

Swiss J Geosci (2010) 103:407–426
DOI 10.1007/s00015-010-0044-y

Distribution, geometry, age and origin of overdeepened valleys and basins in the Alps and their foreland

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Received: 22 February 2010 / Accepted: 14 June 2010 / Published online: 14 December 2010
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Abstract Overdeepened valleys and basins are commonly found below the present landscape surface in areas that were affected by Quaternary glaciations. Overdeepened troughs and their sedimentary fillings are important in applied geology, for example, for geotechnics of deep foundations and tunnelling, groundwater resource management, and radioactive waste disposal. This publication is an overview of the areal distribution and the geometry of overdeepened troughs in the Alps and their foreland, and summarises the present knowledge of the age and potential processes that may have caused deep erosion. It is shown that overdeepened features within the Alps concur mainly with tectonic structures and/or weak lithologies as well as with Pleistocene ice confluence and partly also diffluence situations. In the foreland, overdeepening is found as elongated buried valleys, mainly oriented in the direction of former ice flow, and glacially scoured basins in the ablation area of glaciers. Some buried deeply incised valleys were generated by fluvial down-cutting during the Messinian crisis but this mechanism of formation applies only for the southern side of the Alps. Lithostratigraphic records and dating evidence reveal that overdeepened valleys were repeatedly occupied

and excavated by glaciers during past glaciations. However, the age of the original formation of (non-Messinian) overdeepened structures remains unknown. The mechanisms causing overdeepening also remain unidentified and it can only be speculated that pressurised meltwater played an important role in this context.

Keywords Glaciations · Erosion · Sediments · Overdeepening · Alps · Quaternary

Introduction

“*Mit Butter hobelt man nicht (Butter is not an abrasive agent)*”. This famous statement by Albert Heim (1919), probably one of the most prominent alpine geologists of the late 19th and early 20th century, reflects the scientific point of view regarding the question of whether or not Quaternary glaciations could have caused substantial vertical erosion. At that time, the sedimentary filling of valleys in the Alps was almost exclusively attributed to fluvial action. The thickness of unconsolidated Quaternary sediments in the alpine area was under heavy debate but was mainly expected not to exceed a few tens of metres. This assumption was refuted when during construction works, the front of the Lötschberg Tunnel collapsed on July, 24th in 1908 and buried the whole shift of 28 workers; only 3 survived. The tunnel advance had reached unconsolidated and water-saturated Quaternary sediments about 200 m below Gasterntal (Fig. 1) (Schweizerische Bauzeitung 1908: cit. in Schlüchter 1983).

More than half a century later geophysical surveys, as well as an increasing number of deep drillings, have shown that almost every valley and every major (lake) basin in the Alps and their foreland has a substantial infill of

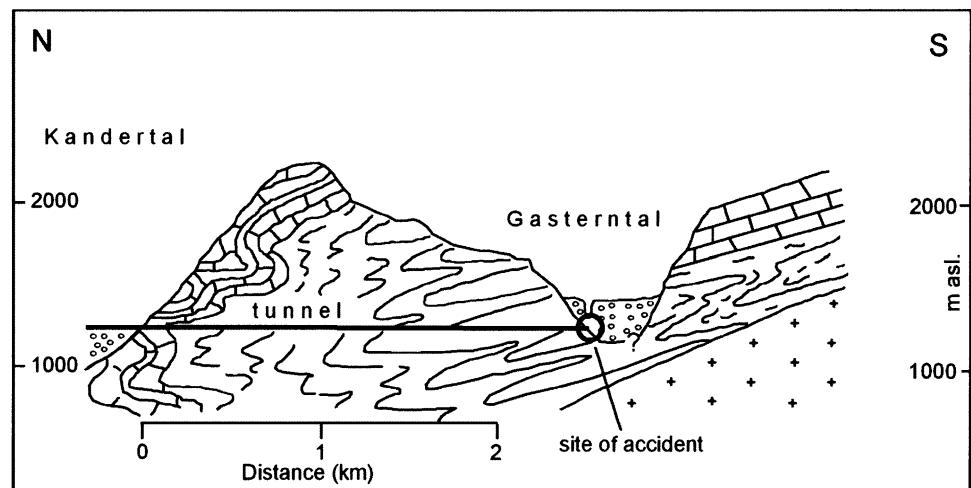
Editorial handling: Markus Fiebig.

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Fig. 1 Cross-section showing the geological situation of the Lötschberg Tunnel (Berner Oberland). On 24th July 1908 the front of the tunnel collapsed and 25 workers were killed. The accident was due to the unexpected presence of unconsolidated sediments about 200 m below the valley bottom of Gasterntal and a sediment and water slurry penetrated into the construction



unconsolidated sediments, reaching a thickness of up to 1,000 m (e.g. Schlüchter 1979; van Husen 1979; Finckh et al. 1984; Weber et al. 1990; Pfiffner et al. 1997).

It is thought that most of the deep erosional features have been formed by glacial processes due to the much higher efficiency of glacial, compared to fluvial, erosion (Montgomery 2002). Deep erosional troughs are inferred to be the result of various subglacial processes including subglacial meltwater action (Menzies 1995; Menzies and Shilts 1996), and this phenomenon is usually referred to as glacial overdeepening. In mountain areas, glacial erosion is considered either to enhance isostatic uplift (Molnar and England 1990; Champagnac et al. 2007) or to at least keep pace with rock uplift. Thus, glacial erosion represents a “buzz-saw” effect, i.e. a climatically controlled limitation of elevation on a large scale (Brozovic et al. 1997; Meigs and Sauber 2000). Additionally, on a smaller scale, glacial erosion results in increased relief (Small and Anderson 1998).

However, for the central and western Alps in particular, the increase in sediment flux during the Miocene to Pliocene is attributed to a shift towards wetter climatic conditions (Cederbom et al. 2004; Willett et al. 2006). Thus, the basal Quaternary unconformity in the Swiss Midlands, with missing upper Miocene to Pliocene strata, is usually interpreted as being the product of non-glacial erosional processes.

The distribution, geometry, age and origin of deep erosional troughs and their sedimentary infill are not only of academic interest but have major relevance in applied geology. While accidents such as the one at the Lötschberg Tunnel are unlikely to occur today, the presence of thick fills of unconsolidated sediments is crucial for the planning of tunnel constructions (selection of best construction technique) and for the management of groundwater resources. In particular the age of and the processes causing overdeepening are relevant for the

siting of radioactive waste disposal sites (e.g. Talbot 1999). For such sites the international agreement is that future generations should not be harmed by any contaminants (IAEA (International Atomic Energy Agency) 1995). As a consequence, repository sites for high-level radioactive waste have to be stable for at least 1 Ma when considering the half-life of the isotopes present in the material. Hence, if the excavation of overdeepened valleys and basins by glaciers and associated processes is a relatively young geological phenomenon, this must be considered when selecting the location of radioactive waste repositories.

The present contribution aims to provide an overview of recent knowledge of the areal distribution and geometry of overdeepened valleys and basins in the Alps and its foreland. This paper reviews constraints on the age of sedimentary fillings, and discusses likely processes that may have formed these deep erosional features. As the exogenic and endogenic processes of an orogen as complex as the Alps are closely linked, the starting point of this review is a short overview of the tectonic setting with respect to morphogenesis.

Tectonic setting and pre-Quaternary landscape evolution

The present alpine and peri-alpine drainage pattern is largely the result of the tectonic development of the Alps during the Neogene, in particular during the Miocene (e.g. Kuhlemann et al. 2001; Kühni and Pfiffner 2001a, b; Schlunegger et al. 2001). In general, the tectonic phase following continental collision during the Neogene was characterised by extensional and strike-slip faulting alternating or coeval with crustal shortening during ongoing continental convergence (cf. Schmid et al. 1996, 2004; Froitzheim et al. 2008).

The modern topographic evolution of the Central and Eastern Alps started during Oligocene times, around 30 Ma ago. First during this period, alluvial fans were deposited in the Molasse zone between Lake Geneva and south-eastern Bavaria (cf. Kuhlemann and Kempf 2002). These fluvial deposits are considered to reflect a growing differential relief within the alpine region due to crustal thickening and uplift (Kuhlemann et al. 2001).

In the Eastern Alps, the collision between the Adriatic indenter and the European plate during the Oligocene resulted in a lateral extrusion towards the East (Ratschbacher et al. 1991), which was largely finalised during the Lower to Middle Miocene. Accompanied faulting resulted in the dissection of the former Oligocene northward drainage, which extended farther to the southwest and south (Frisch et al. 1998; Ortner and Stingl 2001). The longitudinal valleys of the Eastern Alps (e.g. Inn, Salzach, Drau and Enns Valleys) follow the major strike-slip faults of the tectonic pattern and contain Oligocene sediments as well as Miocene syntectonic deposits (Frisch et al. 1998, 2000). GPS data (Grenerczy et al. 2005; Vrabec et al. 2006; Caporali et al. 2009) suggest that the eastward extrusion of the Eastern Alps, which waned in the late Middle Miocene (Frisch et al. 1998), is still active.

