



ARBEITSBERICHT NAB 24-15

Geological long-term evolution: Erosion

November 2024

**Nagra | National Cooperative for the
Disposal of Radioactive Waste**

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Abstract

This report supports the geosynthesis by presenting scenarios of future erosion relevant for site comparison and the safety case. Ensuring repository safety requires that barrier integrity is maintained for 100,000 years for low- and intermediate-level waste (L/ILW) and for one million years for high-level waste (HLW). These barrier properties are depth-dependent, sustained as long as a minimum residual cover remains above the Opalinus Clay. Over time, erosion driven by rock uplift and climate processes reduces this cover. This report addresses two main questions: (i) the expected remaining overburden thickness in each siting region after these timeframes, and (ii) the probability of erosional excavation after one million years.

The erosion assessment builds on understanding from Stage 2 of the Sectoral Plan for Deep Geological Repositories (SGT Stage 2), which considers past landscape evolution as a predictor of future trends. The landscape around the siting regions is shaped by progressive fluvial incision from the Aare and Rhine Rivers, development of low-relief cuesta topography, and deep glacial erosion driven by Alpine foreland glaciations. These processes are largely influenced by glacial/interglacial cycles and long-term rock uplift caused by Alpine tectonics, with generally low erosion rates. Past reorganisation of the Rhine and Danube drainage basins has caused significant episodic incision events; baselevel drops of similar magnitude are not expected in the next million years.

In SGT Stage 3, a comprehensive scientific field investigation programme was launched to analyse the Quaternary evolution of Northern Switzerland, covering the upper 300 meters of the subsurface with seismic data, drill cores, and new geochronology. Numerical models and literature reviews further developed understanding of erosion dynamics. One important outcome was that various anthropogenic CO₂ emissions may result in a delay of the next full glacial for several 100 kyr.

Erosion scenarios developed in SGT Stage 2 were refined to better incorporate and illustrate uncertainties. This includes a new model cascade that allows a combination of numerical simulations and probabilistic calculation cases to estimate future repository overburden and excavation probabilities, thereby integrating expert judgment for quantifying the uncertainties of model parameters. The erosion model resembles a cascading system: Model A (fluvial incision) provides input to Model B (evolution of local topography), and both feed into Model C (deep glacial erosion). Sensitivity analyses were conducted to gauge the impact of parameter uncertainties.

Model A simulates fluvial incision using the stream power incision model (SPIM) based on rock uplift, baselevel drop, erodibility, and drainage scenarios. After one million years, the most likely scenarios indicate that the provisional disposal areas in all siting regions remain well below the fluvial erosion base. However, in some scenarios, the Aare – Rhine River system may incise down to the level of the repository in JO. Accordingly, the JO site depends on the preservation of the additional cover given by the local topography, which is not necessary for Nördlich Lägern (NL) or Zürich Nordost (ZNO). Model B incorporates a 3D geology and simulates local river and hillslope responses to the fluvial incision (output of Model A), accounting for stable drainage networks and alternative channel configurations, particularly in JO. Results indicate that, while JO benefits from additional topographic cover, alternative drainage configurations may reduce the overburden, resulting in a disadvantageous reduction of remaining overburden thickness after one million years. Model C evaluates potential deep glacial erosion by combining fluvial incision (Model A) and drainage scenarios from Model B. Key inputs include depth distributions of glacial overdeepenings, bedrock erodibility, future glaciation scenarios, as well as the spatial occurrence and percentage of glacial overdeepenings within the siting regions.

The results show significant differences in remaining overburden thickness across the three siting regions at the locations of the provisional disposal areas. The most likely scenarios show at least 400 m of remaining overburden thickness above the repository after one million years at all the siting regions. In NL, nearly 600 m of overburden thickness is most likely. At the 5 – 95% probability range, between ~ 125 and 410 m of overburden thickness might be expected in JO, between ~ 350 and 670 m in NL and between ~ 220 and 560 m in ZNO. An excavation of the repository within one million years was estimated to be highly unlikely in all three siting regions. This estimated likelihood is about a magnitude lower for the provisional disposal area in NL than for JO and ZNO.

The erosion assessment shows that, with the geological evolution expected for the period under consideration, the overburden thickness will remain large enough to ensure robust self-sealing within the Opalinus Clay. Regarding less likely erosive events, the NL siting region is the most robust, particularly because of the greater depth of the repository. JO is the least robust, especially towards possible future local river diversions.

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List of Acronyms

a	Year (age)
ADN	Alternative (main) drainage network
<i>azi</i>	Azimuth sample
a.s.l.	Above sea level
<i>BL</i>	Baselevel at Basel
CDF	Cumulative Distribution Function
$\delta^{18}\text{O}$	Composition of stable isotopes of the water molecule in the delta notation. The delta symbol is a way to express the relative difference of isotope ratios between sample and the standard VSMOW (Vienna Standard Mean Ocean Water)
DEM	Digital Elevation Model
DRM	Detrital remanent magnetism
EC	Erodibility classes (grouping of similar rock units, used for the assessment of future erosion)
ECRIS	European Cenozoic Rift System
ELA	Equilibrium Line Altitude
ENSI	Swiss Federal Nuclear Safety Inspectorate
ERCS	Erosion rate from the concentration of solids (a procedure for measuring erosion rates with abrasion mill experiments, as a quantitative indicator of bedrock erodibility)
Fm.	Formation
FMT	Mont Terri rock laboratory
GIA	Glacial Isostatic Adjustment
GNSS	Global Navigation Satellite System
<i>gref</i>	Reference rock uplift gradient used in Model A
GTS	Grimsel Test Site
HDS	Höhere Deckenschotter
HLW	High-level waste (spent fuel assemblies and high-level waste from reprocessing)

HT	Hochterrasse
IRSL	Infrared stimulated luminescence
JO	Jura Ost siting region
<i>K</i>	Erodibility coefficient (parameter of SPIM)
K	Measure of the sensitivity of the critical orbital forcing to the atmospheric CO ₂ concentration
ka	Thousand years (age)
kyr	Thousand years (duration)
L/ILW	Low- and intermediate-level waste
LEB	Local erosion base
LGM	Last Glacial Maximum
LR04 stack	An average of 57 globally distributed benthic d ¹⁸ O records (which measure global ice volume and deep ocean temperature) collected from the scientific literature
<i>m</i>	Parameter of SPIM
Ma	Million years (age)
MEG	Most Extensive Glaciation
MIS	Marine Isotope Stage
Model A	Model component "Fluvial incision" of the model cascade for the erosion assessment
Model B	Model component "Local topography" of the model cascade for the erosion assessment
Model C	Model component "Deep-glacial erosion" of the model cascade for the erosion assessment
MPT	Mid-Pleistocene Transition
MuddPILE	Parsimonious Integrated Landscape Evolution model
Myr	Million years (duration)
n	Number of data points
<i>n</i>	Parameter of SPIM

NAB	Nagra Work Report ('Nagra Arbeitsbericht')
NL	Nördlich Lägern siting region
NT	Niederterrasse
NTB	Nagra Technical Report ('Nagra Technischer Bericht')
OD	Overdeepening
OPA	Opalinus Clay
OSL	Optically stimulated luminescence
P	Percentile (e.g. P25 = 25 th percentile of probability distribution)
pCO ₂	CO ₂ partial pressure
PDF	Probability density function
PgC	Petagrams of carbon (1 PgC is equal to 1 Gigatonne of carbon)
PSHA	Probabilistic Seismic Hazard Assessment
P-PINI	Particle-Pathway Inversion of Nuclide Inventories (Monte Carlo approach to model complex erosion and exposure histories and derived accumulation ages)
QBO	Quaternary borehole
QoI	Quantity of interest
RIO	Rational Impartial Observer (part of SHELF elicitation protocol)
S_{CH}	Characteristic slope angle
SFOE	Swiss Federal Office of Energy
SGT	Sectoral Plan for Deep Geological Repositories
SHELF	Sheffield Elicitation Framework
Sobol-Index (?)	Sensitivity measure to quantify the individual impact of input variables on a model output variance
SPIM	Stream Power Incision Model
SRTM	Shuttle Radar Topographic Mission (of NASA)
SSHAC	Senior Seismic Hazard Analysis Committee
T ₁ ...T _x	Time steps

