Swiss J Geosci (2010) 103:329–344
DOI 10.1007/s00015-010-0045-x

Deep alpine valleys: examples of geophysical explorations in Austria

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Received: 31 December 2007 / Accepted: 5 October 2009 / Published online: 4 December 2010

Abstract Results from geophysical explorations of three deep valleys, selected from different tectonic regimes in the Eastern Alps (Ötz-, Oichten-, and Drau Valley), are presented and discussed. Ongoing tectonic deformation may use tectonic structures related to these valleys. However, seismic activity is low there. During the Würm ice age, the thickness of the ice cover ranged between 300 and 1,500 m above present ground elevation. The geophysical investigations comprised reflection seismology, gravity- and resistivity surveys. The maximum depth down to the erosional base of the valleys varies from ~340 to 700 m. Distinct layer packages of the valley-infill at depths greater than 250 m were termed "old valley-fill". Geophysical parameters and a comparison with the reflection seismic image of an intermediate layer at the recent Pasterze glacier suggest that the top of the "old valley-fill" represents the glacier bed during the decay of the Würm glaciation. Deep erosion is not related to high basal shear stress. The confluence of tributary glaciers is apparently not a significant factor for deep erosion in our examples of deep alpine valleys. We conclude that deep erosion may be related to high water pressure at the glacier bed, supported by specific processes of tectonic weakening.

Keywords Deep alpine valleys · Geophysical explorations · Austria

Editorial handling: Kurt Decker, Stefan Bucher.

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Introduction

Deep valleys are dominant morphological features of the European Alps and other young orogens. They have been incised into lithologically differentiated and tectonically pre-stressed rock mass by fluvial erosion. The shaping of valley cross sections from "V" to "U" occurred predominately through large ice streams during the ice ages. Troughs and basins have been formed by overdeepening, a process connected to glacial erosion. In our study we relate the term "deep" to the degree of overdeepening and the thickness of the present day valley-infill above the erosional base. This implies that we consider examples covered by ice during the ice ages. Deep or overdeepened alpine valleys in the Eastern Alps are, from east to west, the valleys of the rivers Mur, Enns, Traun, Drau, Salzach, Inn, and Etsch (van Husen, 1987, 2000). The main overdeepened valleys in the Central and Western Alps follow the rivers Rhine, Rhone, Isére and Romanche (van der Beek and Bourbon, 2007). The thickness of the present day valley infill reaches several hundreds of metres and the erosional base may be below sea-level in some places (Weber et al. 1990; Weber and Schmid 1991; Besson et al. 1991).

Deep alpine valleys in the sense described above are of great economic and technical importance. Fresh water resources are beginning to dwindle even in alpine areas and groundwater protection is an increasing issue. Deep alpine valleys offer the general possibility of large aquifers protected by sufficiently thick sealing sediment covers (e.g. Tentschert and Schönlaub 1996; Brandecker 1974). Geotechnical projects, like tunnels and galleries crossing valleys, need accurate information about erosion depths and the type of valley fill (Kovári and Fechtig 2004). One well-known example of a disaster due to an unexpected deep valley-fill happened during the construction of the Lötschberg tunnel in 1913 (e.g. Jaeger 1972). Stability considerations of high valley flanks must take into consideration not only the present day slope, but also its continuation below the valley sediments (e.g. Zischinsky 1969). Deep alpine valleys are also of interest for basic geological, geomorphologic and geodynamic research. In many cases they are in close connection to large-scale fault systems. Examples for such fault systems (mainly strike-slip) in Austria are the Inntal-fault, the SEMP (Salzach-Enns-Mariazell-Puchberg), the Mur-Mürz-faults, and the PAL (Periadriatic Lineament) with accompanying faults. These faults have been active since the late Oligocene within an ongoing escape process of crustal patches to the Pannonian basin (Ratschbacher et al. 1991; Linzer et al. 2002).

Geophysical investigations were already carried out using seismic and gravimetric methods in several deep alpine valleys. Examples from Austria are the Inn Valley (Aric and Steinhauser 1977; Weber et al. 1990; Gruber and Weber 2003), the Enns- and Traun Valley (Weber and Schmid 1991; Schmid and Weber 2005; Schmid and Weber 2007). In the Western Alps, De Franco et al. (2006) recorded seismic reflection profiles in Italy and Kissling and Schwendener (1990) determined the sedimentary fill of Alpine valleys by gravity modelling in Switzerland. A high resolution seismic cross section of the Rhone Valley was described by Finckh and Frei (1991), and interpreted by Besson et al. (1993). Pfiffner et al. (1997) summarize and discuss incision and backfilling in the Rhone and Rhine Valleys and in the Ticino region.

During the last decade, the authors carried out geophysical investigations of deep alpine valleys in Austria dedicated to hydrological studies, landslide research, and field exercises with students. Characteristic examples of these investigations will be presented in order to analyse the capacity of the applied geophysical methods, and to discuss the results with respect to morphology of the rock basement and evolution of sedimentation. All examples will be taken from an area covered by ice during the last ice age and glacial erosion will be a special issue. As an instructive comparison, seismic profiles from a recent alpine valley glacier will be shown and included into the discussion.

Characterization of the selected examples of deep alpine valleys

In this study we present the Ötz-, Oichten- and Drau Valley as examples of deep alpine valleys in Austria. Their location is shown on the tectonic map (Fig. 1), which has been generalized after Oberhauser (1980) and Schmid et al. (2004). Tectonics and geology are significantly different

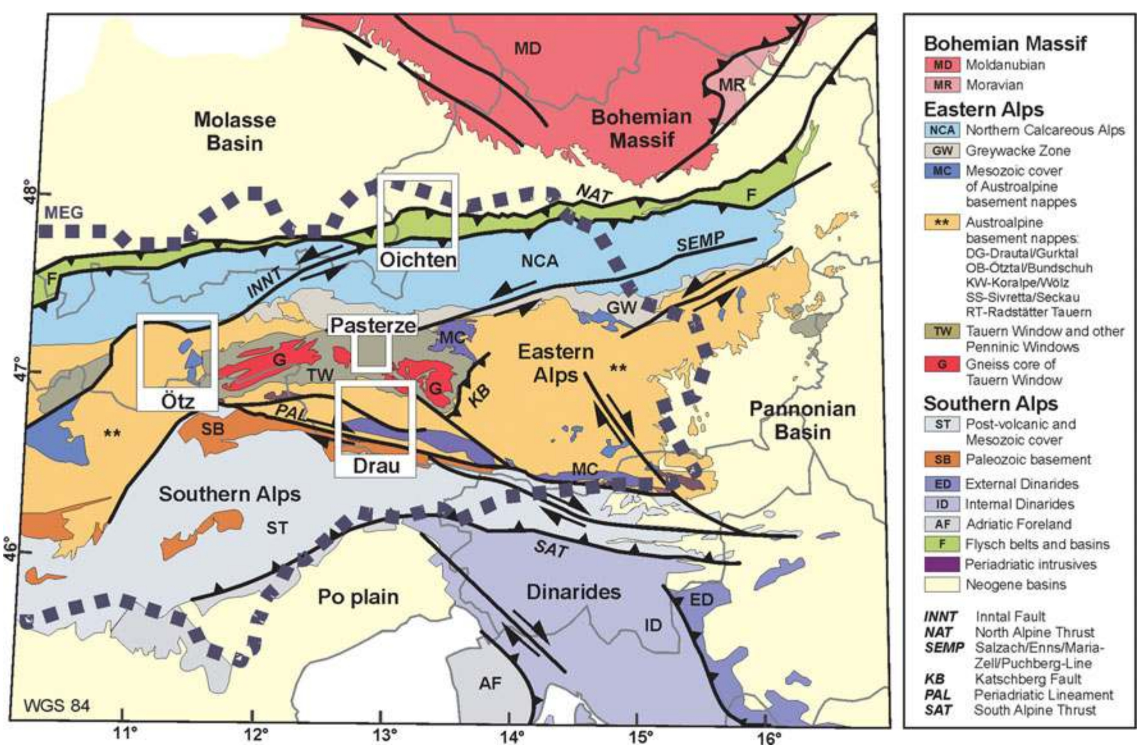

Fig. 1 Tectonic map of the Eastern Alps. Locations of selected examples of deep alpine valleys (Ötz-, Oichten-, and Drau Valley), and Pasterze glacier are marked by *frames*. *Dotted blue line* shows maximum extent of glaciation (*MEG*) during last ice age (Würm)

for these three examples. The Ötz Valley is S–N oriented and located in an extensional regime parallel to the Brenner normal fault. The rock mass consists mainly of ortho- and paragneisses of the Ötztal nappe. The Oichten Valley is in a mainly compressional regime at the North Alpine Thrust fault (NAT), where the Flysch belt overthrusts the Molasse of the Alpine foreland. The part of the Drau Valley under consideration is bounded by the calcareous units of the Drau nappe in the south and crystalline Austro-Alpine nappes in the north. It is located about 8 km north and subparallel to the PAL (see Fig. 1).