Drainage evolution in the Swiss part of the Northern Alpine Foreland Basin (NAFB) was controlled by progradational crustal shortening towards the north. The thrusting of the Swiss Jura Mountains resulted in a passive uplift of the western NAFB and drainage of this area towards the Palaeo–Danube (“Aare–Danube”) between 11 and 6 Ma (cf. Kuhlemann and Kempf 2002; Berger et al. 2005). After 5 Ma, strong erosion of the alpine orogene and in the Swiss NAFB is indicated by highly increased sediment discharge (Kuhlemann et al. 2001; Cederbom et al. 2004). During the Middle Pliocene, the drainage pattern changed towards the northwest in the direction of the tectonically subsiding Bresse Graben (Petit et al. 1996). At the beginning of the Quaternary (2.6 Ma), the rivers Aare and Alpine Rhine were captured by the River Rhine in the Upper Rhine Graben. These major changes in the orientation of the drainage system presumably induced a considerable lowering of the relative base level for the Central and Eastern Swiss Molasse Basin.

During the Messinian, the desiccation of the Mediterranean Sea Basin over about 640 ka resulted in a dramatic base level drop (>1,000 m; Krijgsman et al. 1999). This led to deep fluvial incision which not only affected the Po basin but also the main valleys extending into the Southern Alps (Bini et al. 1978; Finckh et al. 1984). The northward shift of the watershed, and thus the enlargement of the drainage towards the south, continues today, due to the steeper gradient of the rivers flowing towards the Adriatic Sea (Kuhlemann et al. 2001).

Distribution of overdeepened valleys

Based on various reconstructions (van Husen 1987, 2004; Schlüchter 2004, 2010; Ehlers and Gibbard 2004; Fig. 2), it has been shown that alpine glaciations were characterised by a network of connected glaciers, i.e. transection glaciers (Benn and Evans 1998). For this overview the Alps are subdivided into four regions (Danube, Rhine, Rhone, and Po drainage areas) that experienced different changes in relative base level, mainly due to tectonic activity. The largest region is the inner-alpine River Danube drainage area, which is more or less identical with the Eastern Alps. The River Rhine drainage area covers large parts of the Central Alps. The inner-alpine River Rhone drainage area includes small parts of the Central Alps and wide parts of the Western Alps. The drainage area of the River Po and that of all rivers flowing into the Adriatic Sea comprises most of the Southern Alps as well as some parts of the Eastern Alps and the eastern slope of the Western Alps. Figure 2 gives an overview of the distribution of overdeepened valleys and basins in the alpine region and shows the location of sites mentioned in the following text.

Eastern Alps (River Danube drainage system)

Several overdeepened basins are found to the north of the morphological limit of the Eastern Alps and coincide with formerly glaciated areas, especially the position of glacial scour basins. Prominent examples include the area of the former Isar–Loisach Glacier with the basins of Starnberger (Würm) See, Wolfratshausen and Rosenheim (Fig. 2) (Müller and Unger 1973; Veit 1973; Jerz 1979, 1993), all of which are situated in weak Molasse bedrock. The glacial basin of the former Salzach Glacier consists of a main basin situated around the past equilibrium line and radially diverting branch basins down-glacier. According to drilling results, the subsurface bedrock topography of the main basin is spoon-shaped (van Husen 1979; van Husen and Draxler 2009; Fig. 3). Most of the branch valleys such as the linear Oichten Valley are considerably overdeepened (Brückl et al. 2010).

The glacial basin of the Traun Glacier system is located within the Alps and Traunsee, the deepest lake of Austria (191 m), is one of the most prominent examples of an overdeepened basin (Fig. 4). Seismic data in the southern delta area of the River Traun indicate a minimum overdeepening of the whole structure of 350 m (Burgschwaiger and Schmid 2001), incised into carbonatic rocks. To the east of the Traun Valley, glacially shaped basins also occur far beyond the Last Glacial Maximum (LGM) ice limit (e.g. basin of Molln in the Steyr Valley; van Husen 2000) (Fig. 2). A typical example of a glacial basin in the eastward oriented drainage is the Enns Valley

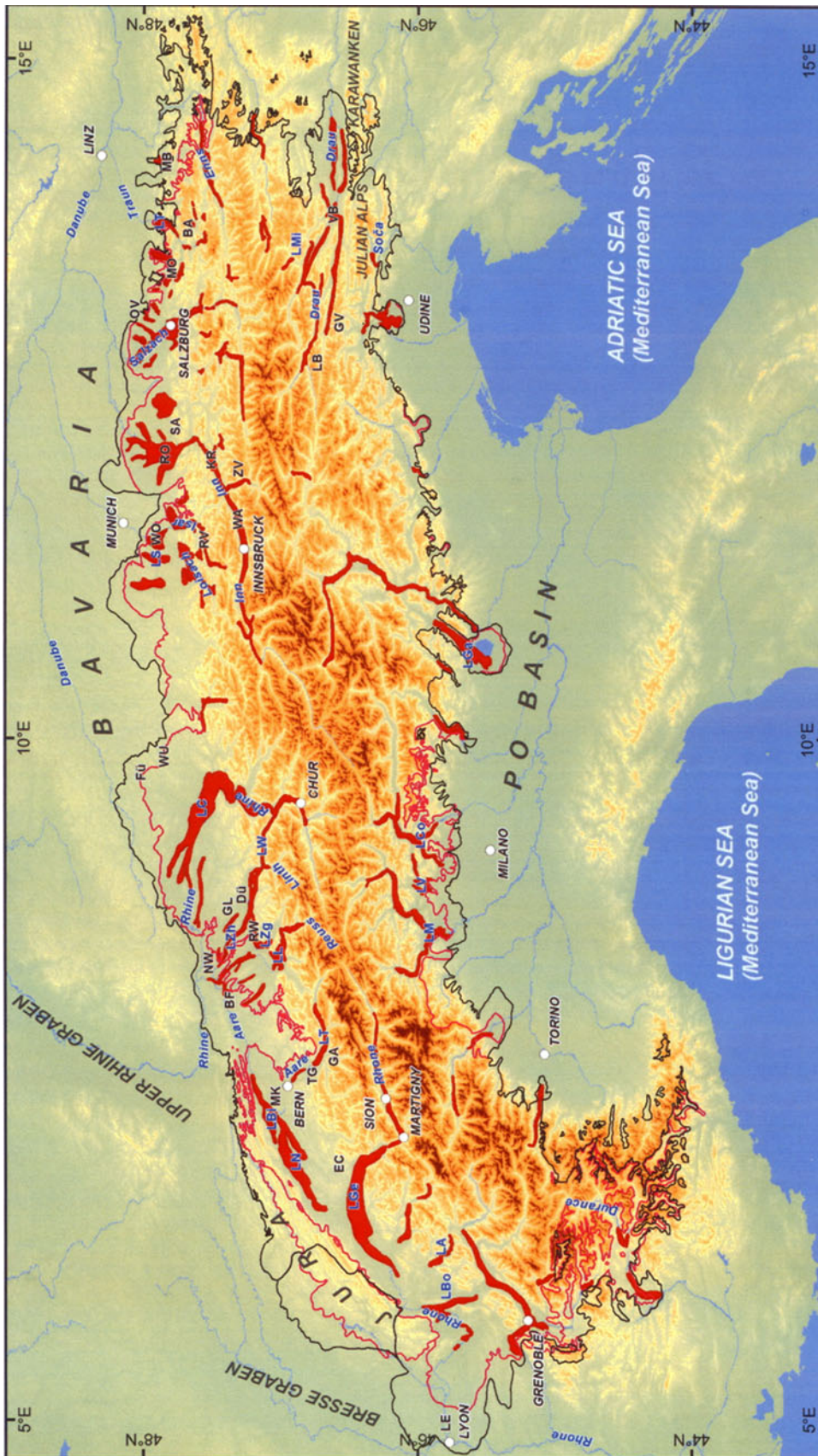


Fig. 2 Relief map of the alpine region showing the limits of the Last Glacial Maximum (yellow line), the maximum limit of Pleistocene glaciation (white line) (both after Ehlers and Gibbard 2004), and the location of overdeepened valleys and basins (based on Jordan et al. 2008 for Switzerland; additional references in text). Topographic background is taken from Jarvis et al. (2008). Abbreviations of lakes and associated basins mentioned in the text (in blue letter; alphabetic order): LA Lac d'Annecy, LBo Lac de Bourget, LC Lake Constance, LCo Lago di Como, LGe Lago di Garda, LLa Lago Lucerne, LM Lago Maggiore, LMi Millstättersee, LN Lake Neuchâtel, LT Lake Thun, LTr Traunsee; LS Stamberger (Würrn) See, LV Lago di Varese, LW Lake Walenstadt, LZg Lake Zug, LZh Lake Zürich. Abbreviations of other sites (in white): BA Bad Aussee, BF Birrfeld, Dir Dürnten, EC Ecoteaux, Fir Fürnmoos, GA Gastern Valley, GV Gail Valley, IV Isere Valley, LE Les Echets, MB Basin of Lienz, LE Les Echets, MB Basin of Lienz, MK Melktrich, MO Mondsee, NW Niederweningen, OV Oichen Valley, RO Rosenheim, TG Thalgut, VA Villach, WA Wattens, WO Wolfrathshausen, WU Wurzsach, ZV Ziller Valley

Fig. 3 Longitudinal cross-section of the Salzach Basin near Salzburg showing the internal structure of a typical glacial tongue basin (modified after van Husen 1979). The River Salzach and its local tributaries (Königsee Ache, Taugelbach, Lammer) are responsible for coarse input into the basin

