Timing of early Quaternary gravel accumulation in the Swiss Alpine Foreland

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Abstract

Deckenschotter ('Cover Gravels') are proximal glaciofluvial gravels located in the northern Alpine Foreland mainly beyond the extent of the Last Glacial Maximum. They cover Tertiary Molasse or Mesozoic bedrock with an erosional unconformity. In Switzerland, Deckenschotter are referred to as Höhere (Higher) and Tiefere (Lower) Deckenschotter based on their topographical positions with a significant phase of incision that separates these two units. For this study, we performed sedimentological analyses to identify the provenance, transport mechanisms and depositional environment of these gravels. In addition, we established the chronology of the Höhere Deckenschotter gravels at Stadlerberg using cosmogenic ¹⁰Be depth-profile dating technique. The inherited ¹⁰Be concentration then allowed estimation of a catchment-wide palaeo-denudation rate. The results from clast fabric investigations indicate that braided rivers within a glaciofluvial environment transported these sediments to the study site mainly as bedload. In addition, the petrographic composition of the deposits shows that a large portion of the gravels was derived through erosional recycling of Miocene Molasse conglomerates. Some material was additionally sourced in the northern Central Alps. We then conclude that gravel accumulation in the Swiss Alpine Foreland was completed at 1.9 ± 0.2 Ma. This age, however, represents a minimum age and the oldest ¹⁰Be depth-profile age ever obtained for a geological unit. Furthermore, a palaeo-denudation rate of c. 0.3-0.4 mm/a was estimated for the catchment of Stadlerberg gravels. Finally, elevation differences between the bedrock underlying the Höhere Deckenschotter and the modern base level imply a long-term regional incision rate of c. 0.12 mm/a.

Keywords: Cosmogenic; ¹⁰Be; depth-profile dating; Deckenschotter; Pleistocene

1. Introduction

In 1909, Penck and Brückner differentiated four Quaternary morphostratigraphic units at sites in the south German Alpine Foreland based on different levels of former outwash plains. At these sites, the higher elevation Deckenschotter units are attributed to the Günz glaciation, lower Deckenschotter units to the Mindel glaciation, Hochterrasse (HT; Higher Terrace) to the Riss glaciation and Niederterrasse (NT; Lower Terrace) to the Würm glaciation. It is important to know that the older sediments are situated at higher elevation while the altitudes of the younger ones are lower. For a long time, the Quaternary stratigraphy of Switzerland was correlated to that of southern Germany (Penck and Brückner, 1909). Based on results of decades of research, a refined stratigraphy was proposed for the northern Swiss Alpine Foreland, which reveals that glaciers advanced at least 13 times into the foreland during the Quaternary (Schlächter, 1986; Graf, 2009a). These include the four to eight Deckenschotter glaciations during the early Pleistocene (Graf, 1993; 2009b), at least five glaciations during the middle Pleistocene (Graf, 2009a) and two glaciations during the late Pleistocene including the Last Glacial Maximum (Schlächter and Müller-Dick, 1996; Keller and Krayss, 2010; Preusser et al., 2011).

1.1 Deckenschotter in Switzerland

In Switzerland, the Deckenschotter units have been divided into Höhere Deckenschotter (HDS; Higher Cover Gravels) and Tiefere Deckenschotter (TDS; Lower Cover Gravels) (Du Pasquier, 1891; Frei, 1912). The stratigraphic relationship of HDS and TDS to the original location in southern Germany is not established yet. Deckenschotter deposits have been mapped as laterally discontinuous isolated relicts on plateaus found between the Jura Mountains in the west and Lake Constance in the east; some outcrops occur even further westward, along the River Rhine (Du Pasquier, 1891; Frei, 1912; Graf, 1993; 2009a; Keller and Krayss, 2010; Preusser et al., 2011) (Fig. 1). Deckenschotter lie unconformably on Tertiary Molasse or Mesozoic carbonate bedrock. The gravel base of the HDS lies approximately between 400 m above sea level (a.s.l.) in the west and 700 m a.s.l. in the east, whereas TDS deposits are found at lower altitudes ranging from 350 m a.s.l. near Basel to 600 m a.s.l. near

Constance (Du Pasquier, 1891; Frei, 1912; Graf, 1993; Keller and Krayss, 2010). The difference in elevation suggests that the two Deckenschotter units are separated from each other by a period of significant incision (Graf, 1993; 2009a,b). Sediments recognised at the various outcrops were interpreted as proximal glaciofluvial gravels with local intercalations of till and overbank fines (Frei, 1912; Suter, 1939; Graf, 1993; Bitterli et al., 2000). These glaciofluvial deposits have been considered to represent braided river sediments, where gravels accumulated in broad channels when the ice released large volumes of coarse sediments that accumulated in close proximity, downstream of the snout of glaciers (Graf, 1993). In addition, these sediments have been interpreted to mark the major drainage network of the northern Swiss foreland during the early Pleistocene (Liniger, 1967; Graf, 1993; Hofmann, 1996; Petit et al., 1996; Kuhlemann and Rahn, 2013). In the period between HDS and TDS deposition, an incision of approximately 80-100 m into formerly deposited gravels and 30-110 m into bedrock are observed. Even deeper bedrock incision was observed after TDS deposition with erosional depths ranging between 140 and 260 m (Graf, 1993; Kuhlemann and Rahn, 2013).

The timing of Deckenschotter accumulations is poorly understood compared to the late Quaternary glaciations (Preusser et al., 2011 and references therein; Heuberger et al., 2014; Ivy-Ochs, 2015). Deckenschotter deposits are thought to be the oldest Quaternary units in the northern Swiss Alpine Foreland known so far. Palaeomagnetic studies were performed at some sites of the Höhere Deckenschotter in Switzerland (see Graf, 1993 for further details). Based on these results, the deposition of the oldest HDS occurred during a period of normal polarity, prior to the Brunhes Chron because some samples showed a normal magnetization with values that do not coincide with the direction of today's magnetic field (Graf, 1993; Bolliger et al., 1996). Thus, the minimum age for their accumulation would have been the Normal Jaramillo event during the Matuyama Reversed Chron (0.915-1.010 Ma; Spell and McDougall, 1992). In the topmost HDS deposits at Irchel, a mammalian faunal assemblage was assigned to Mammal Neogene Zone 17 (MN 17: 1.8 – 2.5 Ma) (Bolliger et al., 1996). Recently, the stratigraphically youngest HDS deposits near Mandach (Fig. 1) were dated with cosmogenic ^{10}Be depth-profile dating to 1.02 ± 0.1 Ma (Akçar et al., 2014). Accordingly, more detailed chronological investigations are needed to improve our understanding.

A first absolute age was obtained by Häuselmann et al. (2007a) who applied burial dating techniques to Deckenschotter deposits at their type locality in southern Germany (Penck and Brückner, 1909). These authors proposed burial ages of $2.35^{+1.08}_{-0.88}$ Ma for HDS sediments at Böhener Feld and $0.68^{+0.23}_{-0.24}$ Ma for TDS deposits at Bad Grönenbach.

Figure 1. Extension of the Valais, Aare, Reuss, Linth and Rhaetian Lobes during the Last Glacial Maximum (from Bini et al., 2009); location of the study site Stadlerberg and distribution of the Deckenschotter in the northern Alpine Foreland (©Federal Office of Topography, swisstopo, CH-3084 Wabern). The black square marks the outline of Figure 2A.

