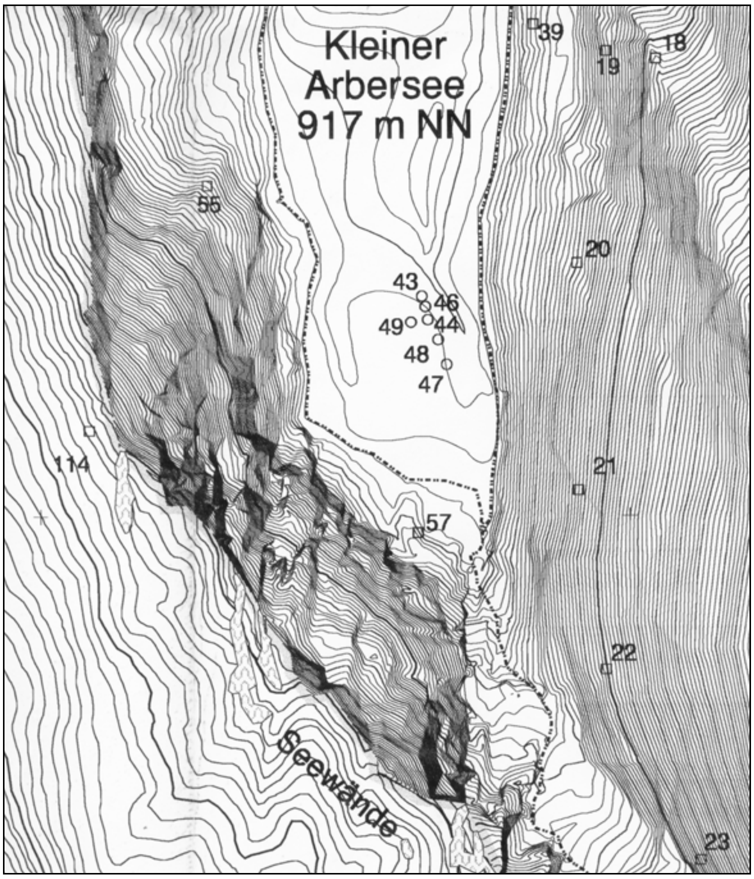


Fig. 13 Clipping of the map published by RAAB (1999) with the section of the 1-m contour lines that were derived in a tachymetric survey and the contour lines from the 1:5,000 map

The 1:5,000 map has an equidistance of 5 m, the map from RAAB (1999) shows contour lines from the tachymetric survey (< 1m accuracy) of 1 m.

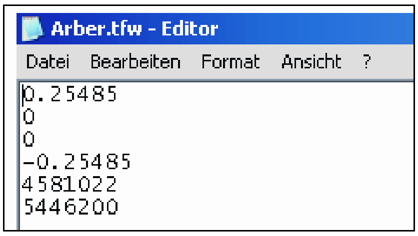


Fig. 14 Tfw-file to georeference the map of RAAB (1999)

The contour lines were digitized on the screen directly in ArcView© and the elevation data was stored in a table. For the glaciated area, the tachymetrically derived contour lines from RAAB (1999) were digitized, for the area outside the outermost moraines, the contour map of the Bavarian Carthographic Survey was digitized with an equidistance of 5 m.

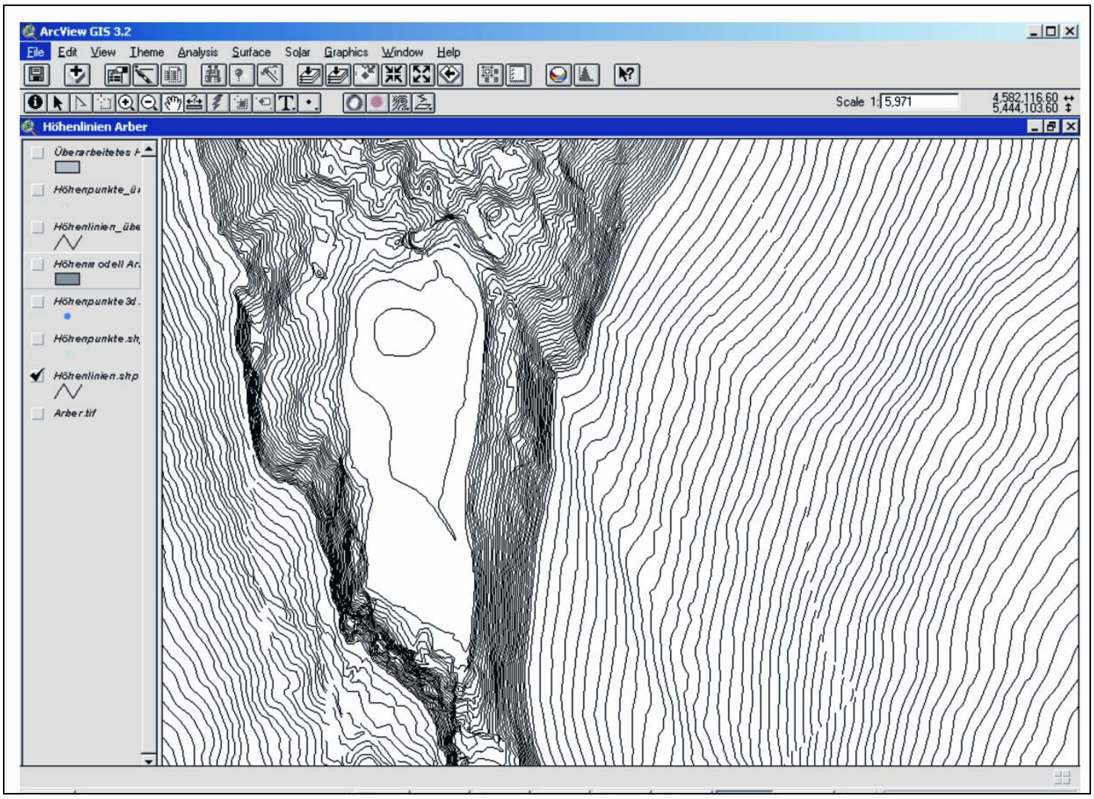


Fig. 15 Digitized contour lines

The map from the inventory of lakes in Bavaria BLLW (1983/1987) shows the bathymetry of the Kleiner Arbersee. The bathymetric contours were digitized for the reconstruction of the glacier bed as the morphometry of the glaciated area at the base of the lake is of particular interest.

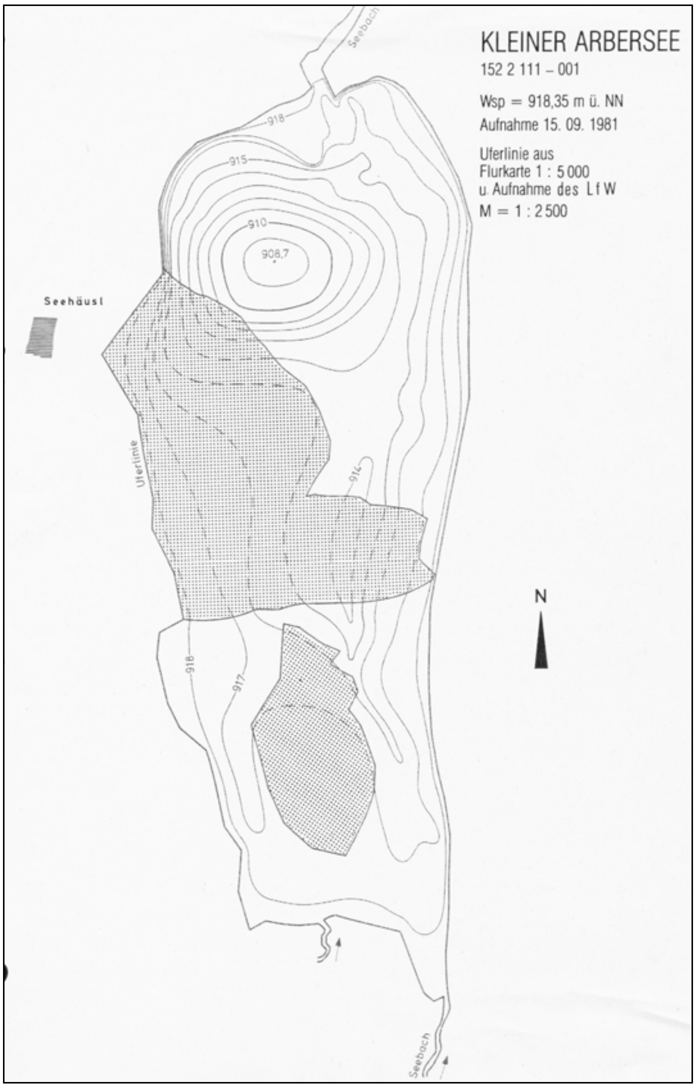


Fig. 16 Bathymetry of the Kleiner Arbersee from BLLW (1983/1987)

As pointed out by RAAB (1999), the accuracy of the 1:5,000 map is not as reliable as the tachymetrically derived elevation data, hence at the point, where the contour lines from the different scales meet, there is a horizontal displacement. This artefact can be seen clearly in the Digital Elevation Model (DEM) (see fig. 17). To remove this effect and to ensure a smooth terrain surface, the contour lines were adjusted manually (see fig. 18).

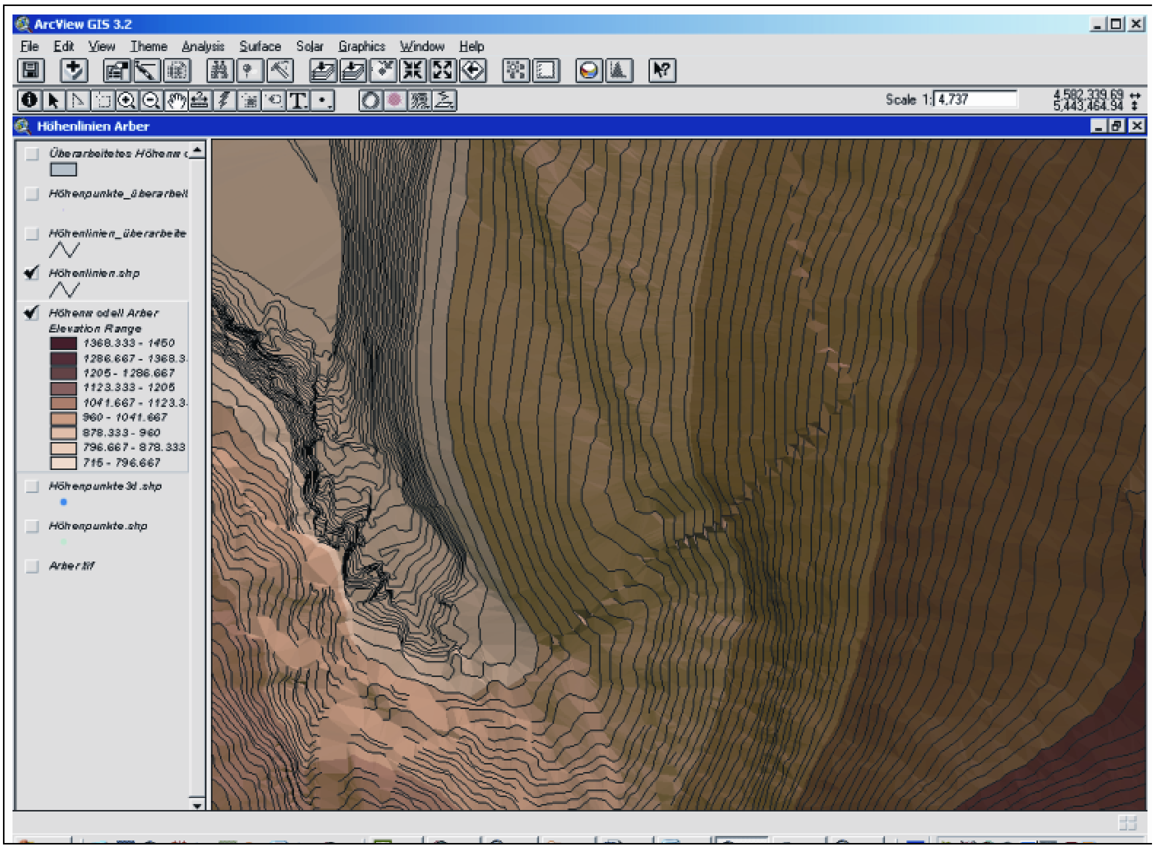


Fig. 17 Discontinuity in the contour lines due to the deviating accuracy of the maps

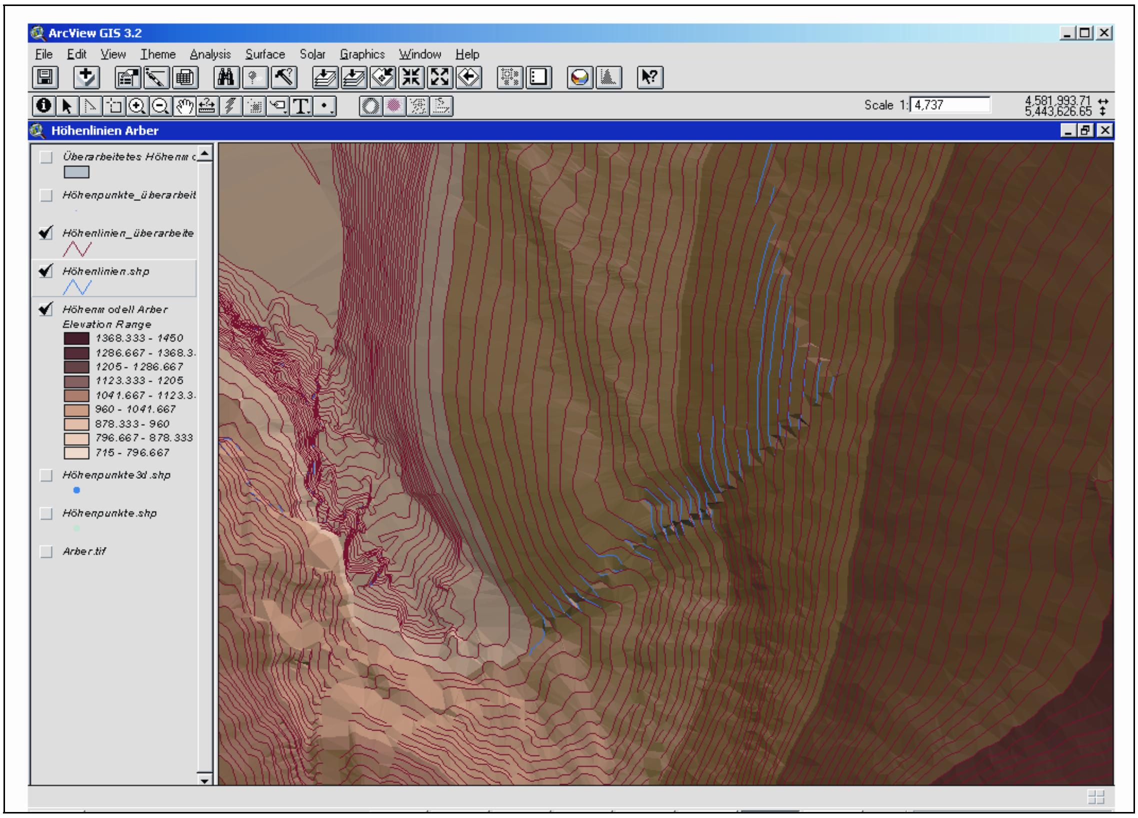


Fig. 18 Correction of the run of the contour lines (blue is the original contour line, red is the corrected contour line; underlying DGM with artefact)

The contour lines were then transformed into points of equal elevation using a free ArcScript written by H. Appel (view.convertToMultipoint[3]).