TBO	Nagra's deep drilling campaign (9 deep boreholes drilled between 2019 and 2022 in the three siting regions of JO, NL, ZNO)
TCN	Terrestrial cosmogenic nuclide(s)
TDS	Tiefere Deckenschotter
<i>U</i>	Rock uplift
<i>UB</i>	Uplift at Basel
<i>UBR</i>	Uplift at Brugg
UCS	Unconfined compressive strength
<i>U grad</i>	Rock uplift gradient, modified as function of the actual azimuth sample (<i>azi</i>) used in Model A
wrt.	with respect to
yr	Year(s) (duration)
ZNO	Zürich Nordost siting region

1 Introduction

As one of the pillars of safety, the "the deep underground location of the waste emplacement rooms in a stable geological environment" is a central aspect of Nagra's safety strategy (Nagra 2024h). This basic requirement aims to ensure the efficient isolation of radioactive waste from the Earth's surface and the functionality and long-term stability of the multiple barrier system regarding geological and climatic evolution. In clay rocks such as the Opalinus Clay, the important parameters for an effective barrier function are very low hydraulic permeabilities, strong radionuclide retention (sorption), and a high self-sealing potential (Nagra 2024e).

Some of these properties are "depth-dependent" and persist as long as a minimum residual cover is maintained above the Opalinus Clay, i.e. above the main barrier of the deep geological repository (Fig. 1-1). With respect to the long-term evolution of the barrier, a minimum overburden thickness of 200 m is considered reliable for self-sealing in the Opalinus Clay. If the overburden thickness is reduced to 50 – 100 m, self-sealing may be limited, and if the overburden thickness is only 20 – 30 m, it may be ineffective (Nagra 2024e).

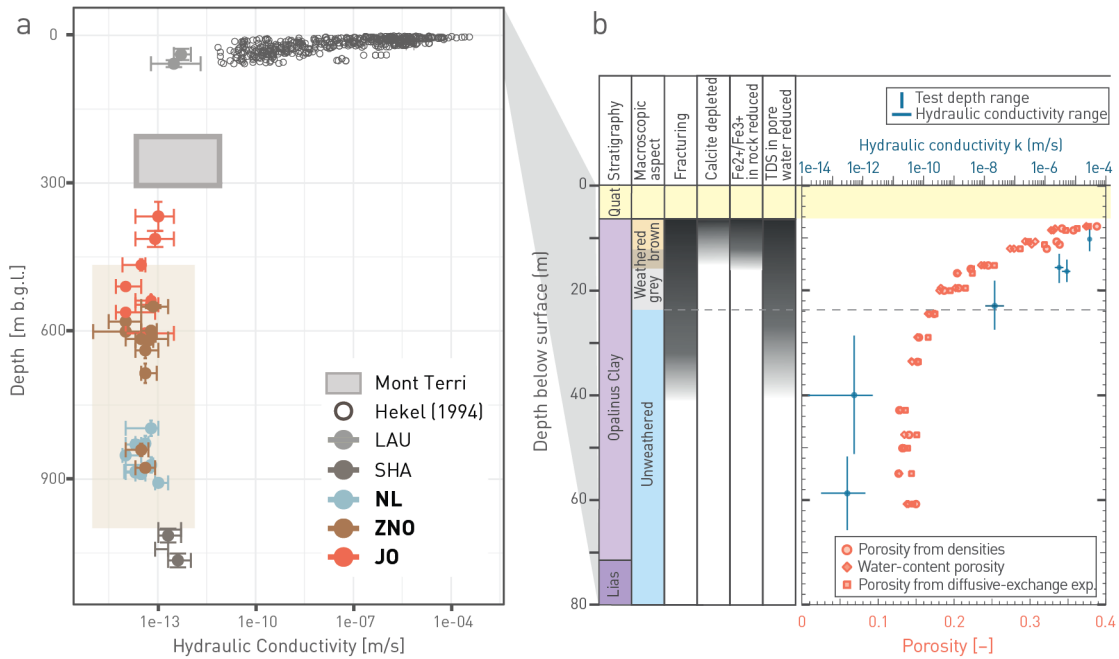


Fig. 1-1: Depth-dependence of barrier properties of the Opalinus Clay

(a) Hydraulic conductivity as proxy for depth-dependence of barrier properties (Hekel 1994, Nagra 2024e) (b) Changes in porewater chemistry, increased porosity, or fracturing in the Opalinus Clay when close to the surface highlight decompaction effects (Mazurek et al. 2023, Nagra 2024e). b.g.l.: below ground level, TDS: Total dissolved solids.

Erosion as a consequence of rock uplift and/or climate evolution results in a reduction of the overburden thickness on long timescales. Future erosion is therefore one of thirteen criteria (criterion 2.2) that must be assessed in relation to safety and technical feasibility of a deep geological repository (Table 1 in SFOE 2008, ENSI 2018). Together with the other three criteria of Group 2 (long-term stability), the erosion criterion addresses the long-term stability of the geological barrier (preservation of the geological barrier function) over the required time period

of one million years for high-level waste (HLW) and 100'000 years for low- and intermediate-level waste (L/ILW). Given these periods under consideration, the first approach to assessing future erosion is to understand similarly long time intervals in the past, i.e. the Quaternary period.

The Quaternary landscape of Northern Switzerland is shaped by the effects of multiple glacial/interglacial cycles (including multiphase deep glacial erosion and deposition), phases of sediment accumulation and fluvial downcutting. A summary of the past landscape evolution as known a decade ago is provided in Nagra (2014b), as well as Heuberger et al. (2014) and Schnellmann et al. (2014).

During SGT Stage 3 (SFOE 2008), a comprehensive scientific field investigation programme was launched to provide a better knowledge base for evaluating the Quaternary evolution of Northern Switzerland. Archives of Quaternary sediments from within the uppermost 300 m of the sub-surface were examined with 67 km of high-resolution 2D seismic reflection lines and 1.8 km of drill cores. Outcrops of fluvio-glacial deposits were analysed in field studies and laboratory tests. A substantial amount of new geochronology was carried out to advance the understanding of the age of the units and the rates of processes.

In addition, GNSS- and precise levelling-based vertical velocities were further assessed to develop an improved understanding of present-day uplift rates (Nagra 2024c). The digital model of the base of the Quaternary sediments (or top bedrock) from 2014 (Pietsch & Jordan 2014) was updated. New datasets of the depth to top bedrock (drilling data, 2D seismic lines) allow the model to be improved, which is one important input basis for evaluating the erosion history. The database for the model update also includes more than 700 drilling data points, 26 high-resolution datasets from 2D seismic investigations of glacial overdeepening structures (Nagra 2018, Gegg et al. 2021), and mapping of areas covered with Deckenschotter deposits (Section 2.4.1). Past studies combined with new investigations help to better assess the geometries of the glacial overdeepenings and their sedimentary fills, which can now be distinguished into typical lithofacies associations (see Section 2.4.2). To better understand the evolution of local topography, the landscapes of the Bözberg area (Jura Ost siting region, low surface-process rates) and the Wutach (high surface-process rates after river capture) were compared based on geomorphic criteria and denudation rates (Section 3.4). An important aspect in evaluating the role of different erosion processes and potential rate changes is the erodibility of different rock types. For this reason, abrasion mill experiments were conducted on rock samples from Northern Switzerland, enabling the distinction of different erodibility classes (Section 3.3.3.3). New geochronological information on fluvio-glacial deposits or fills in glacial overdeepenings allowed an improved and more detailed incision history to be established (see Section 2.4.1.3). These investigations were accompanied by numerical modelling studies (e.g. climate and ice flow, see Nagra 2024b) and data obtained from published work, with the ultimate goal of reaching a state-of-the-art process understanding of the main erosional drivers in Northern Switzerland.