1.2 Goals of this study

Although Deckenschotter deposits in Switzerland have been extensively studied for their lithostratigraphic position, relatively little quantitative information has been published on the chronology of these archives and on the palaeo-denudation rate recorded by them. This paper focuses on the Higher Deckenschotter deposits at Stadlerberg (canton Zurich, northern Switzerland) (Figs. 1, 2). The fabric and petrographic composition of the deposits are studied to identify the provenance of the sediments and to interpret the transport mechanisms and depositional environment. This information then allows inference to be made about the distance to the ice margin at that time. In order to contribute to the understanding of Quaternary landscape evolution in the foreland, it is of high importance to reconstruct the chronostratigraphy of the Swiss Deckenschotter. We use both ^{10}Be depth-profile and ^{10}Be and ^{26}Al isochron burial dating techniques to determine the age when these

sediments were deposited. In addition, we estimate a catchment-averaged denudation rate from the inherited ^{10}Be concentration preserved in these deposits. Finally, based on the reconstructed chronology, we report a post-depositional incision rate into bedrock. In conclusion, the overall goal is to provide, for the first time, quantitative data about the chronology of these archives, the denudation rate preserved in them, and their provenance.

2. Study site

Stadlerberg is located 20 km NNW of Zurich, at an elevation of 620 m a.s.l. (Figs. 1, 2). It is an isolated plateau that stands approximately 150 m above the surrounding topography beyond the extent of the Linth Glacier during the Last Glacial Maximum (LGM). The plateau covers an area of roughly 1 km² and is oriented to the NNW (Fig. 2). The bedrock underlying the Deckenschotter is made up of Miocene Molasse sandstones that formed in shallow marine and terrestrial conditions (Upper Marine and Upper Freshwater Molasse; Suter, 1939; Tarass, 1954). Deckenschotter deposits overlie the Molasse with a sharp unconformity spanning 10-13 Ma (Bürgisser, 1981; Bolliger, 1998). The base of the Deckenschotter is situated at an elevation of 605 m a.s.l. in the south and drops to an elevation of 580 m a.s.l. farther to the north, whereas in the northeast Deckenschotter were mapped at 595 m a.s.l. (Graf, 1993) (Fig. 2). The Deckenschotter at Stadlerberg represent the sedimentary fill of a SE-NW oriented channel as geoelectric measurements have shown (Tarass, 1954). We focused on the abandoned gravel pit Summerhalden situated at 610 m a.s.l., where samples were collected for dating purposes and petrographic analyses

Figure 2. A: Study area Stadlerberg and surrounding Deckenschotter sites. B: Cross section through Stadlerberg (A-B in Fig. 2A) (modified after Tarass, 1954).

3. Methodology

3.1. Sedimentological analyses

The clast suites of the Deckenschotter gravels yield information to allocate the source area of the sediments (Weltje and von Eynatten, 2004 and references therein). Accordingly, we collected approximately 250 clasts at three different levels along a stratigraphic section (Fig. 3). The material was sieved to the gravel fraction (16-63 mm) and counted. A bucket was used to collect the clasts to avoid any biases related to visual preference (after Schlüchter, 1989; Graf, 1993). The gravels were grouped into different lithological classes including: (1) dark grey Alpine limestone, (2) light grey limestone, (3) siliceous limestone, (4) dolomite, (5) vein quartz, (6) sandstone, (7) quartzite, (8) radiolarite, (9) chert/hornstone and (10) metamorphic crystalline rocks. Furthermore, attention was paid to the presence of Permian ‘Verrucano’ (red sandstones and breccias) clasts as they are a key lithology and indicate a well-constrained source area (Linth and Seez Valleys; Hantke, 1980). All samples were plotted on a ternary diagram to delineate possible populations (Schlüchter, 1976). In the diagram, pre-established classes were grouped and the most frequent lithologies were assigned to the three apexes.

Figure 3. Sketch of the abandoned gravel pit at Stadlerberg. The square inset shows the outline of Figure 5. Units 1-3 correspond to the studied stratigraphic section (Fig. 5A). The lithofacies code is based on Miall (1978) and Keller (1996). cGcm: massive, clast-supported gravel with cobbles; bGmm: massive, matrix-supported gravel with blocks; bGcm: massive, clast-supported gravel with boulders.

The transport process of sediments can be traced by studying the clast fabric in the outcrop (e.g. Cailleux, 1947; Sneed and Folk, 1958; Boulton, 1978; Benn and Ballantyne, 1994; Lukas et al., 2013). In addition, clast shape analyses yield useful constraints on whether or not the accumulated material at the sampling site consists of reworked Miocene Molasse material (rounded) or represent rather ‘fresh’ erosional products from the Alps (subangular) (Du Pasquier 1891; Frei 1912). Accordingly, approximately 100 clasts of the gravel fraction from the same lithology, here vein

quartz, were measured, and roundness and flattening index(es) were determined. Morphometric values determined from mixed lithologies have no significance due to their different erosional resistance during transport (e.g. Schlüchter, 1989; Benn and Ballantyne, 1994). The roundness (Z_i) and flattening (A_i) index(es) are calculated as:

$$Z_i = \frac{2r_1}{L} \times 1000 \quad (1)$$

$$A_i = \frac{L+1}{2E} \times 100 \quad (2),$$

where L is the length of the clast, $2r_1$ is the smallest radius of curvature, l is the width and E the thickness of the clast (Cailleux, 1947; Cailleux and Tricart, 1959; Reichelt, 1961). The individual Z_i and A_i values are displayed in a histogram, and related values were grouped from 1-50, 51-100, 101-150, etc. High Z_i values represent rounded clasts and high A_i values indicate flat clasts. The resulting roundness $Md(Z_i)$ and flattening $Md(A_i)$ index(es) of one lithology class is defined by the median and will be compared with values from other Quaternary unconsolidated deposits (Sneed and Folk, 1958; Benn and Ballantyne, 1994).

Here, we preferentially applied the roundness index of Cailleux (1947) for characterizing clast morphologies as most clasts at Stadlerberg have an intermediate roundness (sub-angular to sub-rounded). In contrast, the more widely adopted method developed by Sneed and Folk (1958) works well in glaciated catchments to differentiate between subglacial debris and englacially, supraglacially or fluvially transported sediments (Benn et al., 2004). Roundness categories are assigned to the clasts using qualitative criteria (Sneed and Folk, 1958). The interpretation of the particle form is then based on using either the C_{40} index (percentage of clasts with a $c:a$ ratio ≤ 0.4) vs. RA index (percentage of very angular and angular clasts) or the C_{40} index vs. RWR index (percentage of rounded and well-rounded clasts). The former combines very angular and angular clasts while the latter rounded and well-rounded clasts. It was shown that in some settings the RWR index might be more suitable to identify different transport pathways (Benn et al., 2004; Evans et al., 2010; Lukas et al., 2013). Note that we were aware that it is difficult to accurately determine the Cailleux roundness index, where

assignments of r values (smallest radius of curvature) might yield ambiguous results particular for very angular or angular clasts (Benn and Ballantyne, 1994). No such clasts were found at Stadlerberg.

Finally, imbrication was measured on elongated cobble-sized clasts with a compass to determine the palaeoflow direction at Stadlerberg.