Seismic activity as an indicator of active tectonics is low or below detection level in or near the investigated areas. Epicentres and intensities (I) of felt and/or recorded earthquakes since 1200 are compiled in the map of Zentralanstalt für Meteorologie und Geodynamik, Wien (Erdbeben in Österreich–ein Überblick; http://www.zamg.ac.at/). According to this map, we had three $I = 4$ earthquakes in the investigated area of the Ötz Valley and three $4 < I < 5$ earthquakes in the investigated area of the Drau Valley. No earthquakes are reported from the Oichten Valley. Active tectonics are manifested through crustal deformations, which can be resolved by GPS. According to Grenerczy and Kenyeres (2006) current N–S convergence rates in the Eastern Alps between Europe and the Adriatic micro-plate are 2.3 ± 0.3 mm/year. The extrusion and escape process to the Pannonian basin is an ongoing process with displacement rates of 1–1.8 mm/year. A recent analysis of GPS data in Austria confirms this general trend (Haslinger et al. 2007). However, differential movements at prominent faults cannot be resolved up to now, because of the low density of stations and the relatively short observation periods of these GPS-campaigns.

According to the map of the glacial peak elevations during the last ice age (Würm) (van Husen 1987) the ice surface had an elevation of 2,500 m in the Ötz Valley, 700 m in the Oichten Valley and 2,000 m in the Drau Valley. The present day valley floors have elevations of 1,100–950 m (Ötz), 440 m (Oichten) and 630–610 m (Drau). Maximum ice flux occurs near the equilibrium line at the transition from the accumulation to the ablation area. During the last century this line was at an elevation of 2,600–3,100 m (e.g. Gross et al. 1977; Ohmura et al. 1992) in the Alps, with significant tendencies to higher elevations since 2000 (e.g. Zemp et al. 2006; Prinz 2007). During ice ages, the altitude of the equilibrium line (ELA) dropped in the Western Alps down to about 1,800 m (van der Beek and Bourbon 2007). Estimates by the balance ratio method (e.g. Meier and Post 1962; Benn and Evans 1998) based on the map of the ice surface during the last ice age (van Husen 1987) suggest a significantly lower elevation of the equilibrium line for the Eastern Alps. Therefore, during the maximum extent of the last glaciation (Würm) the investigation areas in the Ötz Valley and Drau Valley were well above the equilibrium line in the accumulation area and the Oichten Valley in the ablation area of the foreland, near the front of the most extensive glaciation.

Methodology of geophysical exploration

The main exploration tool of the deep alpine valleys presented in this paper was a combination of gravimetric investigations and reflection seismology. Resistivity sounding and profiling resulted in 1D-profiles or 2D-cross sections of the specific resistivity and brought additional information about the upper part of the valley-fill. These resistivity surveys followed standard procedures for 1D-sounding and 2D-profiling, described in many text books (e.g. Keary et al. 2002). We therefore omit a comprehensive description of our application of resistivity sounding and profiling. The combination of gravity measurements and reflection seismic studies has also been successfully applied for many exploration targets. Because of its effectiveness for the exploration of deep alpine valleys and some specific aspects we will describe this combination for the Drau Valley, where it was used most extensively (Brückl and Ullrich 2001; Ullrich and Brückl 2004).

The first step was the generation of a map of Bouguer gravity of the investigation area. Gravimetric measurements were carried out with a Scintrex CG3 relative gravimeter at about 1,150 locations, arranged along several profiles covering the whole investigation area of 12 km^2. The gravimetric campaign included the calibration of the instrument at a calibration profile. Instrument drift was determined by repeated measurements at some locations and tidal corrections are programmed in the instrument. Free air anomaly was calculated by subtracting normal gravity at the ellipsoid using Somiliana's formula and consideration of the free air gradient using ellipsoid heights (e.g. Torge 1989). Coordinates of the measurement points were determined by GPS positioning, their elevations were determined by levelling and taking into account geoid undulations (e.g. Erker et al. 2003). The effect of topographic masses was calculated by the use of a digital terrain model, with a grid spacing of 25 m (Bundesamt für Eich- und Vermessungswesen, Wien) in the near field (up to 40 km distance) and 250 m (Institute of Photogrammetry and Remote Sensing, Vienna University of Technology) from 40 to 80 km. A constant density of 2,670 kg/m^3 was assumed and the calculations were carried out by the program of Götze and Lahmeyer (1988). Subtraction of the gravity effect of the topographic masses and a systematic linear local trend from the free air gravity resulted in the (residual) Bouguer anomaly map shown in Fig. 2.

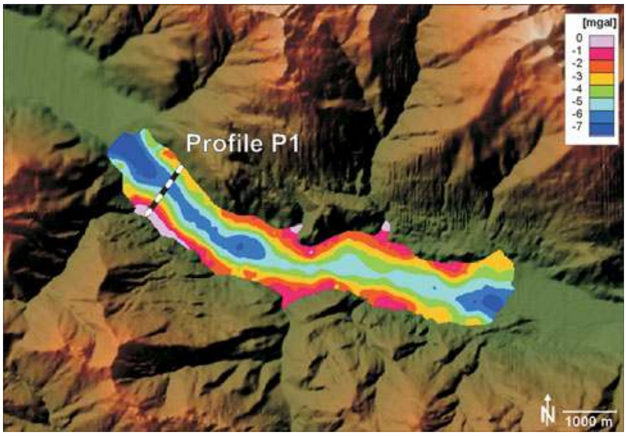

Fig. 2 Bouguer gravity map of the investigation area in the Drau Valley. Gravity data from profile P1 are used for integrated interpretation shown in Fig. 3

The Bouguer gravity map shows the gravity effect of the low density valley-fill and its varying thickness. It may be taken as a qualitative image of the topography of the basement and can support the selection of the most informative seismic profiles. For deep alpine valleys with steeply dipping valley flanks, reflection seismic cross-sections oriented approximately perpendicular to the strike of the valley are most appropriate. The reflection method is used because it is the only surface seismic method which can resolve the topography of the rock basement of a deeply incised valley, and the orientation as cross-section is chosen, because this is the best approximation of a 2D case and seismic rays remain nearly in a vertical cross-section. On the contrary, a longitudinal section would catch rays from the valley floor nearly vertically below, but also from the left and right valley flank and a unique interpretation would be difficult or even impossible. Seismic data acquisition was carried out with a 96-channel OYO-DAS instrument and an accelerated weight drop seismic energy source (VAKIMPAK) (Brückl 1988). Single 30 Hz geophones were used per channel with a spacing of 6 m. Seismic waves were generated at each geophone point with an inline offset of 3 m. The roll-along technique was applied and a nominal 48-fold CMP-coverage was achieved. Reflection seismic processing followed mainly the standard CMP-processing routines (deconvolution and filtering, CMP sorting, static corrections, velocity analysis, dynamic corrections and stacking, post stack migration) described in many text books (e.g. Yilmaz 2001). Static corrections were determined from a refraction seismic evaluation of first arrivals. Simple CMP-stacking may be inadequate, because reflections from the steeply dipping valley flanks and flat reflectors within the valley-infill cause the so-called "stacking conflict". We applied the dip moveout technique to overcome this difficulty. These considerations were relevant for the Drau Valley. In the Ötz and Oichten Valley standard CMP processing and even longitudinal profiles were appropriate, because of the lower depth-to-width ratio of these valleys. The profiles P1 and P2 in the Drau Valley (Figs. 3, 9) (Brückl and Ullrich 2001; Ullrich and Brückl 2004) were reprocessed using a wider frequency range of the seismic signal, applying residual statics and refining the velocity determination.

The next step was 2D-gravimetric modelling along the reflection seismic cross sections. The topography of the valley basement and prominent internal boundaries of the valley fill were taken from the reflection seismic profiles as constraints. Densities of the valley-fill were estimated from seismic P-wave velocities using standard relations (e.g. Gardner et al. 1974; Watkins et al. 1972). For the valley basement, the same density was taken as for the calculation of the Bouguer anomaly (2,670 kg/m^3). A fit between observed and calculated Bouguer anomaly was obtained by interactively adjusting densities of the valley infill within reasonable ranges and the structure of the valley at locations where the reflection seismic profile could not image the valley structure clearly. The GMSYS program (e.g. Campbell 1983; Won and Bevis 1987) was used for interactive 2D-modelling. The simplification of 2D in contrast to 3D is justified in case the structural changes in the direction of the valley are small. The GMSYS program is basically a program for 2D interpretation, but it also allows consideration of cross-sections which do not cross the valley at right angle, and limitation of the spatial extent of the model to the left and right side