Taken together, the key attributes needed for a characterisation of safety-relevant aspects are well explored at all sites. Age determination of the Quaternary geomorphic markers and sediments remains challenging. The remaining uncertainties cannot be further reduced in due time and are therefore sufficiently accounted for in a systematic assessment of future erosion processes. This required updating the different erosion scenarios and argumentation provided during SGT Stage 2.

Following (Schnellmann et al. 2014), the new assessment builds on a vertical division of the landscape into different geomorphic compartments (i.e. local erosion base, local topography, and glacial overdeepenings; see Section 3.1), each reflecting distinct erosion processes that formed them. During SGT Stage 2, however, so-called *base*, *optimistic* and *pessimistic cases* of erosional impact were identified based on evaluation of the Quaternary past. These cases were evaluated separately for fluvial and glacial processes and stacked accordingly (Nagra 2014b). For instance, the glacial base case was added to the fluvial plausible case etc., irrespective of their respective

likelihoods. Conversely, in the new erosion assessment presented here, a new model cascade is provided that allows a combination of numerical simulations and probabilistic calculations to incorporate uncertainties into the forecast of overburden removal. This probabilistic erosion assessment is the major addition of SGT Stage 3 (see Section 3.2).

This report provides an overview of the workflow and results of the assessment of the remaining overburden thickness both in terms of magnitude, and potential for (hypothetical) erosional excavation of the repository in the siting regions after the period under consideration. It summarises the geoscientific basis supporting the argumentation of the Geosynthesis (Nagra 2024e) with respect to erosion for site comparison and the safety case.

The assessment of future erosion potential builds on the general concept of considering processes forms and rates from Quaternary (last 2.6 Myr) as key to evaluating future variability. Knowledge about the interplay and feedback of these processes is important for the spatio-temporal evolution of erosion and associated uncertainties. Accordingly, Chapter 2 provides a summary of the Quaternary geology of Northern Switzerland with the focus on erosion, including the effects of regional uplift, changing past climates, such as ice-house conditions, and landscape evolution. From the Quaternary landscape evolution, the most important influencing quantities are identified as:

- the spatiotemporal distribution of rock uplift,
- the evolution of the local erosion base, local denudation of the topography, and
- deep glacial erosion.

For each of these influence quantities, the main important processes are identified and described in Chapter 3. By applying the new model chain developed in SGT Stage 3, the uncertainties of associated parameters are quantified (using expert elicitation) and their assumed total probabilities are provided together with the main argumentation (which in turn is largely based on the information provided in Chapter 2).

Chapter 4 provides the final calculations of the erosion assessment, i.e. the remaining overburden thickness as well as combined glacial and non-glacial excavation probabilities and their interpretation.

Chapter 5 presents a brief discussion of the plausibility of the results and the main influencing parameters on the erosion assessment that might eventually reduce the remaining overburden thickness.

Finally, the main conclusions drawn from the presented analyses and a rigorous comparison of the siting regions with respect to erosion are outlined in Chapter 6. The report finishes with a "storyboard" for the anticipated erosional long-term geological evolution of Nördlich Lägern (Section 6.2).

2 Pliocene to Quaternary landscape evolution in Northern Switzerland as a basis for assessment of future erosion

Tectonism and climate changes during the Quaternary have shaped the present-day landscape of Northern Switzerland. The main erosion processes that have influenced landscape evolution are the combined action of rivers, especially the drainage evolution of the Aare – Rhine River system, and erosion and sedimentation by repeated extensive Alpine Foreland glaciations. These processes, in turn, control topographic slopes and mass movements, and thus drive denudation of the topography. Consequently, to be able to address the future landscape evolution in Northern Switzerland and the potential influence of erosion on the repository overburden in the siting regions, the Quaternary tectonic and climatic boundary conditions need to be known and contextualised.

The next section starts with the geodynamic constraints that influence the uplift and subsidence pattern along the principal drainages from source to sink (Section 2.1). This is followed by a section on regional Quaternary climate conditions as a driver for erosion (Section 2.2) as well as a summary of processes leading to drainage reorganisation with potentially pulsed high erosion (Section 2.3). Section 2.4 summarises the Quaternary landscape evolution in Northern Switzerland. Importantly, the information presented here is used as basis to inform the model parameters for the assessment of future erosion presented in Chapter 3.

2.1 Uplift and subsidence as driver for erosion: regional consideration

The Aare – Rhine River system connects the Alps with the North Sea as final baselevel (Fig. 2-1a). On its source-to-sink path, the Rhine crosses several tectonic domains that may influence the overall erosive efficiency (Fig. 2-1b and Fig. 2-8). The siting regions are located along the Hochrhein section, between the Bodensee and the Upper Rhine Graben (Fig. 2-1a) and are thus influenced by the neotectonic setting of Northern Switzerland. The main neotectonic forcing factors are the ongoing Alpine orogeny in the south with uplift of the Molasse Basin, transtensional deformation in the Upper Rhine Graben and potential, but subordinate, tectonic activity of the Jura Fold-and-Thrust Belt and the Hegau – Bodensee Graben (see Nagra 2024c). This setting defines the main long-term boundary conditions for fluvial erosive processes that respond to uplift and baselevel drop.

Erosion in Northern Switzerland is further influenced by far-field effects associated with uplift of the Rhenish Massif and the Eifel area (Fig. 2-1c and Fig. 2-8). Uplift in the latter area generates a regional intermediate baselevel of the Rhine and is expected to decouple processes related to sea level changes from the siting regions at the relevant timescale. The main domains are summarised below from source to sink, but with a main focus on Quaternary to recent uplift or subsidence.