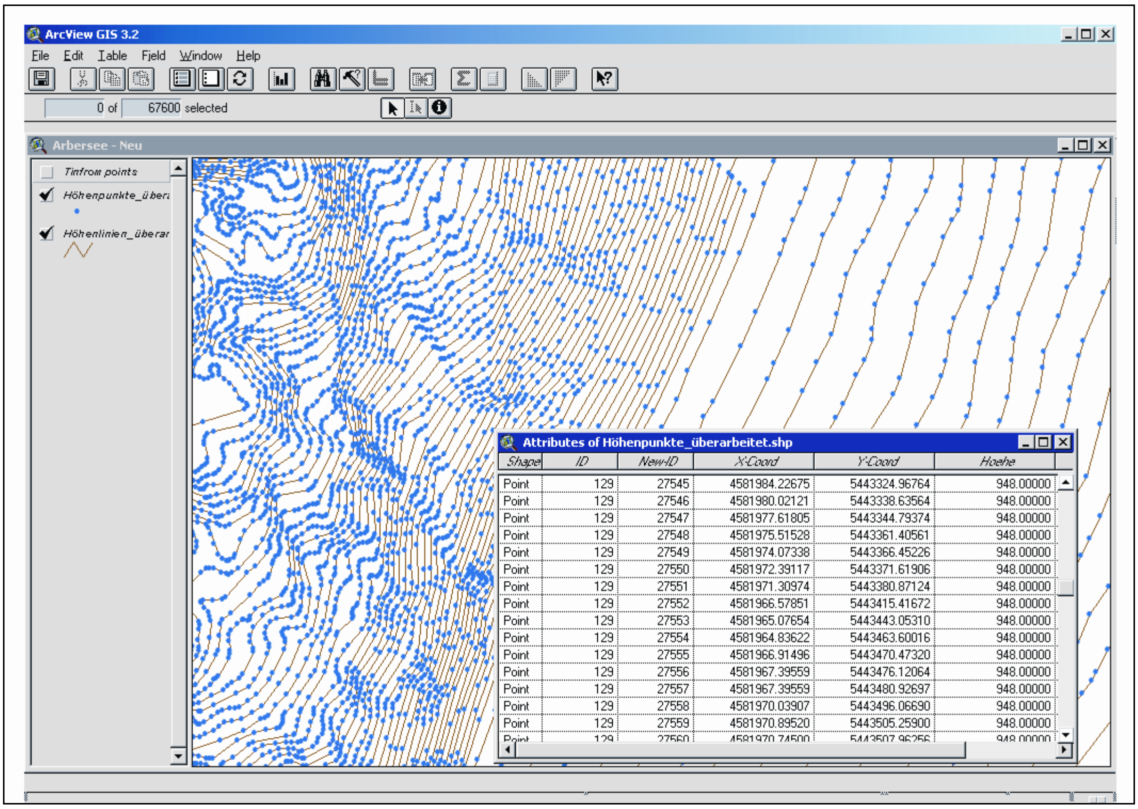


Fig. 19 Contour lines with corresponding points of equal elevation and the data table for the elevation points

From the data points a TIN was calculated and subsequently a GRID could be interpolated. The irregular spacing of the elevation points (from the different contour line equidistance) rendered it impossible to create a single GRID from the data set. The triangulation for the TIN was done by assigning the outer boundary of the elevation points as a hard breakline. The lake side was not defined as a breakline as the bathymetric data from the map of the lake inventory was used for the lakefloor elevation and a smooth topographic surface that reflects the former glacier bed was desirable. To get a better view of the topography, the legend was edited so that 50 classes were created and displayed in a brown to beige elevation range, with beige being the highest elevation (Grosser Arber summit).

[3] http://arcscripts.esri.com/details.asp?dbid=11034 (check for availability on 05.07.04)

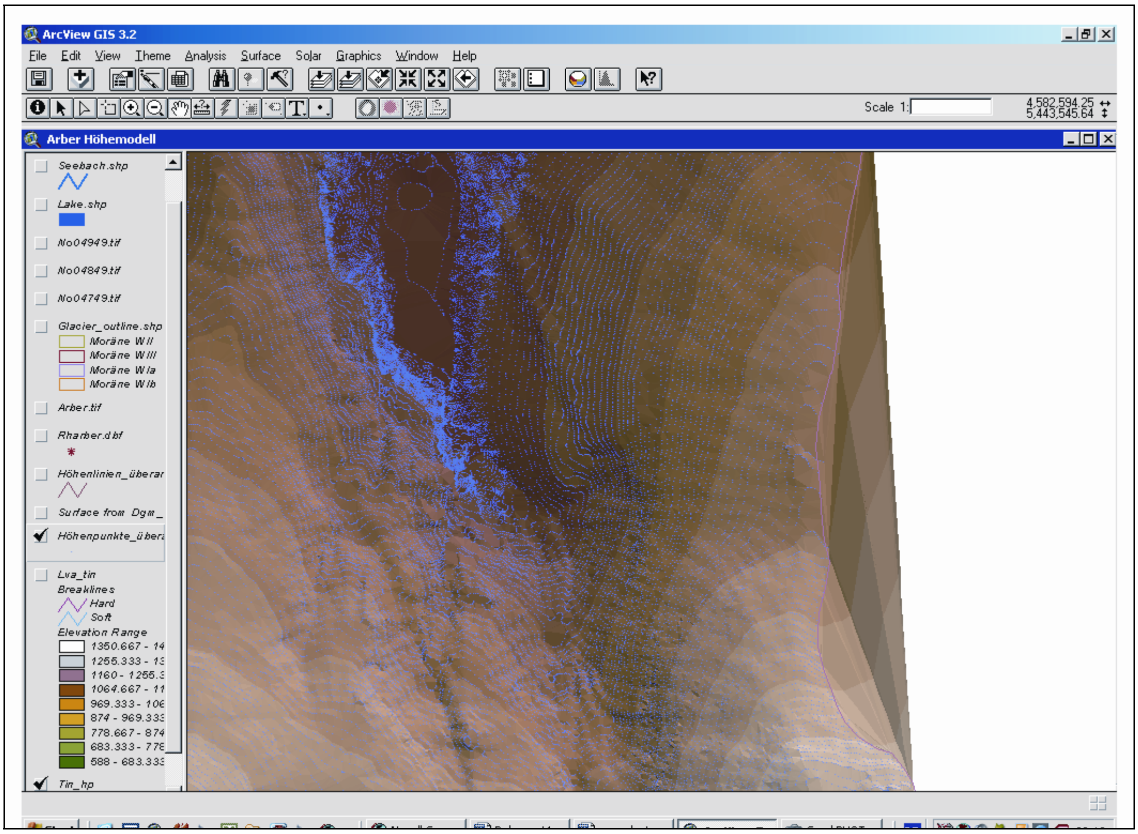


Fig. 20 TIN with the hard breakline and the elevation points used to create the TIN

5.2 Reconstruction of outline and shape of the Kleiner Arbersee glacier

For the ELA-modelling of modern glaciers the outline and the shape of the glacier can be digitized from aerial photographs or even surveyed in the field with GPS. But in this study no modern glacier exists and hence the shape of the former glacier has to be inferred from geomorphological evidences in the field or as visible in the DEM, as well as from general knowledge of glacier mechanics and glaciology.

The reconstruction of the Kleiner Arbersee glacier is conducted by interpreting erosional and depositional features in the landscape. Due to the dense forest cover and the periglacial overprinting of the landscape, the former outline of the glacier is not always unequivocal.

The area of the Kleiner Arbersee is very well studied and mapped in terms of the past glacial environment as described in chapter 2. The well documented study from RAAB (1999) offers a sketch of the lateral and terminal moraines, as does the study from BUCHER (1999). (see fig. 21)

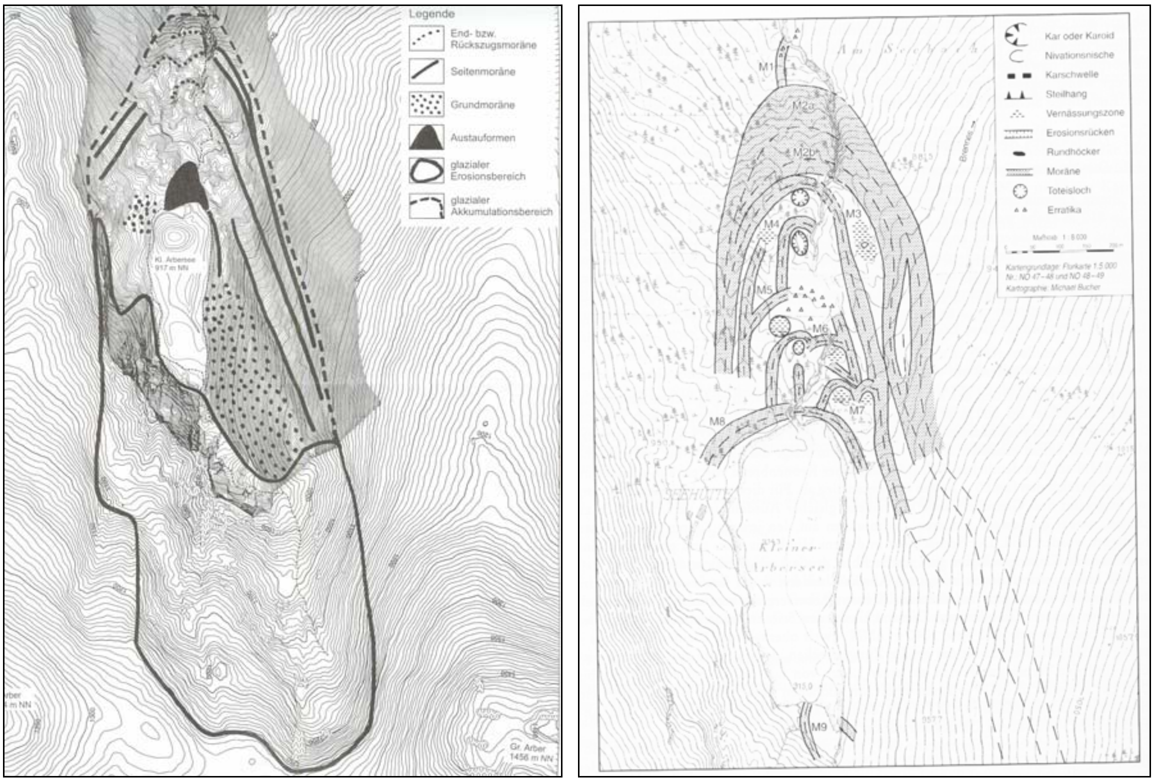


Fig. 21 Sketch of moraine location according to RAAB (1999), left, and BUCHER (1999), right

The DEM of the catchment basin also shows the moraine morphology in the area. Different glacial stages can be distinguished without undue effort. Fieldwork has been done to verify the information from the literature and from the DEM. In combining the different sources of information, a reconstruction of the glacier could be conducted. Different approaches were applied for different zones of the glacier:

Cirque: The outline of the cirque, where the head of the valley cuts into the saddle between Kleiner and Grosser Arber mountain, is apparent in the DEM. The oversteepened slopes of the cirque headwall can well be verified in the field. Problematic for the reconstruction is the transition from the clearly eroded cirque headwalls to the moraine deposits at the slopes that eventually form morphologically visible lateral moraines. This area is overprinted by (periglacial) slope deposits.

Glacier outline east valley side: The outline of the glacier is unequivocally marked by distinct moraines on the east side of the catchment. The moraine ramparts are up to 15 m high and covered with erratic boulders. The moraine morphology can be seen in the DEM, as well as verified in the field. The upper end of the morphologically-visible moraines is blurry.

Glacier outline west valley side: The west side of the valley is not marked by moraines as it consists mainly of glacially eroded bedrock. The elevation of eroding glacier ice is determined by glacial erosion marks. The highest elevation of glacial erosion in different locations on the west valley side were surveyed with a GPS and marked on a map. These marks allow only a very careful interpretation of the glacier outline on the western valley side. Highest glacial erosion allows only the reconstruction of the maximum glacial stage as erosion marks from recessional glacial stages cannot be distinguished.

Glacier tongue: The tongue of the glacier is outlined by moraine deposits. The rampart of the moraine arc at the location of the glacier tongue can be recognized in the field as well as in the DEM.

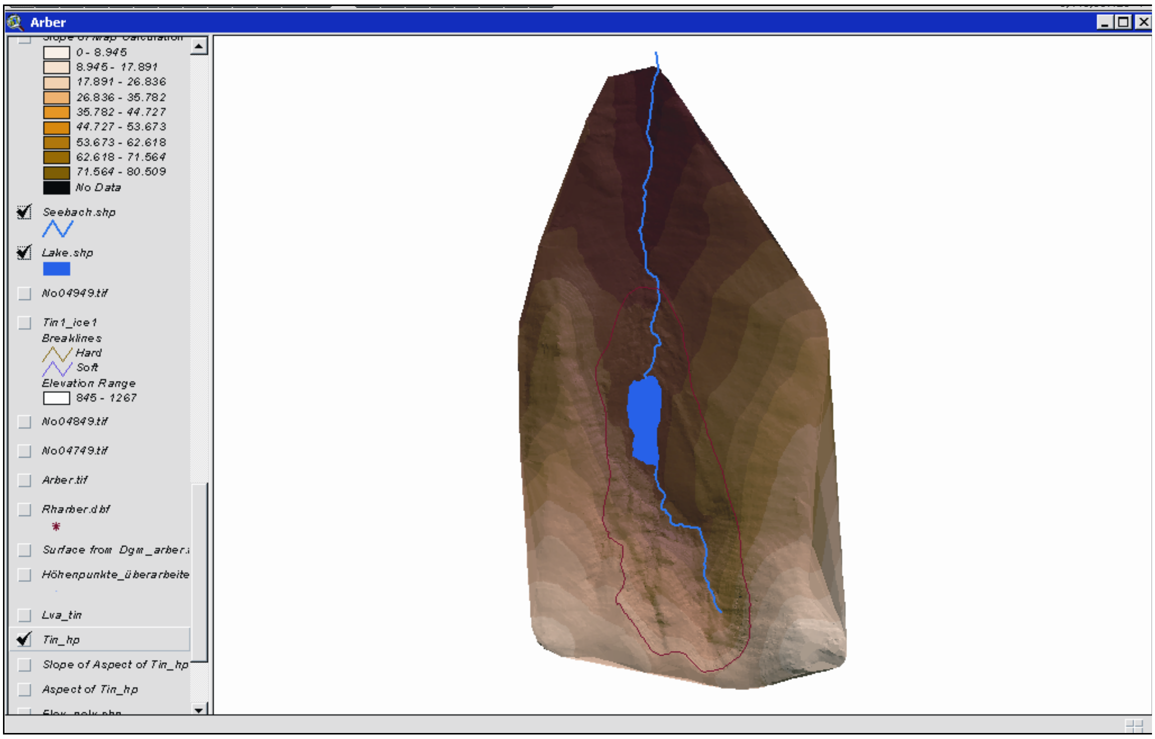


Fig. 22 Outline of the Kleiner Arbersee glacier (red line) for the stadium M2a with underlying DEM, the Kleiner Arbersee Lake and the Seebach River

The outline of the glacial stage M2a for the Kleiner Arbersee glacier was reliably reconstructed for the eastern side, the glacier tongue and the headwall of the cirque. Somewhat subjective is the reconstruction on the western valley side and for the transition zone between the highest morphologically visible moraines and the cirque. To conduct all different ELA-reconstructions the ice margin has to be reconstructed as a closed polygon even for questionable zones.

The reconstruction of the former glacier surface is challenging. The ice-surface topography is reconstructed on the basis of the geometry of the moraine remnants, the limits of the glacial erosion and the cirque headwalls. For simplicity the elevation of the lateral and terminal moraine crests are taken as the ice margin along the lower boundary of the glacier, as well as the cirque headwall for the upper part of the glacier. This simplification is done knowing that the present-day elevation of the moraine rampart is less than at the time of deposition due to solifluction processes and denudation. The inclination of the glacier tongue is hence interpolated between the corresponding moraine elevations on either side of the glacier. This can be conducted fairly easily as the lateral glacier limits are at maximum 830 m apart.

The location of the ice-surface contours is not straightforward as already pointed out by HAWKINS (1985). Generally speaking, the longitudinal profile of a glacier shows a concave shape up-glacier of the ELA and a convex shape below the ELA. This is due to the fact that the internal flow lines point into the glacier in the accumulation zone, are parallel to the ice-surface around the ELA and point out of the glacier in the ablation zone, transporting englacially transported debris to the surface.