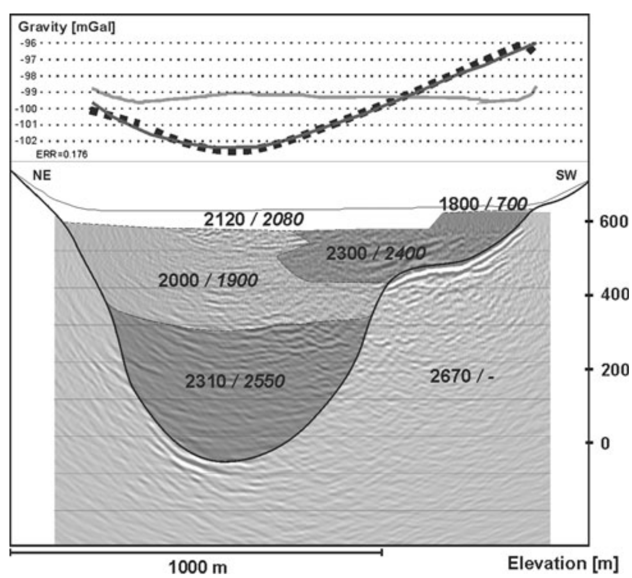

Fig. 3 2D-gravity modelling along reflection seismic cross section (Drau Valley, *profile P1*). Density model is superimposed on depth migrated reflection seismic section. *Numbers* give densities (kg/m^3) and (*italic*) P-wave velocities (m/s). Observed (*dotted line*) and calculated Bouguer-gravity (*solid line*), their differences are shown at the *top*

of the cross-section (2.5–2.75D). Figure 3 shows the result of interactive gravimetric modelling along the seismic profile P1 in the Drau Valley. The location of this profile is shown in Fig. 2. The right valley flank was not imaged by the reflection seismic data because of the limited extent of the seismic profile to the left. The gravimetric modelling proved to be very useful in complementing this missing structural information. There were only little adjustments of densities necessary in order to achieve nearly perfect fit of calculated to observed gravity data.

Gravimetric modelling along the seismic profiles yields density models of the valley-fill. In case these density models vary slowly along the valley and interpolation is justified between the seismic profiles, one can transform the Bouguer anomaly to the topography of rock basement of the valley. This step was done for the gravimetric surveys in the Ötz- and Drau Valley.

Structural models

In the following, we present briefly the field layout and the most important results from the geophysical campaigns. For each valley a map was compiled, showing the topography by contour lines, the local geology by grey code, the location of seismic profiles or resistivity soundings referred to in the text, and the data points covered by the gravimetric survey.

Ötz Valley

The segment of the Ötz Valley under consideration is imprinted by the giant rockslide of Köfels in its central part and by the smaller rockslide of Tumpen at the northern boundary of the investigation area (Fig. 4). The rockslide of Köfels took place 9,800 ± 100 calibrated radiocarbon years BP (Ivy-Ochs et al. 1998) and it is known as the largest rockslide in crystalline rocks of the Alps. Melting of rock during the sliding phase led to the generation of pumice and hyalomylonite (Preuss 1971; Erismann et al. 1977; Erismann 1979). This observation has drawn earth scientists' attention to this mass movement since the nineteenth century. A review on the Köfels rockslide has been given by Heuberger (1994). The Tumpen rockslide is a later event dated to 3,000 years BP (Poscher and Patzelt 2000).

The target of most geophysical campaigns in this area was the Köfels rockslide. Because the present day rockslide mass extends from the western to the eastern valley flank, these surveys also cover the Ötz Valley. Structural models of the base of the rockslide, the rock base of the valley and the top of an "old valley-fill" overthrusted by the rockslide mass were constructed by the use of seismic data, as well

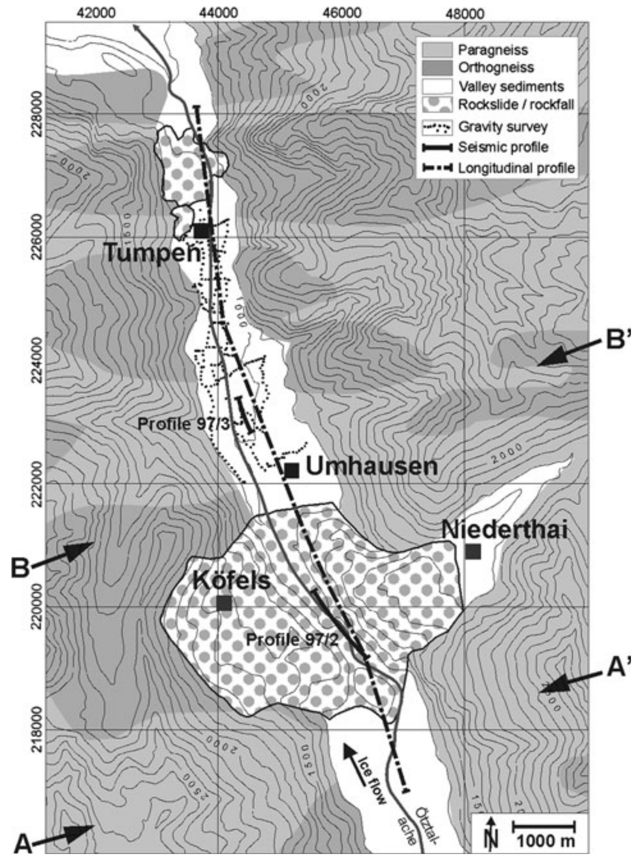

Fig. 4 Ötz Valley—topography and geology at the investigation site. Reflection seismic profiles shown in Fig. 5 and gravity data points are plotted. *Arrows* and *letters* *A*, *A*′, *B*, *B*′ mark representative cross sections shown in Fig. 11

as information from boreholes and an exploration edit (Brückl et al. 2001). The investigations at the Köfels rockslide also covered the adjoining basins in the south (Längenfeld) and north (Umhausen). Additional seismic investigations were carried out near the Tumpen rockslide (Poscher and Patzelt 1996). These investigations targeted sinkholes within the valley-fill. In 2001, a gravimetric survey was carried out by the authors between the Umhausen basin and the Tumpen rockslide to get information about the valley structure in this segment. This gravimetric data has not been published.

Figure 5 shows a longitudinal section along the axis of the Ötz Valley from the Längenfeld basin, over the Köfels rockslide, through the Umhausen basin to the Tumpen rockslide. A significant feature with relevance to deep alpine valleys is the deepening of the rock base of the valley by ~300 m just at the southern edge of the Köfels rockslide. The reflection seismic profiles were acquired and processed similarly to profile P1 and P2 in the Drau Valley. They also show the "old valley-fill" overthrusted by the Köfels rockslide mass (profile 97/2), which can be correlated downstream into the Umhausen basin (profile 97/3).

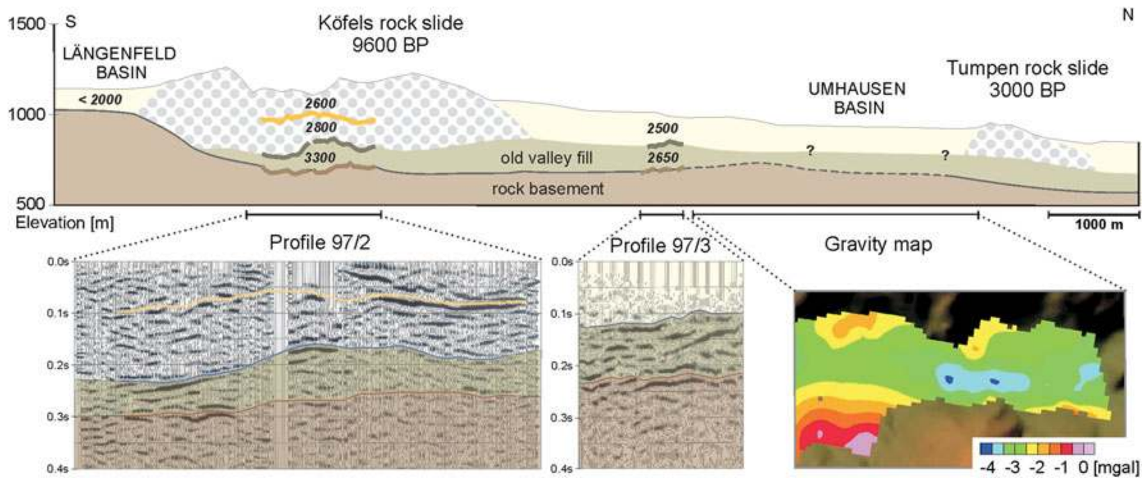

Fig. 5 Ötz Valley: longitudinal section. *Numbers* in cross section are P-wave velocities (m/s). Time sections of reflection seismic profiles 97/2 and 97/3, and Bouguer gravity map are shown below

Total thickness of valley-fill in the Längenfeld basin is ~100 m, in the Umhausen basin and at the Tumpen rockslide ~400 m. A residual Bouguer gravity map was calculated by the use of the gravity data acquired in the area between the Köfels and Tumpen rockslide, applying the same methodology as described for the Drau Valley. The residual Bouguer gravity indicates a sill in the longitudinal profile. Downstream the sill there is no clear evidence if "old valley-fill" exists or not. P-wave velocities of the strata classified as "old valley-fill" were derived from optimum stacking velocities and range from 2,650 m/s in the Umhausen basin to 3,300 m/s below the Köfels rockslide. Several boreholes were drilled in the Umhausen basin showing mainly gravel and boulders down to an elevation of 900 m (P. Hacker, personal communication). They did not reach the top of the "old valley-fill". In the Längenfeld basin fine grained sediments alternate with sand, gravel and boulders (Klebelsberg 1951). P-wave velocities of the valley-fill in the Längenfeld basin are below 2,000 m/s and we did not discern "old valley-fill" in this area.