3.2 Cosmogenic nuclide dating

Cosmogenic nuclides are produced through spallation reactions and muon capture in minerals of rocks and sediment at or near the Earth's surface (Gosse and Philips, 2001). ^{10}Be and ^{26}Al can be applied to determine an age since deposition of a geological layer using either their accumulation (depth-profile dating) or their radioactive decay (burial dating) (Granger and Smith, 2000; Granger and Muzikar, 2001; Wolkowinsky and Granger, 2004; Dehnert and Schlüchter, 2008). Depth-profile dating uses the fact that the build-up of cosmogenic nuclides decreases with depth following the known physical principles (Gosse and Philips, 2001). This method requires sampling of geological layers in artificial outcrops, preferably with a flat-topped landform in order to guarantee that the uppermost surface of the deposit remains as unmodified as possible and is related to a defined geomorphologic process. The method was first introduced by Repka et al. (1997) to date fluvial terraces of the Fremont River in Utah. On the other hand, the burial dating technique uses the difference in half-lives of ^{10}Be (1.387 Ma; Korschinek et al., 2010; Chmeleff et al. 2010) and ^{26}Al (0.705 Ma; Norris et al. 1983) and thus the ^{26}Al versus ^{10}Be ratio to determine the burial time, when the pre-burial and post-burial concentrations are known or estimated (e.g. Granger and Muzikar, 2001). When the post-burial production by muons cannot be ignored, a recently introduced application of the burial dating technique, referred to as the isochron burial dating method, is used (Balco and Rovey, 2008). In this method, samples from either the same stratigraphic horizon (assuming to represent the same timeline) or from different depths within a depositional sequence are collected. The methodology in either case works if the collected material has different inherited nuclide concentrations but the same post-burial production (Balco and Rovey, 2008; Darling et al., 2012; Erlanger et al., 2012; Balco et al., 2013; Çiner et al., 2015). When the measured ^{26}Al concentrations are plotted against ^{10}Be concentrations,

they should fall on a line (Balco et al., 2008; Erlanger et al., 2012). The slope of the regression line is used to determine the $^{26}\text{Al}/^{10}\text{Be}$ ratio at the time of burial and thus the depositional age (for further details see Balco and Rovey, 2008; Darling et al., 2012; Erlanger et al., 2012; Balco et al., 2013, Çiner et al., 2015).

A recent collapse in the southeastern part of the abandoned gravel pit at Stadlerberg gives access to a fresh outcrop. In addition, this HDS site meets the methodological requirements: it is an artificial outcrop with a flat-topped morphology. Only a thin soil layer (c. 30 cm at the sampling profile location) is covering the gravels and the quartz content of gravel deposits is sufficient for dating with cosmogenic ^{10}Be . We collected seven sediment samples for depth-profile dating along a vertical profile of 5.2 m (Fig. 4, Table 1). In a depth between 5.2 and 5.5 m, nine quartz clasts were collected from the same lithostratigraphic horizon for isochron burial dating (Fig. 4). The quartz clasts for isochron burial dating were crushed followed by sieving. The sediment samples from the depth-profile were wet-sieved. For both sediment and clast samples, the size fraction between 250 and 400 μm was used for further processing.

Figure 4. Field photograph showing the abandoned gravel pit at Stadlerberg. Samples STDL-DP-1 to 7 were collected for depth-profile dating, while isochron burial dating was applied on samples STDL-1 to 9.

The collected samples were processed in the Surface Exposure Laboratory of the Institute of Geological Sciences at the University of Bern following the lab protocol described in Akçar et al. (2012). The $^{10}\text{Be}/^9\text{Be}$ and $^{26}\text{Al}/^{27}\text{Al}$ AMS measurements were then performed at the ETH tandem facility in Zurich (Christl et al., 2013). The long-term weighted average $^{10}\text{Be}/^9\text{Be}$ ratio of $(2.51 \pm 0.53) \times 10^{-15}$ was used for full process blank correction. A depth-profile age was then modelled with MATLAB® using Monte Carlo simulations developed by Hidy et al. (2010; available online at <http://geochronology.earthsciences.dal.ca/downloads-models.html>) based on exposure age, erosion rate and inheritance. Table 2 shows the input parameters for the simulations. We applied no correction factor for topographic shielding and cover (e.g. snow). We therefore used the default value 1 as input. The ^{10}Be half-life with a value of 1.387 ± 0.012 Ma was utilized (Korschinek et al., 2010; Chmeleff et al., 2010). The local production rate was scaled to the Lal (1991) / Stone (2000) scheme using a reference production rate due to spallation of 3.93 ± 0.19 at/g/a (Northeastern North America production rate calibration dataset; Balco et al., 2009; CRONUS-Earth calculator update from v. 2.1 to v. 2.2 published by Balco in October 2013 after Balco et al., 2008; <http://hess.ess.washington.edu/>). Thus a site specific spallogenic production rate of 7.18 ± 0.15 at/g/a was obtained. Calculation of the

muonic production was fit to a depth of 6 m. We determined a bulk density between 1.8 and 2.3 g/cm³ (Manger, 1963) for the sediment samples and assumed that it did not vary with depth. We applied an attenuation length of 160 ± 5 g/cm² (Dunne et al., 1999). Finally, to model a depth-profile age we simulated 100,000 profiles and used a χ^2 cut-off value of less or equal than 5.

Table 1. Sample details and cosmogenic nuclide data of the samples from Stadlerberg (N 47.5337, E 8.4458, 610 m a.s.l.)

Sample name	Sample depth (cm)	Sample type	Amount of quartz after leaching (g)	Amount of Al after leaching (ppm)	Quartz dissolved (g)	⁹ Be spike (mg)	¹⁰ Be/ ⁹ Be (x10 ⁻¹⁴)	Relative uncertainty (%)	¹⁰ Be concentration (10 ⁴ atoms/g)	Al (mg)	²⁶ Al/ ²⁷ Al (x10 ⁻¹⁴)	Relative uncertainty (%)	²⁶ Al concentration (10 ⁴ atoms/g)	²⁶ Al/ ¹⁰ Be
STDL-DP-1	30 - 40	sediment	66	339.26	49.9134	0.1251	61.89	4.1	10.32 ± 0.43	16.93	n.d.	n.d.	n.d.	n.d.
STDL-DP-2	58 - 68		77	216.10	49.9273	0.1183	41.63	4.8	6.55 ± 0.32	10.79				
STDL-DP-3	111 - 121		80	101.07	50.4359	0.1246	35.50	5.1	5.82 ± 0.30	5.10				
STDL-DP-4	160 - 170		77	114.34	50.1104	0.1254	23.55	5.7	3.90 ± 0.22	5.73				
STDL-DP-5	233 - 243		81	81.65	49.7356	0.1253	19.31	7.2	3.21 ± 0.23	4.06				
STDL-DP-6	358 - 368		47	222.70	46.9806	0.1250	13.46	8.1	2.35 ± 0.19	10.46				
STDL-DP-7	518 - 528		86	184.44	50.7258	0.1253	13.96	6.3	2.26 ± 0.15	9.36				
STDL-1	518 - 548	vein quartz clast	38	15.05										
STDL-2			73	14.24	52.9437	0.1265	9.71	10.5	1.51 ± 0.16	0.75	35.15	5.7	11.17 ± 0.64	7.40 ± 0.90
STDL-3			75	22.21	56.6565	0.1272	13.34	10.0	1.96 ± 0.20	1.26	21.09	7.4	10.46 ± 0.77	5.32 ± 0.67
STDL-4			76	9.33	54.6962	0.1270	10.47	9.9	1.59 ± 0.16	0.51	53.62	5.7	11.16 ± 0.64	7.04 ± 0.82
STDL-5			46	204.83										
STDL-6			68	38.21										
STDL-7			48	68.67	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
STDL-8			53	13.41										
STDL-9			31	5.65										

Reported blank ratio and concentration are referenced to S2007N (Kubik and Christl, 2010). AMS measurement errors are at 1σ level, including the statistical (counting) error and the error due to normalization of the standards and blanks. The error weighted average ¹⁰Be/⁹Be full process blank ratio is (2.51 ± 0.53) x 10⁻¹⁵.

n.d.: not determined

3.2 Palaeo-denudation rates

Cosmogenic nuclides can be used to infer basin-averaged erosion rates in the past (e.g. Granger et al., 2013). For instance, Granger et al. (1997, 2001) were the first to report palaeo-erosion rates that cover the past few million years. Furthermore, cosmogenic nuclides can also be used to determine palaeo-erosion rates recorded by fluvial deposits. Palaeo-erosion rates in the Meuse River catchment in the Netherlands increased after 0.7 Ma probably due to changes in climate and tectonic conditions (Schaller et al., 2004). A study by Charreau et al. (2011) illustrated that in Central Asia high erosion rates were recorded between 3.5 and 1.7 Ma with a peak at 2 Ma, which corresponds to the onset of Quaternary glaciations. The previous examples demonstrate that estimating palaeo-denudation rates at different timescales sheds light on possible variations in rates of landscape change in response to climate and tectonic driving forces (Whipple and Tucker, 1999; Bekaddour et al., 2014).