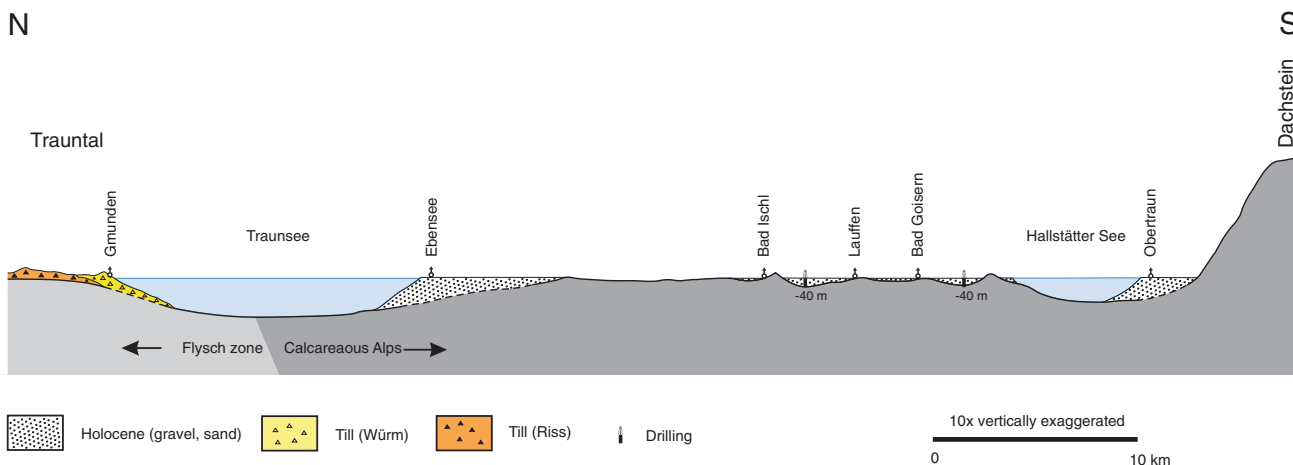
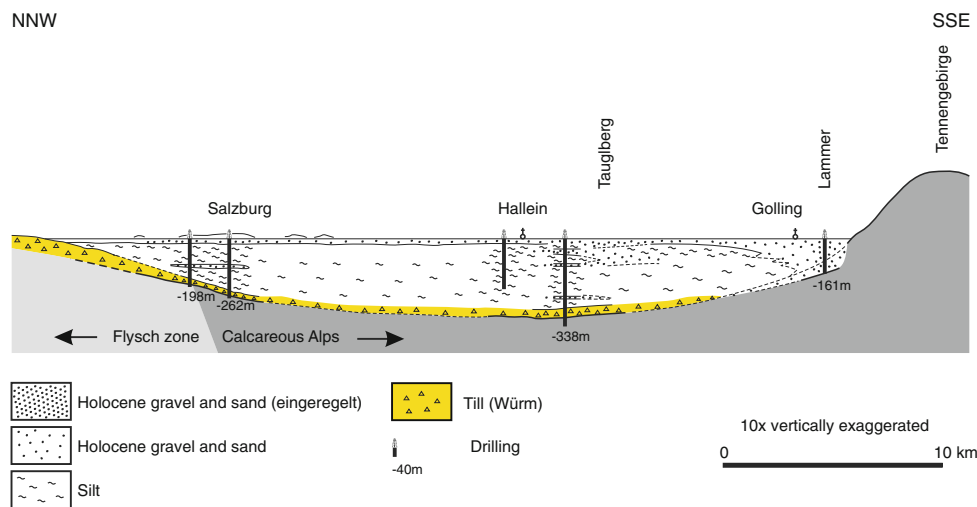


Fig. 4 Longitudinal cross-section of the Traun Valley showing glacial tongue basins finally shaped during the LGM (Traunsee) and during the Lateglacial Gschnitz Stadial (Hallstätter See). Both lakes

are presently being filled by delta sediments of the River Traun (modified after van Husen 1979)

with a minimum bedrock depth of 192 m observed in a drillhole (van Husen 1979), and probably as deep as 480 m according to seismic data (Schmid et al. 2005). In contrast, the depth to bedrock in the area of the former Drau Glacier, just in front of the emerging Karawanken Mountains, is only in the range of 100 m according to drillholes (Nemes et al. 1997; Spendingwimmer and Heiss 1998) (Fig. 2).

Overdeepening in inner-alpine valleys, located within the accumulation zone of Pleistocene glaciers, are usually associated with specific situations in the former ice flow pattern. Overdeepening features are found at former glacial confluences coinciding with broadening valley situations, for example, in the Gail Valley (Carniel and Riehl-Herwirsch 1983) (Figs. 2, 5a). Such a setting in the basin of Lienz, located within faulted crystalline rocks (Walach

1993), continues to its narrow down-flow prolongation in the Upper Drau Valley (Brückl and Ullrich 2001) (Figs. 2, 5b). Major overdeepened features linked to confluences as a result of complex dendritic ice-flow patterns are also found in narrow valleys within limestone areas, such as the Riss Valley (Frank 1979) (Figs. 2, 5c). However, overdeepened basins related to former ice diffluences are also known, such as the lake of Millstättersee, cut into metamorphic rocks (Penck and Brückner 1909) (Fig. 2).

Other principal occurrences of overdeepened troughs are linked to longitudinal valleys (e.g. Inn, Enns and Salzach Valleys) or areas affected by Miocene tectonic movements (e.g. Villach Basin) (Fig. 2). However, the amount of overdeepening is still a matter of debate as the presence of Tertiary sediments complicates the interpretation of geophysical data (in particular seismic and gravimetric surveys).