Oichten Valley

Geophysical investigations in the Oichten Valley were carried out along one cross-section as part of a geophysical fieldwork tutorial at Vienna University of Technology in 1998. Results were not published before. Processing and modelling followed the general procedures described before. Figure 6 shows the location of the geophysical transect over the Oichten Valley together with the topography, the main geological units and the location of the one borehole used for interpretation. The profile is located just at the NAT, where the Flysch overthrusts the Molasse. Mass movements in the Flysch unit on the south-eastern valley flank are another geological observation relevant for the evolution of the Oichten Valley. The current flow direction of the Oichten creek's current is from NE to SW. However, during the ice ages the Salzach glacier spread radially into the northern foreland (e.g. Weinberger 1955) and its flow direction was opposite to the present day flow of the Oichten creek. We regard the NE direction of the ice stream as the representative flow direction of the Oichten Valley.

The reflection seismic profile started at the slope of the Haunsberg and extended over approximately 2/3 of the valley to WNW. Spacing of receiver locations was 5 m and a 24-fold nominal CMP-coverage was achieved. All other acquisition parameters were the same as for the Drau-Valley. The whole profile comprised 265 receiver locations resulting in a total length of 1,320 m. Processing followed the procedure described above. Figure 7 shows the final depth migration. Interval velocities, derived from optimum stacking velocities, are assigned to the different seismic units. The gravity survey was carried out along the seismic profile using a SCINTREX CG3 gravity meter. The measurements were taken at a spacing of 50 m. The gravimetric profile extends the seismic profile over the whole valley and has a total length of 2,160 m. Bouguer gravity was calculated using a simplified 2D-terrain model and assuming a density of 2,380 kg/m^3, in spite of the standard 2,670 kg/m^3. A local linear trend was also removed from the Bouguer anomaly before modelling. The geophysical profile crosses the Oichten Valley obliquely, which was considered by the modelling procedures with GMSYS. Structures revealed by the seismic profile were extended over the whole valley by the gravimetric modelling, and densities estimated from seismic velocities were adjusted (see Fig. 7). Resistivity measurements using Schlumberger and dipole–dipole electrode configurations, as well as electromagnetic mapping,

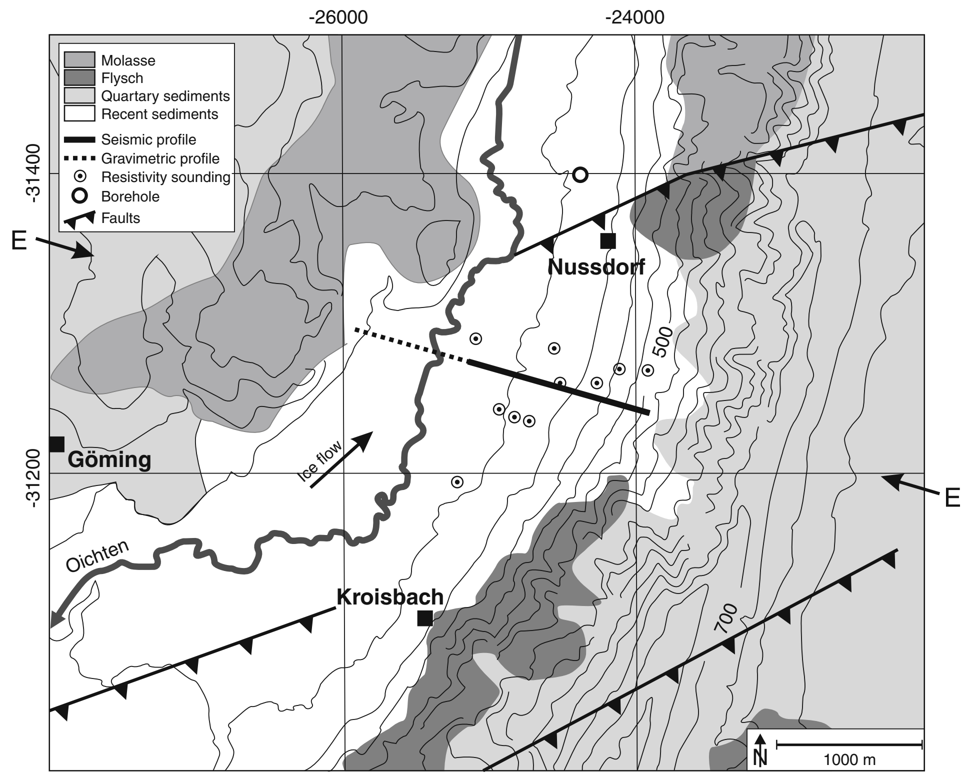

Fig. 6 Oichten Valley: topography, geology, and tectonic structures at the investigation site. Combined reflection seismic and gravimetric profile, centre points of resistivity soundings, and location of borehole (referred to in text) are plotted. *Arrows* and *letters E*, *E′* mark representative cross section shown in Fig. 11

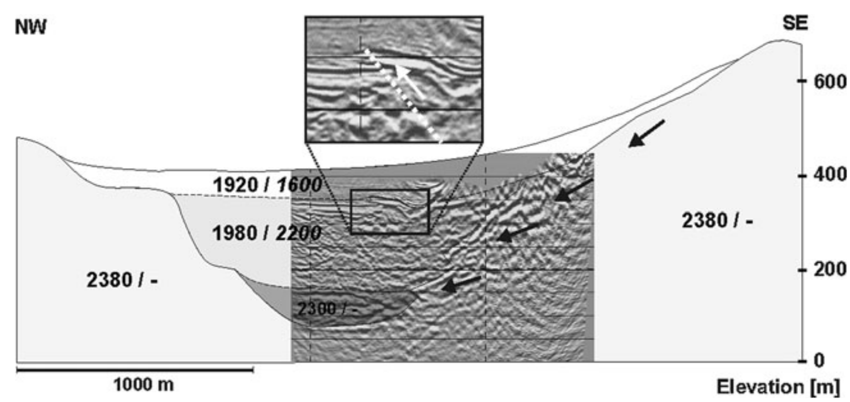

Fig. 7 Oichten Valley: gravimetric and reflection seismic cross section. *Numbers* give densities (kg/m^3) and (*italic*) P-wave velocities (m/s). *Arrows* on *right-hand* slope indicate mass movements. *Inset* enlarges structure interpreted as thrust fault

were carried out in the vicinity of the seismic and gravimetric profile within the course of the University fieldwork.

The combined reflection seismic and gravimetric profile (Fig. 7) begins at the Haunsberg slope in the SE, dips gently to the flat valley floor at an elevation of about 410 m, and reaches the outcropping Molasse in the NW. The depth migration of the seismic data shows several characteristic horizons. At the right valley flank, discontinuous reflections indicate mass movements in the Flysch slope of the Haunsberg. The uppermost unit of the valley-fill shows a succession of flat and mainly undisturbed reflections with high frequency content, which are typical for unconsolidated, lacustrine deposits. Average values for P-wave velocities and densities are 1,600 m/s and 1,920 kg/m^3. At an elevation of about 330 m, a band of strong and flat reflectors marks the transition to more compacted sediments, with an irregular, incoherent reflection pattern. The average values for P-wave velocities and densities increase to 2,200 m/s and 1,980 kg/m^3. Resistivities of the uppermost valley sediments range from 10 to 100 Ohm m, in the deeper unit from 30 to 70 Ohm m. The relatively low resistivities of the valley-infill indicate a high clay and silt content. This interpretation is supported by the nearby borehole Oichtental-9 (courtesy RAG, Rohöl-Aufsuchungs Aktiengesellschaft, Wien; for location see Fig. 6), where sediments with high clay and silt content dominate, but alternate with gravel at greater depths.

At an elevation of $\sim$150 m another strong reflector, together with the gravity data, indicates a distinct sedimentary unit. The P-wave velocities of this deepest valley-infill were estimated at around 2,600 m/s, but not resolved accurately. However, the gravimetric interpretation also yields a significant increase in density compared to the shallower units (2,300 vs. 1,980 kg/m^3). The deepest reflector at a minimum elevation of about 70 m was correlated with the erosional base of the valley. We label the sediments between 70 and 150 m elevation as "old valley-fill". The borehole Oichtental-9 had a maximum depth of 243 m and did not reach the "old valley-fill". The interpretation of the gravity data allowed for the extension of the main structural features resolved by the reflection data to the NW margin of the Oichten Valley, where Molasse forms a gentle slope.