Here, we estimated palaeo-denudation rates through the consideration of the deepest measured ^{10}Be concentration in our depth-profile (STD-L-DP-7; 518-528 cm). We proceeded by correcting the measured concentration for radioactive decay and for post-burial isotope accumulation, thereby yielding the nuclide concentration at the time of deposition (Schaller et al., 2002). The exact reconstruction of the catchment is nearly impossible as the gravels at the study site show a provenance distributed over a large area of the Central Alps. We therefore compared the obtained concentration to normalised ^{10}Be concentrations measured in modern river sediments in the Central Alps of Switzerland and their inferred modern denudation rates (e.g. Wittmann et al., 2007; Chittenden et al., 2014). In these studies, denudation rates were calculated using an average basin elevation and thus a mean catchment production rate. We therefore normalised the measured ^{10}Be concentrations published in these studies to sea-level and high latitude (SLHL) for comparison purposes. We used the scaling scheme of Stone (2000) (St) for spallation with a high latitude, sea level production rate of 3.93 atoms/g (Northeastern North America production calibration dataset; Balco et al., 2009).

Table 2. Input parameters for the Monte Carlo Simulator in Matlab (Hidy et al., 2010).

Parameter	Value
Latitude (degree)	47.533
Longitude (degree)	8.446
Altitude (m)	610
Strike (degree)	0
Dip (degree)	0
Shielding correction factor	1
Cover correction factor	1
Uncertainty of ^{10}Be Half-life (%)	1
Local spallogenic production rate ($\text{at.g}^{-1}.\text{a}^{-1}$)	7.18
Error in local spallogenic production rate ($\text{at.g}^{-1}.\text{a}^{-1}$)	± 0.15
Depth of muon fit (m)	6
Error in total production rate (%)	5
Density (g.cm^{-3})	1.8 – 2.3
χ^2 value	5
Number of profiles	100,000
Age (a)	1,000,000 – 3,000,000
Erosion rate (cm.ka^{-1})	3 – 7
Total erosion threshold (cm)	7500 – 11,000
Inheritance (at.g^{-1})	0 – 40,000
Attenuation length (g.cm^{-2})	160 ± 5

A half-life of 1.387 ± 0.012 Ma for ^{10}Be (Korschinek et al., 2010; Chmeleff et al., 2010) is considered for the calculations. Local spallogenic production rate is calculated with the CRONUS-Earth exposure age calculator (<http://hess.ess.washington.edu/math/>; v. 2.2; Balco et al. (2008) and update from v. 2.1 to v. 2.2 published by Balco in October 2009) and constant production rate Lal (1991)/Stone (2000) scaling model using the NENA (Balco et al., 2009) production rate calibration dataset.

4. Results

4.1 Clast petrography and morphology

The outcrop at the abandoned gravel pit Summerhalden is c. 6 m high and 30 m wide (Fig. 3). No contact with Molasse bedrock or till is observed. In general, the deposits are massive yellowish, greyish glaciofluvial gravels, poorly-sorted and clast-supported. The cobbles have sub-angular to sub-rounded shapes, an average size of 10 cm and are often cemented. The northwestern part of the outcrop contains larger blocks with sizes ranging from 23 to 28 cm and partly openwork structure; whereas the southeastern part is characterized by a coarsening upward sequence where the lower part

contains cobbles with an average size of 8-10 cm, while boulders are up to 25 cm large in the upper section. Furthermore, in the central part of the pit, a gravel bed with large rounded blocks up to 60 cm in diameter is observed (Fig. 3). Throughout the outcrop, fractured clasts are found as well as disc-shaped clasts. The measured palaeoflow direction in the gravel pit is towards SSW (207°).

A lithostratigraphic section was logged through the southeastern part of the gravel pit showing the coarsening-upward sequence (Figs. 3, 5A). The 5.6 m-long section consists of a single massive sequence. We divided the section into three different parts based on grain size. The lower part (5.6-4.6 m from top) is 1 m thick (the lower contact is not exposed) and is composed of gravel and cobble with minor sand. In this part, the majority of the clasts are still intact, but crystalline clasts are strongly weathered and break apart when removed. Furthermore, a few holes resulting from clasts are observed. The petrographic investigation of the first level showed that approximately 70% of the gravels consist of limestone clasts (Fig. 5B). Vein quartz clasts represent the second largest group (11%), whereas quartzite occurs infrequently (<1%). In addition, some crystalline (9%) and chert clasts also occur (6%). Only very small quantities of dolomite, sandstone and radiolarite were encountered (1-2%).

The middle part (4.6-2.1 m from top) of the profile is 2.5 m thick and contains cemented cobbles with an average diameter of 20 cm. In some locations, pore spaces are filled with sand (Fig. 5C). Secondary calcite is sometimes observed. In this part, the crystalline clasts are more strongly weathered and break apart. Some clasts are fractured. Moreover, the surface of numerous calcareous clasts is weathered, indicating dissolution by groundwater. Clast holes are not uncommon. The petrography reveals that the limestone components represent the largest group (50%) while crystalline constituents represent the second largest group (20%) (Fig. 5C). Vein quartz and chert clasts are represented with 25%. The relative abundance of radiolarite, sandstone, dolomite and quartzite constituents remain constant in this layer. Finally, a few calcareous tuff gravels are encountered (1%).

The uppermost part (2.1-0.3 m from top) is characterized by 1.8 m of blocks that have an average diameter of 25 cm in a sandy-clay matrix. The matrix displays a beige-brownish colour due to iron

oxides and hydroxides (Blume et al., 2010) (Fig. 5D). Intact clasts are scarce. Most clasts are strongly weathered and decomposed, leading to numerous clast holes (30%). Clast counts show that limestone constituents have disappeared except for the most resistant siliceous limestone components (7%). The most common lithological group is made up by crystalline constituents (36%). The hard components including quartzites, cherts and radiolarites as well as sandstones and calcareous tuffs have slightly higher percentages than in the middle part of the section. It is, however, not ideal to use such a weathered horizon (the uppermost part in this case) for provenance tracing as dissolved clasts could add a bias in the analysis (Graf, 1993). Therefore, we discard the counting results of the uppermost part (2.1-0.3 m from top) for further discussion.

Figure 5. A: Field photograph showing an overview of the sampled stratigraphic profile. The profile is subdivided into three different levels (B, C and D). B-D: Petrographic composition of the different parts of the section. The white squares show the location of the depth-profile samples.

The results of the petrographic investigations of the three different horizons are summarized in a ternary diagram (Fig. 6). The limestone and dolomite components are grouped and plotted on top, while quartzite, vein quartz, radiolarite, and chert constituents plot on the bottom right apex. The crystalline components are finally presented in the bottom left apex of the diagram. The small amount of limestone gravels, an increase in vein quartz, quartzite and chert components and multiple clast holes indicate a transition from Deckenschotter to soil (e.g. Conradin, 1991) (Fig. 6B).

Figure 6. Results of the clast petrography at Stadlerberg. A: Stratigraphic column of the profile. The lithofacies code is based on Miall (1978) and Keller (1996). cGcm: massive, clast-supported gravel with cobbles; bGmm: massive, matrix-supported gravel with blocks; Fmp: massive fine sediments with pedogenic features. B: Ternary diagram illustrating the results.