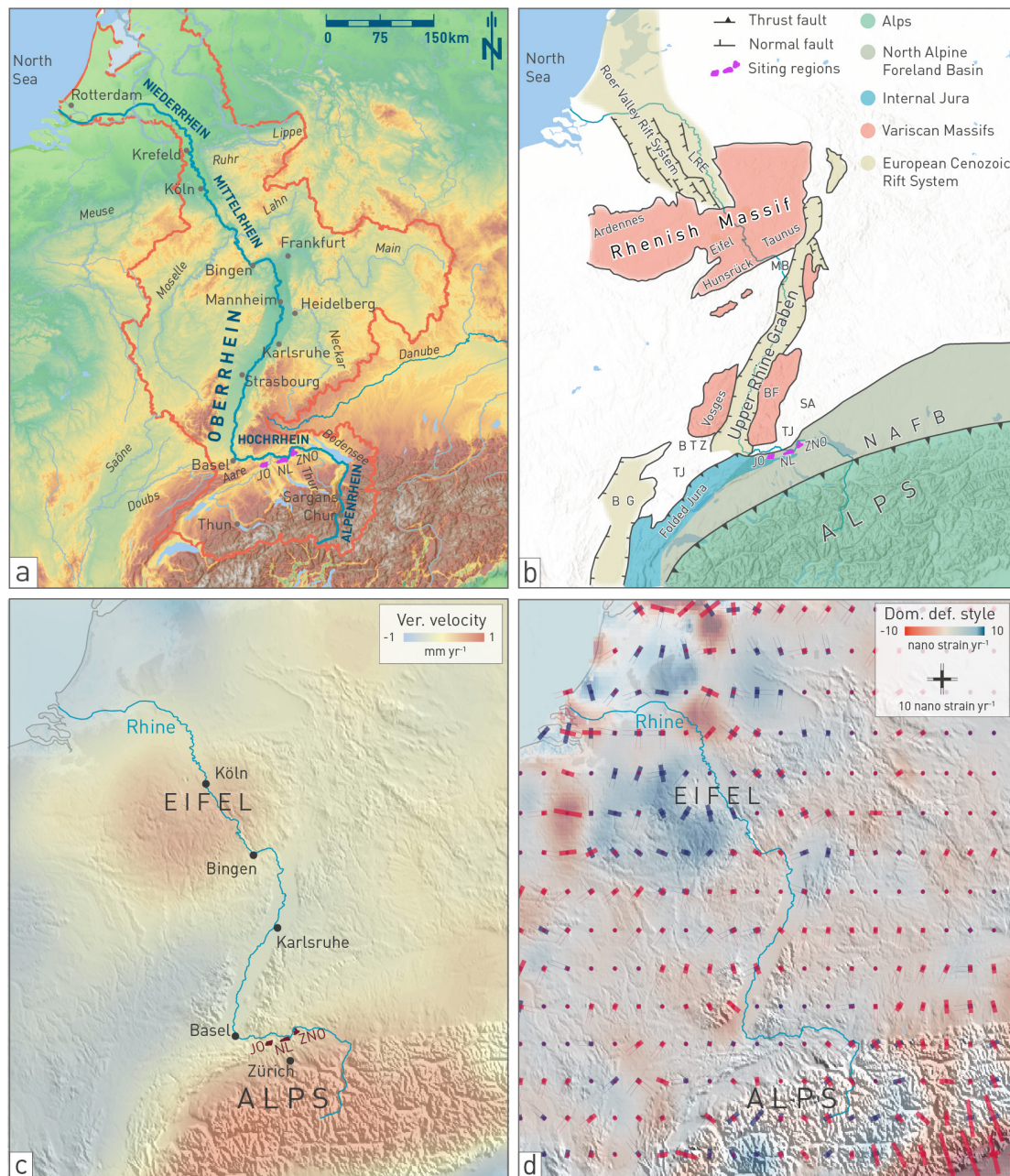


Fig. 2-1: Present-day Rhine catchment in a morphotectonic and geodynamic context for fluvial incision and sediment accumulation

(a) Rhine catchment. The course of the Rhine River is highlighted with a dark blue line. (b) Morphotectonic domains. BF: Black Forest, BG: Bresse Graben, BTZ: Bresse Transfer Zone, TJ: Tabular Jura, Ks: Kaiserstuhl, LRE: Lower Rhine Embayment, MB: Mainz Basin, NAFB: North Alpine Foreland Basin, SA: Swabian Alb. Data source: EU-DEM (European Environment Agency), rivers based on the EU-Hydro River Network Database 2006-2012 (vector), Europe, <https://land.copernicus.eu/en/products/eu-hydro/eu-hydro-river-network-database>. (c) Geodetic velocities (redrawn from Piña-Valdés et al. 2022). Vertical motion as a driver for erosion is highlighted by red and blue colours (uplift and subsidence, respectively). The Alps and the Eifel hotspot are prominent uplift regions that influence the profile of the Rhine River. (d) Strain rates (redrawn from Piña-Valdés et al. 2022) with red pins showing contraction and blue pins showing extension) are generally low within the Northern Alpine Foreland.

The Alps

Overall, the geodynamic setting of Central Europe is influenced by the convergence between the African and Eurasian plates (e.g. DeMets et al. 1990, Serpelloni et al. 2022, Nagra 2024c).

Based on global navigation satellite system (GNSS) measurements, the vertical velocity field of the Alps is in the order of ~ 2 mm/yr (local maxima at 2.5 mm/yr) along the northward-convex border of the Alps, increasing from the west to the central Alps with a generally positive correlation with topography (Sánchez et al. 2018, Piña-Valdés et al. 2022, Serpelloni et al. 2022, Pintori et al. 2022; see Fig. 2-1c for the general picture). Total decadal-scale uplift is generally higher in the areas that undergo less horizontal shortening, i.e. it is higher in the western and central Alps compared to the eastern part (Sternai et al. 2019). A tectonic contribution of shortening to rock uplift and exhumation appears to be pronounced in the Eastern and eastern Southern Alps, where ~ 2 mm/yr between the Adriatic plate and stable Europe are accommodated (Sánchez et al. 2018, Sternai et al. 2019, Serpelloni et al. 2022, Piña-Valdés et al. 2022).

Denudation often occurs as an erosional response to rock uplift. Basin-wide denudation rates from the Central Alps (Fig. 2-2) seem to correlate with rock uplift, inferred from geodetic measurements and are on average 0.27 ± 0.14 mm/yr for the internal Alpine Foreland and 0.9 ± 0.3 mm/yr for catchments within the Central Alps (Wittmann et al. 2007). However, the correlation is weaker in the Alpine catchments compared to regions within the internal Alpine Foreland. In a more recent compilation of denudation rates across the European Alps based on catchment-wide cosmogenic nuclide measurements, Delunel et al. (2020) evaluate first-order influences of topographic metrics, rock uplift, and environmental aspects on denudation rate. These authors find that, along strike, denudation rates increase from the western and eastern margins towards the centre of the Alps, where the highest rates are observed. Correlation with rock uplift is weak, and uncorrelated for steep catchments (slope $> 30^\circ$). In general, Delunel et al. (2020) find that mean rock uplift rates based on geodetic measurements exceed millennial denudation rates, and actual surface uplift occurs across most of the Alps. Along strike, however, it appears that the Western Alps are closer to equilibrium, the Central Alps gain elevation, while the Eastern Alps lose topography. Delunel et al. (2020) refer to deep crustal/mantle processes to explain such a pattern.

Going further back in time, Fox et al. (2015b, 2016) compiled and inverted thermochronological data of the Alps to reveal spatiotemporal exhumation patterns. As exhumation requires rock uplift, such patterns may reflect variability in uplift through time. Here, the focus is on the past four million years (Fig. 2-3), which represents the relevant neotectonic period for the assessment (see Nagra 2024c). Exhumation rates vary across the Alps but are in general below 1 km/Myr ($= 1$ mm/yr) between four and two million years ago (Fig. 2-3a). Fox et al. (2015b) inferred a two- to threefold increase in exhumation rates across the Western Alps and a small increase in the Central Alps approximately two million years ago (Fig. 2-3b; note that this is the temporal resolution, the actual increase may have occurred later, e.g. during the Mid-Pleistocene Transition). These authors relate the significant increase in exhumation of the Western Alps to a response to the progressive slab detachment, which is assumed to have affected only the western sector of the orogen (Lippitsch et al. 2003). This slab breakoff would have resulted in an isostatic response, further driving larger areas into the accumulation zone of large glaciers and thus enhancing glacial erosion.

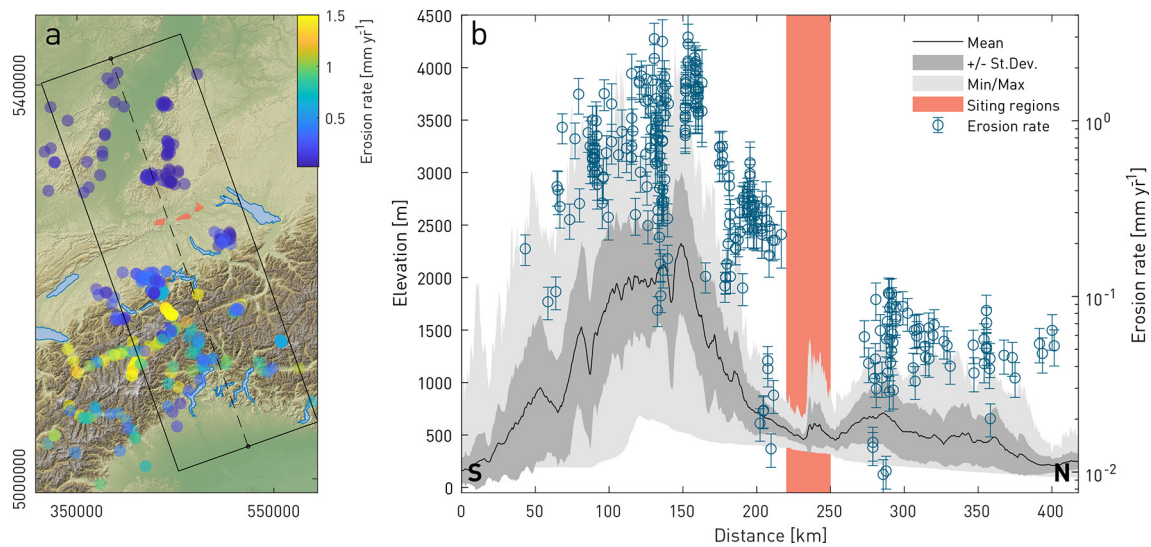


Fig. 2-2: Basin-wide erosion rates in the Alps, Alpine Foreland, Black Forest and Vosges

(a) Map showing spatial distribution of colour-coded erosion rates and outline of data swath profile shown in (b). Siting regions are highlighted in orange. (b) Data swath across the Alps, Alpine Foreland, Black Forest and Vosges, showing basin-wide erosion rates superposed on topography. Highest erosion rates correspond to areas of highest topography in the Alpine interior. Erosion rates in the Alpine Foreland, Black Forest and the Vosges are more than an order of magnitude lower than the highest Alpine rates. The data were acquired from the OCTOPUS database v.2.2 (Codilean et al. 2022) and comprise data from varying sources.