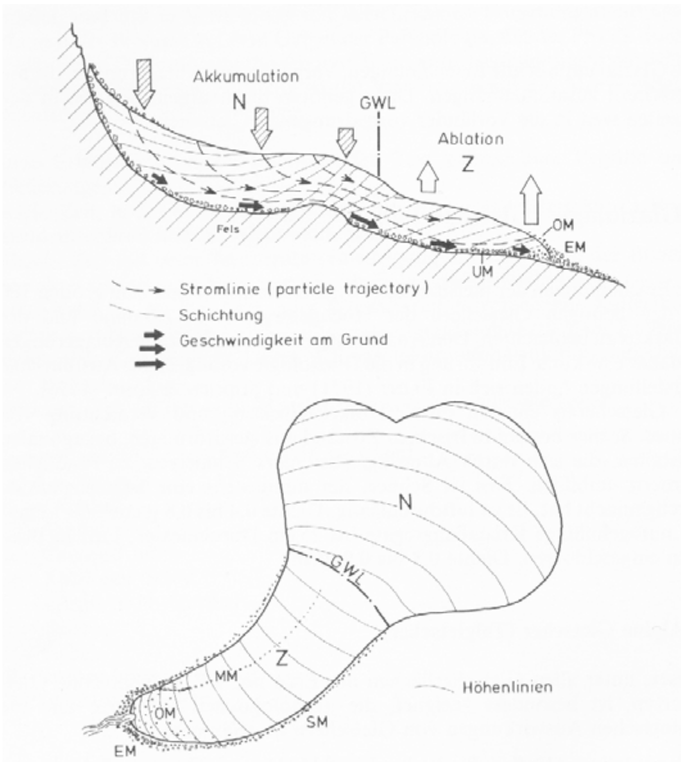


Fig. 23 Schematic longitudinal profile of a glacier (from SCHREINER 1997)

Contours are accordingly normal to the valley walls in the vicinity of the ELA, with a down-glacier curvature in the ablation zone and a concave curvature in the accumulation zone. This leads to a shallow depression in the topography of the upper glacier and an arching of the glacier surface of the zone of the glacier tongue. This generalized concept of the ice-surface topography in the longitudinal profile was first pointed out by HESS (1904) and is still used to approximate the ELA for modern glaciers (see Chapter 4.7). Contours in the ablation zone usually show a rapid change in curvature from the bedrock to the glacier ice due to high ablation at the ice-margin, contours in the accumulation zone show simply a smooth change in direction from the bedrock to the ice-surface contour. (see fig. 24)

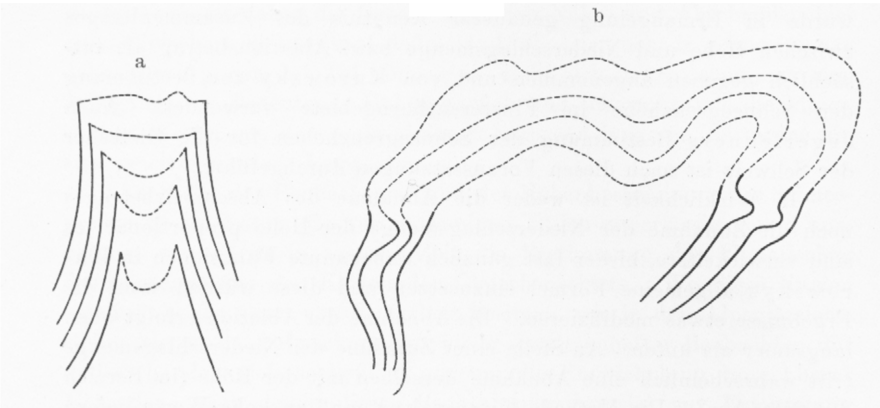


Fig. 24 Shape of contour lines in the ablation zone (left) and the accumulation zone (right), according to HESS (1904)

This can be verified by looking at ice-contour maps of glaciers. Figure 25 shows an example of the Vernagtferner glacier, Austria from the time of a positive mass balance. The above described shape of the contours can well be recognized on the map. The map was used as a guide to apply the appropriate amount of surface concavity to the reconstructed ice-surface.

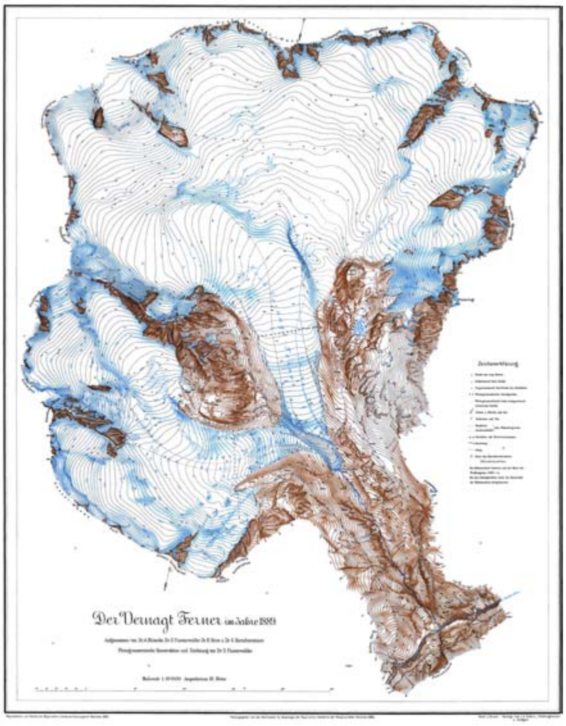


Fig. 25 Map of Vernagtferner 1889 from Finsterwalder, published by the Kommission für Glaziologie, München[4]

Along the latitudinal axis the contour lines are bulging, with the maximum thickness of the glacier-ice in the middle of the glacier. The reason for the thinning of the ice thickness towards the valley sides is the higher ablation in this zone due to higher friction and shading from the cold ice-preserving catabatic winds that flow in the middle of the glacier.

For simplification the depth contour of the modern catchment relief is also the line of maximum thickness of the former glacier. The zone of maximum ice-thickness is assumed to be the zone of maximum glacial scouring. All ice-surface contours were reconstructed normal to the depth contour. The depth contour was reconstructed by applying the “Steepest-Path-tool” of the 3D Analyst© and following the flow path of the Seebach river.

[4] http://www.lrz-muenchen.de/~a2901ad/webserver/webdata/vernagt/vernagt.html (checked for avail-ability 07.07.04)

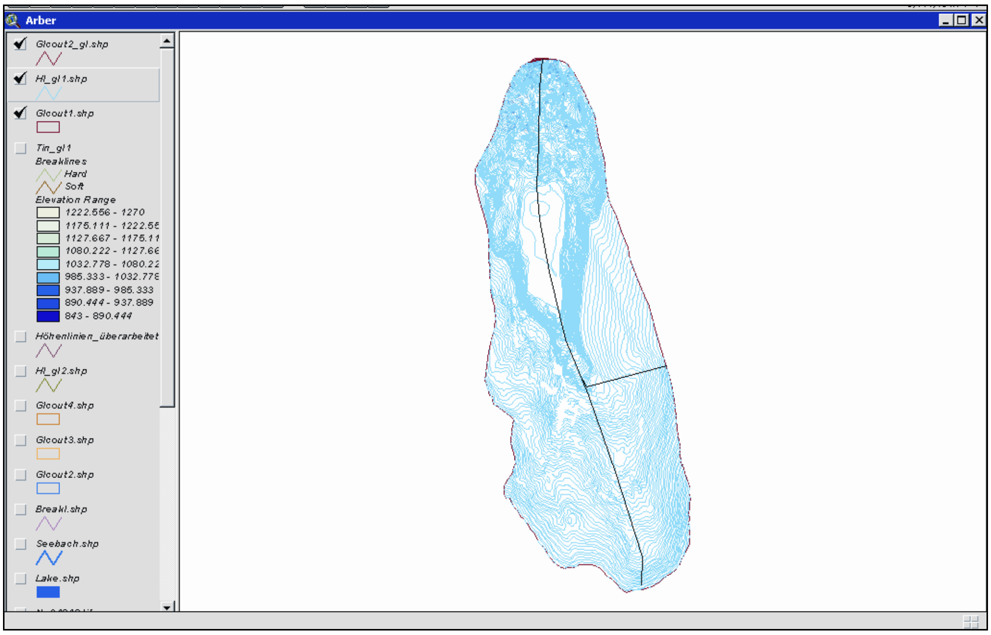


Fig. 26 Depth contour of the catchment and flow path of the Seebach river

For the reconstruction of the ice-surface contours the ELA was approximated by the maximum elevation of lateral moraines, the cirque floor altitudes and references in the literature (ERGENZINGER 1967, HAUNER 1980). It is approximated at an elevation of 1055 m asl. (see fig. 27)

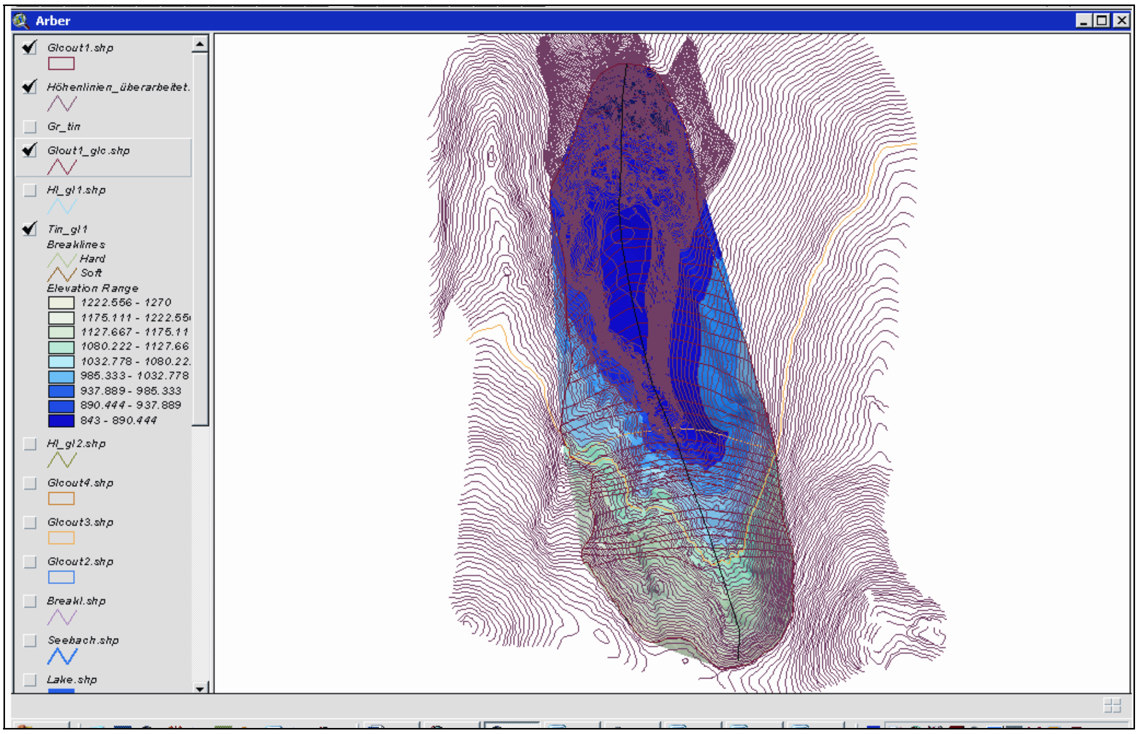
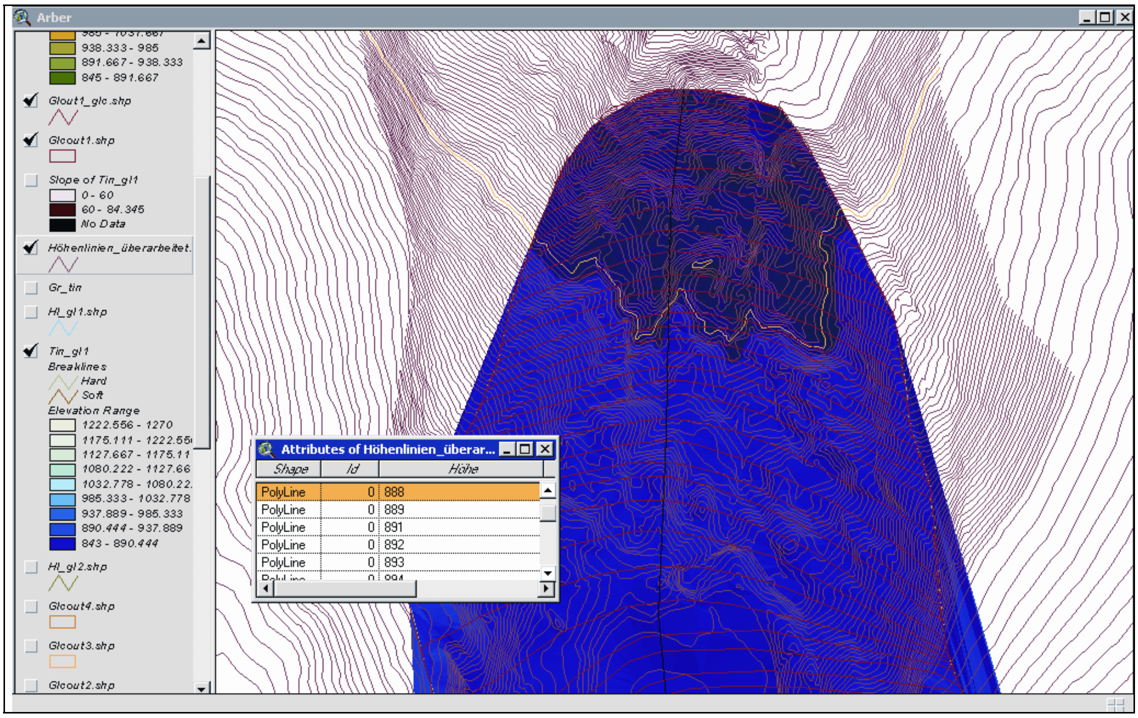


Fig. 27 Initial ELA-approximation for the determination of the ice-surface contours. The yellow high-lighted, curved line is the relief contour of 1055 m and the fairly straight line is a first 1055 m ice-surface contour.

The curvature of the ice-surface contours was drawn starting with a fairly straight line of 1060 m at the ELA and progressively more convexly curved contours below the ELA and more concavely shaped contours above the ELA, taking the cirque morphology into account.

The glacier-surface can simply be an approximation of the glacier as the curvature of the contours is a function of the state of mass balance of a glacier. A glacier with a positive mass balance that is surging has different ice-surface topography than a glacier that melts down. The tongue of a surging glacier is more vaulted than for a melting glacier.

The reconstruction of the glacier M2a in this study is based on the assumption that the glacier is in steady-state. The ice-surface contours were drawn by connecting points of equal elevation on the opposing lateral moraines or the elevation of glacial erosion marks. The maximum curvature of the contour line was drawn at the depth contour. This results in a maximum thickness of the glacier along this line for the ablation zone, and a depression along this line for the accumulation zone. Contours for the glacier tongue were drawn at 2.5 to 5 m contours equidistance, in the cirque they were drawn with an equidistance of 5 to 10 m. The ice-margin was reconstructed fairly steep by drawing a 5 m rampant glacier side and a smoother curvature for the actual ice-surface.



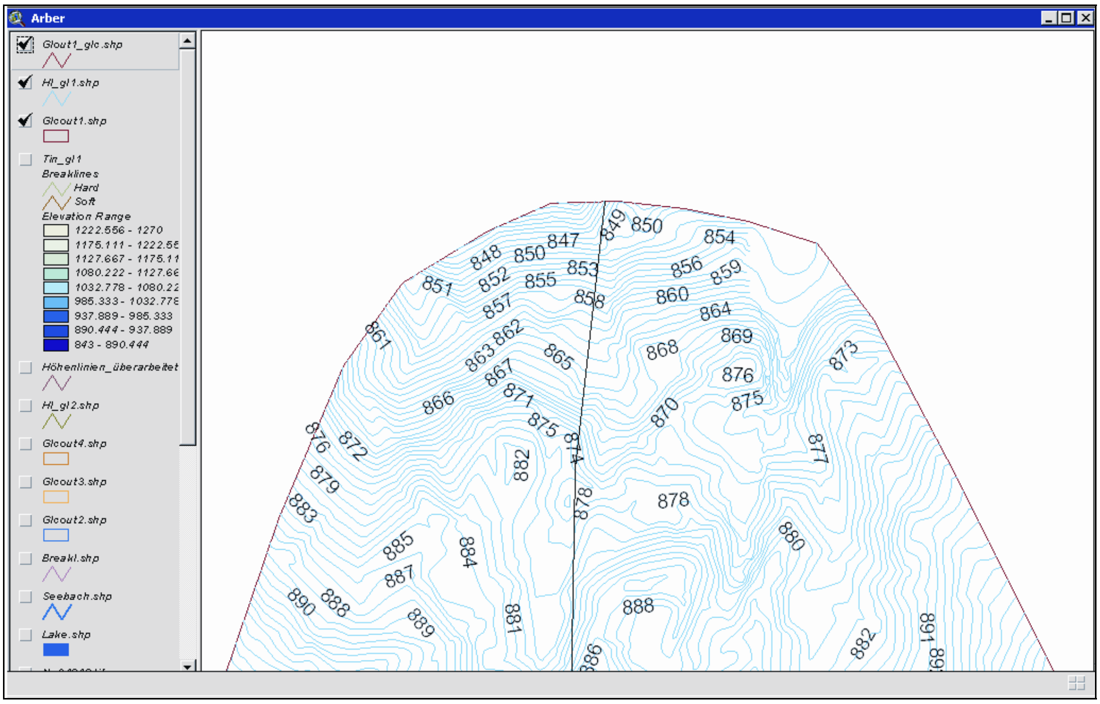


Fig. 28 Reconstruction of ice-surface contours done by highlighting a relief contour line and connecting the control points with a maximum curvature at the depth contour for the glacier tongue.