We used the shape of the clasts to determine whether the accumulated gravels at Stadlerberg consist of reworked Miocene Molasse material rather than ‘fresh’ erosional products of the Alps. We

proceeded by calculating roundness and flattening index(es) from vein quartz clasts. The results, illustrated in Figs. 7A and B, reveal roundness indices (Z_i) that range from 100 to 600 with a median value $Md(Z_i) = 272$. A bimodal distribution is observed in the histogram, with a primary mode between 150 and 250 and a secondary mode between 300 and 400. The flattening indices (A_i) vary from 100 to 400 with a few outliers between 600 and 650. The median has a value of $Md(A_i) = 188$.

Figure 7. Histograms showing the results of the clast morphometric studies on vein quartz. A: Roundness histogram. The roundness index $Md(Z_i)$ is given by the median with 272. B: Flattening histogram. The flattening index $Md(A_i)$ is given by the median with 188.

4.2 Cosmogenic nuclide dating

AMS-measured $^{10}\text{Be}/^9\text{Be}$ and $^{26}\text{Al}/^{27}\text{Al}$ ratios (with uncertainties) as well as calculated ^{10}Be , ^{26}Al concentrations and $^{26}\text{Al}/^{10}\text{Be}$ ratio for each sample are shown in Table 1. Results of leaching tests to reduce aluminium are also shown. The concentrations of the seven sediment samples vary from $2.26 \pm 0.15 \times 10^4$ atoms/g for the lowermost sample to $10.34 \pm 0.43 \times 10^4$ atoms/g for the uppermost sample. In Fig. 8, the ^{10}Be concentrations together with 1σ uncertainties are plotted against depth. They display an exponential decrease with depth.

Figure 8. Measured ^{10}Be concentrations including the 1σ uncertainties of the STDL-DP samples plotted against depth.

The results of the Monte Carlo simulation are given in Table 3 and modal output with the frequency distributions and χ^2 values of exposure age, erosion rate and inheritance are shown in Fig. 9. The best fit through the seven data points is illustrated in Fig. 10A, whereas the possible solution space with a χ^2 cut-off value of ≤ 5 is shown in Fig. 10B. The simulation yields a best-fit solution to the measured nuclide concentrations for: a modal depth-profile age of 1857^{+193}_{-177} ka, a modal top erosion rate of $5.0^{+1.2}_{-1.1}$ cm/ka and for an inheritance of $19'000^{+12'000}_{-18'500}$ atoms/g (Table 3). The modal values are similar to the mean and median values of the simulation, thus the errors of the modal values are based on the minimum and maximum values generated by the simulation. The minimum χ^2 value for the best fit reveals an age of 1771 ka with an erosion rate of 4.7 cm/ka, and an inheritance of 23'000 atoms/g (Table 3). Note that the Monte Carlo simulation code requires a constraint on the net erosion on the top of the section as modal input parameter to calculate an age. This parameter is adjusted iteratively within a range of values (Hidy et al., 2010).

Table 3. Results of the Monte Carlo simulations with Matlab

¹⁰ Be	Age (ka)	Inheritance (10 ⁴ atoms/g)	Erosion rate (cm/ka)
Mean	1867	1.8	5.0
Median	1869	1.8	5.0
Mode	1857	1.9	5.0
Minimum χ^2	1771	2.3	4.7
Maximum	2050	3.1	6.2
Minimum	1680	0.5	3.9

Total number of simulated profiles is 100'000. For the minimum and maximum values a χ^2 cut-off value of ≤ 5 is applied. Please note that erosion rate statistics cannot be used for reporting a value for erosion rate (Hidy et al., 2010).

Figure 9. Modal output of the Monte Carlo simulations showing frequency distributions and χ^2 values for exposure age, erosion rate and inheritance.

Figure 10. Output of the Monte Carlo depth-profile age simulation. A: illustration of the best fit through the samples for the lowest χ^2 value. B: possible solution space with a χ^2 cut-off values of ≤ 5 .

Only 3 of the 9 quartzites and vein quartz clasts that were collected for isochron burial dating provided the necessary low total aluminium concentrations (< 30 ppm) and > 50 g pure quartz for further chemical treatment and subsequent $^{26}\text{Al}/^{27}\text{Al}$ AMS-measurements (see Table 1). It made no sense to further process samples STDL-5 and STDL-7 as they showed a too high (> 70 ppm) total

aluminium concentrations (Table 1). This would result in low $^{26}\text{Al}/^{27}\text{Al}$ ratios and as a consequence lead to large uncertainties (Çiner et al., 2015). The measured ^{26}Al concentrations are plotted versus ^{10}Be concentrations including 1σ uncertainties (Fig. 11). Unfortunately, it is not possible to calculate an isochron-burial age from these three samples as they show no scatter in data (see Balco and Rovey, 2008; Erlanger et al., 2012 for details). Furthermore, they do not plot on one line, which is due to the fact that they probably have similar pre-burial histories. Given that no age can be determined from these samples, isochron burial dating is not further discussed in this paper.

Figure 11. Measured ^{26}Al concentrations plotted vs. ^{10}Be concentrations of the isochron burial dating STDL samples. The errors represent 1σ uncertainties. The black line illustrates the surface production rate ratio of 6.75 (Balco et al., 2008).

4.3 Palaeo-denudation rates

The ^{10}Be concentration of the lowermost sample (STDL-DP-7) from the depth profile is $2.26 \pm 0.25 \times 10^4$ atoms/g (Table 1). Note that at a depth of 5.2 m, the post-burial production is c. 1% and therefore negligible (after Granger et al., 1997). This measured nuclide concentration was corrected for radioactive decay during 1.9 Ma, which yielded a value of $6.18 \pm 0.41 \times 10^4$ atoms/g. This concentration was then normalised to SLHL, which corresponds to $3.25 \pm 0.22 \times 10^4$ atoms/g.

For comparison purposes, we used the concentrations of the largest catchments located in the northern part of the Central Alps (Fig. 12). In addition, we gave preference to those streams that have similar lithologies in their headwaters as the source area inferred for the deposits at Stadlerberg: Sense, Emme, Kleine Emme a and b, and Reuss a and b basins (Wittmann et al., 2007) (Table 1). We also included the results from basins N3 and N8 situated on either side of the Niesen mountain (Chittenden et al., 2014) (Fig. 12; Table 4). ^{10}Be concentration of $(3.25 \pm 0.22) \times 10^4$ atoms/g is similar to the concentrations measured and normalised to SLHL in the Sense basin where a denudation rate of 0.3 mm/a has been reported (Table 4). The two catchments at the Niesen, where modern denudation rates

are 0.4 and 0.3 mm/a, respectively, reveal also comparable concentrations to STDL-DP-7 (Table 4). The normalised ^{10}Be concentrations of samples Emme, Kleine Emme a and b, and Reuss a and b are too small to be comparable to STDL-DP-7. Based on these comparisons, we infer a palaeo-denudation rate in the order of 0.3 to 0.4 mm/a for the sedimentary records at Stadlerberg.

Figure 12. The ^{10}Be concentration at Stadlerberg at time of deposition was compared to similar ^{10}Be concentrations of Wittmann et al. (2007) and Chittenden et al. (2014) measured in river sediments in the Central Alps of Switzerland (see text for details). The values near the catchments represent denudation rates in mm/a. All measured ^{10}Be concentrations were normalised to sea-level, high latitude (SLHL) using the scaling scheme of Stone (2000) (St) for spallation with a production rate of 3.93 atoms/g (Balco et al., 2009).

Table 4. Cosmogenic ^{10}Be data and derived denudation rate data.