The reflection band at 330 m elevation and its overburden show interesting details near the right valley flank. One detail is a succession of approximately circular reflections, which have the appearance of older river beds. Another detail, enlarged by an inset in Fig. 7, is interpreted as a thrust fault with a throw of about 15 m. The fault movement interpreted from the seismic section must have taken place after the ice age, during the latest stage of the development of the Oichten Valley, when the uppermost lacustrine strata were deposited. The question is, if this fault has been generated by active tectonics at the NAT, where the Flysch nappe overthrusts the Molasse (see Fig. 6), or through the push of the instable Flysch at the Haunsberg slope. As geologic investigations indicate that N-directed thrusting of Northern Calcareous Alps ceased during the Miocene (Linzer et al. 2002; Ortner et al. 2006) and geodetic data shows no current uplift of the Flysch belt in this area (Höggerl 2001), we favour the latter interpretation.

Drau Valley

No deep boreholes are found in the selected section of the Drau Valley. An extensive reflection seismic and gravimetric investigation has been carried out, supplemented by resistivity profiling. Geophysical data acquisition and processing of these investigations were addressed above, at the occasion of the description of the methods. A complete description has been given by Brückl and Ullrich (2001), Ullrich and Brückl (2004). The extent of the gravimetric survey and the location of the two seismic lines discussed in this paper are delineated in Fig. 8.

Figure 9 shows the depth migrated section of the reflection seismic profile P2. Bedrock and different units of the valley-fill are colour coded and P-wave velocities derived from the reflection seismic velocity analysis. Densities used for gravimetric modelling are also added. A short section of a resistivity profile has been superimposed on the seismic profile in order to support interpretation. An inset enlarges a segment of the seismic section, which is important for the interpretation. On top of the seismic profile, there is a view of the landscape from the geophysical profile to the west. The topography of the profile P1 (Fig. 9) is nearly constant at an elevation of 610 m. The geophysical parameters (P-wave velocity, density, resistivity) of the near surface layers correspond to fluvial deposits (gravel and sand). A change in seismic facies and a resistivity decrease at an elevation of ~570 m indicate a transition to silt and lacustrine clay at greater depth. Like in the Oichten Valley, high frequency continuous reflections (see inset in Fig. 9) are typical for this lacustrine deposition and support the interpretation. This pattern changes at lower elevations

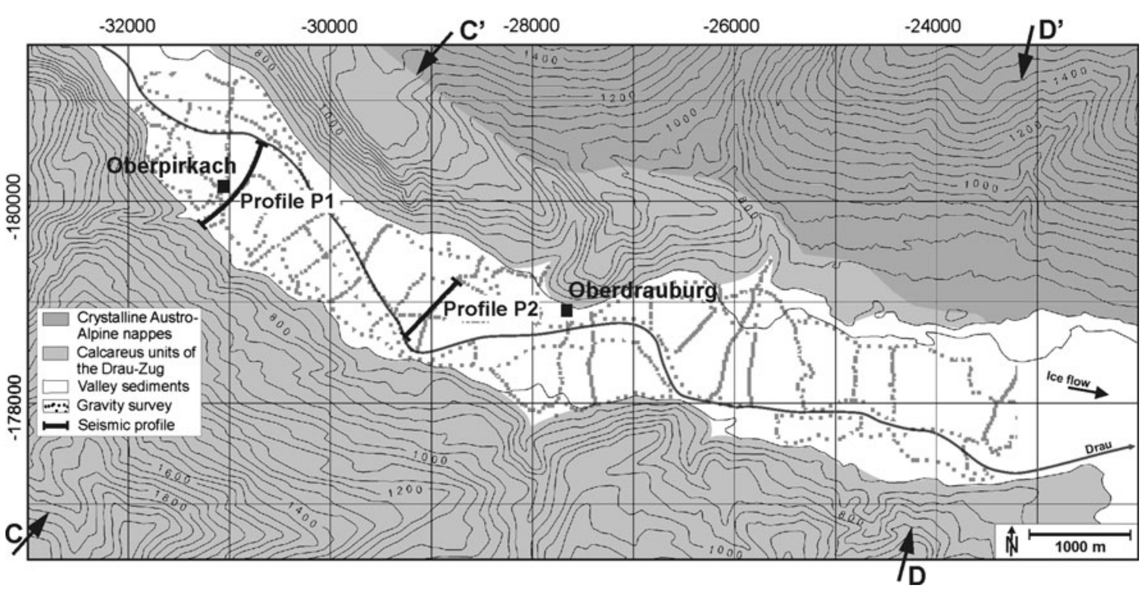

Fig. 8 Drau Valley: topography and geology at the investigation site. Reflection seismic profiles shown in Figs. 3 and 9, and gravity data points are plotted. *Arrows* and *letters* C, C', D, D' mark representative cross sections shown in Fig. 11

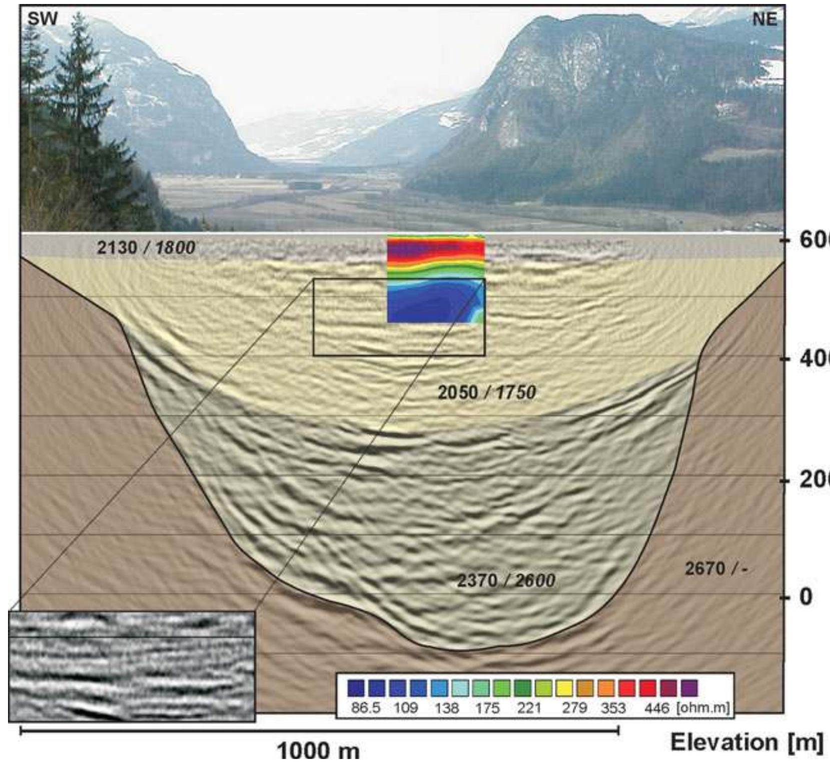

Fig. 9 Drau Valley: gravimetric and reflection seismic cross section of profile P2. *Numbers* give densities (kg/m^3) and (*italic*) P-wave velocities (m/s). A short section of a resistivity profile has been superimposed on the seismic profile in order to support interpretation. On *top* of the seismic profile, a view of the landscape from the geophysical profile to the west has been placed. The inset enlarges the part of the seismic section marked by a *rectangle*

(~450 m) to more distinct, sub-horizontal reflectors, indicating a change of the conditions of deposition. At elevations of ~400 m at the valley flanks and ~280–300 m in the centre of the valley, reflection seismic data shows a significant boundary to a deeper unit, which we address as "old valley-fill". Because of its curved shape, we assume that this boundary represents the glacier bed during the decay phase of the Würm glaciation. Reflectors within the "old valley-fill" at 100–200 m elevation indicate that its deposition was a multi-phase process. The rock basement has been eroded about 100 m below sea level and the total thickness of the infill is ~700 m. The steep dip of the valley flanks fits to the dip of the rock faces exposed at the surface in the west of the profile.

The internal structures of the valley-fill of profile P1, central part (see Fig. 3), correspond to profile P2 (Fig. 9). The P-wave velocities for the "old valley-fill" at profiles P1 and P2 are 2,550 and 2600 m/s and represent a significant contrast to the hanging deposits with 1,900 and 1,750 m/s. In the southern part of profile P1 there is a gently dipping portion of the erosional base at about 150 m depth. The sediments in this area are interpreted as an alluvial fan of a tributary meshed together with sediments of the river Drau. A similar geologic situation may be the fan of the Frauenbach about 5 km upstream, described by Poscher and Patzelt (1995) on the basis of borehole and geophysical data. Gravel and sand alternate with fan sediments down to a depth of about 40 m. Below 40 m also quiescent area sediments were found. The facies transition at that depth agrees with the interpretation of the geophysical profiles P1 and P2. Maximum borehole depth was around 50 m. C14 dating indicates a period of high sedimentation rate around 11 Kyr BP. A map of the erosional base of the Drau Valley derived from seismic and gravity data will be shown and discussed later.