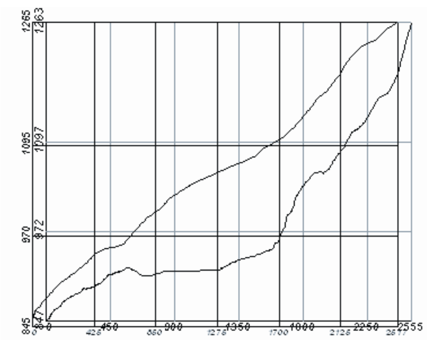


Fig. 29 Longitudinal profile of modern relief, which is an approximation of the former glacier bed and the reconstructed ice-surface of the Kleinen Arbersee glacier for stadium M2a

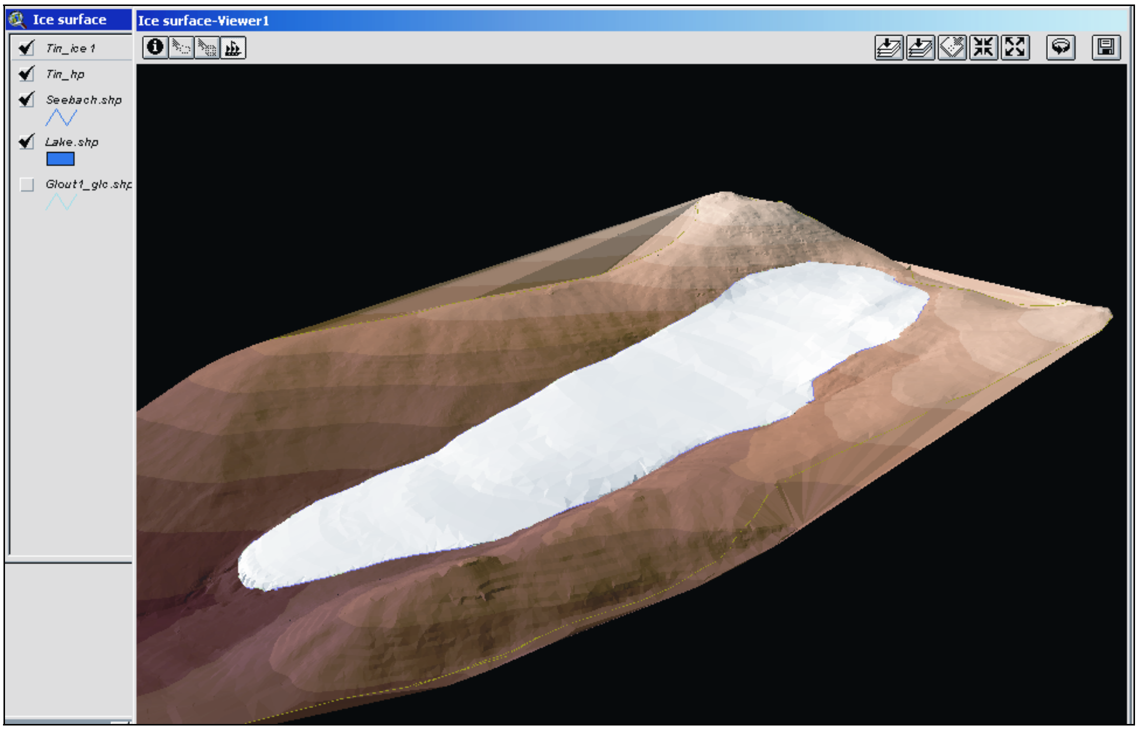


Fig. 30 Reconstructed ice-surface for the glacier at the stage M2a. The glacier tongue is underlain by the moraine rampart.

This reconstruction can simply be an approximation as there are so many unknowns, but the restricted dimension of the Kleiner Arbersee glacier and the closely spaced control points on either side of the glacier permit less subjectivity in contouring.

5.3 Reconstruction of the cirque floor elevation for the Kleiner Arbersee glacier

The Kleiner Arbersee glacier left behind a staircase of cirque floors as visible in the DEM as well as even in photographs. RATHSBURG (1928) was the first to point out the staircase of cirque floors.

Fig. 31 Picture of cirque floor from the Kleiner Arbersee glacier from RATHSBURG (1928)

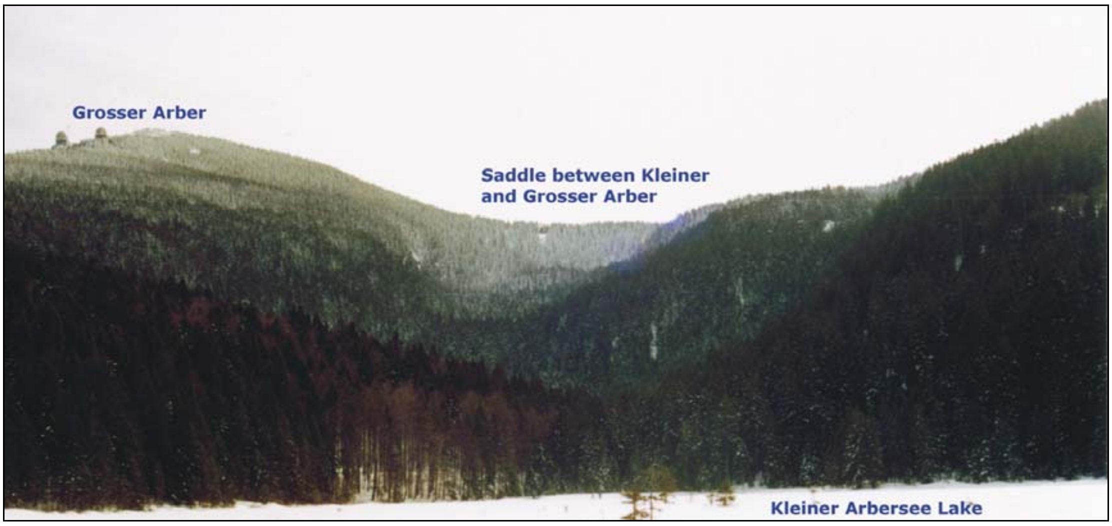


Fig. 32 Overview picture of the catchment, where the cirque floors can be recognized

The late-glacial Kleiner Arbersee glacier supposedly consisted of four different cirque glaciers, each leaving behind a staircase of cirque floors. BUCHER (1999) plotted the longitudinal profiles of the four different lobes; the figure displays the arm-chair appearance of the different cirque floors. As can be seen in the figure 33 and as pointed out in the literature (e.g. MANSKE 1989, ERGENZINGER 1967) all of the four lobes showed 2-3 cirque floors with the elevations around 1050 m asl and 1100 m asl (and 1250 m asl). The cited altitudes were verified in the field by GPS and altimeter measurements.

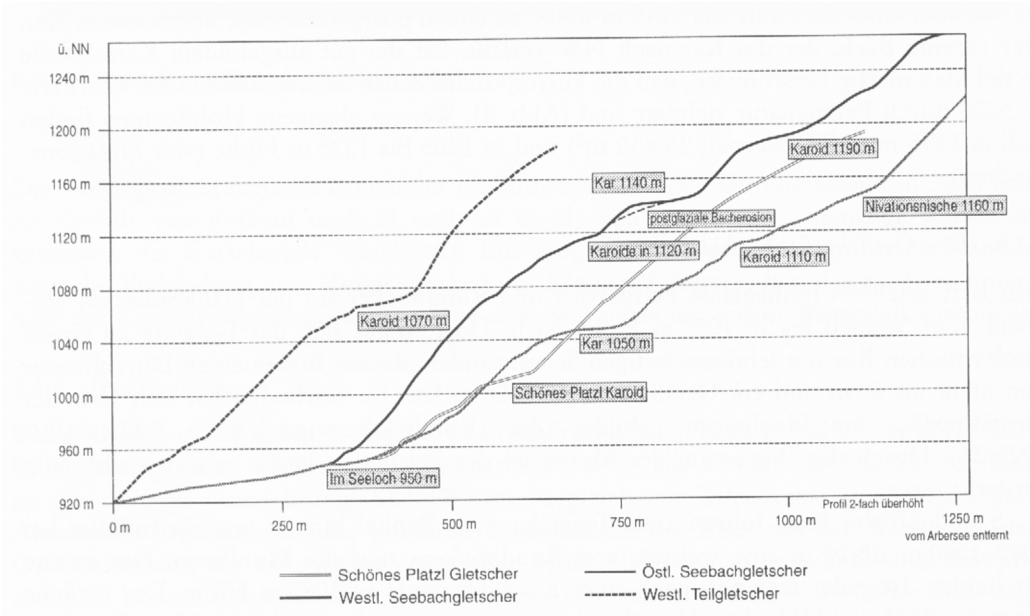


Fig. 33 Depth contours of the modern relief of the four different cirques of the Kleiner Arbersee glacier that show the altitude of the cirques (Kar) or cirque-like depressions (Karoid) in the longitudinal profile (from BUCHER 1999).

The lowest cirque is found at an altitude of 950 m asl. The best developed cirques with a flat cirque floor and an arm-chair like shape is located at 1050 and at 1100 m asl, this can as well be seen from the contour lines. Earlier studies (e.g. MANSKE 1989, BUCHER 1999, ERGENZINGER 1967) agree on the assumption that the development of the staircase of cirques and the four different lobes was late-glacial erosion. For the high-glacial situation, when the M2a moraine was deposited, they assume one large cirque.

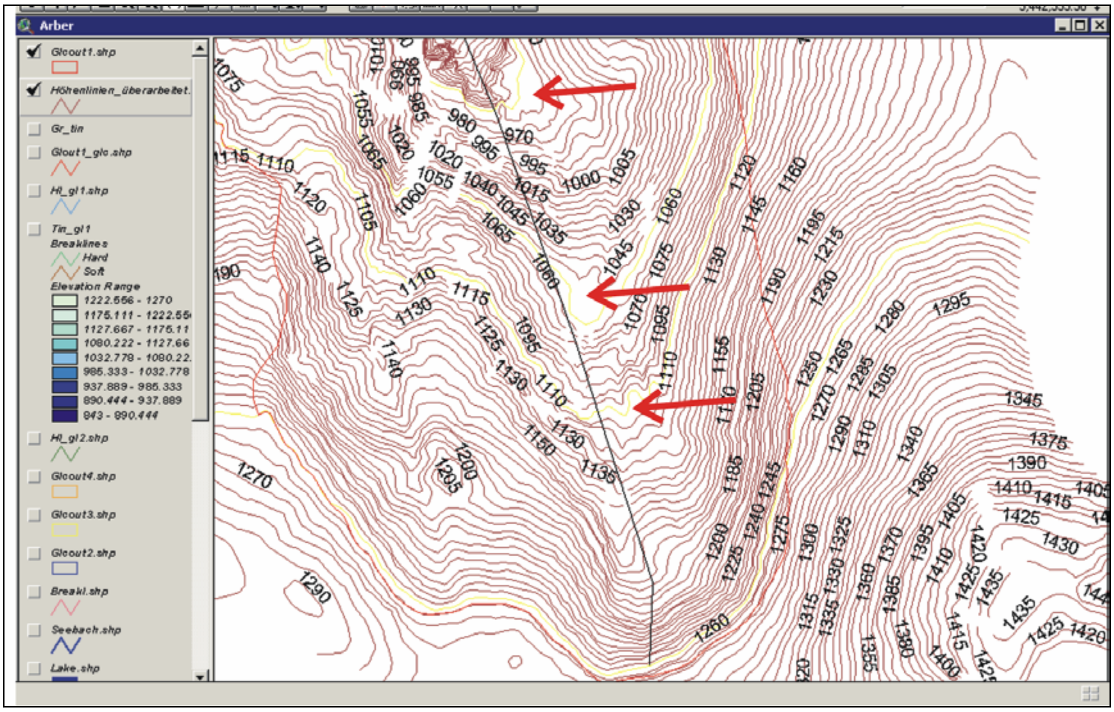


Fig. 34 Contour lines of the upper catchment of the Kleiner Arbersee glacier with high-lighted contours, where the cirque-floors were located (see text).

5.4 Reconstruction of the upper glacial limit of the Kleiner Arbersee glacier

To apply the THAR-method, the elevation of the bergschrund of the Kleiner Arbersee glacier has to be determined. As pointed out earlier (chapter 4.3) the bergschrund is in some cases not easy to locate. The threshold of a 60°-slope is not reached in the catchment (see fig. 35).

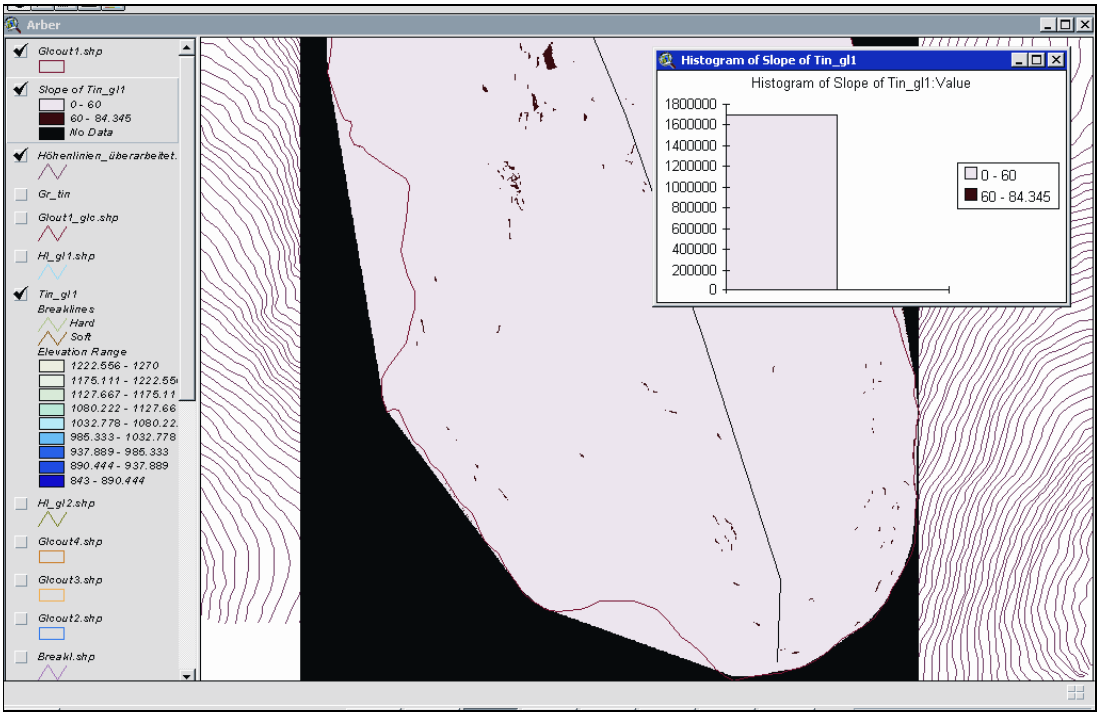


Fig. 35 Slope grid of the catchment classified in slope < 60° and slope > 60°.

Instead of the bergschrund, it was proposed to use the highest glacier limit to apply the THAR-method of the ELA reconstruction. In this case it seems more appropriate to use the glacier limit instead of the bergschrund. The upper glacier limit for the Kleiner Arbersee glacier is supposedly at 1270 m asl for the stadium M2a.