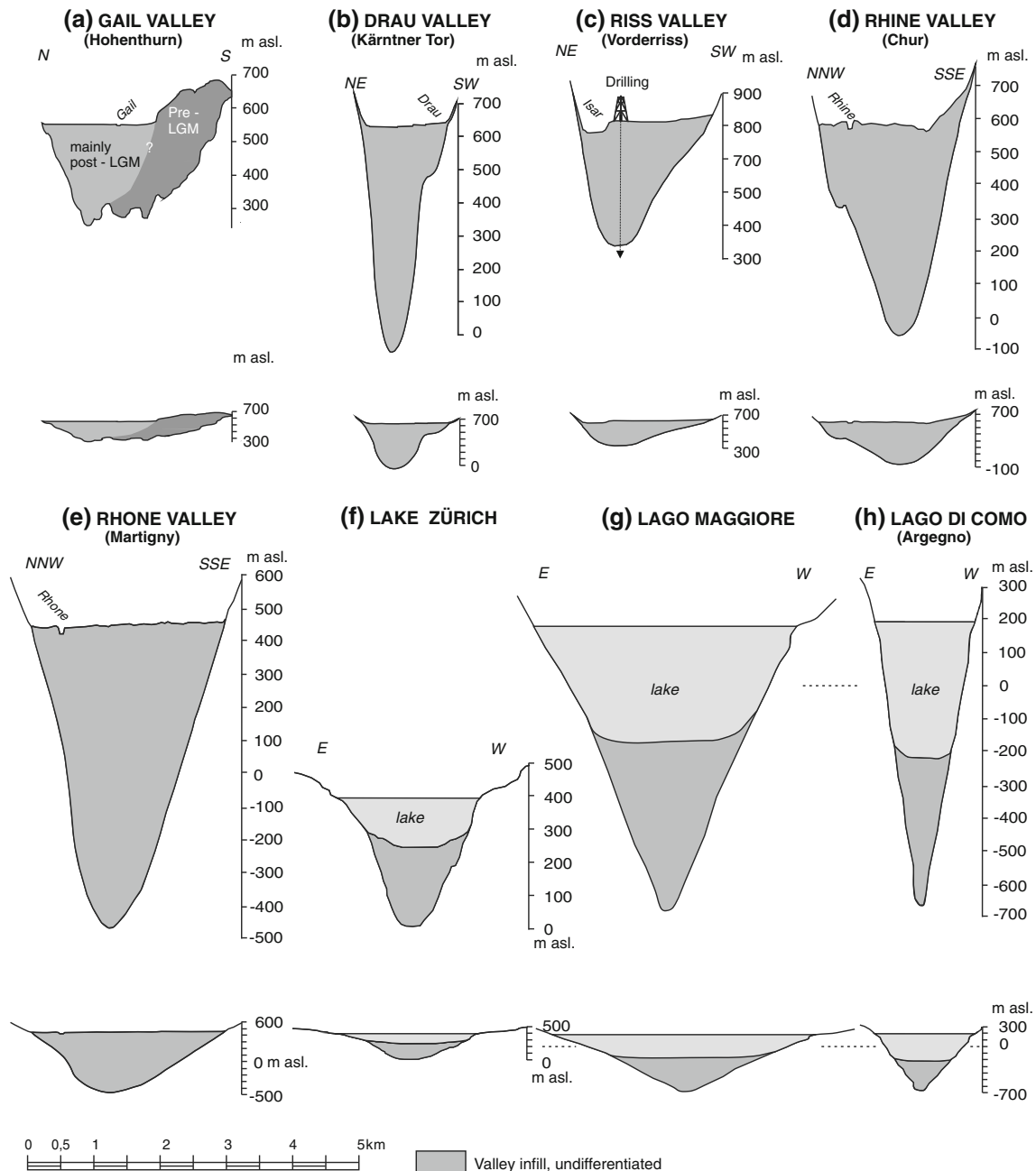


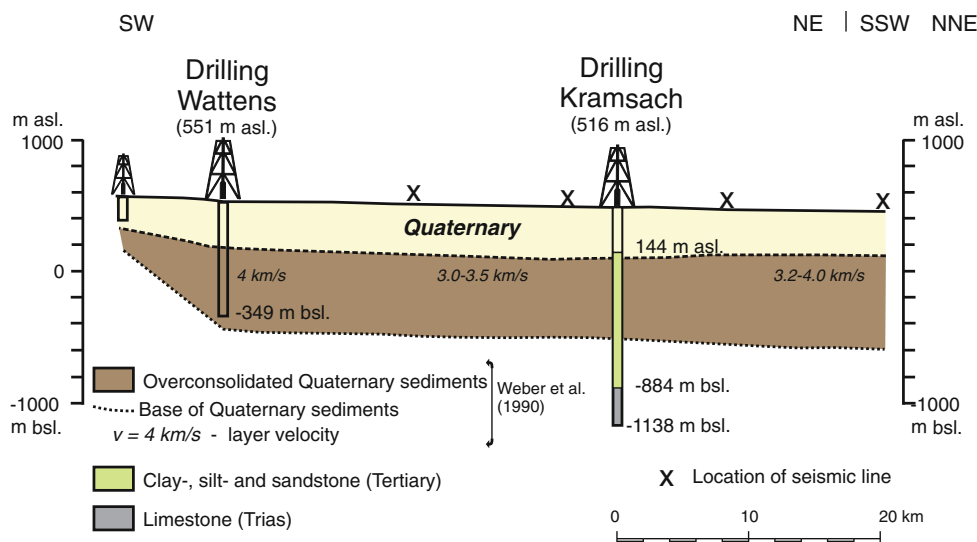
Fig. 5 Typical cross-sections of overdeepened alpine valleys shown with $\times 5$ vertical exaggeration and to scale modified after: **a** Carniel and Riehl-Herwirsch 1983; **b** Brückl et al. 2010; **c** Frank 1979; **d–e** Pfiffner et al. 1997; **f–h** Finck et al. 1984

An example in this context is the Inn Valley east of Innsbruck (Fig. 6), where seismic surveys indicate a depth of the bedrock surface of 1,000 m confirmed by the counter-flush drillhole near Wattens (Weber et al. 1990). Interestingly, the change to remarkably high velocities in sonic logs in the drillhole (up to $4,000 \text{ m s}^{-1}$) from 350 m downward has been related to Quaternary sediments (Weber et al. 1990). In accordance with a consistent reflection in the range between 300 and 400 m in the Inn Valley, this lower package was interpreted as “over-consolidated” Quaternary sediments.

An additional drillhole (Kramsach) showed that the base of the Quaternary is at a depth of 372 m, whereas Tertiary sediments reach down to 1,400 m (G. Gasser, pers. comm.).

Similar problems in differentiating Quaternary and pre-Quaternary rocks through geophysical surveys are evident in the Upper Inn Valley west of Innsbruck, where the top of “over-consolidated” sediments is at a depth of 400 m, whereas the supposed base of the Quaternary appears to occur between 700 and 900 m (Gruber and Weber 2004; Poscher 1993) (Fig. 6). With this in mind, the reported geophysically

Fig. 6 Longitudinal cross-section of the Inn Valley west of Innsbruck revealing the discrepancy between expected Quaternary infill deduced from seismic surveys and evidence of the counter-flush drilling of Wattens (*brown layer*, Weber et al. 1990) compared to the results of the spot-cored drilling of Kramsach (Gasser, pers. comm.)



estimated glacial erosion of 920 m in the most prominent tributary of the Inn Valley, the Ziller Valley (Weber and Schmid 1992), should be taken with caution. Examples similar to the Inn Valley are known from the Villach Basin with a “younger” sedimentary cycle on top of a thick package most probably Neogene sediments (Schmöllner et al. 1991).

The most extraordinary example of overdeepening with respect to its geometry and lithology is reported from Bad Aussee (NW Styria). In a narrow basin a minimum depth of the bedrock surface of 880 m has been revealed by drilling in an area made up by evaporitic bedrock (van Husen and Mayer 2007).

Central Alps (Rhine drainage system)

Lake Constance (Bodensee) is the most prominent glacial basin within the Rhine Glacier system (Müller and Gees 1968; Finckh et al. 1984). It is located just to the north of the alpine front in Molasse bedrock. While the outer limit of the LGM glaciation reflects a typical piedmont lobe of the Rhine glacier, the overdeepened part of the NW–SE oriented Lake Constance trough is apparently controlled by the tectonic setting (Schreiner 1979). This is in contrast to observations made further to the east, where overdeepening is oriented in the direction of the main glacier flow (e.g. Salzach glacier; Fig. 2).

Overdeepened structures can be traced from Lake Constance up-valley via Hohenems (Oberhauser et al. 1991; Schoop and Wegener 1984) and into the LGM accumulation area (Pfiffner et al. 1997) (Figs. 2, 7a, 5d). The overdeepened structure of Lake Walenstadt represents the course of an older Rhine Valley (Finckh et al. 1984; Müller 1995), down-valley joining the River Linth drainage (Wildi 1984), and finally leading into the

elongated basin of Lake Zürich (Hsü and Kelts 1984) (Fig. 7f).

Similar elongated basins cut into relatively weak bedrock (Molasse sediments) are present in the ablation area of the former Reuss Glacier with the fjord-like Lake Lucerne and the relatively shallow Lake Zug (Finckh et al. 1984). The overdeepened feature continues into the narrow depression of the Reuss Valley (Wildi 1984) (Fig. 2). A major overdeepened structure with the same orientation is found in the Glatt Valley Basin (Haldimann 1978) (Fig. 2). These and most other overdeepened structures in the foreland of the central Swiss Molasse Basin (e.g. Reuss and Linth glacier) strike perpendicular to the Alps. A prominent exception is the deep trough of Richterswil, which runs approximately parallel to the alpine front and connects the Lake Zürich Basin to the Reuss Valley (Wyssling 2002; Blüm and Wyssling 2007; Fig. 8). Some of the overdeepened valleys in the Central Swiss Midlands are present beyond the LGM (e.g. Birrfeld; Nitsche et al. 2001) and are evidence of multiphase glacial and fluvial erosion during the Quaternary. Interestingly, and in contrast to other areas, overdeepening in the foreland of the Swiss Alps is restricted to Molasse bedrock.

A sequence of deep basins in bedrock extending from the ablation area of the LGM glacier into the inner-alpine sector is evident in the Aare Valley (Schlüchter 1979; Pugin 1988). The amount of overdeepening varies between 605 m for Lake Thun (Finckh et al. 1984) to 250 m for the inner-alpine Gastern Valley (Schlüchter 1979) (Fig. 2). A prominent example of an overdeepened basin in a confluence situation in the Central Alps is found at Innertkirchen (Bernese Oberland; overdeepening ca. 120 m), where three glacier lobes joined during past glacial periods (Sustenletscher, Aareletscher, Gauligletscher).

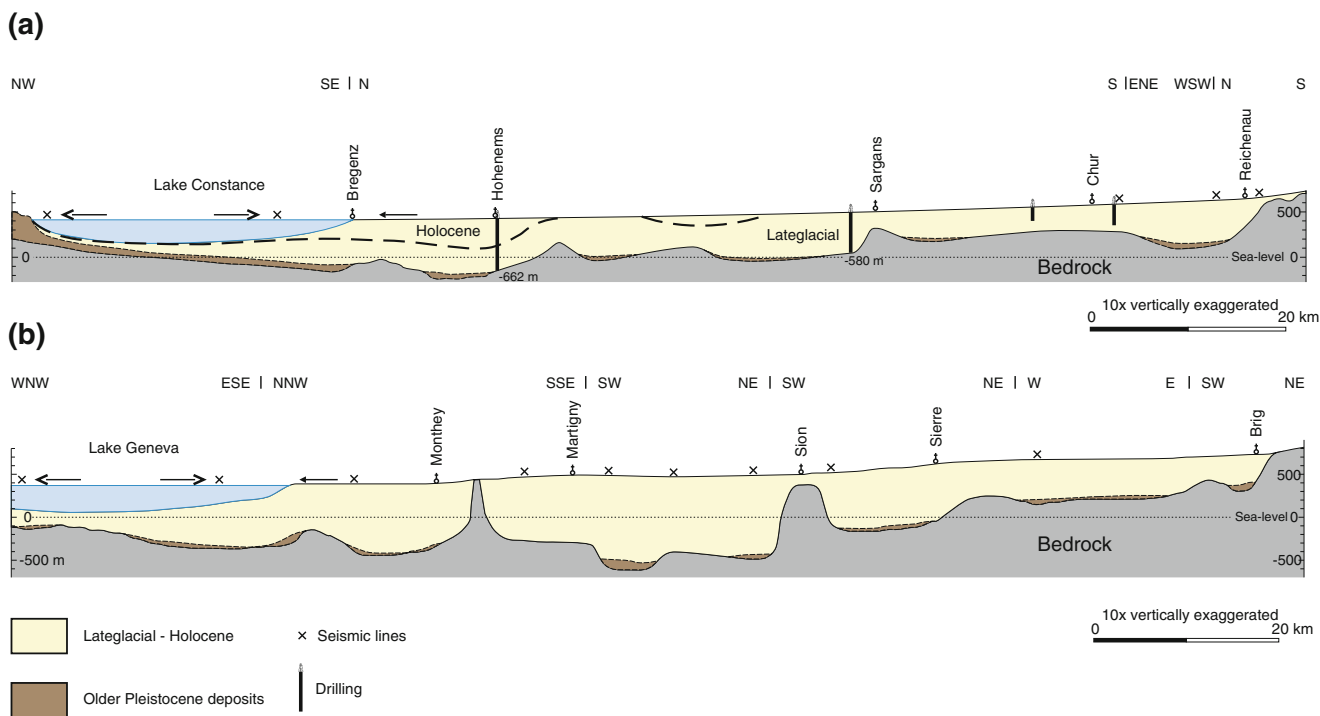


Fig. 7 Schematic longitudinal section through the Rhine and Rhone Valley (modified after Hinderer 2001)