Sample	Latitude (°N)	Longitude (°E)	Altitude	^{10}Be concentration	Production rate	Scaling factor	^{10}Be concentration normalised to SLHL	Denudation rate
name	(DD.DD)	(DD.DD)	(m a.s.l.)	(10^4 atoms/g)	(atoms/g.a)		(10^4 atoms/g)	(mm/a)
STDL-DP-7	47.5337	8.4458	610	6.18 ± 0.41	7.47	1.90	3.25 ± 0.22	$0.3-0.4^*$
Sense ¹	46.8251	7.3207	547	3.99 ± 0.75	6.99	1.78	2.24 ± 0.42	0.3
Emme ¹	47.0319	7.6360	600	3.54 ± 0.42	7.49	1.91	1.86 ± 0.22	0.3
Kleine Emme a ¹	47.0517	8.2217	470	2.43 ± 0.48	6.64	1.69	1.44 ± 0.28	0.4
Kleine Emme b ¹	47.0517	8.2217	470	1.88 ± 0.46	6.64	1.69	1.11 ± 0.27	0.5
Reuss a ¹	46.6350	8.8483	453	1.45 ± 0.43	5.89	1.50	0.97 ± 0.29	1.3
Reuss b ¹	46.6350	8.8483	453	1.00 ± 0.42	5.89	1.50	0.67 ± 0.28	1.9
N3 ²	46.6074	7.6527	877	4.68 ± 0.19	9.24	2.36	1.99 ± 0.08	0.4
N8 ²	46.6321	7.5631	842	5.95 ± 0.36	9.04	2.30	2.59 ± 0.16	0.3

Measured ^{10}Be concentrations were normalised to sea-level, high latitude (SLHL) using the scaling scheme of Stone (2000) (St) for spallation with a production rate of 3.93 atoms/g (Northeastern North America production calibration dataset; Balco et al., 2009).

*Denudation rate estimated based on comparisons with the catchments Sense, Emme, N3 and N8.

¹Data is from Wittmann et al. (2007).

²Data is from Chittenden et al. (2014).

5. Discussion

In the next section, we will use clast fabric analysis to characterize the sedimentary environment at Stadlerberg and to infer the transport mechanism. In addition, results of clast shape investigation will build the basis to evaluate if the sediments were derived from the Alps or if they represent material recycled from Miocene Molasse conglomerates in the foreland. The results of the petrographic analyses of the clasts will serve to allocate the provenance of the material. As final steps, we will place the depth-profile age inferred for the Deckenschotter at Stadlerberg into a general framework of landscape evolution that also considers the estimate palaeo-denudation rate recorded by these deposits.

5.1 Sedimentary environment

The massive fabric of the Stadlerberg Deckenschotter together with the poorly-sorted, mainly clast-supported framework (where the matrix consists of finer sand fractions, silt and clay) suggests limited winnowing and short transport distance. Such a fabric has commonly been described for material that was derived from a glacier (Edwards et al., 1975; Etienne et al., 2003). These observations suggest that the material was deposited on a braided stream (Miall, 1977; 1996) in the forefield of a glacier (Benn and Evans, 2010). Furthermore, imbricated clasts are encountered, which are a characteristic feature of gravel bed river systems (Johnston, 1922; Miall, 1977; 1996; Bluck, 1979; Collinson, 1996). Additionally, the coarsening-upward sequence in the southeastern part of the gravel pit can be explained either through downstream bar migration, which might be related to a glacier advance upstream in the catchment; or by an increase in water velocity with time (Miall, 1985; 1977). In either case, these observations suggest that the depositional environment of this Deckenschotter deposit is a glacier proximal, braided river system with longitudinal bars.

In addition, the presence of clasts of cellular dolomite (*Rauhewacke* in German) is an indicator for very short fluvial transport distance, i.e. the clasts from the Alps were brought by the glaciers to the Foreland and then deposited by the rivers. Any longer fluvial travel distance would have destroyed these delicate clasts. The upstream distance to a glacier's snouts can be inferred from the average

sizes of the clasts. In this context, an average grain size of 10 cm at Stadlerberg would correspond, for example, to a distance of 200 m from the ice margin of the Lewis River on Baffin Island (Canada) (after Church, 1972), 20 km from the source of the Skeidarar River in Iceland (after Boothroyd and Nummedal, 1978) or 6-18 km from the Younger Dryas end moraines of the Scandinavian Ice Sheet in southern Norway (Andersen, 1960). Thus, the grain size population at Stadlerberg suggests that the upstream distance to the palaeo-ice margin might range between a few hundreds of metres to few tens of kilometres. In either case, this implies that the glaciers have stretched down to the foreland at that time. It is important to note that the decrease in grain size downstream is also influenced by the change of the stream gradient, the lithology of the rock and the initial clast size (Maizels, 2002). It is also proportional to the length of the river system (Bluck, 1987).

Clast morphometric analysis of the gravels at Stadlerberg provides supporting information for the depositional environment and gives information on the origin of the sediments. The median of the roundness index ($Md(Z_i)$) from vein quartz clasts of 272 supports the inferred glaciofluvial environment (e.g. Cailleux and Tricart, 1959; Schlüchter, 1976). In contrast, crystalline clasts of ‘Nagelfluh’ (Molasse conglomerates) are commonly characterized by Z_i values of around 400 (Schlüchter, 1976). Based on these results, the secondary peak between 300 and 400 of the bimodal distribution in our Z_i value diagram most likely represents the better rounded, reworked material from the Molasse, whereas the first peak rather represents sub-angular clasts with a provenance from the Alps (Fig. 7). In addition, information can be gained by looking at the flatness of the clasts whether they are equant or flat (Schlüchter, 1989). In our study, the combination of sub-angular and equant clast shapes indicates glaciofluvial transport. The outliers with A_i values between 600 and 650 are identified as flat disc-shaped clasts and are an indicator of fluvial transport (Sneed and Folk, 1958).

5.2 Provenance of the material

Provenance studies enable to trace the history of sediment from the initial source area to the accumulation area (Weltje and von Eynatten, 2004). In the Alps, Quaternary sediments, which were derived directly from the mountains to the Alpine Foreland bypassing the sedimentary traps of the

alpine marginal lakes, can only be correlated with a glacial advance (Schluchter, 1989). As the gravels at Stadlerberg were deposited close to the snout of the glacier, key lithologies were analysed in order to identify the palaeoglacier system. The few Verrucano (red breccia clasts) point to transport by the Linth palaeoglacier (Du Pasquier, 1891; Frei, 1912; Hantke, 1980). Key lithologies that are characteristic for the Reuss and Valais palaeoglaciers were not found (e.g. Windgällen Porphyry, Niesenbrekzie, Mont Blanc Granite or Arolla Gneiss; after Hantke, 1980). In the same sense, because of the absence of ophiolites at Stadlerberg, the Grisons in the east and Valaisian Alps in the southwest can be excluded as a provenance, given that ophiolites are particularly encountered in the Arosa zone or Platta nappe (Steinmann, 1905) and in the Zermatt-Saas Fee unit (Bearth, 1967) (Fig. 13).

Quartzites are found in situ in the St. Bernhard nappe in the Valaisian Alps and in the Arosa zone in the Grisons Alps (Richter, 1930; Sartori et al., 2006) (Fig. 13). In addition, they occur as clasts within the Miocene Molasse conglomerates in the Alpine Foreland (Pavoni, 1957; Matter, 1964). Given that the Grisons and Valaisian Alps have already been excluded as source terrains because of the absence of ophiolites as explained above, the quartzites were most likely derived through erosional recycling of the Miocene Molasse conglomerates. The crystalline clasts encountered at Stadlerberg are strongly weathered and decomposed and it was not possible to recognise the colour. Therefore, these clasts could have been derived either from Miocene Molasse conglomerates (Matter, 1964), or from the external massifs such as the Aar Massif. The same interpretation can be made for the numerous limestone clasts that could have been derived from the Helvetic nappes or Penninic thrust nappes north of the Aar Massif where these lithologies are abundant (McCann, 2008) or from the Molasse conglomerates where these clast types have been frequently encountered (Matter, 1964) (Fig. 13). The coarse limestone clasts with a diameter of 60 cm are probably directly derived from the Alps as no clasts of this size have been reported from Molasse conglomerates (Matter, 1964). Taking into account the observations made before, the resulting catchment extent of the HDS deposits at Stadlerberg probably comprised the Miocene conglomerates in the Molasse Basin as well as some parts of the Alps including the Aar and Gotthard Massifs and their northern side as shown in Fig. 13.