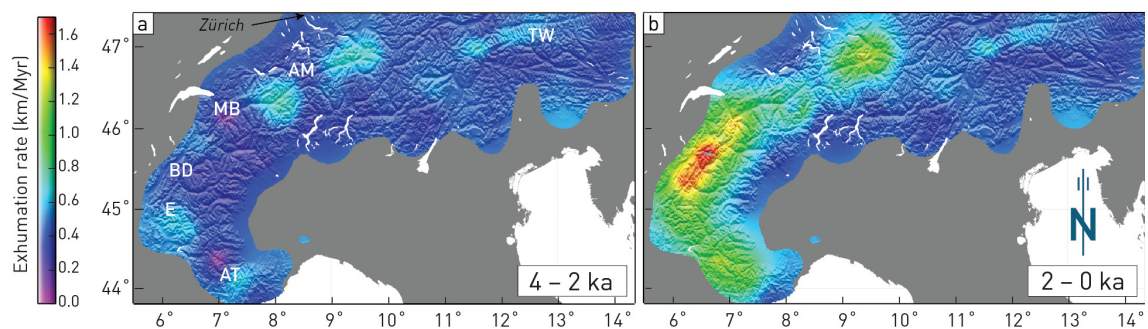


Fig. 2-3: Pattern and magnitude of Alpine exhumation during the past 4 Myr

Alpine exhumation rates over the past four million years (a) and in two-million-year intervals (b), inferred from analysis of thermochronometric data. Figure is redrawn and modified from Fox et al. (2015b), see there for temporal data resolution. Arrow marks the approximated location of the city of Zürich for orientation. Labels highlight important geological domains and massifs, as TW: Tauern Window, AT: Argentera, E: Ecrins, BD: Belledonne, MB: Mont Blanc, AM: Aar Massif. The main signal highlighted in the study of Fox et al. (2015b) is the high exhumation rate in the Western Alps in the past 2 Myr (b).

The tectonic drivers of ongoing Alpine uplift are superposed by the effects of significant buoyancy forces associated with isostasy (e.g. Sternai et al. 2019, Piña-Valdés et al. 2022). Isostatically driven processes, such as postglacial (e.g. Mey et al. 2016) or erosional rebound/unloading (Champagnac et al. 2009), progressive slab detachment (e.g. Fox et al. 2015b), as well as positive feedbacks due to the formation of dynamic topography and the establishment of horizontal gradients associated with gravitational potential energy, are considered to play a role – yet with different amounts in western and central Switzerland when compared to eastern Switzerland (Sternai et al. 2019; see Fig. 2-4). The proportion of glacial isostatic adjustment (GIA) versus other uplift-contributing processes in the Alps is currently debated, with authors attributing GIA only (e.g. Mey et al. 2016) and others including effects of erosion (e.g. Champagnac et al. 2009) or delamination (see Sternai et al. 2019 or Mazzotti et al. 2020 for a summary). However, buoyancy forces in response to deglaciation or erosion and sediment redistribution might cause crustal uplift in the order of 10 – 30% and 10 – 20% in the western and central Alps, respectively, and up to 50% and 30 – 40% in the eastern Alps (Sternai et al. 2019). It is important to note that GIA effects are highly exponential through time, with rapid uplift shortly after deglaciation and a subsequent slowing down of uplift (e.g. Turcotte & Schubert 2014, Grosset et al. 2023). Accordingly, the highest rates of crustal uplift would be expected immediately after ice retreat.

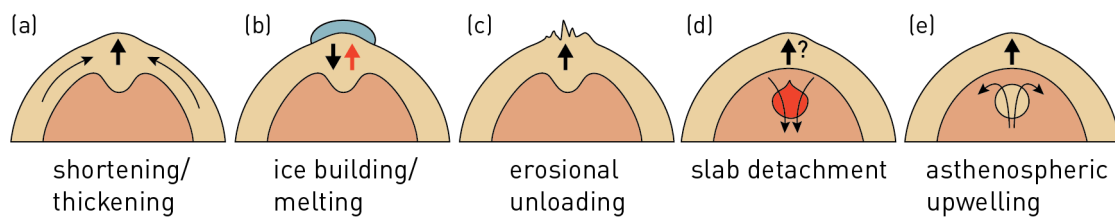


Fig. 2-4: Schemes of different models driving Alpine uplift

Models are redrawn from (Sternai et al. 2019) and illustrate the different models assumed to drive Alpine uplift.

Conclusion: The Alpine orogeny is the primary long-term tectonic driver for the topographic evolution of the Alpine mountains and the associated foreland. Orogeny is ongoing at low rates and, together with superposed buoyancy effects, results in spatiotemporally shifting loci of rock uplift. Deformation, topographic gradients, and associated erosion rates decrease significantly from the Alps towards the foreland. Geodetically derived short-term vertical motions are higher than rates based on long-term uplift proxies. This pattern is expected to continue in the future.

Alpine Foreland

The dominance of Alpine-related uplift (including both tectonically and isostatically driven processes) results in an uplift gradient oriented perpendicular to the Alpine front with rates that decline towards the NNW (Champagnac et al. 2009, Mey et al. 2016). Such a trend is visible in interpolations of proxy data for long-term thermochronology (Fox et al. 2015b; see Fig. 2-3), basin-wide erosion rates (Wittmann et al. 2007, Delunel et al. 2020; Fig. 2-2) and measurements of present-day motion (GNSS, precise levelling; see Fig. 2-5a and b). The roughly SSE to NNW trend is also reflected by the principal flow direction of the larger rivers traversing Northern Switzerland, suggesting a long-term tectonic control on the geomorphological evolution.

Using exhumation or fluvial incision rates as proxy for rock uplift (which is a main principal quantity used in this report) is a simplification that requires the assumption of steady-state topography. Although such a concept requires uniform erosion and tectonic stability, it is a valuable first-order measure on long timescales, when short-term fluctuations level out. However, because of the complexities in tectonically active regions that are also influenced by climate fluctuations, such an assumption needs to be treated with caution. Past exhumation of the Molasse Basin is summarised in Nagra (2024e). The amount of Miocene/Pliocene exhumation is generally estimated to be around 1 km (e.g. Mazurek et al. 2006, Villagómez Díaz et al. 2021, Omodeo-Salé et al. 2021, Schegg & Leu 1998). The onset of exhumation, however, is debated, with estimations ranging between 12 Ma and 5 Ma (von Hagke et al. 2015, Looser 2022, Mazurek et al. 2006, Villagómez Díaz et al. 2021). Under the simplest assumption of linear exhumation, inferred exhumation rates range between ~ 0.08 mm/yr and 0.2 mm/yr. However, several studies indicate heterogeneous exhumation rates in the past ~ 10 Myr.