Western Alps (Rhone drainage)

The most prominent overdeepened basin within the catchment of the River Rhone is Lake Geneva, located in the ablation area of the Valais Glacier (formerly often referred to as Rhone Glacier; Kelly et al. 2004) and cut into Molasse bedrock with a depth of the bedrock surface approximately 300 m below sea level (Finckh et al. 1984, Fig. 5e). Beside the Lake Geneva trough, some secondary troughs related to the Valais Glacier are found in the Swiss Plateau. The SW–NE trending lakes of Neuchâtel and Biel are further examples of basins within the Molasse zone. These lakes are situated at the northern edge of the Molasse Basin right at the foot of the Jura Mountains. The most important overdeepening is located in an old valley to the south of the Lake Neuchâtel/Lake Biel system (Fig. 2). The basin morphology is associated with the north-eastern branch of the former Rhone Glacier (Finckh et al. 1984), which in the Lake Neuchâtel/Lake Biel area flowed almost parallel to the Jura Mountains in a northeastward direction.

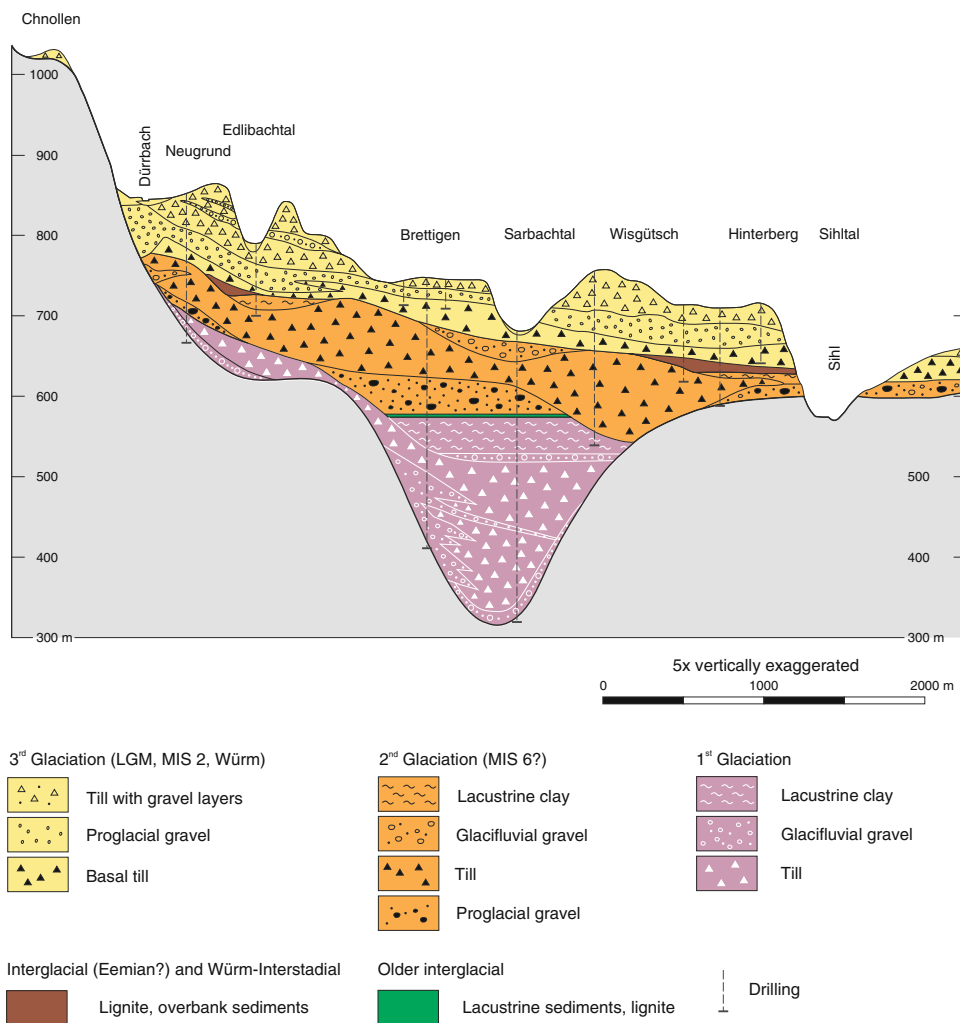
All valleys located between Lake Geneva and the Isère Valley that were glaciated during the LGM display overdeepened sections separated by bedrock ridges of hard limestone (Nicoud et al. 1987). Lac d'Annecy and Lac du Bourget (van Rensbergen et al. 1998, 1999) resemble glacial basins within the ablation area of the Isere Glacier. Major overdeepened features have been reported from the upper Isere and Durance Valleys (Fourneauux 1976, 1979).

The deepest of these basins are Lake du Bourget (bedrock surface approximately 110 m below sea level inferred from reflection seismic data; Finckh et al. 1984) and the Grenoble Basin (bedrock surface at least 177 m below sea level verified by drilling; Fourneauux 1976). Nicoud et al. (1987) attributed the basin fill, essentially consisting of lacustrine facies associations, to a series of different proglacial lake systems that formed immediately after the LGM ice collapse.

Southern Alps (Mediterranean drainage system)

The complex nature of the filling of alpine valleys to the south of the Alps was already recognised by Sacco (1888: cit. in Cadoppi et al. 2007). Later, it has been shown that most of the glacial basins of the former large piedmont glaciers of the River Po catchment area are indicated by deep lakes such as Lago di Garda (350 m water depth), Lago Maggiore (370 m) and Lago di Como (410 m) (Figs. 2, 5g–h). Seismic surveys show that the bedrock surface is up to 400 m below the bottom of the lakes (Finckh et al. 1984). The depth of the bedrock surface of ≥ 300 m below sea level and the mainly V-shaped nature of these overdeepened structures are interpreted as resulting from incision during the Messinian crisis (Finckh 1978; Bini et al. 1978). Based on seismic facies interpretations, the infill may comprise a substantial part of Neogene sediments, related to the marine transgression following the

Fig. 8 The complex sedimentary filling of the trough of Richterswil, central Switzerland, indicates deposition during at least four individual glaciations (modified after Wyssling 2002)



Mediterranean Sea low-stand (Pffiffer et al. 1997; Felber and Bini 1997). These overdeepened valleys resulting from the Messinian down-cutting extend to areas up-valley of the prominent lakes as well (Felber et al. 1998). However, the basin of Lago d’Iseo is apparently not of Messinian origin, and it appears that the morphology of the lake was mainly shaped prior to the last glaciation (Bini et al. 2007).

In comparison to other formerly glaciated catchment areas of the River Po, the reported depth of the bedrock surface of less than 200 m below surface within the Piave Valley, NE Italy (Pellegrini et al. 2004), and in the River Soča area, located in the Slovenian part of the Julian Alps (Bavec et al. 2004), is rather limited.

Approaches to constrain the age of overdeepening

There are several options to approach dating of glacial overdeepening, however, all of these are related to the

sediment filling of the erosional trough. As glaciers may have reached an area several times during the Quaternary and removed older sediments, this has often produced highly complex cut-and-fill relationships. As a consequence, all dating results reflect only the minimum age of initial bedrock erosion.

A combination of sedimentology and palynology allows to distinct between glacial and non-glacial sediments, thus providing information on whether a basin fill was caused by just one or by multiple glaciations. Palynology also permits identification of characteristic vegetation patterns that are specific to certain periods of the past. Probably the best suited palynostratigraphic marker in the present context is the Holsteinian Interglacial that is characterised by abundant *Fagus* (beech) and the last occurrence of *Pterocarya* (wingnut) in Central Europe (cf. Beaulieu et al. 2001; Tzedakis et al. 2001). Correlation of this interglacial with marine stratigraphy is still controversial, and the period either corresponds to MIS 9 (ca. 300 ka; Geyh and

Müller 2005; Litt et al. 2007) or MIS 11 (ca. 400 ka; e.g. Beaulieu et al. 2001).

Palaeomagnetism has the potential to establish chronological information for valley and basin infills (cf. Hambach et al. 2008). However, the previously often-used assumption of reverse magnetisation in Quaternary sediments representing an age below the Brunhes/Matuyama boundary is not robust, as it has been shown that (at least) 16 short-lived excursions of the magnetic field occurred during the past 700 ka (Lund et al. 2006).

The problem of numerical dating is that few methods reach beyond the last glaciation and in this age range are often in an experimental state. Well-established numerical dating methods such as radiocarbon, K/Ar and Ar/Ar are of limited use because either the deposits are too old or suitable material (i.e. tephra) is rare.