Based on clast morphometry and the catchment extent, the reworked part from the Molasse Basin makes up approximately 50%.

Figure 13. Map showing the tectonic units (©Federal Office of Topography, swisstopo, 2005) from which the main lithologies found at Stadlerberg could be derived. Aa: Aar Massif, AR: Aiguilles Rouges Massif, GM: Gotthard Massif, MB, Mont Blanc Massif; Sil: Silvretta Nappe, Su: Suretta Nappe; Ar: Arosa zone, Pl: Platta Nappe, Zs: Zermatt-Saas Fee zone. The black line illustrates the possible catchment extent of the HDS deposits at Stadlerberg.

5.3 Age of the deposits

Deckenschotter deposits are topographically distinct and occur as discontinuous units of variable ages (Frei, 1912; Graf, 1993; 2009a). Recent analysis of the top surface of the Höhere and Tiefere Deckenschotter units indicate that there is: (i) a general topographic dip towards Basel in both HDS and TDS surfaces and (ii) a relatively large relief that developed between both units (Heuberger and Naef, 2014). Therefore, the reversed stratigraphic relationship can only be used in the sense of relative age assignments. The minimum age of 1.9 ± 0.2 Ma for the site Stadlerberg at 610 m a.s.l. lies within the time interval reported from mammal teeth for the site Irchel at 675 m a.s.l.: 2.5-1.8 Ma (MN 17; Bolliger et al., 1996). The age at Stadlerberg is older than the Deckenschotter deposits at Mandach at 510 m a.s.l. with 1.02 ± 0.1 Ma determined from cosmogenic depth-profile dating (Akçar et al., 2014)

(Fig. 1). So far, the reconstructed chronologies seem to support the attributed morphostratigraphy. In addition, the ages agree within the errors with those reported by Kuhlemann and Rahn (2013). In particular, these authors compared magnetic periods with the variation of the oxygen isotope record (Lisiecki and Raymo, 2005) and estimated a likely time span for Deckenschotter deposition of 2.05 to 2.50 Ma.

The onset of glaciations in the northern hemisphere is assumed to have occurred at approximately 2.7 Ma (Maslin and Ridgwell, 2005). From the reconstructed age of 1.9 ± 0.2 Ma at Stadlerberg, we cannot directly answer whether the onset of glaciations in the Alps was synchronous with global cooling; yet we conclude that it took place at, or before 1.9 ± 0.2 Ma. Hence, the glaciofluvial deposits at Stadlerberg with a distance of few hundred metres to few tens of kilometres to a palaeoglacier potentially documents one of the first glacier advances in the Swiss Alpine Foreland.

5.4 Palaeo-denudation rates and geomorphic implications

Estimating palaeo-erosion rates at different timescales allows identifying landscape changes in response to climate and tectonics (Whipple and Tucker, 1999). The estimated palaeo-denudation rate at Stadlerberg between 0.3 and 0.4 mm/a is comparable to the mean denudation rate of c. 0.3 mm/a calculated for modern river sediments of the Alpine Foreland (Wittmann et al., 2007; Norton et al., 2008; Büchi et al., 2014). These rates have been related to rather low relief, altitude, hillslope gradients and low uplift rates (same authors). Accordingly, we interpret that the moderate palaeo-denudation rate inferred from ^{10}Be at Stadlerberg points to a landscape where a large portion of the hillslopes were at failure, or less inclined (Ouimet et al., 2009). Morphometric analyses have illustrated that such conditions currently prevail in the foreland surrounding the Miocene conglomerates which were a source area for the deposits at Stadlerberg. For instance, in the Napf area, modern ^{10}Be -based denudation rates range between 0.3 and 0.5 mm/a, while in the Hörnli area farther to the east, they have average values of 0.35 mm/a (Norton et al., 2008; Büchi et al., 2014). In the Alps, however, the current picture is different where U-shaped valleys with oversteepened flanks, steep shoulders and multiple inner gorges predominate (Norton et al., 2010; Schlunegger and Norton,

2013). In these valleys, basin-averaged denudation rates range between 0.4 and 1 mm/a and more, most likely because of the frequent occurrence of mass wasting processes on oversteepened steep hillslopes and inner gorge channels (Norton et al., 2010; Kober et al., 2012; Schlunegger et al., 2013). Accordingly, for the time of c. 2 Ma when the Stadlerberg gravels accumulated, we infer a landscape that was most likely characterized by smoother hillslopes than at present. This is in agreement with findings of other authors, e.g. Häuselmann et al. (2007b). Note, however, that our estimate is a lower bound of the palaeo-erosion rate and we estimate a minimum ^{10}Be integration time of 3000 years for this erosion rate (Granger et al., 1996). Note also that the here reported basin-averaged denudation rates correspond to estimates based on independent methods such as compilations of apatite fission track-based exhumation rates, and estimates of foreland basin inversion amounts (Mazurek et al., 2006; Cederbom et al., 2011; Baran et al., 2014).

5.5 Landscape evolution after 2 Ma

Fluvial incision into underlying bedrock can be estimated when absolute ages of terraces are available (e.g. Burbank et al., 1996; Brocard et al., 2003; Schaller et al., 2005; Häuselmann et al., 2007b). Thus, the depth-profile age of 1.9 ± 0.2 Ma allowed quantifying a post-depositional incision rate for the surrounding of the Stadlerberg site. From drillings and interpolations the vertical elevation difference between the base of the Deckenschotter at the study site and the bedrock below the Rhine River measures c. 220 m (Heuberger and Naef, 2014 and references therein). The consideration of this vertical height difference and age yields a bedrock incision rate of 0.12 mm/a (Fig. 14). This value is compared to incision rates in the northern Alpine Foreland reported by others (e.g. Häuselmann et al., 2007b, Kuhlemann and Rahn, 2013; Akçar et al., 2014) (Fig. 14). Akçar et al. (2014) dated the HDS deposits at Mandach to 1 Ma and estimated a bedrock incision rate of 0.1 mm/a for post-HDS times by using the depth of incision of 110 m into bedrock (after Graf, 1993). Kuhlemann and Rahn (2013) revealed an incision in the period between 2.05 Ma and c. 1.2 Ma of 30-110 m into bedrock resulting in an average rate between 0.04 and 0.13 mm/a. From 0.8 Ma until present, a higher rate ranging from 0.18-0.33 mm/a (Kuhlemann and Rahn, 2013) was obtained, which is explained by an acceleration of the erosion in the Alps (Kuhlemann et al., 2002). A similar value than the one at Stadlerberg (0.12

mm/a) results when considering an average rate from their study for the time span between 2.05 Ma until the present. In contrast to the two previous studies, Häuselmann et al. (2007b) observed that glacial valley incision rates abruptly increased from 0.12 mm/a to 1.2 mm/a at 0.8-1.0 Ma, which corresponds to the timing of the Mid-Pleistocene Revolution (MPR; e.g. Schlüchter and Müller-Dick, 1996; Maslin and Ridgwell, 2005). However, caution has to be taken when comparing the values of Häuselmann et al. (2007b) with the one estimated at Stadlerberg. The site studied by Häuselmann et al. (2007b) is located within the core of the Alps, where the chronology of valley lowering was different than in the foreland without an explicit link. The northern Swiss Alps show a more pronounced valley topography where valley glaciers were dominant in contrast to the Alpine Foreland with advancing piedmont glaciers. High rates of several millimetres per year seem to be typical for eroding valley glaciers, as observed e.g. in the Swiss Alps, Scotland, Norway and Alaska (Hallet et al., 1996; Hebdon et al., 1997; Riihimäki et al., 2005). A possible explanation for the high rates could be that since 0.8 Ma, subglacial meltwater under pressure compared to fluvial incision, played a major part in the incision process (Kuhlemann and Rahn, 2013). To summarize, the quantified incision rate at Stadlerberg lies in the same order of magnitude as rates reported by Akçar et al. (2014) and Kuhlemann and Rahn (2013).