5.5 Reconstruction of the maximum elevation of lateral moraines of the Kleiner Arbersee glacier

The highest elevation of the lateral moraines can only be determined on the east side of the former glacier. As pointed out earlier, there are no morphologically visible moraine deposits on the west side of the glacier.

The highest elevation of lateral moraines for the glacier stadium M2a is blurry. The morphologically clearly visible moraine rampart of the lateral moraine (stadium M2a) fades out at an elevation of around 1055 m asl. This can be inferred from the slope of the contours as well as verified in the field. This elevation is in accordance to BUCHER (1999) and RAAB & VÖLKEL (2003). To achieve the altitude of the EL across the glacier from the maximum elevation of the lateral moraines on the east side of the glacier across the ice-surface, it is assumed that the ELA is normal to the depth contour (maximum thickness of glacier, see above). The drawback is that the lateral moraines can only be determined for one side of the glacier.

5.6 Determination of the elevation of the surrounding mountain ridges, the highest summit in the catchment and the lowermost glacier limit

The highest summit of the catchment is the Grosser Arber mountain with an elevation of 1456 m asl. Part of the mountain ridge that surrounds the cirque is the Kleiner Arber mountain and the saddle between Kleiner and Grosser Arber.

Grosser Arber summit	1456 m asl
Kleiner Arber summit	1384 m asl
Saddle between Kleiner and Grosser Arber	1270 m asl

Tab. 2 Summit elevation of mountains that surround the catchment of the Kleiner Arbersee glacier

One of the modified-Höfer approaches utilizes the median elevation of the surrounding mountain ridges. To determine this elevation the grid of the catchment was reclassified into three classes, (1) being the elevation below the former glacier surface (<1270 m asl), (2) being the elevation above the former glacier surface (>1270 m asl) and (3) no data class. A polygon was digitized that outlines the area above the former glacier surface, only with cells that drain into the catchment. From the Spatial Analyst© menu the option of "summarize by zones" was chosen and the statistics of the elevation values of the grid cells were calculated (see fig. 36).

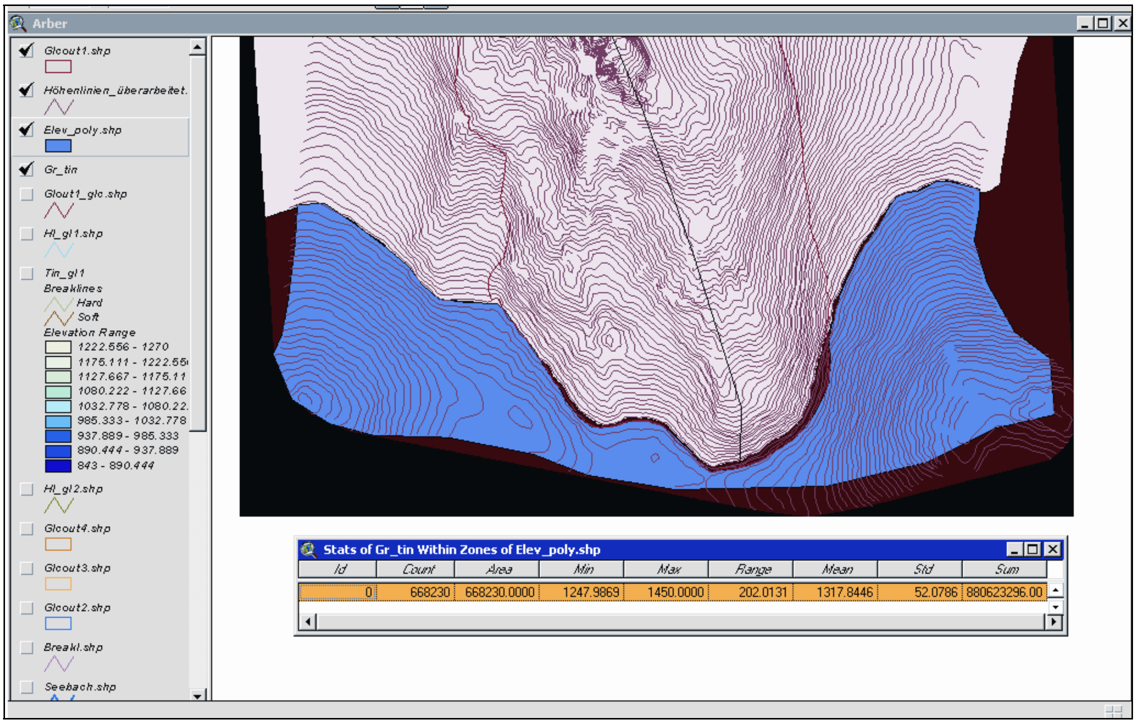


Fig. 36 Statistics for grid cell values of the catchment area that surmount the former glacier surface

Accordingly, the median elevation of the mountain ridge surrounding the cirques of the Kleiner Arbersee glacier is 1318 m asl. If the elevations of the ridges in the catchment (see tab. 2) are simply used for calculation, a median elevation of 1370 m asl results due to the overrating of the summit peaks.

The lower-most glacier limit for the stadium M2a can be located at 845 m asl.

6 ELA modelling for the glacial stage M2a

In this study six different methods and their modifications of estimating pELAs were applied to the glacial stage M2a of the Kleinen Arbersee glacier. All methods require some form of glacier reconstruction or topography interpretation. The realization of the ELA-modelling by utilizing a GIS-software is presented here. The algorithms or technical approaches that were used are described.

6.1 AAR-methods and modifications

The ELA for the Kleiner Arbersee glacier determined using the AAR-method and its modifications (Kurowski-method, Brückner-method) was conducted by planimetry of the reconstructed glacier surface of the Kleiner Arbersee glacier. The ice-surface contours were interpolated by triangulation into a TIN with the outline of the glacier as a hard breakline. The TIN was subsequently converted into a Grid-file. The floating point grid file was transformed to an integer grid file using the [aGrid].int request. The int-request truncates the floating point values at the decimal point. No data was lost as the elevation values are without decimal places. In the grid table the number of grid cells for each elevation value of the glacier was summed. The calculation of the area could be done by calculating with the number of grid cell, because all cells are of the same size. A total of the number of cells for the ice-surface was calculated and amounts to 1,591,096. The table with the elevation value and the respective grid cell counts was exported as a dbf-file and imported into MS-Excel©. In Excel© the elevation values were sorted descending and the number of grid cells was summed. The altitude for different AA-ratios was calculated and the respective number of cells for the ablation and accumulation area could be inferred from the cumulative number of grid cells. In the Excel© spreadsheet the respective altitude for the grid cells could be read off. (see fig. 37).

AA-ratio	**Sum of grid cells of ablation area**	**Altitude (asl)**
0.75 Brückner-approach	397,774	990 m
0.67	525,062	1025 m
0.6	636,438	1046 m
0.5 Kurowski-approach	795,548	1067 m
0.4	954,658	1092 m

Tab. 3 EL-altitudes for different AA-ratios for the Kleiner Arbersee glacier

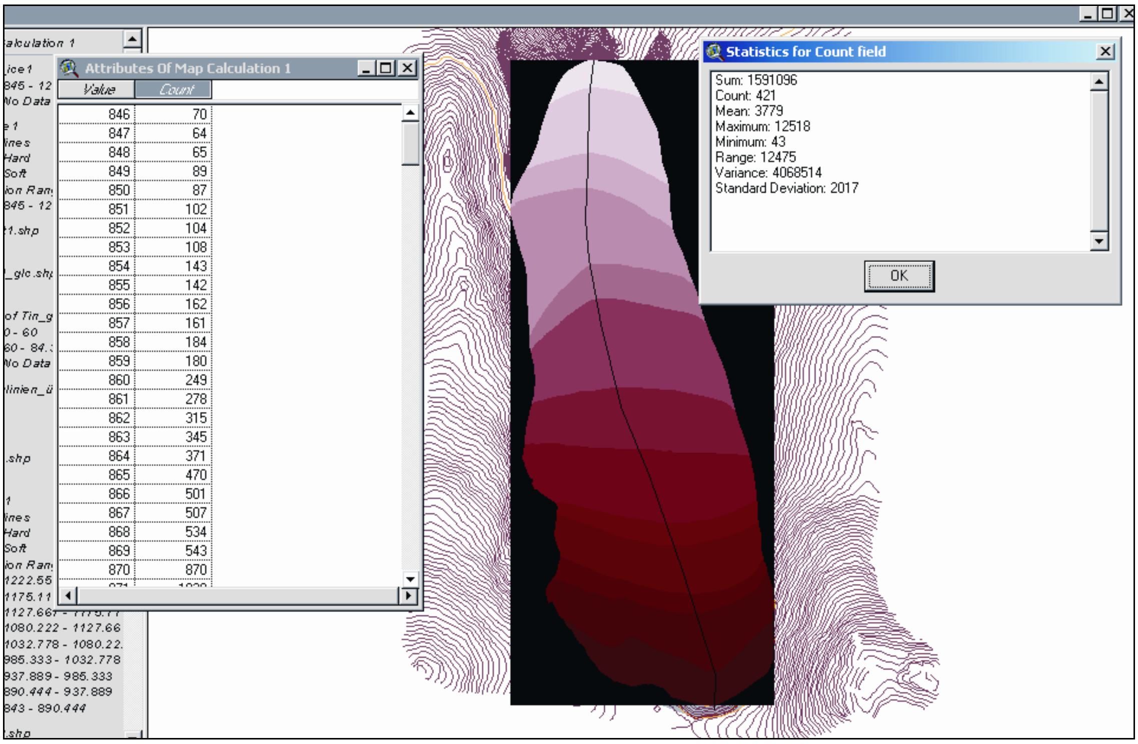


Fig. 37 Grid of ice-surface with the table of the integer elevation values and the respective statistics

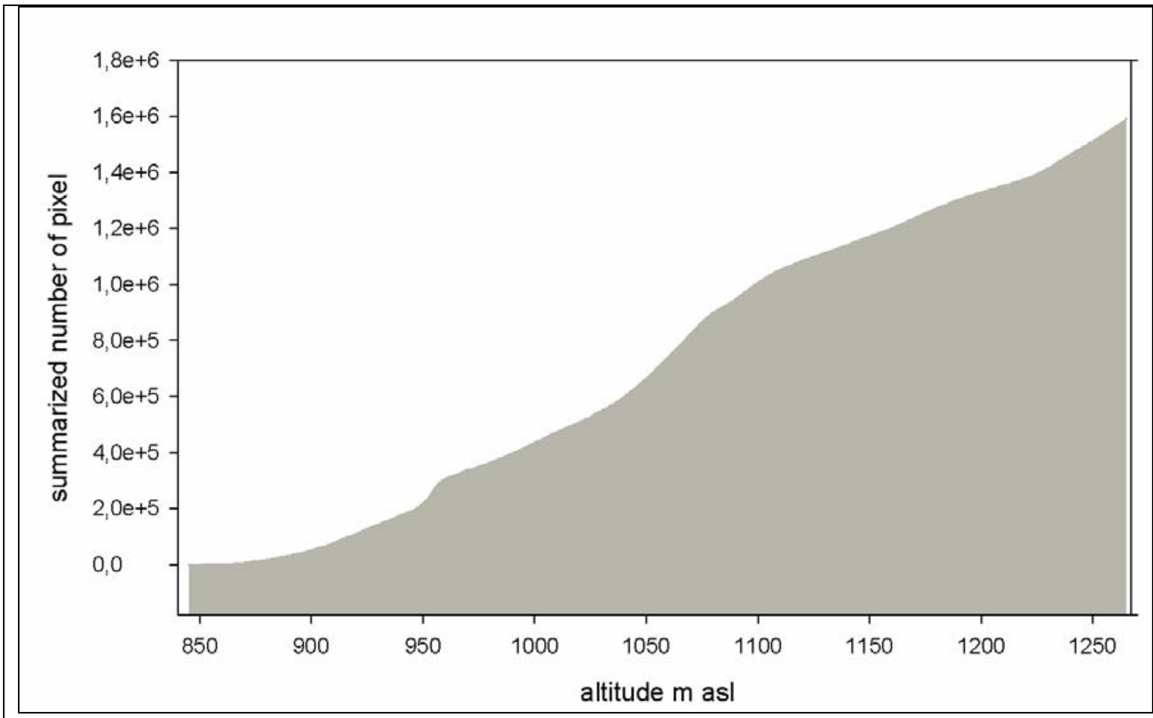


Fig. 38 Hypsographic curve of the summarized grid cells of the ice-surface, ELA can be read off the cumulative curve.

The AA-ratio for the altitude of 1055 m asl on the glacier can be calculated to be 0.58-0.59. The range was calculated for the summed number of grid cells for this altitude (657,731 to 665,331 cells lay at the elevation of 1055 m asl).

The balance-ratio method cannot be applied in the sense of FURBISH & ANDREWS (1984) because the mass balance of the glacier is unknown. But for the case that the Kleiner Arbersee glacier is in equilibrium (accumulation and ablation cancel out in the equation) and the ELA is located around the 1055 m contour line, the balance-ratio can be calculated. If the mass balance terms cancel out the ratio of the extent of the accumulation zone (925765 cells) to the ablation zone (665331 cells) is 1.3. This result does not reflect the findings of the study from FURBISH & ANDREWS (1984), where they found a BR of 2 to be representative for maritime, mid-latitude glaciers. As there are two unknown in the calculation above (mass budget information, location of the ELA), it is not possible to evaluate the result, no ELA can therefore be inferred.

If the glacier is situated in an area of high avalanche activity, the steep cliffs surrounding the glacier have to be included in the accumulation area. Reasonable assumptions about the debris cover of the glacier have to be made.

6.2 Median bedrock method

The median bedrock approach is similar to the AAR-method. The same methodology was applied, only a grid from the glacier bed, as represented by the modern valley relief, was used. To determine the pELA according to this approach the median elevation (arithmetic mean) for the valley relief surface was calculated applying a ratio of 0.5. The total grid cell number is 1,703,042. Accordingly the median elevation of the surface of the former glacier bed is located at an altitude of 1000 m asl.
The ratio of the area above the ELA to the area below the ELA for a possible pELA at 1050 m asl, as suggested by other approaches, can be inferred to be 0.63.

6.3 Modified-Höfer methods

The different modifications of the Höfer-approach allow a fairly quick estimation of the ELA.
The **modified-Höfer method** that is based on the median elevation of the ridges that surmount the glacierized zone has the largest uncertainty in the way of determination the median elevation. A possible approach was presented in Chapter 5 by taking all grid cells into account. This yields a significantly lower median elevation as does the simple averaging of the summit and ridgeline elevations. The cell number-approach yields a median elevation of 1318 m asl and the averaging of summit elevations yields a median elevation of 1370 m asl. BUCHER (1999) averaged the summit elevations for the catchment and also derived 1370 m asl for the ELA-modelling in his study. The lowermost elevation of the glacier tongue for the glacial stadium M2a of the Kleinen Arbersee glacier is 845 m. Consequently, the ELA for the median elevation of 1318 m is 1082 m; the ELA for the median elevation of 1370 m lies at 1108 m, if a ratio of 0.4 is applied it can be calculated to be 1055 m.

The **THAR- or bergschrund method** calculates the ELA with an upper limit at 1270 m (see chap. 5) and the arithmetic mean of the vertical distance to the glacier tongue. The corresponding ELA can be inferred to lie at an altitude of 1058 m.

The **TSAR-method** utilizes the summit elevation of the Grossen Arber (1456 m) to calculate the altitude of the EL. The calculated ELA with a ratio of 0.5 is accordingly highest, with an altitude of 1151 m, a ratio of 0.4 seems more appropriate (see tab. 4).

Method	ratio 0.3	ratio 0.4	ratio 0.5	ratio 0.66	ratio 0.7
Modified-Höfer Median elevation 1318 m	987 m (0.76)	1034 m (0.63)	1082 m (0.43)	1158 m (0.25)	1176 m (0.21)
Modified-Höfer Median elevation 1370 m	1003 m (0.72)	1055 m (0.56)	1108 m (0.34)	1192 m (0.18)	1213 m (0.15)
THAR-method	973 m (0.78)	1015 m (0.69)	1058 m (0.54)	1126 m (0.31)	1143 m (0.28)
TSAR-method	1028 m (0.66)	1089 m (0.41)	1151 m (0.26)	1284 m (0.06)	1273 m (-)

Tab. 4 Paleo-ELAs for the Kleinen Arbersee glacier from calculation according to the Höfer-methods with different ratios. The dark shaded values are ruled out due to the minimum assumption of the MELM and the maximum assumption of the cirque-floor altitude. The ratios in brackets are the calculated equivalent AARs for the “Höfer-elevations”.