Luminescence dating allows determining the deposition age of silty and sandy sediments (cf. Preusser et al. 2008, 2009). The major limitation of this approach, apart from methodological uncertainties, is limited experience in dating sediments older than 150 ka. Initial studies imply that the method may be used up to 250 ka (Preusser et al. 2005; Preusser and Fiebig 2009), and possibly beyond. However, this will need to be validated by systematic methodological investigations.

A successful test in applying U/Th dating to Pleistocene sediments is presented by Spötl and Mangini (2006), who dated carbonate precipitates in gravel deposits near Innsbruck ("Höttinger Brekzie"). On the other hand, a study by Kock et al. (2009) on carbonate precipitates in gravel deposits from River Rhine terraces reveals overestimation of U/Th ages compared to luminescence dating and geological constraints. Apparently, this is due to methodological problems associated with carbonate precipitation caused by bacteria. If and to what extent U/Th methodology can be used to date carbonate precipitates in deeply buried sediments remains unknown, especially as the interaction with groundwater may cause open system behaviour (cf. Scholz and Hoffmann 2008). Another option would be U/Th dating of peat but this is also associated with several methodological problems (cf. Geyh 2008). In principle, U/Th dating can date back to 500 ka but its potential is still poorly explored in the context of dating deeply buried sediments.

An innovative approach in dating sediment burial is the use of cosmogenic nuclides ^{10}Be and ^{26}Al (Dehnert and Schlüchter 2008). However, while this method has been successfully used to date cave deposits in the Swiss Alps (Häuselmann et al. 2007), its reliability and accuracy for dating fluvial deposits from the alpine foreland leaves room for further improvement of the methodology (Häuselmann et al. 2007; Dehnert et al. 2010). Nevertheless, the potential in using cosmogenic nuclides for burial dating is important

although this would require some methodological breakthroughs.

Sedimentary filling of basins after the last glaciation of the Alps

The filling of glacial basins is controlled by factors such as the size of the basin and the relation between the main-river and its tributaries with respect to water and debris discharge (van Husen 2000). Large basins with a strong main-river and small tributaries often consist of thick layers of fine-grained bottom-sets intercalating with coarse-grained delta deposits (e.g. basin of Salzach Glacier, van Husen 1979). In contrast, the infill of valleys with strong tributaries appears more complex and inhomogeneous. In addition, the basal beds may show typical glacio-lacustrine features such as drop-stones bearing mud and diamicton ("water-lain till") as indicated in some drillholes (e.g. Lister 1984). In many cases sedimentation in glacially eroded basins started immediately with the down-wasting of ice masses, and evidence of (dead) ice contact such as contorted delta beds (e.g. van Husen 1985) and intra-formational faults are quite common. Additionally, basal gravel beds may resemble former meltwater channels (Pffner et al. 1997; Moscariello et al. 1998).

Mass movements of various scales are an integral part of valley fills, and re-sedimentation in the form of slumping has to be deciphered, a process that may lead to a repetition of sequences and thus stratigraphic pitfalls. Deep seated gravitational movements (i.e. 'sackungen') can result in substantial infilling (Brückl et al. 2010), which in some cases completely altered the glacially influenced shape of the valley and masked overdeepened valley sections (e.g. Zischinsky 1969). 'Sturzstrom' and other deposits of fast landslides are rarely recognised by drilling (Gruber et al. 2009) or in onshore seismic facies interpretations (Gruber and Weber 2004). However, high-resolution seismic imaging in lakes shows that mass movement deposits are important depositional processes in glacially overdeepened basins (e.g. Schnellmann et al. 2005).

It is shown that for the last glaciation of the Alps the above mentioned processes resulted in an early back-fill of most alpine valleys starting with the phase of ice decay at the very beginning of Termination I (c. 21–19 ka ago) (Klasen et al. 2007; Reitner 2007). In some inner-alpine valleys the final backfill of overdeepened basins started after the last glacial erosion event during the Gschnitz Stadial (c. 16 ka), and in most cases this process was completed before the Holocene (van Husen 1977, 1979). The reduced vegetation cover during the early phase of the Lateglacial is considered as a major precondition for high sedimentation rates at that time and the early infill of

troughs (Müller 1995). The progradation of the Rhine delta within the inner-alpine valley towards its modern position at the southern end of Lake Constance is the best example for ongoing infill (Eberle 1987).

Pre-LGM sedimentary fillings of overdeepened structures in the alpine region

Geophysical data imply a complex history of erosion and infilling of valleys (e.g. Bader 1979; Finckh et al. 1984; Pfiffner et al. 1997; Nitsche et al. 2001; Reitner et al. 2010). However, some relevant discrepancies between seismic interpretations and drillholes (Müller 1995; van Husen and Mayer 2007) indicate the importance of integrated studies combining geophysical methods and borehole/core investigations. Nevertheless, several records prove the existence of sedimentary fills older than Termination I (e.g. Fiebig 2003; Herbst and Riepler 2006). Many of these records feature lake deposits with pollen successions that are attributed to the Last Interglacial, the Eemian, as the equivalent of Marine Isotope Stage (MIS) 5e in the alpine region (cf. Preusser 2004). Examples for such archives are, from east to west, Mondsee (Drescher-Schneider 2000), Samerberg (Grüger 1979, 1983), Wurzach (Grüger and Schreiner 1993), Füramoos (Müller et al. 2003), Dürnten (Welten 1982), Niederweningen (Anselmetti et al. 2010), and Les Echets (Beaulieu and Reille 1984) (Fig. 2). The pollen in sediments from below the Eemian deposits

show a similar warming trend as in Termination I pollen records (e.g. Wohlfarth et al. 1994), implying an antecedent glacial period. As a consequence, it is often assumed that these basins were occupied and at least partly excavated by glaciers during MIS 6.

The Wolfratshausen Basin was formed by the former Isar-Loisach Glacier that was fed via transfluences from the Inn Glacier system. The general geometry of the basin and the successions of sediment, attributed to three major glaciations, indicate changes in glacial erosion (Fig. 9; Jerz 1979). The first glacial advance apparently formed the configuration of the basin, followed by a major down-cutting during the next glaciation. Glacial erosion during the youngest event, representing the Würm Glaciation (LGM; MIS 2), was limited (Fig. 9). The Gernmühler Basin (Samerberg) shows a similar structure, with the most severe erosion by a branch of the Inn Glacier during the glaciation prior to the Holsteinian (Grüger 1979, 1983). Erosion during the following glacial periods was less pronounced (Jerz 1983).

In the northern part of Switzerland, valleys are cut into the so called “Swiss Deckenschotter”, glaciofluvial sediments of Early Pleistocene age (Graf 1993, 2009; Bolliger et al. 1996). In the Birrfeld area, a basin outside the LGM limit, lithostratigraphy corresponding to the overall subsurface geometry shows four till beds with basal unconformities separated by glaciolacustrine sediments (Nitsche et al. 2001). The Richterswil trough is filled by sediments attributed to at least three glaciations (Fig. 8; Wyssling 2002; Blüm and Wyssling 2007). However,

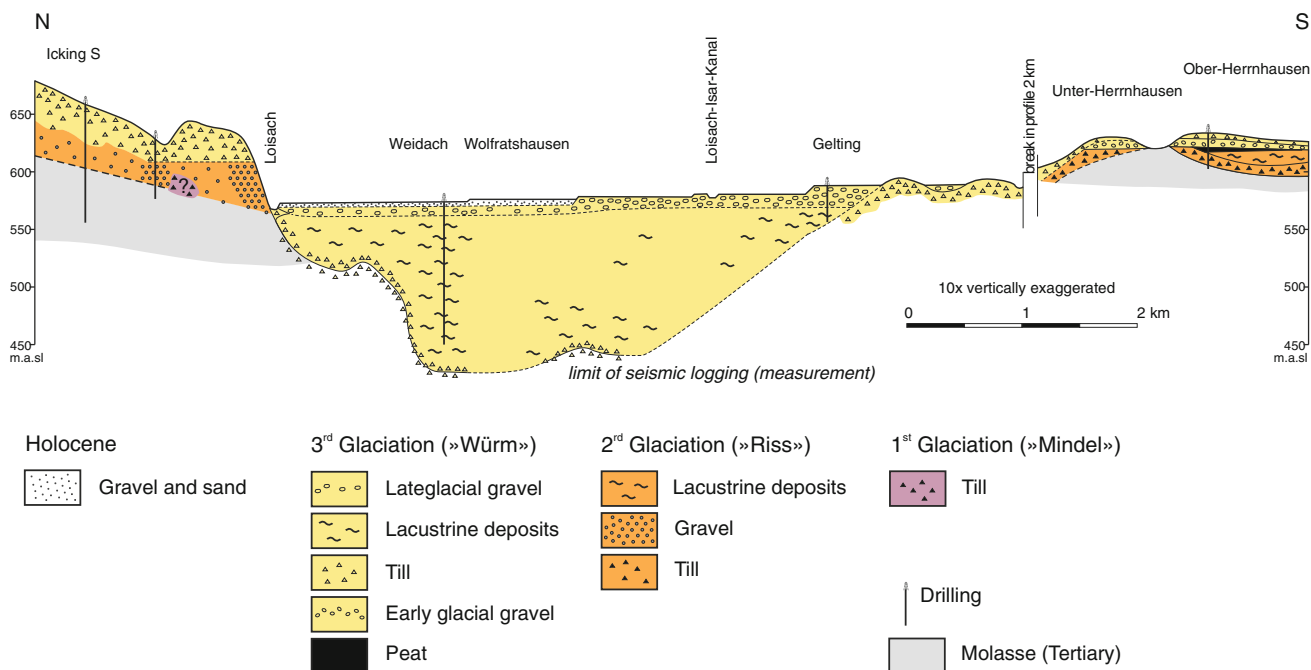


Fig. 9 Cross-section through the Basin of Wolfratshausen, Bavaria, revealing that the basin was occupied by at least three glaciations during the Quaternary (modified after Jerz 1979)

independent age control is missing in both these areas. In contrast, pollen analyses of the succession of Thalgut (Welten 1988; Schlüchter 1989a, b) shows deep erosion by a glacier that must be older than Holsteinian. At Meikirch, evidence is available for glacial erosion just prior to MIS 7 (Preusser et al. 2005).