Bedrock incision rates based on Quaternary geomorphic markers that are often used as proxy for rock uplift cannot be directly estimated within Northern Switzerland. Here, effects of uplift and exhumation are entangled with the impact of baselevel lowering resulting from changes in drainage routing (see Section 2.3). Nevertheless, a first-order assessment of bedrock incision can be calculated from bedrock elevation changes and the age of the Deckenschotter sediments (see Section 2.4.1). Two approaches were used: firstly, applying mean ages of 1.38 ± 0.31 Ma for the Höhere Deckenschotter (HDS) and 0.9 ± 0.13 Ma for the Tiefere Deckenschotter (TDS) and the lowest elevation for the respective Deckenschotter outcrops (Fig. 2-5c and d) to calculate incision rates with respect to the deepest fluvially eroded bedrock level (local erosion base, see Section 3.1). Associated incision rates are mainly between ~ 0.1 and ~ 0.2 mm/yr (Fig. 2-5c to e). Note that the incision signal not only includes a share from drainage reorganisation but is also a mixed signal stemming from rock uplift of the Alpine Foreland and baselevel fall at the transition to the Upper Rhine Graben. This is probably reflected in the interpolated isolines of HDS and TDS incision in Fig. 2-5c and d, which are slightly rotated to a more NW-SE trend compared to levelling data and GNSS measurements (Fig. 2-5a, b). Secondly, valley-wise calculation of incision rates (e.g. along the Limmat Valley or Lower Aare Valley, see Appendix A.2) including uncertainties of the Base Deckenschotter elevations revealed slightly higher rates (mean rates HDS to LEB ~ 0.14 to 0.25 mm/yr and mean rates TDS to LEB ~ 0.2 to 0.32 mm/yr).

Present-day surface uplift rates as derived from geodetic measurements (precise levelling and GNSS) are considerably higher compared to past exhumation or fluvial incision rates. From the Alps towards the northern foreland, GNSS-based vertical velocities decrease to positive sub-millimeter and even to stable zero mm/yr rates (Sánchez et al. 2018, Henrion et al. 2020, Pintori et al. 2022; Fig. 2-1c). Low vertical velocities are observed in the Jura Mountains and the Molasse Basin. Northern Switzerland is currently uplifting at rates between ~ 0.2 and 0.5 mm/yr within a European framework (Fig. 2-5a, b, and e; see also Nagra 2024c).

Conclusion: The geomorphic characteristics of the Northern Alpine Foreland are influenced by the long-term tectonic control of the Alpine orogeny. Long-term uplift proxies based on exhumation and average Quaternary fluvial incision reveal rates below ~ 0.3 mm/yr, while short-term geodetically derived surface uplift rates can be up to 0.5 mm/yr, when viewed within a stable Europe framework.

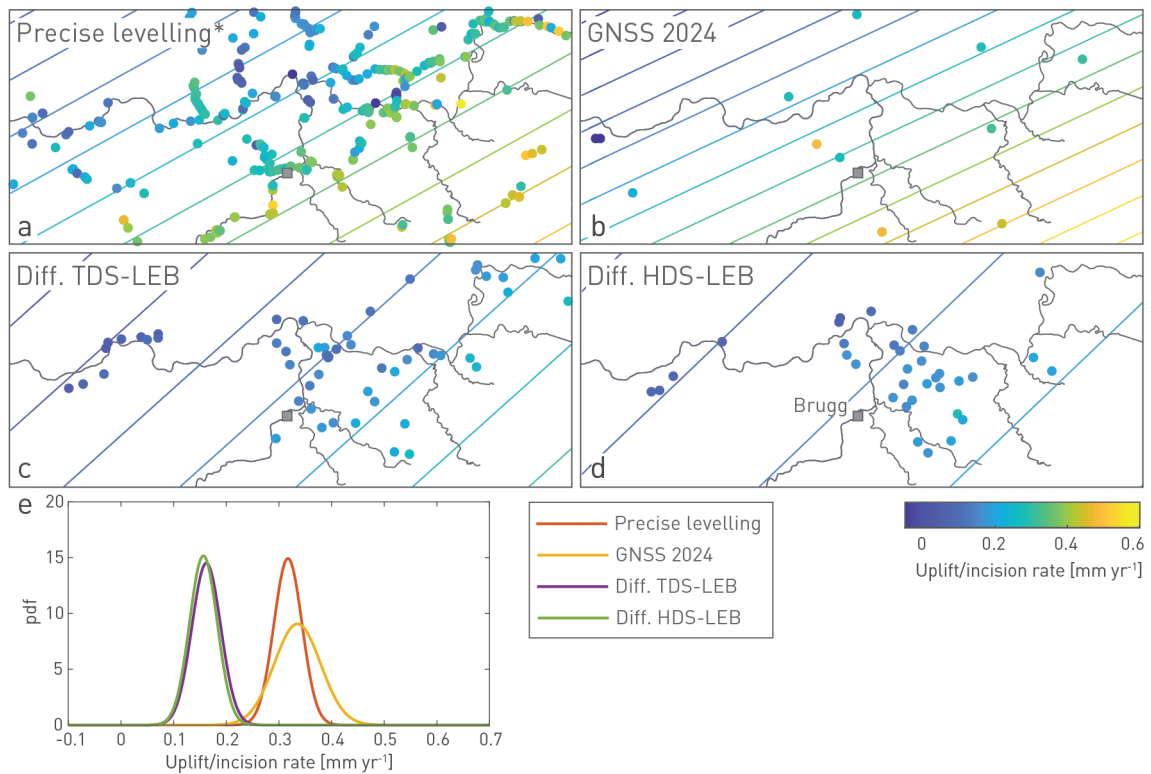


Fig. 2-5: Pattern and rates of surface uplift and incision proxies in Northern Switzerland on different timescales

The uplift planes in (a) to (d) are calculated from point measurements based on a regression using the coordinates as independent variables. Data points are colour-coded based on their uplift or incision rate. Main rivers are drawn for orientation, the grey square marks the location of the city of Brugg. The contours show isolines of constant uplift rate from a planar fit to the data. (a) The interpolation of precise levelling data shows a gradient at an azimuth of $\sim 153^\circ$, perpendicular to the Alpine front. Vertical rates are measured with respect to Laufenburg. Laufenburg is currently uplifting at a rate of ~ 0.28 mm/yr with respect to stable Europe and this value was added for better comparison with GNSS (see Nagra 2024c for more detail on precise levelling and GNSS data) (b) Interpolation of the vertical GNSS data shows a similar pattern to precise levelling data at an azimuth of $\sim 157^\circ$. (c) Interpolation of height differences between the geomorphic marker represented by the local erosion base (LEB, see Section 3.1 for definition) and the base of the Tiefer Deckenschotter (TDS) using a mean age of 0.9 Myr (d) Similar to (c) but using the Höhere Deckenschotter (HDS) as a marker and a mean age of 1.38 Myr. Note that the elevation differences are based on the lowest elevation of each mapped Deckenschotter occurrence (Heuberger & Naef 2014). (e) Probability density function of all uplift or incision proxy data shown in the respective maps above, calculated as an example for the location of the city of Brugg (square in maps). For the TDS age, a standard deviation of 0.13 Myr was considered, for the HDS age with standard deviation of 0.31 Myr. Note the higher rates of present-day geodetic surface uplift signals wrt. the longer-term incision rates.

Black Forest Massif and Vosges

The Vosges and Black Forest are part of the southern Upper Rhine Graben rift shoulders (see Fig. 2-1). The Black Forest bounds the Hochrhein to the north. Both areas started to be exhumed due to regional updoming approx. 20 Ma ago (Ziegler & Dèzes 2005, 2007).

Volcanism associated with mantle upwelling might have caused thermal weakening prior to the initial graben formation and regional uplift of the Black Forest and Vosges due to lithospheric folding (Timar-Geng et al. 2006 and references therein). Depending on the inferred geothermal gradient, total denudation of the Black Forest and Vosges since the Eocene might have accumulated to 1.0 – 1.7 or 1.3 – 2.2 km (Timar-Geng et al. 2006).

Exhumation expressed by the effects of catchment-wide denudation reaches rates between ~ 0.01 and 0.1 mm/yr (Morel et al. 2003, Timar-Geng et al. 2006, Meyer et al. 2010, Glotzbach et al. 2014).