6.4 Maximum elevation of lateral moraines

As described in chapter 5, the maximum elevation of the lateral moraines for the stadium M2a of the Kleinen Arbersee glacier can only be reconstructed on the east side of the valley. The elevation is inferred from DEM-interpretation, fieldwork as well as from different literature sources to be located at an altitude of about 1055 m.

As the glacier is not wide (830 m) at this point it can be convincingly argued that the MELM on the east side yields enough evidence to assume that the elevation of 1055 m is true for the whole glacier. It has to be born in mind that the maximum elevation of the lateral moraines is a minimum assumption for the ELA (see chapter 4.4).

6.5 Cirque-floor altitude approach

The cirque-floor altitude for the stadium M2a for the Kleinen Arbersee glacier cannot be reconstructed with confidence, as the visible cirques might also be of later origin. As discussed in chapter 5 the best developed cirque-floor level can be found in accordance with the literature and field observations to be at 1050 m asl and at around 1100 m. If the mean of the cirque-floor elevations is taken as an estimate of the pELA as MÜLLER (1978) proposed, the mean elevation is around 1038 m asl (950 m, 1050 m, 1100 m, 1250 m). Following CHARLESWORTH (1937) the pELA is represented by the lowest cirque of a staircase of cirques, the pELA is located at 950 m.

6.6 Glaciation limit approach

The glaciation limit approach calculates the altitude of the EL as the arithmetic mean between the highest non-glaciated and the lowest-glaciated mountain of a region. This method is complicated to apply to the Bavarian forest. It is very subjective of whether a mountain was glaciated or not in this mountain range. On different mountains nivation hollows give evidence of perennial snow but not of a full glaciation. The Kleiner Arber summit does not show any sign of glacial overprinting, but the slopes of the Kleiner Arber show cirques and have experienced glaciation. For this case the Kleiner Arber would be the lowest glaciated mountain, but taking the summit elevation for the glaciation limit-approach would not reflect reality. To derive a sensible estimate of the ELA it seems more appropriate to use the saddle between Kleiner and Grosser Arber as the lowest glaciated elevation. Determination of the highest non-glaciated mountain is almost as complicated. The Spitzberg mountain (1058 m asl) and the Sollerberg mountain (886 m asl) do not show distinct signs of glaciation, even though that they might have carried perennial snow or firn patches at high-glacial climatic conditions. The arithmetic mean between the altitudes yields an estimate of the pELA for the Spitzberg of 1164 m asl and of the Sollerberg of 1078 m asl.

6.7 Approaches that are not applicable to the Kleiner Arbersee glacier

From the above summarized approaches of ELA-modelling the ones that require information on the glacier mass balance are not applicable to former glaciers. Determination of the ELA by the contour-line method of HESS (1904) cannot be applied to areas where no glacier is present. As pointed out above the balance ratio method cannot be used for study areas where no information on the mass budget exists.

7 Discussion and Interpretation of the different approaches and their results

The valley of the former Kleiner Arbersee glacier is well suited for the reconstruction of the Wuermian or late-glacial glaciation. The glacier meets most of the preconditions that exist to conduct pELA-modelling. The well-preserved erosional and depositional glacial features allow a fairly reliable reconstruction of the glacier limits, reducing the uncertainty of the geometric data that the pELA modelling is based on. Additionally the geometry, the location and the aspect of the Kleiner Arbersee glacier rules out many drawbacks. The glacier is N-NE exposed (see fig. 39 with an average slope of 10.2° of the reconstructed glacier surface and 19.9° of the glacier bed with a gentle sloping subglacial topography with no steep steps (see fig. 40 and fig. 29). The simple, symmetrical geometry and the restricted extent of the glacier allow the application of most of the described methods (chap. 4) without further adjustment as discussed above.

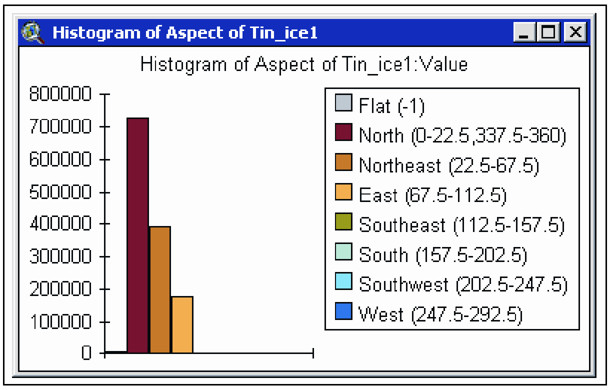


Fig. 39 Histogram of aspect values for the Kleinen Arberseeglacier

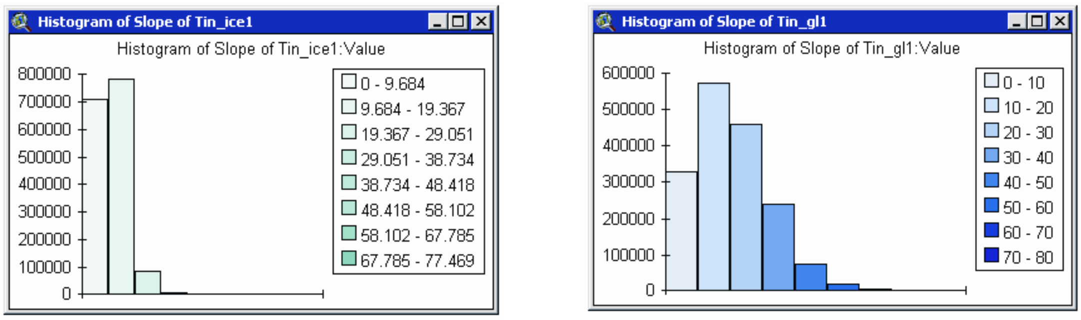


Fig. 40 Histograms of the slope of the reconstructed glacier surface (left) and the glacier bed (right)

Avalanche nourishment of the glacier can be neglected as the areal extent of the slopes above the upper glacier limit is very restricted. The calculation of the median elevation of the surface above the glacier limit yielded only a mean additional zone of 48 m vertical meters. The question of possible debris-cover is more complicated to approach. BENN & LEHMKUHL (2000) point out that extensive debris-cover of a former glacier can be inferred if prominent lateral moraines and hummocky moraines were deposited by the glacier. Even though the relief is not very hummocky, the lateral moraine ramparts are very distinct and well developed. On the other hand, steep cliffs from which rocks could fall off are not existent in the drainage. As the former debris cover cannot be reconstructed it is also neglected, well knowing that this might be a source of uncertainty for the reconstruction of the pELAs.

The evaluation of the modelling results is complicated and cannot be carried out with much confidence. Other studies that compared different ELA approaches (e.g. MEIERDING 1982, HAWKINS 1985, CARRIVICK & BREWER 2004, TORSNES et al. 1993) compared the results to modern glaciers or to other reconstructed glaciers in the study area. As neither of this is possible for this work, assumptions had to be made. The **MELM** (1055 m asl) is a reliable indicator of the minimum location of the pELA, as glaciological principles do not allow deposition of moraines above the ELA. The MELM is hence used as the minimum threshold value, any method that yield pELAs lower than the MELM is excluded. The maximum altitude of the EL is more complicated to constrain. Any pELA that is above the upper glacial limit is ruled out for this study. Even though theoretically the ELA could be located above the upper glacial limit for a glacier with negative mass balance, this consideration does not hold true for former glaciers, which are defined to be in a steady state (discussion see chap. 3). To narrow the altitudinal zone of possible pELAs, the **cirque-floor elevation** (1100 m asl) for the best developed cirque is taken as the maximum elevation of the pELA of the Kleiner Arbersee glacier. Any pELA determined to be located above this elevation is also considered erroneous. This assumption might be questionable as the cirque-floors are not unequivocally determinable and underlying principles allow deviations (discussion in chap. 4.5).

The methods of pELA determination can be divided in methods that (1) account for mass balance dependency and are based on glaciological principles as do the different AAR-approaches, the MELM-approach and the contour-line method, and (2) the methods that are simply based on topographic indicators are the different Höfer-approaches, the cirque-floor elevation method, the MBE-method and the Glaciation-limit approach. The methods that incorporate mass budget relations are more reliable than the simplified geometric determination.

The **AAR-method** can readily applied to the Kleiner Arbersee glacier as uncertainty is low in the surface reconstruction of the glacier and the surface has its largest extent in the vicinity of the ELA, which means that small departures do not influence the ratio significantly (KERSCHNER 1990). The AAR for the minimum elevation of pELA (1050 m asl) can inferred to be 0.59; the respective AAR for the maximum elevation of the pELA (1100 m asl) can inferred to be 0.37. These values are lower than the widely used ratio of 0.67. Following GROSS et al. (1977), that would hint towards humid climatic conditions, i.e. maritime climates. This conclusion should not be drawn before a whole series of pELAs for different glaciers in the Bavarian Forest were determined, as specific local climatic conditions may lead to this low AAR. The **Kurowski**-ratio of 0.5 (1067 m asl) lies in this possible range and the Kleiner Arbersee glacier reaches all prerequisites formulated by KUROWSKI (1891): (1) simple geometry, (2) consistent aspect, (3) longitudinal profile and (4) uniform slope. The **Brückner**-ratio of 0.75 is out of the possible range. The **balance-ratio method** is only suitable for modern glaciers, where mass budget information is available as pointed out earlier. If the assumption of steady-state mass balance for the integrated period of a glacial stadium (e.g. M2a) holds true, the mass budget values always cancel out of the equation. As FURBISH & ANDREWS (1984) propose a balance-ratio of 2, this means they draw the line of the ELA for a steady-state glacier at the elevation that equals the AAR of 0.67.

The results from the three different **Höfer-approaches** scatter considerably. Table 4 shows the ratios that were used for the calculation. Most of them can be ruled out according to the above described minimum and maximum elevations that are admitted for the M2a-stage of the Kleinen Arbersee glacier. It is not possible to evaluate the best method of determining the median elevation of the surrounding ridges above the glacial limit. The conventional method of averaging only the summit altitudes is as proposed by the literature. It seems, however, more sensible to account for the actual sum of elevations, than for single, possibly high towering peaks.

Generally speaking, the **TSAR-method** seems to overestimate the pELA and only a ratio of 0.4 (1089 m asl) lies in the allowed altitudinal range. The **THAR-method** only yields a valid result, when applying the ratio of 0.5 (equal to an AAR of 0.54, 1058 m asl). The **modified-Höfer-approach** gives valid results for a ratio of 0.5 (median elevation 1318 m asl) (ELA: 1082 m asl). The modified-Höfer approach with a median elevation of 1370 m asl yields for the ratio of 0.5 a result close to the defined maximum (1108 m asl) and for a ratio of 0.4 a result close to the defined minimum (1055 m asl).

The **cirque-floor** elevation as proposed by CHARLESWORTH (1937) (lowest cirque in case of many cirques) can be excluded as well as the median cirque-floor elevation as proposed by MÜLLER (1978), if the minimum criterion of the MELM elevation is applied. Tertiary erosion of the cirques as proposed by BENN & LEHMKUHL (2000) can be ruled out as the Quaternary morphodynamic in the drainage overprinted the Tertiary relief considerable. This suggests that the best developed cirque-floor might reflect the ELA for the maximum glacial stadium M2a for the Kleinen Arbersee glacier.
The **glaciation-limit method** is too spongy and subjective to be seriously applied for a single glacier. It might be appropriate for a first crude approximation of pELAs for a whole area that lacks reliable topographic information. The calculations for the glaciation-limit with the Spitzberg mountain as the lowest non-glaciated mountain is out of the range, the glaciation-limit for the Sollerberg mountain falls with a value of 1078 m asl into the allowed range.
The **MBE-method** supposedly accounts better for the drainage topography, but does not seem appropriate as the resulting pELA is below the minimum threshold. It can be adjusted by using a different ratio than 0.5, but this needs more confirmation.

Generally speaking, the best results for the pELA-reconstruction of the Kleinen Arbersee glacier can be achieved with an AAR of 0.5 (1067 m asl), a THAR-ratio of 0.5 (1058 m asl) and a modified-Höfer method with the median elevation derived from the averaged grid cell values of 0.5 (1082 m asl), the average pELA can be calculated to be located at 1069 m asl with a standard deviation of 12.1 m. But again, the AAR-method should be preferred to the Höfer-methods due to its higher reliability as it integrates the surface area and is based on glaciological laws. Table 5 summarizes the calculated results and gives the arithmetic mean and the standard deviation. As a result of the discussion of the different methods and the fact that the defined thresholds are included in the arithmetic mean the above given value of 1069.1 m asl (σ = 12 m) should be used.

AAR (ratio 0.5)	1067 m asl
THAR (ratio 0.5)	1058 m asl
Modified Höfer (mean elevation 1318 m asl)	1082 m asl
TSAR (ratio 0.4)	1089 m asl
Cirque floor elevation	1100 m asl
MELM	1055 m asl
Glaciation limit (Sollerberg)	1078 m asl
Arithmetic mean of all methods	**1076 m asl**
Standard deviation	**16.5 m**

Tab. 5 Compilation of the calculated pELA values for the Kleinen Arbersee glacier (stadium M2a)

For the Kleinen Arbersee glacier no ELA-depression can be calculated, because no reference level can be unequivocally defined. There are a set of recessional, well-preserved moraines that could serve as reference, but absolute dating would be the prerequisite in order to interpret the determined ELA-depression.

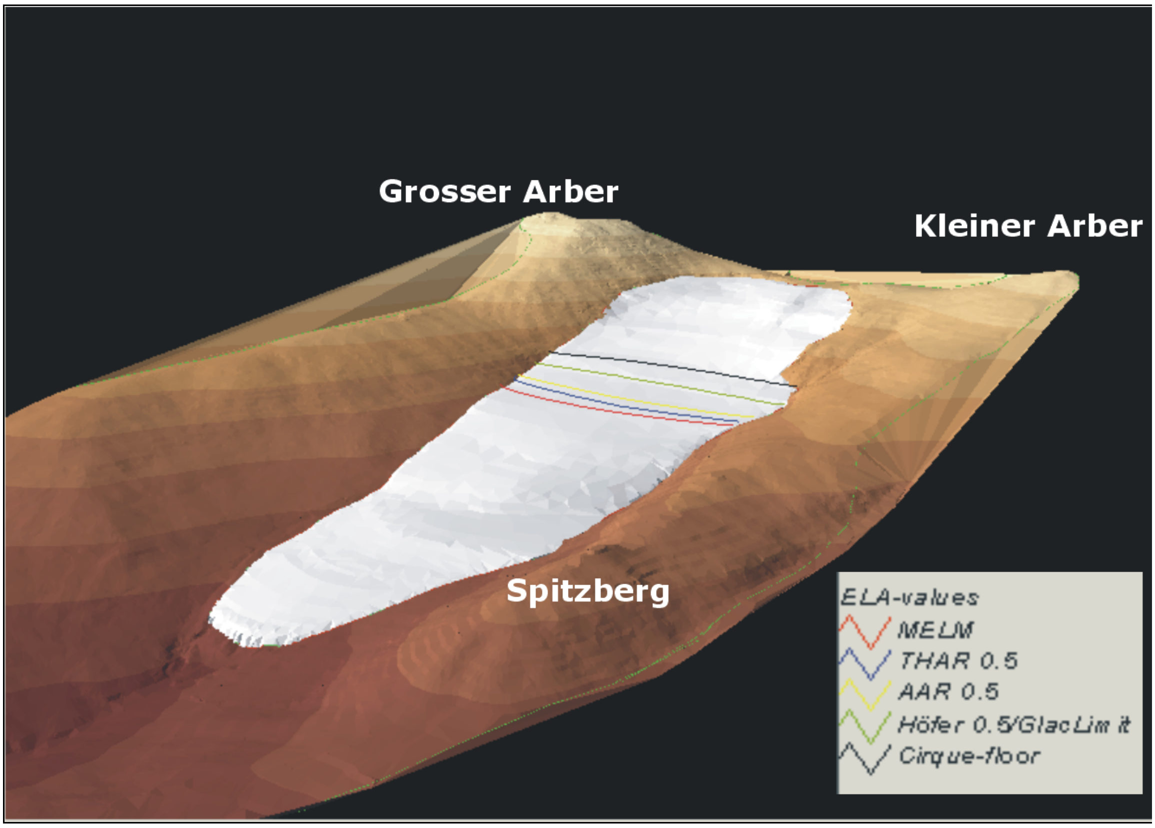


Fig. 41 ELA-contours on the 3D Model of the glacier (not optical illusion of area, see fig. 42)