The Ecoteaux Basin resembles two different sediment successions each beginning with subglacial sediment and the whole sequence being covered by LGM till (Pugin et al. 1993). The pollen record of the lowermost sequence, located on top of a till, indicates formation during an interglacial older than Holsteinian. The palaeomagnetic signal with an inverse polarity was interpreted as part of the Matuyama Epoch, but this interpretation should be treated with caution (see discussion above). Interestingly, based on cosmogenic nuclide burial dating, Haeuselmann et al. (2007) deduced an increase in erosion in the inner-alpine part of the Aare Valley around 0.8–1.0 Ma, probably caused by rapid glacial incision. This important break in morphogenetic conditions coincides with the change in periodicity of glacial cycles from 41 to 100 ka (Haeuselmann et al. 2007), commonly referred as the Middle Pleistocene transition (cf. Head and Gibbard 2005).

In the River Po drainage area the oldest dated sedimentary successions of glacial origin on top of marine deposits of Early Pliocene age (Zanclean) have been described from the Lago di Varese area (cf. Bini and Zuccoli 2004). A carbonate cement of conglomerates of glaciofluvial origin (Ceppo dell' Olona unit) gives a minimum age estimate of 1.5 Ma (Bini 1997). Based on the maximum age of underlying sediments the two lowermost till beds are attributed to the Early Pleistocene (MIS 96–100; Uggeri et al. 1997). However, based on the magnetostratigraphic record in the perialpine Po Basin, as well as pollen and sequence stratigraphy, the onset of major glaciations and thus sediment excavation occurred in MIS 22 (0.87 Ma; Middle Pleistocene transition) (Muttoni et al. 2003).

Deep glacial erosion outside the alpine realm

For an understanding of deep glacial erosion it is important to note that this phenomenon is not limited to the Alps, but is found in many other regions that were glaciated during the Quaternary such as North America, Northern Europe, and the British Isles. Interestingly, there is also evidence for the formation of deep erosional channels during pre-Quaternary glaciations (cf. Le Heron et al. 2009), for example, for the Late Palaeozoic (Carboniferous–Permian) glaciation that covered the West Australian Shield (Pilbara Ice Sheet) (Eyles and de Broekert 2001).

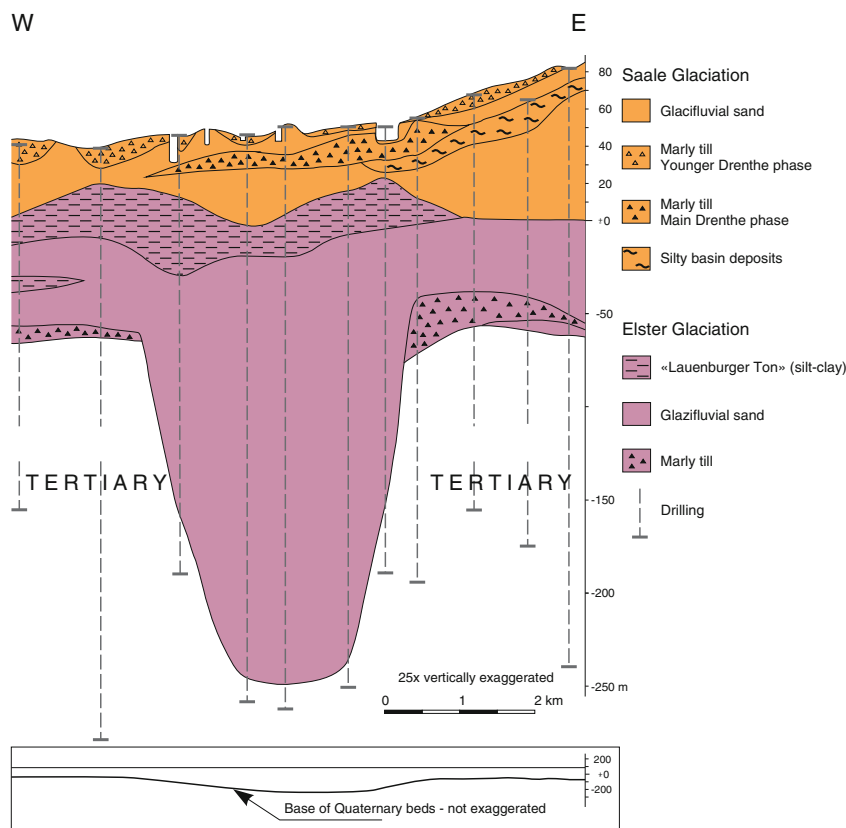
The southern margin of the former Scandinavian Ice Sheet, especially in Northern Germany and in Denmark, is

particularly well investigated with regard to this issue. In this region of Northern Europe, buried valleys cut into Tertiary and Quaternary sediments, often referred to as tunnel valleys, are widespread (cf. Huuse and Lykke-Andersen 2000; Kluiving et al. 2003; Jørgensen and Sandersen 2009; Lutz et al. 2009). The channels are up to 500 m deep, in some cases more than 150 km long (cf. Ehlers et al. 1984; Huuse and Lykke-Andersen 2000; Stackebrandt 2009), and show to some extent cross-cutting relationships (Kristensen et al. 2007). Palynostratigraphy indicates that the oldest phase of tunnel valley formation occurred prior to the Holsteinian (Fig. 10). It is hence usually assumed that the first and most pronounced glacial overdeepening occurred during the Elster Glaciation.

Most authors assume that subglacial meltwater was responsible for the formation of tunnel valleys (e.g. Grube 1979; Hinsch 1979; Kuster and Meyer 1979; Ehlers et al. 1984; Habbe 1996; Huuse and Lykke-Andersen 2000). Mooers (1989) suggests that the dominant source of the water responsible for tunnel-valley formation along the Southern Laurentide Ice Sheet was seasonal meltwater from the glacier surface that reached the bed through moulins and crevasses. The apparent continuity of the valleys resulted from the headward development of the englacial drainage system during ice retreat. By contrast, Hooke and Jennings (2006) propose that the formation of tunnel valleys was caused by catastrophic releases of meltwater that were produced by basal melting and stored for decades in subglacial reservoirs at high pressure. Piotrowski (1994) advocates that at the time of the first Weichselian ice advance, a large subglacial water reservoir developed in the area of the Baltic Sea Basin and caused a rapid, surge-like ice movement. As the ice sheet advanced out of the Baltic Sea Basin, drainage of the water reservoir was prevented by the ice toe overriding the permafrost on the Saalian highlands. During the ice retreat, frozen ground was left beyond the ice margin and subglacial meltwater catastrophically drained through the tunnel valleys.

Sandersen et al. (2009) demonstrate that the incision of tunnel valleys in Vendsyssel, Denmark, that reach 180 m below sea level, was related to the main Weichselian advance and a later re-advance of the Scandinavian Ice Sheet. Luminescence dating of sediment in which the valley was eroded and of its sedimentary fill (Krohn et al. 2009) shows that nine generations of individual tunnel valleys formed between *c.* 20 and 18 ka ago. Thus, the formation of each individual tunnel valley must have occurred in less than a few hundred years. Sandersen et al. (2009) suggest that the process behind tunnel valley formation are repeated out-bursts of meltwater that eroded narrow subglacial channels. With the decrease of water pressure after each outburst, the channels were closed by ice or re-filled with glacial sediments and gradually broadened and widened by glacial erosion.

Fig. 10 Cross-section through the “Wintermoorer Rinne” near Buxtehude, northern Germany, as an example of a tunnel valley formed during the Elster Glaciation at the southern margin of the Scandinavian Ice Sheet (modified after Kuster and Meyer 1979). The “Lauenburger Ton” unit is, at its type location, overlain by deposits of the Holsteinian (Meyer 1965)



Potential processes of deep erosion in the alpine region

The effects of tectonic movements on deep erosion have been discussed since the early days of research, but theories on the genesis of peri-alpine lakes based on large-scale uplift (Heim 1919) are nowadays considered unlikely. Nevertheless, tectonic analyses, in particular in the Eastern Alps, show that the modern setting is quite similar to that of the Miocene with still active strike-slip faults (Plan et al. 2010), potentially enabling ongoing pull-apart basin formation along longitudinal valleys. But beyond some assumptions on Quaternary graben formation in the Inn Valley (Ortner 1996; Ortner and Stingl 2003) there are no hard facts supporting such direct tectonic hypotheses for the formation of overdeepened features. However, particularly in inner-alpine regions, rivers often follow tectonic lineaments (higher erodability) and therefore tectonics played an important role in predetermining the flowpath of the ice.