Unequivocal indications of neotectonic differential movement between the Folded and Tabular Jura and the Black Forest have not been observed. In spite of active seismicity (Cloetingh et al. 2006), the region appears to be stable as a tectonic block, including the Black Forest and its Mesozoic cover, i.e. the Folded and Tabular Jura area (Verderber 2003). However, gentle movement of the entire area, for example in the form of regional tilting, cannot be excluded (Becker et al. 1987, Egli et al. 2017). Demoulin et al. (1998), Schlatter et al. (2005) and Fuhrmann et al. (2013, 2014) document a slight uplift of the graben shoulders at the Black Forest and Vosges Massifs of up to 0.3 mm/yr (relative to the reference point at the location of Freudensstadt in the eastern Black Forest Massif). A current visual inspection of precise levelling data and GNSS-based vertical velocities obtained at permanent stations does not reveal a distinct signal related to the Black Forest uplift with respect to the Hochrhein area (Nagra 2024c).

Conclusion: The long-term tectonic evolution and uplift history of the Black Forest and Vosges Massifs is closely linked to early mantle upwelling, superseded by exhumation related to the graben shoulders of the Upper Rhine Graben. Present-day uplift at a low rate is revealed with respect to Upper Rhine Graben subsidence.

Upper Rhine Graben

The Upper Rhine Graben (Fig. 2-1) is an integral part of the European Cenozoic Rift System (ECRIS) and is bordered to the northwest by the uplifting Rhenish Massif and to the south by the Jura Fold-and-Thrust-Belt. Within this tectonic domain, the Rhine (Oberrhein) drains a tectonic domain characterised by long-term intermittent subsidence, accompanied by erosion of the uplifting rift shoulders. The Upper Rhine Graben is bounded by the Vosges Mountains in the west and the Black Forest in the east. Closely related to the Miocene to present-day tectonic evolution of the Upper Rhine Graben is the development of the principal Alpine and peri-Alpine rivers (Rhine, Rhone, Aare, Doubs, Danube; Giamboni et al. 2004, see Section 2.4.2).

Eocene and Quaternary subsidence in the 350 km-long and up to 50 km-wide SSW-NNE-striking Upper Rhine Graben was episodic and associated with tectonic and non-tectonic phases (e.g. compaction of sediments, see Fig. 2-6b for Quaternary sedimentation rates and c for net sediment accumulation during the Quaternary (Ellwanger et al. 2008). Sediment isopach maps suggest that the Upper Rhine Graben evolved in a sequence of distinct subsidence phases and that subsidence centres migrated as result of a changing ambient stress field acting on inherited pre-rift structures (Schumacher 2002).

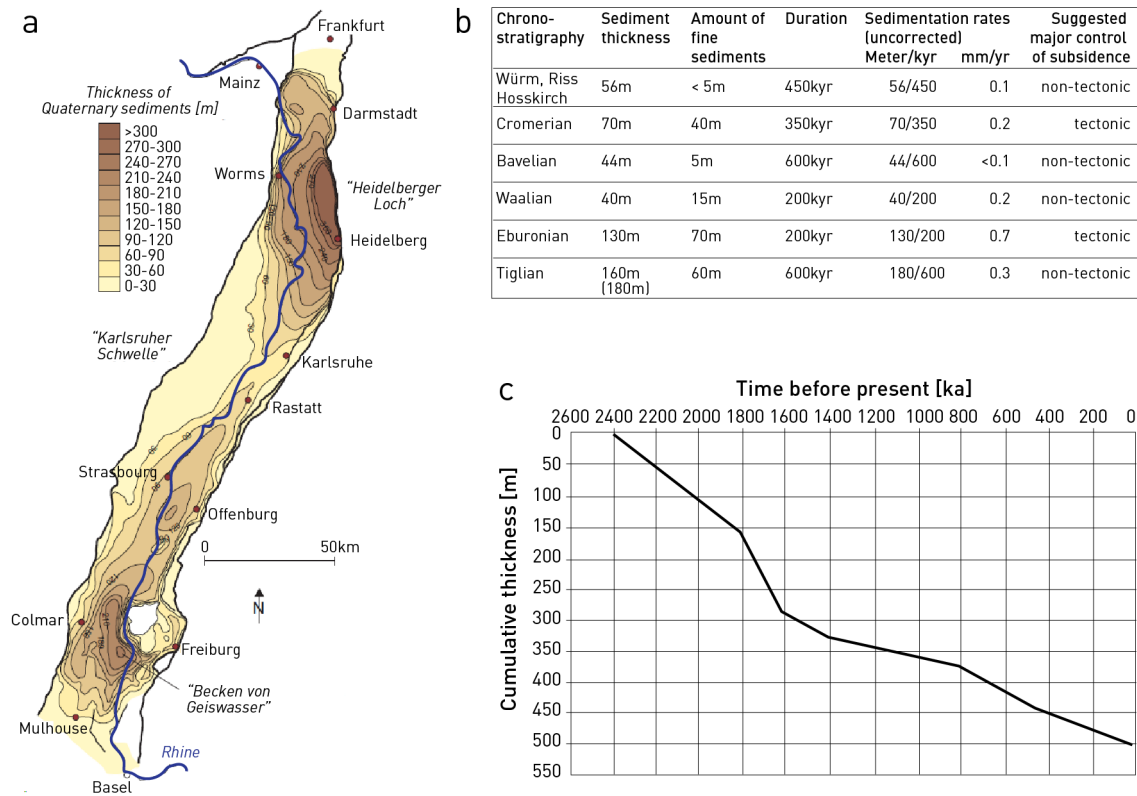


Fig. 2-6: Quaternary subsidence history of the Upper Rhine Graben

(a) Quaternary sediment thickness in the Upper Rhine Graben (modified from Hagedorn & Boenigk 2008), (b) Average sedimentation rates in the Heidelberg Basin ("Heidelberger Loch" in (a)) based on chronostratigraphic ages (Ellwanger et al. 2008). According to these authors, tectonic control of basin subsidence is suggested, if the accommodation space cannot be associated with compaction of sediments alone. (c) Net sediment accumulation during the Quaternary, based on the rates provided in (b) (Ellwanger et al. 2008).

The late-stage (at least Pliocene-Quaternary, probably starting in Early Miocene) shift to a NW-SE-directed maximum horizontal stress led to left-transensional to left-lateral reactivation with establishment of the present-day depositional centres (Heidelberg and Geiswasser Basins, see Fig. 2-6a). The central part of the Upper Rhine Graben comprises a restraining bend in such a setting (Schumacher 2002), explaining the low subsidence and low sedimentation observed in this area ('Karlsruher Schwelle').

The subsidence centres are generally located closer to the eastern flank of the Upper Rhine Graben. In the northern Upper Rhine Graben (Heidelberg Basin, exceeding 1'000 m in the northern part), asymmetric subsidence and sediment accumulation are generally higher compared to the southern Upper Rhine Graben (Geiswasser Basin, see Fig. 2-6a; Gabriel et al. 2013, Edel et al. 2007, Hagedorn & Boenigk 2008, Rotstein & Schaming 2011).

Reconstructing the Plio-Pleistocene evolution of the Upper Rhine Graben is further complicated by the fact that the southern part mainly received coarse Alpine deposits during these periods. These deposits underwent significant fluvial reworking, sorting, and admixing with debris from the graben shoulders (Vosges, Black Forest). Quaternary subsidence rates are in the order of 0.1 – 0.3 mm/yr, increasing substantially towards the northern Upper Rhine Graben (Berger et al. 2005, Hagedorn & Boenigk 2008, Peters & van Balen 2007, Preusser et al. 2021). Ellwanger et al. (2008) distinguish two phases of predominantly tectonically controlled subsidence for the

northern Upper Rhine Graben: between approximately 1.8 to 1.6 Myr ago locally at a rate of 0.7 mm/yr and between approximately 800'000 to 350'000 yr ago at a rate of 0.2 mm/yr (see Fig. 2-6b and c; note that 0.7 mm/yr represents by far the highest rate observed during the Quaternary).