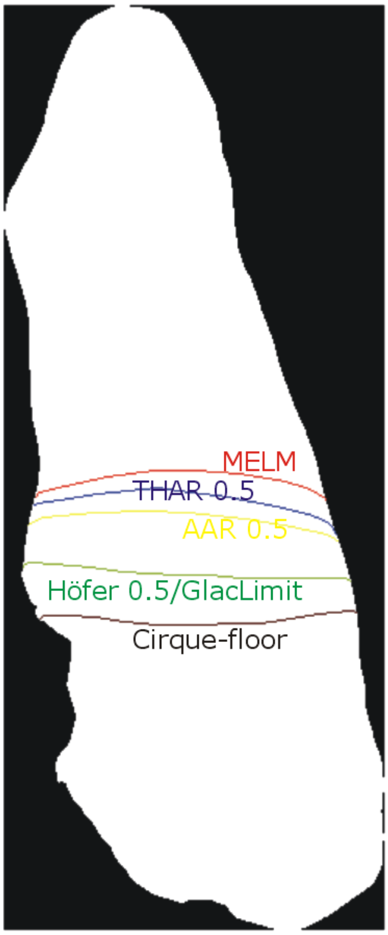


Fig. 42 ELA-contours on the glacier surface from a topview

8 Conclusion and perspective

The drainage of the former Kleiner Arbersee glacier is well studied and mapped. The glacier was a small, north-northeast facing cirque glacier with a glacier tongue that excavated the Kleinen Arbersee lake. It is a typical glacier for the Wuermian glaciation of the Bavarian Forest and left behind a text-book-like glacial setting. An altitudinal range of 55 m of a possible location of the pELA could be constrained by defining thresholds. 1069 m asl seems to be the most reliable estimate of the pELA. Generally speaking, the ice-surface integration approach of the AAR-method yields the most reliable results. For studies where only the gross morphology of a glacier can be reconstructed the Höfer-approaches ensure reasonable results. The determination of the maximum and minimum pELA threshold by the cirque-floor method and the MELM method, respectively, has shown to be beneficial. Any of the other approach should only be applied in the above discussed special situations and the results have to be interpreted carefully. For succeeding studies the attempt of taking a certain fraction of the total glacier volume as an indicator for its mass balance (KLEIN & ISACKS 1998, and MEIER & BAHR 1996 and BAHR et al. 1996 therein) and the more geometrically orientated approaches of pELA-determination from KUHLE (1986) should be further developed and evaluated.

The pELA as an important parameter for glacier mass balance could be modelled for the study area applying GIS-software. The GIS approach allows conducting all kinds of spatial analyses for the glacier, the drainage and the surrounding ridgelines. As the data basis for this study is of high resolution, the spatial analyses are far more reliable and accurate than traditional map interpretation. After completing the time-consuming production of the DEM, the spatial queries and analyses can efficiently be done; especially the planimetry of the glacier and the cumulative hysographic curves can readily be achieved. Traditional map interpretation has many shortcomings, resulting in inaccurate parameterization as CARRIVICK & BREWER (2004) convincingly showed in their study on ELA-determination for Northern Scandinavia. For study areas that are not accessible for field work or for which no maps of satisfying resolution are available, remote-sensing techniques allow a first approximation of the pELAs as shown by DUNCAN et al. (1998) for Central Nepal. In any case, the accuracy of the base map or the image is the key to the accuracy of each ELA-modelling approach as well as a possible errors. Reliability of the ELA-reconstruction is improved if the geometry of the studied glacier fulfils the requirements discussed in chapter 3. The validity of the method can only be evaluated by looking at the mass balance of modern glaciers. This however leads to the problem that most modern glaciers are presently at a negative mass balance, allowing only limited extrapolation of the measured ratios. The best way of utilizing ELA-values is the correlation of pELAs only for climatologically comparable regions; the pELAs to be compared have to be derived from the same method. In this case any uncertainty inherent in the method cancels out and only relative values are correlated.

For the Kleinen Arbersee glacier no correlation or accuracy tests can be conducted as only one data set exists. Further studies should aim to reconstruct pELAs for the same glacial stage in different catchments of the Bavarian Forest. This leads to the need of reliable absolute dating of the glacial stages. Not before absolute data is available reliable correlations between catchments can be realized. The dating approach with cosmogenic nuclides seems most appropriate to yield the desired absolute ages for moraine deposition. A step further will then be the calculation of trend-surfaces for the pELA of a substantial number of different catchments in the Bavarian Forest. The interpretation of the slope of such a trend-surface reveals regional trends of glacier development and hence allows deducing paleoclimatic conclusions as precipitation gradients and evidence for the atmospheric circulation pattern. Good examples for trend-surface analyses for ELAs over multiple catchments are for example the studies of MEIERDING (1982), HAWKINS (1985) and CARRIVICK & BREWER (2004). As most of the glaciers of the Bavarian Forest are of a similar shape and glacier type (cirque and restricted valley glaciers) this could be done with some confidence. To compare the paleoclimatic situation in the Bavarian Forest to other mountain range no absolute values of pELA can be used, as they reflect the local topographic situation. For this case the determination of the ELA-depression between different glacial stages, for instance the depression from LIA to LGM glaciation, is needed. ELA-depressions again require the unequivocal age determination of the glacial deposits that are used in the calculation. The ELA-depression is a powerful tool for comparative mountain research and allows inference of possible synchrony or asynchrony between the paleoclimatic conditions in different regions of the world. If the ELA-depression is combined with other climatic information such as the timberline or information on rock glaciers as shown by KERSCHNER (1985) and KERSCHNER et al. (2000) for the Tyrolean Alps, even more reliable climatic reconstructions can be achieved.

The presented proceedings of ELA-determination can readily be transferred to other glacial settings.

References

AMMANN, C., JENNY, B., KAMMER, K. & B. MESSERLI (2001): Late quaternary glacier response to humidity changes in the arid andes of Chile (18-29°S). – Palaeogeography, Palaeoclimatology, Palaeoecology, **172**, pp. 313-326.

ANDREWS, J.T. (1975): Glacial Systems. An approach to glaciers and their environments. – California, 183 p.

ANDREWS, J.T., BARRY, R.G. & L. DRAPIER (1970): An inventory of the present and past glacierization of Home Bay and Okoa Bay, East Baffin Island, NWT, Canada, and some climatic and palaeoclimatic considerations. – Journal of Glaciology, **57**, pp. 337-362.

ANONYMOUS (1969): Mass balance terms. – Journal of Glaciology, **52**, pp. 3-7.

BAUMGARTNER, A. (1970): Klima und Erholung im Bayerischen Wald. – In: Makowsky, H. & H. Weinzierl (ed.): Natur, Freizeit und Erholung. – Verhandlungen Deutscher Beauftragter für Naturschutz und Landschaftspflege, **17**, pp. 39-53.

BAYBERGER, F. (1886): Geographisch-geologische Studien aus dem Böhmerwalde. die Spuren alter Gletscher, die Seen und Thäler des Böhmerwaldes. - Petermanns Geographische Mitteilungen, **81**, 63 p.

BAYKLIMFO (1996): Klimaatlas von Bayern. – München, 114 p., 57 maps.

BENN, D.I. & A.M.D. GEMMELL (1997): Calculating equilibrium-line altitudes of former glaciers by the balance ratio method: a new computer spreadsheet. – Glacial Geology and Geomorphology, tn01 (http://ggg.qub.ac.uk/ggg/papers/full/1997/tn011997/tn01.html)

BENN, D.I. & F. LEHMKUHL (2000): Mass balance and equilibrium-line altitudes of glaciers in high-mountain environments. – Quaternary International, **65**, pp. 15-69.

BLWW (BAYERISCHES LANDESAMT FÜR WASSERWIRTSCHAFT) (1983/1987) : Verzeichnis der Seen in Bayern mit einem Kartenteil. – München, 566 p., 2 volumes.

BRAITHWAITE, R.J. & F. MÜLLER (1978): On the parameterization of glacier equilibrium line altitude. – World Glacier Inventory, IAHS-AISH Publication No. **126**, Proceedings of the Workshop at Riederalp, Switzerland, 17.-22.9.1978, pp. 263 – 271.

BRAITHWAITE, R.J. (1984): Can the mass balance of a glacier be estimated from its equilibrium-line altitude? – Journal of Glaciology, **30**, pp. 364-368.

BRÜCKNER, E. (1886): Die Vergletscherung des Salzachgebietes. – Geographische Abhandlungen, Vol. **1**, 183 p.

BUCHER, M. (1999): Der glaziale Formenschatz im Gebiet um den Kleinen Arbersee (Bayerischer Wald, Deutschland). – Zeitschrift für Geomorphologie, NF, **113**, S. 45-58.

BURBANK, D.W. & K.J. CHENG (1991): Relative dating of Quaternary moraines, Rongbuk Valley, Mount Everest, Tibet: Implications for an ice sheet on the Tibetan Plateau. – Quaternary Research, **36**, pp. 1-18.

CARRIVICK, J.L. & T.R. BREWER (2004): Improving local estimations and regional trends of glacier equilibrium line altitudes. – Geografiska Annaler, **86A**, pp. 67-79.

CHARLESWORTH, J.K. (1957): The Quaternary Era. Glaciology, glacial geology. – London, 1700 p.

CHINN, T.J. & I.E. WHITEHOUSE (1978): Glacier snow line variations in the Southern Alps, New Zealand. – World Glacier inventory, IAHS-AISH Publ. No. **126**, pp. 219 – 228.

CLARK, D.H., CLARK, M.M. & A.R. GILLESPIE (1994): Debris-covered glaciers in the Sierra Nevada, California, and their implications for snowline reconstruction. – Quaternary Research, **41**, pp. 139-153.

DORNBUSCH, U. (2000): Pleistocene glaciation of the dry western Cordillera in southern Peru (14°25' – 15°30' South). – Glacial Geology and Geomorphology, (http://boris.qub.ac.uk/ggg/papers/full/2000/rp012000/rp01.html)

DUNCAN, C.C., KLEIN, A.J., MASEK, J.G. & B.L. ISACKS (1998): Comparison of late Pleistocene and modern glacier extents in Central Nepal based on digital elevation data and satellite imagery. – Quaternary Research, **49**, pp. 241-254.

ELLING, W, BAUER, E., KLEMM, G. & H. KOCH (1987): Klima und Böden. Waldstandorte. – Nationalpark Bayerischer Wald, Schriftenreihe des Bayerischen Staatsministeriums für Ernährung, Landwirtschaft und Forsten, Vol. **1**, 255 p.

ERGENZINGER, P. (1967): Die eiszeitliche Vergletscherung des Bayerischen Waldes. – Eiszeitalter und Gegenwart, **18**, pp. 152-168.

FINSTERWALDER, R. (1952): Zur Bestimmung der Schneegrenze und ihrer Hebung seit 1920. – Sitzungsberichte der Bayerischen Akademie der Wissenschaften, Mathematisch-Naturwissenschaftliche Klasse, **6**, pp. 51-54.

FLINT, R.F. (1970): Glacial and Quaternary Geology. – New York, London etc., 892 p.

FURBISH, D.J. & J.T. ANDREWS (1984): The use of hypsometry to indicate long-term stability and response of valley glaciers to changes in the mass transfer. – Journal of Glaciology, **30**, pp. 199-211.

GROSS, G. (1983): Die Schneegrenze und die Altschneelinie in den österreichischen Alpen. – Innsbrucker Geographische Studien, **8**, pp. 59-83.

GROSS, G., KERSCHNER, H. & G. PATZELT (1977): Methodische Untersuchungen über die Schneegrenze in Alpinen Gletschergebieten. – Zeitschrift für Gletscherkunde und Glazialgeologie, **7**, pp. 223-251.

GÜMBEL, K.W. (1894): Geologie von Bayern. Geologische Beschreibung von Bayern. – 2 Vol., Cassel, 1184 p.

HASTENRATH, S. (1971): On snow line depression and atmospheric circulation in the tropical Americas during the Pleistocene. – South African Geographical Journal, **53**, pp. 53- 69.

HAUNER, U. (1980): Untersuchungen zur klimagesteuerten tertiären und quartiären Morphogenese des Inneren Bayerischen Waldes, Rachel-Lusen, unter besonderer Berücksichtigung pleistozän kaltzeitlicher Formen und Ablagerungen. - Nationalpark Bayerischer Wald, **6**, 198 p.

HAWKINS, F.F. (1985): Equilibrium-line altitudes and paleoenvironment in the Merchants Bay Area, Baffin Island, N.W.T., Canada. – Journal of Glaciology, **31**, pp. 205-213.

HESS, H. (1904): Die Gletscher. – Braunschweig, 426 p.

HEUBERGER, H. (1966): Gletschergeschichtliche Untersuchungen in den Zentralalpen zwischen Sellrain- und Ötztal. – Wissenschaftliche Alpenvereinshefte, **20**, 126 p.

HÖFER, H. (1879): Gletscher- und Eiszeitstudien. – Sitzungsberichte der Akademie der Wissenschaften. Mathematisch-Naturwissenschaftliche Klasse, Vol. 1, Nr. **79**, pp. 331-367.

HOINKES, H. (1970): Methoden und Möglichkeiten von Massenhaushaltsstudien auf Gletschern. – Zeitschrift für Gletscherkunde und Glazialgeologie, **6**, pp. 37-90.

HUMLUM, O. (1986): Mapping of glaciation levels: comments on the effect of sampling area size. – Arctic and Alpine Research, **18**, pp. 407-414.

IVANOV, N.N. (1959): Izwest.Wsesoj.Geogr. Obschtsch. **91**, pp. 410-423. (russian), primary source not visited, see SCHÖNWIESE (1994) or Web-source: www.klimageographie.de.vu

JERZ, H. (1993): Das Eiszeitalter in Bayern. – Stuttgart, 243 p.

KERSCHNER, H. (1978): Zur Rekonstruktion eines spätglazialen Gletscherstandes mit Hilfe eines rechnerisch ermittelten Zungenlängsprofiles. – Zeitschrift für Gletscherkunde und Glazialgeologie, **14**, pp. 119- 123.

KERSCHNER, H. (1980): Outlines of the climate during the Egesen advance (Younger Dryas, 11000 – 10000 BP) in the Central Alps of the Western Tyrol, Austria. – Zeitschrift für Gletscherkunde und Glazialgeologie, **16**, pp. 229-240.

KERSCHNER, H. (1985): Quantitative palaeoclimatic inferences from lateglacial snowline, timberline and rock glacier data, Tyrolean Alps, Austria. – Zeitschrift für Gletscherkunde und Glazialgeologie, **21**, pp. 363-369.

KERSCHNER, H. (1990): Methoden der Schneegrenzbestimmung. – In: Liedtke, H. (ed.) : Eiszeitforschung. - Wiss. Buchgesellschaft, Darmstadt, pp. 299-311.

KERSCHNER, H., KASER, G. & R. SAILER (2000): Alpine Younger Dryas glaciers as palaeo-precipitation gauges. – Annals of Glaciology, **31**, pp. 80-84

KLEIN, A.G. & B.L. ISACKS (1998): Alpine glacial geomorphological studies in the central Andes using Landsat Thematic Mapper images. – Glacial Geology and Geomorphology, rp01/1998. (http://ggg.qub.ac.uk/GGG/papers/full/1998/rp011998/rp01.htm)

KNOCH, K. (1952): Klima-Atlas von Bayern. – Bad Kissingen, 78 maps, 83 p.

KÖPPEN, W.P. & R. GEIGER (1972): Handbuch der Klimatologie. – Nendeln, 5 vol., reprint from edition 1930-38.

KUHLE, M. (1986): Schneegrenzberechnung und typologische Klassifikation von Gletschern anhand spezifischer Reliefparameter. – Petermanns Geographische Mitteilungen, **1**, pp. 41-51.

KUHN, M. (1980): Die Reaktion der Schneegrenze auf Klimaschwankungen. – Zeitschrift für Gletscherkunde und Glazialgeologie, **16**, pp. 241-254.

KUHN, M. (1983): Die Höhe der Schneegrenze in Tirol, berechnet aus Fliris klimatischen Profilen. - Innsbrucker Geographische Studien, **8**, pp. 85-91.