Pasterze

In 1997 a reflection seismic survey was carried out at the Pasterze, the largest glacier in the Austrian Alps, situated in the Glockner range, near the Drau Valley investigation area (Fig. 1). The number of channels was 48 and a nominal CMP-coverage of 24-fold was achieved. All other acquisition parameters were the same as for the Drau Valley. Figure 10 shows a map of the investigation area with the location of the reflection seismic profiles (a), and depth migrated sections of the longitudinal (b) and transversal (c) profiles. Maximum ice thickness is about 150 m at these profiles. At the transverse profile and the corresponding part of the longitudinal profile we identify a distinct, deeper reflection horizon, which indicates an over-deepening of the erosional base by about 100 m during an earlier glacial stage. The material between the present day and earlier glacier beds shows internal structure and may consist of moraine and sediments deposited during a periglacial stage of this area. C^{14} dating of wood and peat samples released at the front of the currently retreating tongue of the Pasterze glacier indicate that the glacier bed was ice free in

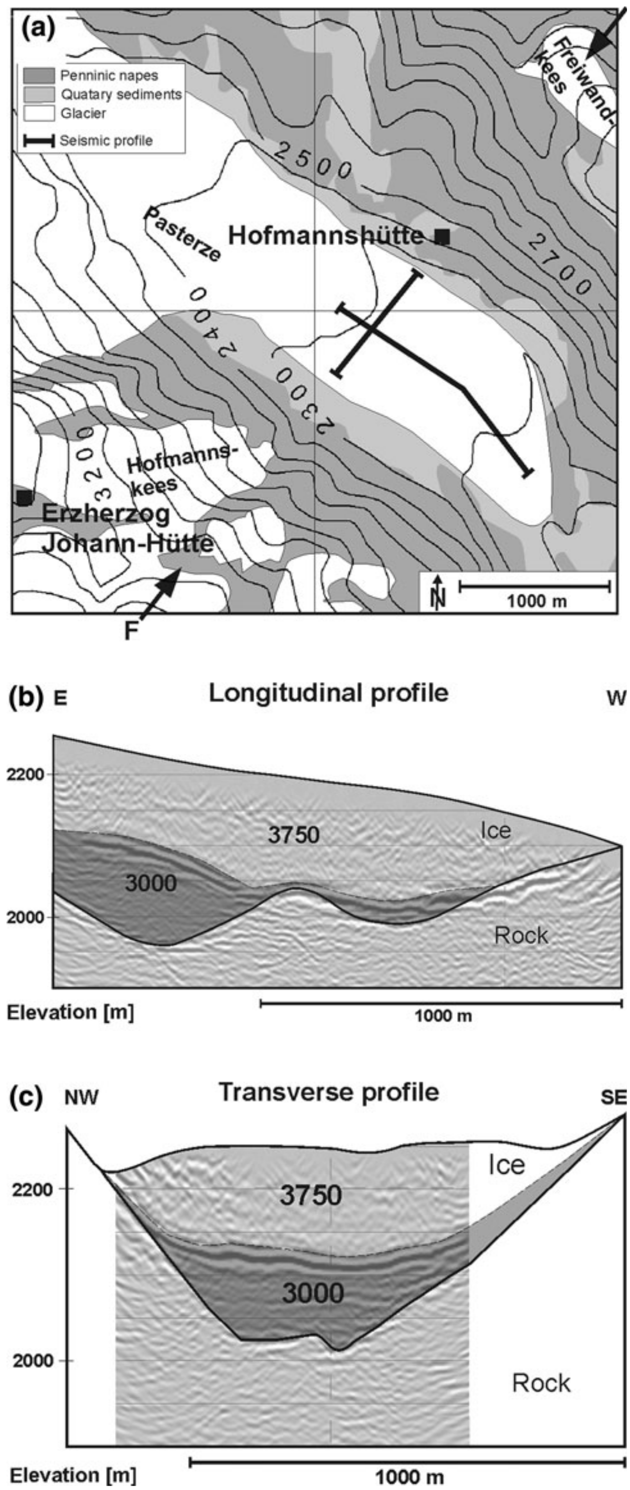

Fig. 10 Pasterze glacier: geology of investigation area with location of the reflection seismic profiles (**a**), and depth migrated sections of the longitudinal (**b**) and transversal (**c**) profiles. Numbers give P-wave velocities (m/s). *Arrows* and *letters F*, *F′* mark representative cross-section shown in Fig. 11

this area during several phases between 10 and 8 Kyr BP (Slupetzky 1990; Nicolussi and Patzelt 2000). There is also a remarkable similarity between the "old valley-fill" in the Drau Valley and the sedimentary layer at the bed of the Pasterze glacier (compare Fig. 10b with Figs. 3, 9).

Interpretation of "old valley-fill"

Observations of valley sediments comparable with our "old valley-fill" were made in the Inn-valley west of Innsbruck (Gruber and Weber 2003) and the Inn and Ziller Valley at their confluence (Weber et al. 1990; Weber and Schmid 1991). The sediments of the Inn Valley reach a thickness of about 700 m west of Innsbruck and a strong internal reflector marks the transition to more compacted layers, interpreted as a moraine. Sediments at the confluence of Inn and Ziller reach a thickness of about 900 m. A strong and continuous reflector in the depth range of 150–300 m separates unconsolidated sediments (P-wave velocities 1,500–1,800 m/s) from more compacted ones at greater depth (P-wave velocities 2,400–2,800 m/s). The deeper strata are tentatively interpreted as glacial and interglacial sediments.

Reflection seismic sections in the Rhone Valley, Western Alps, especially the lines Vétroz and Martigny revealed a detailed stratigraphy and a maximum depth to bedrock of about 800 m (Finckh and Frei 1991; Besson et al. 1993). The valley sediments resemble, from surface to bedrock, deltaic units, lacustrine and glacio-lacustrine deposits (P-wave velocities 1,650–2,000 m/s), meltout and reworked till (1,900–1,950 m/s), lodgement till (2,300–2,400 m/s) and subglacial deposits (2,050 m/s) as the deepest unit. One single deglaciation event can explain such a sedimentary sequence (Pfiffner et al. 1997). That means that the last (Würm) glaciation completely excavated the valley sediments of the last interglacial. On the other hand, we also find examples where sediments deposited during the last interglacial were not completely eroded by the Würm glaciers. One prominent example from the Eastern Alps is the Inn-terrace of Baumkirchen near Innsbruck, which contains a sequence of stratified clay and gravel, topped by basal till. C^{14} dating reveals ages between 32 and 27 Kyr BP (e.g. van Husen 2000 and references therein). An example from the Western Alps is the Birrfeld area. On the basis of borehole information and reflection seismic profiles, infilling sediments deposited by five different glaciations or glacial advances with a total thickness of about 200 m were interpreted (Nitsche et al. 2001).

To the knowledge of the authors there is no information available from boreholes or other direct access about the sediments labelled as "old valley-fill" in the Ötz-, Oichten- and Drau Valley. The only constraint on the minimum age of these sediments exists in the Ötz Valley, where the Köfels rockslide overthrusted the "old valley-fill" 9,800 yr BP (Ivy-Ochs et al. 1998; Brückl et al. 2001). We therefore attempt a characterization of the old infill on the basis of the geophysical evidence. The thickness of the "old valley-fill" in the centre of the valleys reaches from 80 m in the Oichten Valley to a maximum of 400 m in the Drau Valley. P-wave velocities of the "old valley-fill", derived from the reflection seismic velocity analysis, are >2,500 m/s and densities, constrained by the gravimetric data, are $\geq$2,300 kg/m^3. These values are significantly higher than for the more shallow strata of our examples and for fluvial and lacustrine sediments and meltout till in the Inn- and Rhone Valley and the Birrfeld area (Weber and Schmid 1991; Pfiffner and 1997; Nitsche et al. 2001). An exception is the Umhausen basin, where the sediments overlying the "old valley-fill" have an average P-wave velocity of 2,500 m/s. However, these sediments are derived from the Köfels rockslide and are a particular case. Further, strong reflectors mark the top of the "old valley-fill". As pointed out before, transverse profiles, especially profile P2 in the Drau Valley (Fig. 9), show a concave shaping of these horizons and a similarity to the transverse profile at the Pasterze (Fig. 10b). Both the intermediate layer at the Pasterze and the "old valley-fill" show internal reflectors, indicating a complex deposition process. From this geophysical evidence we come to the following generalized interpretation of the internal structure of the valley-fill in our investigation area. The sediments above the "old valley-fill" comprise fluvial sediments, lacustrine deposits and meltout till. Mass movements (especially the Köfels rockslide) as well as alluvial fans modify this deposition scheme locally. The top of the old infill represents the most recent glacier bed. In the Oichten Valley this was at the maximum extent of the Würm glaciation, which reached about 10 km to the north-east beyond the location of the geophysical profile (Weinberger 1955). In the Ötz- and Drau Valley the intermittent glacier advances of the Late Upper Würm (van Husen 2000) gave the top of the "old valley-fill" its final shape. "Old valley-fill" in the Ötz- and Drau Valley could have been completely eroded during the high-stand of the Würm glaciation. However, on the basis of the geophysical data alone we cannot exclude the possibility that the main body of the "old valley-fill" was deposited before the ice build-up of the Würm glaciation.

Why deep erosion?