Figure 14. Variation of the erosion and incision rates during the last 5 Ma in the Alps and Alpine Foreland. The black curve shows the variation of the global marine oxygen isotope record (Lisiecki and Raymo, 2005). The Mid-Pleistocene Revolution (MPR), shown by the dashed line, marks the beginning of the transition between 41 and 100 ka cycles and is attributed to 0.95 Ma (Maslin and Ridgwell, 2005). The Neogene-Quaternary boundary is after Gibbard et al. (2010).

6. Conclusions

We studied the Höhere Deckenschotter outcrop in the abandoned gravel pit at Stadlerberg in the northern Swiss Alpine Foreland where we collected information on the genesis and timing of these deposits. We concluded that the gravels were transported to the study site mainly as bedload in braided rivers in a glaciofluvial system. The average clast diameter of 10 cm at Stadlerberg was compared to the clast size of recent proglacial outwash sediments and their distance from the ice margin. This comparison led to the conclusion that the gravels at our study site were deposited within a distance between a few hundred metres to few tens of kilometres from the feeding up-valley palaeoglacier. In addition, sedimentological analyses showed that the gravels at Stadlerberg are both reworked from the Molasse conglomerates as well as derived from the Alps. The better rounded clasts originated from the Molasse Basin whereas the more angular ones are rather fresh from the Alps. To conclude, the resulting catchment extent probably comprised the Hörnli and minor parts of the Napf

alluvial fans in the Molasse Basin as well as the Aar and Gotthard Massifs, Penninic thrust nappes and Helvetic nappes in the Alps.

^{10}Be depth-profile dating of seven sediment samples at Stadlerberg yielded a minimum age of 1.9 ± 0.2 Ma. Two of three $^{26}\text{Al}/^{10}\text{Be}$ ratios measured in quartz clasts for isochron burial dating were above the surface ratio, thus more clasts with ratios lower than surface ratio are needed to calculate an isochron-burial age. In conclusion, gravel deposition in the Swiss Alpine Foreland likely occurred at 1.9 ± 0.2 Ma. The HDS deposits at the study site document one of the first glacial advances into the Alpine foreland.

We estimated a catchment-wide palaeo-denudation rate of c. 0.3-0.4 mm/a for the study site based on inherited ^{10}Be concentrations, which agrees with published data (e.g. Baran et al., 2014). From the palaeo-denudation rate we inferred that c. 2 Ma ago, when the gravels at Stadlerberg were deposited, the landscape was most likely made of smoother hillslopes than at present. Finally, we quantified a long-term incision rate of 0.12 mm/a based on reconstructed chronology and a vertical incision of 220 m between the base of HDS and bedrock.

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8. References

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9. Figure captions

Figure 1. Extension of the Valais, Aare, Reuss, Linth and Rhaetian Lobes during the Last Glacial Maximum (from Bini et al., 2009); location of the study site Stadlerberg and distribution of the Deckenschotter in the northern Alpine Foreland (©Federal Office of Topography, swisstopo, CH-3084 Wabern). The black square marks the outline of Figure 2A.

Figure 2. A: Study area Stadlerberg and surrounding Deckenschotter sites. B: Cross section through Stadlerberg (A-B in Fig. 2A) based on geoelectric sounding (from Tarass, 1954).

Figure 3. Sketch of the abandoned gravel pit at Stadlerberg. The square inset shows the outline of Figure 3. Units 1-3 correspond to the studied stratigraphic section (Fig. 5A). The lithofacies code is based on Miall (1978) and Keller (1996). cGcm: massive, clast-supported gravel with cobbles;

bGmm: massive, matrix-supported gravel with blocks; bGcm: massive, clast-supported gravel with boulders.

Figure 4. Field photograph showing the abandoned gravel pit at Stadlerberg. Samples STDL-DP-1 to 7 were collected for depth-profile dating, while isochron burial dating was applied on samples STDL-1 to 9.

Figure 5. A: Field photograph showing an overview of the sampled stratigraphic profile. The profile is subdivided into three different levels (B, C and D). B-D: Petrographic composition of the different parts of the section. The white squares show the location of the depth-profile samples.

Figure 6. Results of the clast petrography at Stadlerberg. A: Stratigraphic column of the profile. The lithofacies code is based on Miall (1978) and Keller (1996). cGcm: massive, clast-supported gravel with cobbles; bGmm: massive, matrix-supported gravel with blocks; Fmp: massive fine sediments with pedogenic features. B: Ternary diagram illustrating the results.

Figure 7. Histograms showing the results of the clast morphometric studies on vein quartz. A: Roundness histogram. The roundness index $Md(Z_i)$ is given by the median with 272. B: Flattening histogram. The flattening index $Md(A_i)$ is given by the median with 188.

Figure 8. Measured ^{10}Be concentrations including the 1σ uncertainties of the STDL-DP samples plotted against depth.

Figure 9. Modal output of the Monte Carlo simulations showing frequency distributions and χ^2 values for exposure age, erosion rate and inheritance.

Figure 10. Output of the Monte Carlo depth-profile age simulation. A: illustration of the best fit through the samples for the lowest χ^2 value. B: possible solution space with a χ^2 cut-off values of ≤ 5 .

Figure 11. Measured ^{26}Al concentrations plotted vs. ^{10}Be concentrations of the isochron burial dating STDL samples. The errors represent 1σ uncertainties. The black line illustrates the surface production rate ratio of 6.75 (Balco et al., 2008).

Figure 12. The ^{10}Be concentration at Stadlerberg at time of deposition was compared to similar ^{10}Be concentrations of Wittmann et al. (2007) and Chittenden et al. (2014) measured in river sediments in the Central Alps of Switzerland (see text for details). The values near the catchments represent denudation rates in mm/a.

Figure 13. Map showing the tectonic units (©Federal Office of Topography, swisstopo, 2005) from which the main lithologies found at Stadlerberg could be derived. Aa: Aar Massif, AR: Aiguilles Rouges Massif, GM: Gotthard Massif, MB, Mont Blanc Massif; Sil: Silvretta Nappe, Su: Suretta Nappe; Ar: Arosa zone, Pl : Platta Nappe, Zs: Zermatt-Saas Fee zone. The black line illustrates the possible catchment extent of the HDS deposits at Stadlerberg.

Figure 14. Variation of the erosion and incision rates during the last 5 Ma in the Alps and Alpine Foreland. The black curve shows the variation of the global marine oxygen isotope record (Lisiecki and Raymo, 2005). The Mid-Pleistocene Revolution (MPR), shown by the dashed line, marks the beginning of the transition between 41 and 100 ka cycles and is attributed to 0.95 Ma (Maslin and Ridgwell, 2005). The Neogene-Quaternary boundary is after Gibbard et al. (2010).

10. Table captions

Table 1. Sample information and cosmogenic nuclide data of the samples from Stadlerberg.

Table 2. Input parameters for the Monte Carlo Simulator in Matlab (Hidy et al. 2010).

Table 3. Results of the Monte Carlo simulations with Matlab.

Table 4. Denudation rate data for the northern Alps derived from cosmogenic ^{10}Be .