Examples of peri-alpine lakes and valleys in the River Po area, which were deeply incised during the Messinian event and then partly refilled during the Pliocene, show a clear fluvial morphological signature in the lower part of the valley cross-sections (Piffner et al. 1997). Glacial erosion, expected to be the major driving factor behind

deep incision of valleys and basins did not, in these cases, reach bedrock, either vertically or laterally.

Our understanding of the mechanisms of glacial erosion is limited by the restricted access to the subglacial domain. According to theoretical and empirical studies, abrasion and quarrying are regarded as the main mechanisms of subglacial erosion, with their rates increasing with sliding speed or higher abrasion coefficients (Iverson 1995; Menzies and Shilts 1996).

Penck (1905) observed a principal connection between the position of glacial basins and the location of the Equilibrium Line Altitude (ELA) during past glaciations, where the highest ice flow velocities occur. He proposed that the cross-section of a glacial valley is naturally adjusted to allow the movement of the ice provided by the mass balance up-valley. Thus, the depth and width of a glacial trough must increase to a maximum around the ELA and decrease to zero at its terminus. However, in the case of glacial basins the highest flow velocities at the ELA are most probably accompanied by basal debris rich ice (van Husen 2000).

Hooke (1991) highlights in particular the role of quarrying, triggered by changes in subglacial water pressure. According to this model, a minor convexity in the glacial bed results in crevassing at the glacier surface, leading to a localised water input and thus to subglacial erosion.

As erosion acts on the down-glacier side, further progress results in an amplified convexity, followed by more crevasse leading to a positive feedback at the head of the overdeepening.

The typically gentle reverse slopes of glacial basins are regarded to result from the balance between erosion and sedimentation (Alley et al. 2003). If the subsurface reverse slope is substantially steeper than that of the glacier surface, subglacial water with temperatures near the melting point freezes, and sediment transport stops due to supercooling and freezing of subglacial meltwater. Resulting till sedimentation may in turn cause a reduction of the reverse subglacial slope angle, enabling again transport by water.

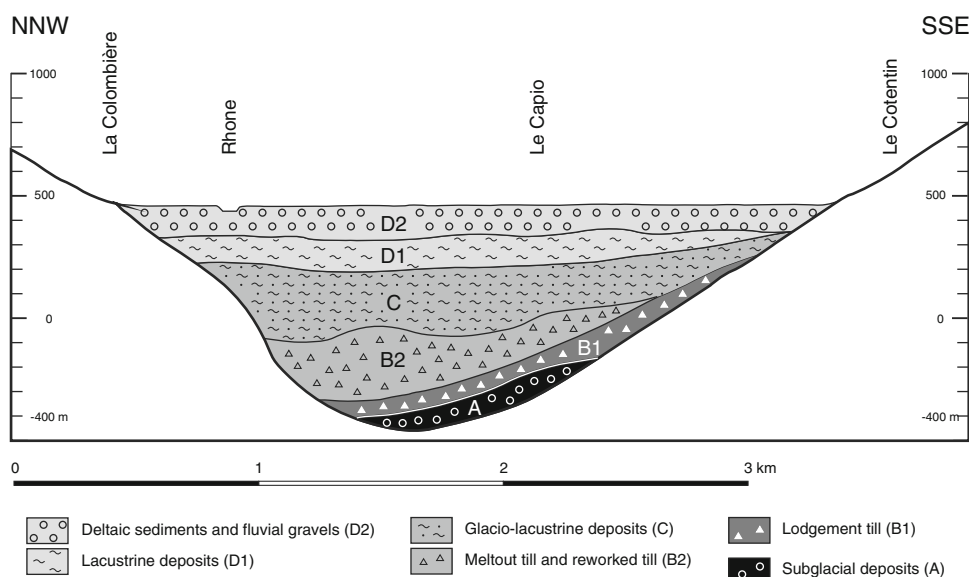
Based on the evaluation of 3D seismic and geophysical borehole data, Praeg (2003) and Kristensen et al. (2008) suggest that supercooling resulted in ice-marginal basal refreezing of subglacial meltwater, causing substantial erosion of sediment along subglacial channels. These authors hypothesise the possibility of large-scale sediment erosion, transport and deposition leading to the formation of tunnel valleys and their fillings by subglacial “conveyor-belt” systems. While many parts are filled by sediment during the formation of the tunnel valleys, the remaining open parts of the troughs will be filled with glaciofluvial and glaciolacustrine deposits.

Penck (1905) applied his principal rule for glacier confluences for valleys within the former accumulation area. It is argued that such a situation should have first increased the sliding velocity, which finally resulted in an increase of cross sectional area of the glacier, in order to accommodate the added discharge of ice. The plausibility of such a theoretical approach has been tested by modelling the long profile of glacial valleys (Anderson et al.

2006). In addition models iteratively show how the transformation from an initial V-shaped to a finally U-shaped valley can occur (Harbor 1992). The role of erosion by subglacial pressurised meltwater action is a matter of discussion in the formerly glaciated Alps, but is often considered to be of only local importance (Iverson 1995). Prominent examples of narrow channels formed by local subglacial erosion are the gorges of the River Aare (Müller 1938; Hantke and Scheidegger 1993) and River Emme (Haldemann et al. 1980). However, geometries similar to tunnel valleys are found in many areas of the Alps, for example, the trough of Richterswil (Wyssling 2002), linear depressions in the ablation area of the former Salzach Glacier (Salcher et al. 2010), and small channel structures in the bedrock topography of the Gail Valley (Carniel and Riehl-Herwirsch 1983; Fig. 5a). The presence of gravel deposits between basal tills and bedrock (Hsü and Kelts 1984; Pfiffner et al. 1997; Fig. 11) further confirms that the action of sediment-laden turbulent subglacial meltwater was probably of major importance in the overdeepening of valleys and basins.

Another important factor controlling erosion is the lithology of the subglacial bed. Most of the pre-existing valleys in the Alps follow faults or other tectonic structures, where bedrock lithology is partly fractured by joints and hence prone to quarrying. Nevertheless, overdeepened structures exclusively linked to the occurrence of weak lithology are exceptional within the Alps, such as the “deep hole” (>880 m) in the upper Traun Valley (van Husen and Mayer 2007). The highest degree of deep glacial erosion reported from Martigny (Pfiffner et al. 1997) may be explained by a combination of a tectonically weakened zone and ice confluence.

Fig. 11 Complex basin sequence of the Rhone Valley near Martigny. The presence of gravel deposits between basal tills and bedrock is interpreted to indicate the action of sediment-laden turbulent subglacial meltwater (modified after Pfiffner et al. 1997)



Conclusions

Overdeepened valleys and basins are common features in the alpine region. In summary, overdeepened features are related to the following situations:

- inner-alpine longitudinal valleys pre-defined by tectonic structures and/or weak lithologies
- ice confluence and partly also diffluence situations in inner-alpine locations
- elongated buried valleys, mainly oriented in the direction of former ice flow
- glacially scoured basins in the ablation area of glaciers
- valleys generated during the Messinian crisis (the Mediterranean Sea low-stand)
- high erodability of lithology

In detail, glacier flow, as controlled by mass balance and ice dynamics (confluences and diffluences), is considered a major control in the formation of overdeepening. The occurrence and morphology of buried depressions is often the result of more than one glacial cycle, as shown by the lithostratigraphic record and dating evidence. At least some glacially scoured basins and troughs were repeatedly occupied and excavated during most major glaciations since the Middle Pleistocene. The increase in inner-alpine valley incision, as shown by Swiss caves and by sediment supply to the River Po Basin, exemplifies major glacial erosional events. These events were probably of relatively short duration but glacial erosion rates are expected to exceed those of fluvial action by several magnitudes (Hallet et al. 1996). In addition, mass movements, as a result of glacial over-steepening of slopes, at least facilitated removal of large broken-up masses by the following glaciation.

The time-transgressive nature of overdeepening is evident far upstream from the position of the ELA during the LGM in the inner-alpine valleys. Breaks in the subsurface topography (basin-and-riegel morphologies) do not only indicate confluences but may also point to recurrent positions of the glacier front during phases of limited glaciation. The palaeogeography during Termination I, in particular during the Heinrich I event, may define such a situation, which may have occurred more often in the sense of average glacial conditions (Porter 1989), but at least during multiple phases of ice build-up as well as ice decay. This demonstrates the importance of the full range of climatic conditions in understanding subsurface topography.

Overdeepened features along the Scandinavian and Laurentian Ice Sheets were likely formed by outbursts of subglacial meltwater within very short periods of time (maximal a few hundred years; Krohn et al. 2009). Although subglacial erosional features are also clearly identified in the alpine area, present knowledge does not

yet allow a final judgement of whether this process was also the major factor in the formation of overdeepened valleys and basins. Inspiration for future research is provided by the simple fact that the origin and age of overdeepened valleys and basins remains a central and unsolved problem in the evolution of alpine morphology.

Acknowledgments We thank Helene Pfalz-Schwingsenschlögl for drawing most of the figures in this manuscript. Andreas Dehnert, Christoph Janda, and Johannes Reischer provided technical support to create Fig. 2. Information on the location of overdeepened features was kindly provided by Gerhard Doppler and Ernst Kroemer for Bavaria, and by Giovanni Monegato for Northern Italy. The authors are indebted to Urs Fischer, Wilfried Haerberli, Mads Huuse, Oliver Kempf, Sally Lowick, John Menzies, and Michael Schnellmann for their constructive comments and suggestions on earlier versions of the manuscript.

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