Deepening of alpine valleys has been described (e.g. Penck 1905) and numerically simulated (e.g. Anderson et al. 2006) to occur at the confluence of headwater tributaries. Increased deepening of the Ötz Valley occurs just at the southern edge of the Köfels rockslide, but also coincides with the confluence of the Horlach Valley from the east. This may be an argument for the special glacier-dynamic situation being linked to the deep erosion. On the other hand, about 5 km upstream in the Längenfeld basin, the Gries Valley (a larger tributary than the Horlach Valley) joins the Ötz Valley. The geophysical exploration did not extend to this part, but borehole data (e.g. Klebelsberg 1951) gave evidence that no enhanced deepening of the Ötz Valley occurred at that place. Confluence of tributary valleys does not exist or correlate with deep erosion at the Oichten- and Drau Valley. Therefore, we conclude that the confluence of tributary glaciers to the main glacier stream is a process, which is of no or only minor importance for excess deepening at our examples of alpine valleys.

Glacial erosion generally causes widening of valleys into characteristic U-shaped troughs (Kirkbride and Matthews 1997; Anderson et al. 2006). A look at representative cross-sections through the valleys in our investigation areas (Fig. 11) shows that the erosional base of the Oichten Valley best resembles the classical U-shaped valley. The cross-section has only been slightly modified by mass movements that occurred in the Flysch of the right valley flank. The cross-sections of the Ötz- and Drau Valleys suggest that U-shaping of the valleys occurred less the deeper the rock base has been eroded. There is no evidence for significant modification of these cross-sections by Holocene mass movements. The following analysis has been motivated by this observation and considers dynamics of the main ice streams in the valleys quantitatively and as a factor supporting deep erosion. The effects of subglacial water erosion are not adequately considered by this analysis.

An important factor increasing the erosion potential of an ice stream is its sliding velocity at the glacier bed (Andrews 1972; Hallet 1979, 1996; Humphrey and Raymond 1994). Sliding can only occur when melting temperatures are reached at the glacier bed. One physical quantity controlling sliding velocity is basal shear stress (Weertman 1957). Another is effective normal stress at the sliding surface, which mobilizes friction and reduces the sliding velocity. Effective normal stress increases with ice thickness and decreases with water pressure at the basal surface (Bindschadler 1983; Paterson 2002). Factors favouring high water pressure could be a position near or below the equilibrium line, great thickness of the ice cover, and a low gradient or over-deepened longitudinal profile of the valley base.

Next we consider basal shear stress as a quantity related to basal sliding and glacial erosion. We estimate average basal shear stress at five cross-sections of our deep alpine

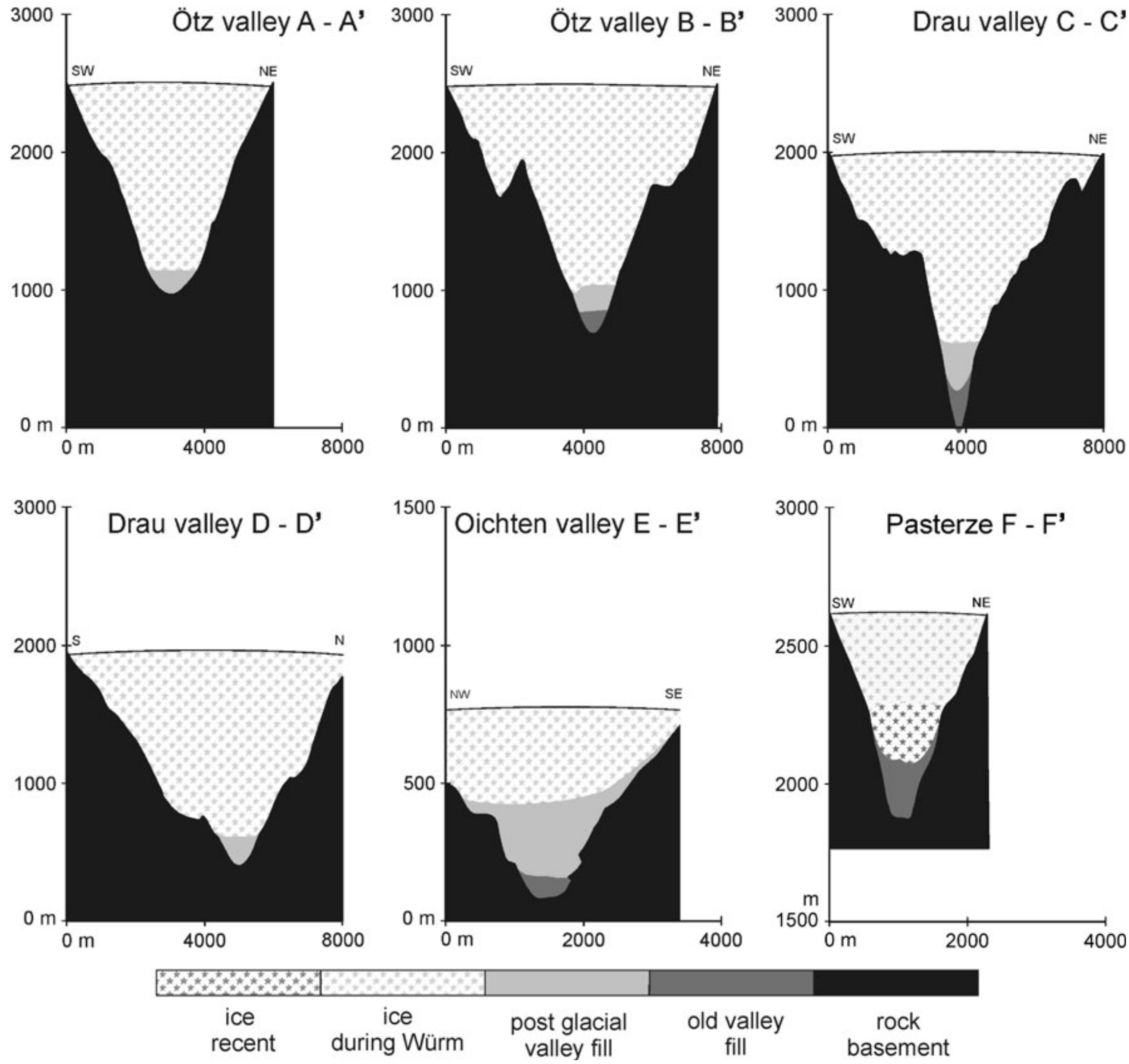

Fig. 11 Representative cross-sections of Ötz-, Oichten-, Drau Valley and Pasterze glacier. Ice cover during ice age, recent glacier ice, post glacial sediments, old valley-fill, and rock base are assigned by *grey scales* (see legend)

valleys and at Pasterze glacier by the formula (e.g. Paterson 2002):

$$\tau = d \cdot g \cdot r \cdot \sin(\alpha) \tag{1}$$

with τ basal shear stress, d density of ice (900 kg/m^3), g gravity acceleration (9.81 m/s^2), r hydraulic radius, α ice surface inclination in flow direction.

The hydraulic radius of a cross-section through the ice stream is the ratio of the cross section area (A) and the perimeter (p) in contact with ice. The quantities A and p are determined from the cross sections shown in Fig. 11, the dip of ice surface was taken from the map of the last ice age (van Husen 1987). The location of the cross-sections is shown in Figs. 4, 6, 8, 10. The variable input parameters to Eq. 1 and the results are compiled in Table 1. Basal shear stress during the ice age has been calculated for two scenarios. The first is that the glacier bed is represented by the bedrock. The second alternative is that the top of the "old valley-fill" represents the glacier bed. "Old valley-fill" was identified in the three investigation areas only where over-deepening exceeds about 300 m (Figs. 3, 5, 7, 9, 12). It is absent on profiles Ötz Valley A–A′ and Drau Valley D–D′. At the Pasterze glacier we also calculate basal shear stress for the present day situation.

Van Husen's (1987) map represents the highstand of the last (Würm) glaciation. Elevation and dip of the ice surface during early and late stages of glaciation differed considerably from the high-stand. Therefore, our calculations of basal shear stress relate only to the maximum extent of glaciation. Another point is that the ice surface during the last ice age is related to present day markers of ice on the surface and does not consider glacial rebound. Therefore,

Table 1 Calculated average basal shear stress with either old valley-fill or rock base as the glacier bed at selected cross sections and input data

	Ice surface elevation (m) NN	Dip of ice surface (°)	Hydraulic radius (m); old valley fill/rock base	Basal shear stress (KPa); old valley fill/ rock base
Ötz Valley A–A′	2,500	0.55	–/742	–/63
Ötz Valley B–B′	2,500	0.55	809/815	69/69
Oichten Valley E–E′	770	0.92	236/353	33/50
Drau Valley C–C′	2,000	0.44	649/633	44/43
Drau Valley D–D′	2,000	0.44	–/777	–/53
Pasterze F–F′	2,700	2.0	269/282	83/87
Pasterze-today F–F′	2,250	4.6	91/–	65/–