KUHN, M. (1984): Mass budget imbalances as criterion for a climatic classification of glaciers. – Geografisca annaler, **66A**, pp. 229-238.

KULKARNI, A.V. (1992): Mass balance of Himalayan glaciers using AAR and ELA methods. – Journal of Glaciology, **38**, pp. 101-104.

KUROWSKI, L. (1891): Die Höhe der Schneegrenze mit besonderer

Berücksichtigung der Finsteraarhorn-Gruppe. – Pencks geographische Abhandlungen, **V**, Wien., pp. 119-160.

LEHMKUHL, F. (1995): Geomorphologische Untersuchungen zum Klima des Holozäns und Jungpleistozäns Osttibets. – Göttinger Geographische Abhandlungen, **102**, 200 p.

LICHTENECKER, N. (1938): Die gegenwärtige und eiszeitliche Schneegrenze in den Ostalpen. – In: Götzinger, G. (ed.): Verhandlungen der III. Internationalen Quartärkonferenz, Wien, pp. 141-147.

LOCKE, W.W. (1990): Late Pleistocene glaciers and the climate of Western Montana, USA. – Arctic and Alpine Research, **22**, pp. 1-13.

LOUIS, H. (1955): Praktische Hinweise und Hilfsmittel zur landeskundlichen Arbeit. Schneegrenze und Schneegrenzbestimmung. – Geographisches Taschenbuch, pp. 414-418.

MAHR, A. (1998): Lockerbraunerden und periglaziale Hangsedimente im Bayerischen Wald. Untersuchungen zur Paläoumwelt und Geomorphodynamik im Spätglazial und ihrem Einfluss auf die Pedogenese. – Regensburger Geographische Schriften, **30**, 327 p.

MAISCH, M. (1987): Zur Gletschergeschichte des alpinen Spätglazials: Analyse und Interpretation von Schneegrenzdaten. – Geographica Helvetica, **42**, pp. 63-71.

MANLEY, W.F., KAUFMANN, D.S. & J.P. BRINER (1997): GIS determination of Late Wisconsin equilibrium line Altitudes in the Ahklun Mountains of Southwestern Alaska. – Geological Society of America Abstracts, **29**, p. A-33

MANSKE, D. (1989): Zur eiszeitlichen Überprägung des Arbergebietes (Bayerischer Wald). – Acta Abertina Ratisonensia, **26**, pp. 183-198.

MEIER, M.F. (1965): Glaciers and Climate. – In: Wright, H.E. & D.G. Frey (eds.): The Quaternary of the United States, pp. 795-805.

MEIERDING, T. C. (1982): Late Pleistocene Glacial Equilibrium-Line Altitudes in The Colorado Front Range: A Comparison of Methods. – Quaternary Research, **18**, pp. 289-310.

MÜLLER, F. & K. SCHERLER (1980): Some comments on the Swiss glacier inventory. – World Glacier Inventory, IAHS-AISH Publication No. **126**, Proceedings of the Workshop at Riederalp, Switzerland, 17.-22.9.1978, pp. 185-195.

MÜLLER, F. (1978): Present and Late Pleistocene equilibrium line altitudes in the Mt. Everest region – an application of the glacier inventory. - World Glacier inventory, IAHS-AISH Publ. No. **126**, pp. 75-89.

MÜLLER, F., CAFLISCH, T. & G. MÜLLER (1976): Firn und Eis der Schweizer

Alpen. Gletscherinventar. – Publ. No. 57 of the ETHZ, Zürich, 174 p.

MÜLLER-HOHENSTEIN, K. (1973): Die naturräumlichen Einheiten auf Blatt 165/166 Cham. – Bundesforschungsanstalt für Landeskunde und Raumordnung (ed.): Naturräumliche Gliederung Deutschlands. Geographische Landesaufnahme 1:200 000., Bonn, 86 p.

MÜLLER-WESTERMANN, G. (1996): Klimadaten von Deutschland. Zeitraum 1961-1990. – Offenbach a.M., 431 p.

OHMURA, A., KASSER, P. & M. FUNK (1992): Climate at the equilibrium line of glaciers. – Journal of Glaciology, **38**, pp. 397- 411.

OSMASTON, H. (1969): Glaciers, glaciation and equilibrium line altitudes on Kilimanjaro. – In: Mahaney, W.C. (ed.): Quaternary and Environmental Research on East African Mountains, pp. 7-30.

ØSTREM, G. (1966): The height of the glaciation limit in southern British Columbia and Alberta. – Geografisca Annaler, 48A, pp. 126-138.

PARTSCH, J. (1882): Die Gletscher der Vorzeit in den Karpathen und den Mittelgebirgen Deutschlands. – Breslau, 198 p.

PATERSON, W.S.B. (1994): The physics of glaciers. – Oxford, 250 p.

PÉWÉ, T.L. & R.D. REGER (1972): Modern and Wisconsinan Snowlines in Alaska. – In: Fyles, J.G. (ed.): Quaternary Geology – Géologie du Quaternaire, Section 12, Int. Geological Congress Proceedings, No. 24, S. 187-197.

PFAFFL, F. (1989): Der Bayerische Wald im Tertiär und Quartär. – Geologische Blätter Nordost-Bayerns, **39**, pp. 1-38.

PFAFFL, F. (1997): Zur Geologie des Arbergebietes. – In: Bayerisches Landesamt für Umweltschutz (ed.): Die Naturschutzgebiete am Arber. Schriftenreihe: Aus den Naturschutzgebieten Bayerns, München, **144**, pp. 9-14.

PORTER, S. C., PIERCE, K. L. & T.D. HAMILTON (1983): Late Pleistocene mountain glaciation in western United States. – Porter, S.C. (ed.): Late Quaternary environments of the United States. Volume 1, the late Pleistocene. - Minneapolis, pp. 71-111.

PORTER, S.C. (1975A): Equilibrium-line altitudes of Late Quaternary glaciers in the Southern Alps, New Zealand. – Quaternary Research, **5**, pp. 27-47.

PORTER, S.C. (1975B): Glaciation Limit in New Zealand’s Southern Alps. – Arctic and Alpine Research, **7**, pp. 33-37

PORTER, S.C. (1977): Present and past glaciation threshold in the Cascade Range, Washington, USA: Topographic and climatic controls, and paleoclimatic implications. – Journal of Glaciology, **78**, pp. 101-116.

PRIEHÄUSSER, G. (1927): Der Bayerische Wald im Eiszeitalter. I. Glaziale Spuren in der Umgebung des Großen Arbersees. – Geographisches Jahrheft, **40**, pp. 133-150.

PRIEHÄUSSER, G. (1930): Die Eiszeit im Bayerischen Wald. – Abhandlungen der Geologischen Landesuntersuchung des bayerischen Oberbergamtes, München, pp. 5-47.

RAAB, T. & J. VÖLKEL (2003): Late Pleistocene glaciation of the Kleiner Arbersee area in the Bavarian Forest, south Germany. – Quaternary Science Reviews, **22**, pp. 581-593.

RAAB, T. (1999): Würmzeitliche Vergletscherung des Bayerischen Waldes im Arbergebiet. – Regensburger Geographische Schriften, **32**, 327 p.

RATHSBURG VON, A. (1928): Die Gletscher des Böhmerwaldes zur Eiszeit. – Bericht der naturwissenschaftlichen Gesellschaft zu Chemnitz, **22**, pp. 65-161.

RICHTER, E. (1888): Die Gletscher der Ostalpen. – Stuttgart, 306 p.

ROHRMÜLLER, J., MIELKE, H. & D. GEBAUER (1996): Gesteinsfolge des Grundgebirges nördlich der Donau und im Molasseuntergrund. – In: Freudenberger, W. & K. Schwerd (ed.): Erläuterungen zur Geologischen Karte von Bayern 1:500.000. München, pp. 16-54.

SCHEUERER, M. (1991): Flora und Vegetation des Naturschutzgebietes „Kleiner Arbersee" im Hinteren Bayerischen Wald. – Hoppea, Denkschrift Regensburger Botanische Gesellschaft, **50**, pp. 233-289.

SCHEUERER, M. (1997): Das Klima des Arbergebietes. – In: Bayerisches Landesamt für Umweltschutz (ed.): Die Naturschutzgebiete am Arber. Schriftenreihe Aus den Naturschutzgebieten Bayerns, München, **144**, pp. 15-17.

SCHREINER, A. (1997): Einführung in die Quartärgeologie. – Stuttgart, 257 p.

SELTZER, G.O. (1994): Climatic interpretation of Alpine snowline variations on millenial time scale. – Quaternary Research, **41**, pp. 154-159.

SISSONS, J.B. & D.G. SUTHERLAND (1976): Climatic interferences from former glaciers in the South-East Grampian Highlands, Scotland. – Journal of Glaciology, **76**, pp. 325-346.

SLUPETZKY, H. (1974): Untersuchungen zur Massenbilanz eines Hanggletschers. Ergebnisse und Schlussfolgerungen aus der Messreihe 1963/64 – 1970/71 vom Stubaier Sonnblickkees (Hohe Tauern). – Zeitschrift für Gletscherkunde und Glazialgeologie, **10**, pp. 11-69.

TORSNES, I., RYE, N. & A. NESJE (1993): Modern and Little Ice Age equilibrium-line altitudes on outlet valley glaciers from Jostedalsbreen, Western Norway: an evaluation of different approaches to their calculation. – Arctic and Alpine Research, **25**, pp. 106-116.

TROLL, G. (1967a): Bau und Bildungsgeschichte des Bayerischen Waldes. – in: Bayerisches Geologisches Landesamt (ed.): Führer zu geologisch-petrographischen Exkursion im Bayerischen Wald. – Geologica Bavarica, **58**, pp. 15-33.

TROLL, G. (1967b): Der Arber-Kaitersberg Zug. – in: Bayerisches Geologisches Landesamt (ed.): Führer zu geologisch-petrographischen Exkursion im Bayerischen Wald. – Geologica Bavarica, **58**, pp. 53-66.

UNESCO/IASH (1970): Perennial ice and snow masses. – Technical Papers in Hydrology, **1**, 59 p.

VISSER, C. (1938): Glaziologie. – IN: VISSER, C. & J. VISSER-HOOFT (ed.): Karakorum und die angrenzenden Gebiete in den Jahren 1922, 1925, 1929/30 und 1935, Vol. **2**, 216 p.

VÖLKEL, J. (1995): Periglaziale Deckschichten und Böden im Bayerischen Wald und seinen Randgebieten. – Zeitschrift für Geomorphologie, NF, Supplementband **96**, 301 p.

VÖLKEL, J. (1999): Charakterisierung saprolithischer Verwitterung im südwestlichen Moldanubikum durch röntgendiffraktometrische Untersuchungen. - Zentralblatt für Geologie und Paläontologie, **5-6**, pp. 289-304.

WILLIAMS, V.S. (1983): Present and former equilibrium-line altitudes near Mount Everest, Nepal and Tibet. – Arctic and Alpine Research, **15**, pp. 201-211.

WISSMANN V., H. (1959): Die heutige Vergletscherung und Schneegrenze in Hochasien. – Abhandlungen der Mathematisch-Naturwissenschaftlichen Klasse, **14**, 331 p.

Appendix

App. 1 Compilation of AA-ratios applied in different study locations (selected references)

Author	(applied) AAR	Region	Time period
AMMANN et al. (2001)	0.67	Andes of Chile (18-29°)	Late Pleistocene glaciers
Braithwaite & Müller (1978)	0.67*	Worldwide	Modern glaciers
Braithwaite (1984)	0.67 and 0.5*	Worldwide	Modern glaciers
Brückner (1886)	0.75	Salzach-glacier	LGM glaciers
Burbank & Cheng (1991)	0.67*	Tibetan Plateau	Modern and late Quaternary glaciers
CLARK et al. (1994)	0.15 (0.2 - <0.1)	Sierra Nevada	Modern, debris-covered glaciers
Dornbusch (2000)	0.75	Peru	Late Pleistocene glaciers
GROSS (1983)	0.62	Austrian Alps	Modern glaciers
GROSS et al. (1977)	0.67 (values between 0.68 and 0.60)	Eastern Alps	Modern glaciers (mean from 24 years)
HAWKINS (1985)	0.65	Baffin Island	Modern glaciers, Late Pleistocene glaciers
HOINKES (1970)	0.6 (0.25-0.81)	Austrian Alps	Modern glaciers
Kulkarni (1992)	0.47 and 0.43 (average 0.44)	Western Himalaya	Modern glaciers
Lehmkuhl (1995)	0.67*	East Tibet	Late Pleistocene glaciers
LOCKE (1990)	0.65*	Montana, USA	Late Pleistocene glaciers
MAISCH (1987)	0.67*	Switzerland	Late Pleistocene glaciers
Meierding (1982)	0.65	Colorado, USA	Modern and Pleistocene glaciers
Müller & Scherler (1978)	0.4	Switzerland	Modern glaciers (negative mass balance)
MÜLLER (1978)	0.49 for N, 0.41 for NE, 0.41 for E, 0.36 for SE, 0.38 for S, 0.38	Mount Everest Region	Modern glaciers (Strong avalanche nourishment and ice-preserving influence of frequently extensive debris

	for SW, 0.44 for W, 0.38 for NW		cover)
MÜLLER et al. (1976)	0.40 (0.50 proposed for steady state)	Swiss Alps	Glaciers from 1973 (mass budget is negative)
Osmaston (1969)	0.6*	Kilimanjaro	Late Pleistocene glaciers
Porter (1975)	0.65, 0.57, 0.6	New Zealand	Modern glaciers
Slupetzky (1974)	0.70 (0.59 – 0.79	Austrian Alps	Modern glaciers (steady-state)
TORSNES et al. (1993)	0.6*	Norway	Modern glaciers, LIA glaciers
Unesco (1970)	0.28, 0.33, 0.24, 0.38, 0.30	Axel Heiberg Island	Glaciers from 1958 and 1959
VISSER (1938)	0.50, 0.58, 0.6, 0.67, 0.71, 0.81	Karakorum	Glaciers from 1925
Williams (1983)	0.60	Mount Everest area	Late Pleistocene glaciers (including slopes due to avalanching)
Wissmann (1959)	0.50 *	Asian mountain ranges	Modern glaciers

* Ratios just applied, not inferred

App. 2 Compilation of ratios of the Höfer-methods in different study locations (selected references)

Author	(applied) Höfer method	Region	Time period
Braithwaite (1984)	0.5* (THAR)	Worldwide	Modern glaciers
CLARK et al. (1994)	0.7 (0.6-0.8) (THAR)	Sierra Nevada	Modern, debris-covered glaciers
DUNCAN et al. (1998)	0.5* (THAR)	Central Nepal	Modern and late Pleistocene glaciers
Heuberger (1966)	0.5* (TAS)	Eastern Alps	Modern and late Pleistocene glaciers
Klein & Isacks (1998)	0.45 (THAR)	Andes	Pleistocene glaciers
Lehmkuhl (1995)	Louis, Höfer*	East Tibet	Late Pleistocene glaciers
LOCKE (1990)	0.40 (THAR)*	Montana, USA	Late Pleistocene glaciers
Meierding (1982)	0.35 and 0.40	Colorado, USA	Modern and Pleistocene glaciers
Péwé & Reger (1972)	THAR of 0.66	Alaska	Modern glaciers
PORTER (1975a)	0.5 within 50 m of ELA	North America	20 modern glaciers with normally distributed area-altitude curves
PORTER et al. (1983)	0.4 to 0.5	Western US	Late Pleistocene glaciers
TORSNES et al. (1993)	0.4* (0.35-0.4 best results)	Norway	Modern and LIA glaciers
Wissmann (1959)	0.5* (original Höfer)	Asian mountain ranges	Modern and last glacial glaciers

Eidesstattliche Erklärung

Hiermit erkläre ich an Eides statt, die vorliegende Masterarbeit mit dem Thema

„Constraining glacial equilibrium lines of altitude (ELA) by different geometrical approaches applying a Geographic Information System (GIS). A case study for the Kleinen Arbersee glacier, Bavarian Forest, Germany.“

selbständig verfasst zu haben und keine anderen als die angegebenen Quellen und Hilfsmittel benutzt zu haben.

Die Arbeit wurde bisher in gleicher oder ähnlicher Form keiner anderen Prüfungsbehörde vorgelegt und auch noch nicht veröffentlicht.

____

Regensburg, 09. Juli 2004