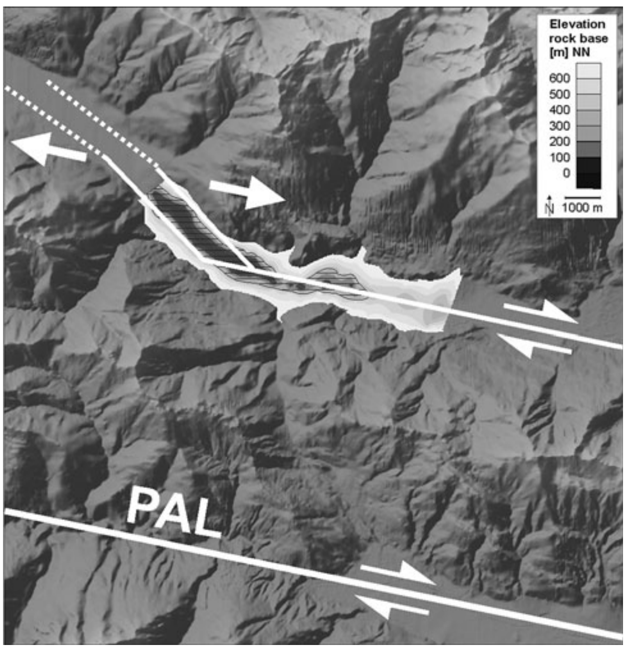

Fig. 12 Drau Valley: digital terrain model and scheme of a pull-apart mechanism supporting deep erosion in the NW–SE striking part of the valley. Main strike-slip direction is parallel to PAL. Elevation of rock base of the valley is shown by *colour* code. Hatching marks areal extent of old valley fill

elevation and dip of the ice surface (α) was lower than shown on the map. Total uplift of the Fennoscandian ice sheet is <700 m, tilting of terraces near the boundary of glaciation is mostly <0.05°, and present day uplift rates are <10 mm/year (e.g. Benn and Evans 1998). The present day uplift rate of the central Eastern Alps is 1–2 mm/year (Haslinger et al. 2007), but one cannot be certain how much can be attributed to isostatic rebound and how much is caused by ongoing tectonic processes. Persaud and Pfiffner (2004) carried out a quantitative study considering ice load and elastic thickness of the lithosphere and arrived at 167 m maximum deflection in the centre of Swiss Alps. Maximum tilting occurs near the border of the ice sheet and the foreland and is in the same range as observed in Fennoscandia. These figures suggest that the dip of the ice surface in the Eastern Alps derived from the map is not more than 10% too high, and we neglect the effect of crustal deflection during the ice age in the following.

In Table 1 we see that the basal shear stress ranges from 43 to 69 kPa for our examples of deep valleys during the highstand of the Würm glaciation and the recent Pasterze glacier. A value of about 100 kPa is considered as a good estimate for alpine glaciers world wide (e.g. Paterson 2002). The amount of over-deepening does not positively correlate with basal shear stress. For example, in the Drau Valley (Fig. 11, profile Drau C–C′), the highest amount of over-deepening (~700 m) coincides with the lowest basal shear stress (43 kPa). There is also no significant difference for basal shear stress, either the “old valley-fill” or the rock base constitutes the glacier bed. From this evidence we conclude that excess shear stress at the glacier bed is not an inescapable condition for deep erosion. Furthermore, erosion deeper than the top of the “old valley-fill” did not change the hydraulic radius of the ice streams at the Würm highstand significantly. That means, the material below the top of the “old valley-fill” does not represent an obstacle for efficient ice transport through the valley cross-sections. We consider high water pressure and low effective normal stress at the glacier bed to cause high sliding velocities and to enhance erosion, especially at locations of very strong overdeepening and later deposition of “old valley-fill”. A low valley gradient and overdeepening may favour high water pressure, thus enabling positive feedback between deepening of the basement and increase of water pressure.

Our three examples of deep alpine valleys have been selected with respect to different specific tectonic regimes. The valleys follow large scale fault systems and tectonic predesign is evident. However, the amount of overdeepening changes considerably. Therefore, valley structures will be analysed to see if particular tectonic weakening could be related to deep erosion. The Ötz Valley has been selected because of its situation in an extensional regime. The Längenfeld- and Umhausen basins may be interpreted as extensional basins or graben (Ampferer 1939). However, deep erosion started just at the southern edge of the Köfels rockslide and comprises only the Umhausen basin and not the Längenfeld basin. There may be a mutual relation in the sense that tectonic weakening of the rock base found its expression in deep erosion, as well as destabilizing the valley flank at the giant Köfels rockslide. The Oichten Valley is located at the NAT, where Flysch overthrusts Molasse. As discussed before, we do not assume ongoing tectonic movements at this fault. However, topographic features and the geophysical cross-section give evidence for mass movements at the right valley flank, formed by the Flysch of the Haunsberg (Fig. 7). Glacial erosion and mass movements may form a positive feedback loop, favouring overdeepening in this area. The Drau Valley is in a dextral strike-slip regime sub-parallel to the PAL (e.g. Ratschbacher et al. 1991; Linzer et al. 2002). Figure 12 shows the topography of the rock base derived from reflection seismic profiles and gravity mapping (Brückl and Ullrich 2001; Ullrich and Brückl 2004). In the western part of the valley, erosion is about 300 m deeper than in the eastern part; likewise, in the Lienz basin west of our investigation area the erosion base is significantly higher (Walach and Posch 1990). Furthermore, the deeply eroded part of the valley strikes about NW–SE, whereas the other parts strike WNW-ESE, nearly parallel to PAL. We introduce a pull-apart mechanism, which may have weakened the rock base and supported the very deep erosion in the NW–SE striking part of the valley (see Fig. 12).

Conclusions

Results from geophysical explorations of three alpine valleys in different tectonic regimes have been presented. The combination of 2D reflection seismic profiles, gravity data and integrated interpretation proved to be an efficient tool for the construction of structural models of the valleys. The erosional base of the valleys was mapped and significant internal structuring of the valley sediments has been resolved. Distinct layer packages in the lower part of the valley sediments at depths greater than 300 m were addressed as "old valley-fill". No evidence from boreholes or other direct access exists on the nature of these strata. P-wave velocities and densities of the "old valley-fill" are significantly higher than for the shallower layers, which comprise fluvial sediments, lacustrine deposits and meltout till. Especially the profiles in the Drau Valley have a concave shape of the strong reflectors representing the top of the "old valley-fill". These findings and a striking similarity with the reflection seismic image of an intermediate sedimentary layer at the recent Pasterze glacier suggest that the top of the "old valley-fill" built the most recent glacier bed. In the Oichten Valley, where the geophysical profile is located about 10 km behind the maximum extent of the Würm glaciation, it must have been near the highstand of the Würm glaciation. In the Ötz- and Drau Valley intermittent advances during the Late Upper Würm represent the most recent glaciers, which overthrust and shaped the "old valley-fill".

A qualitative judgement of the shape of cross-sections shows that the more they deviate from a U-shape, the deeper the erosion. A quantitative analysis shows that ice dynamics—characterized by basal shear stress—were lower during the ice ages than observed at recent alpine glaciers. Furthermore, the top of "old valley-fill" as a glacier bed and the corresponding cross-sections support efficient ice flux equally well as the deeper erosional rock base. We conclude that basal shear stress is not positively correlated with depth of erosion. However, deep erosion and especially overdeepening may be related to high water pressure at the glacier bed, which supports high sliding velocities. The confluence of tributary glaciers is apparently not a significant factor for deep erosion in our examples of deep alpine valleys.

Like other deep alpine valleys, our examples are associated with tectonic structures, active during Tertiary (e.g. Linzer et al. 2002). Current seismicity is relatively low for the Ötz- and Drau Valley, and below detection level for the Oichten Valley. Tectonics as an essential factor enabling or enhancing deep erosion is supported by some arguments in the Ötz Valley. In the Oichten Valley structures correlated with active mass movements are imaged by the seismic profile. There may be a positive feedback loop between slope destabilization by mass movements and increased overdeepening. In the Drau Valley a pull-apart mechanism consistent with the general tectonic regime in this area (Fig. 12) could explain the excessive deep erosion in its western part.

Valuable information on structure and development of deep alpine valleys can be drawn from the geophysical data we present here, as even the investigations have reconnaissance character. Further explorations of this kind and systematic compilation of existing data could improve our knowledge of deep alpine valleys considerably. However, 3D high resolution seismic data in combination with deep scientific boreholes will be inevitable to answer questions like dating, and details of sedimentation processes.

Acknowledgments The geophysical investigations were funded by the Austrian Academy of Sciences (Ötz Valley), the Vienna University of Technology (Oichten Valley), the former Austrian Federal Ministry of Education, Science and Culture, and the Geocenter, Rohstoffforschung, Carinthia (Drau Valley). We further give our thanks to Verbund, Austrian Hydro Power AG, Vienna, and Pöyry Infra GmbH, Salzburg for permission to show the seismic profiles from Pasterze glacier, to P. Hacker for the permission to integrate the gravity data from the Ötz Valley into our study, and to RAG–Rohöl Aufsuchungs AG, Vienna for the permission to use the borehole data from the Oichten Valley. We further thank Dirk van Husen and an anonymous reviewer for their valuable remarks and hints for an improvement of this paper.

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