Fuhrmann et al. (2014) infer that the Upper Rhine Graben is mainly subsiding at rates between -0.2 and -0.5 mm/yr, which they attribute to a combination of tectonic subsidence and sediment load (e.g. sediment compaction and isostatic effects).

Using GNSS data, Henrion et al. (2020) also reveal a weak vertical velocity signal that extends into the area of the Upper Rhine Graben shoulders (the Black Forest and Vosges Mountains). This is associated with observed subsidence within the Upper Rhine Graben north of Basel and west of the Upper Rhine Graben towards the Bresse Graben (Sánchez et al. 2018, Henrion et al. 2020, Pintori et al. 2022).

Conclusion: The Upper Rhine Graben is currently deforming in a transtensional sense of motion, which has implications for the centres of ongoing low-rate subsidence. Importantly, it seems that the Karlsruhe Schwelle as a restraining bend at least partly buffers the drainage of the Oberrhein. The contrasting subsidence rates are expected to flatten the Oberrhein profile (as seen today), which in turn would lead to further deposition of coarse Alpine-derived bedload in the southern Upper Rhine Graben.

Rhenish Massif and Eifel

With respect to the Atlantic Ocean basin, the adjacent shelf and delta areas, as well as the Lower Rhine Graben, the Rhenish Massif constitutes the first tectonic domain that is actively uplifting. This uplift coincides with bedrock incision of the Rhine River (Mittelrhein, see Fig. 2-1a and Fig. 2-1b).

The Variscan Rhenish Massif has been steadily uplifted since at least the Oligocene (Illies & Greiner 1979, Illies et al. 1979, Garcia-Castellanos et al. 2000, Ziegler & Dèzes 2005, 2007). Extensive volcanism in the area, spanning from the Oligocene (Westerwald) until the Late Quaternary (East and West Eifel; Schmincke 2007), is closely connected with asthenospheric upwelling associated with a plume (Ritter & Christensen 2007).

Presently, the Rhenish Massif forms a low mountain range with elevations up to ~ 700 m, which is dissected by several rivers (Garcia-Castellanos et al. 2000). A succession of staircase river terraces (Fig. 2-7) used as strain markers document uplift-related incision that has apparently accelerated since the Pliocene (Fuchs et al. 1983) or the latest Early to Middle Pleistocene (Meyer & Stets 1998).

Kreemer et al. (2020) speculated that uplift has either increased during the Late Quaternary or that the loading/unloading effect related to glacial periods causes the net vertical movements to vary considerably over time. Garcia-Castellanos et al. (2000) compiled Quaternary terrace data and present an uplift gradient that increases towards the centre of the spatial uplift distribution. The highest amount attained during the past ~ 800 kyr (up to 250 m) was located at the left bank of the Rhine in the Eifel area, whereas the lowest amounts (< 50 m) are reported west of the Maas River in the neighbouring Ardennes.

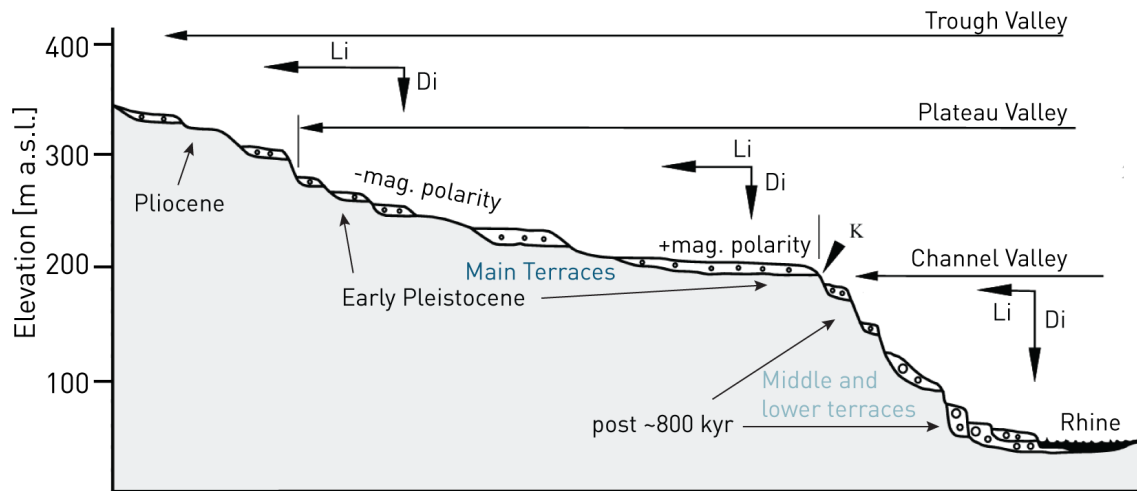


Fig. 2-7: Cross-section of the Mittelrhein Valley showing the terraces and the valley form

Figure is redrawn after (Meyer & Stets 1998). The pre-Pliocene wide trough valley began incising during the Pliocene. Between the Pliocene and Early Pleistocene, lateral scouring (Li) dominated over downward incision (Di). Uplift and subsequent incision led to the observed staircase pattern. The principal terraces ('Hauptterrassen', Main Terraces) still represent a Plateau Valley. Their sediments are characterised by mainly negative magnetic polarities (Meyer & Stets 1998). The change to positive polarities associated with the Brunhes-Matuyama boundary is traditionally assigned to the end of Main Terrace aggradation. The strong incision pulse represented by the Channel Valley and the Middle and Lower Terraces is thus thought to be younger than approximately 800 kyr. mag.: magnetic.

Demoulin & Hallot (2009) suggested that part of the total uplift might be related to knickpoint retreats in tributaries rather than regional tectonic deformation. Based on this interpretation, the total uplift may be reduced to 140 m, yielding a Quaternary long-term uplift rate (since ~ 800 ka) of ~ 0.2 mm/yr (Demoulin & Hallot 2009). These authors also favoured lithospheric folding as a reasonable uplift mechanism over the sole effect of the local Eifel plume. Importantly, the present-day elevations of post-Middle Miocene geomorphic markers (intramontane plains, fluvial terraces) are much higher than what could be expected as a position for these areas in response to baselevel changes associated with eustatic sea-level changes alone (Meyer et al. 1983).

An ellipsoidal uplift pattern can be observed in the present-day vertical strain estimated from the 3D GNSS velocity field as shown by Piña-Valdés et al. (2022) and Henrion et al. (2020). These short-term geodetic measurements of the Eifel area amount to a mean value of 1.17 ± 0.43 mm/yr, respectively (Fig. 2-1c, d), which the authors relate to mantle activity.

Conclusion: Combined, these data covering different timescales suggest long-term and active present-day uplift of the Rhenish Massif-Eifel areas at amounts that exceed eustatic sea-level changes and have potentially increased in the past ~ 800'000 years. It is therefore possible that these processes reflect a long-term signal and will probably continue to influence the evolution of topography and the drainage network in the future. Under this assumption, the Mittelrhein section would constitute an intermediate baselevel that is able to decouple downstream processes and reduce upstream migrating signals of downstream changes along the length of the longitudinal profile.

Lower Rhine Graben and delta

Because of the decoupling of fluvial processes in the Mittelrhein region, incision triggered by processes beyond the Eifel region is not expected for areas farther south, including the Aare and Hochrhein. Nevertheless, for completeness, the river profile sections northwest of the Rhenish Massif and Eifel uplift are briefly summarised here.

Downstream from the cities of Bonn and Köln, the Rhine meanders through the Lower Rhine Graben (Niederrhein, Fig. 2-1). Approximately until the city of Emmerich, the Rhine River has gradually incised into older strata during Quaternary times. Middle and Late Pleistocene terrace deposits have been correlated with the terrace record of the Mittelrhein (Brunnacker 1978, Boenigk 1981, Boenigk & Frechen 2006, Lauer et al. 2011, 2017). Downstream of Emmerich, sediments occur in normal, buried stratification.

The Lower Rhine Graben is characterised by active normal faulting. Positive GNSS-based vertical velocities related to the Eifel area transition into the southern Lower Rhine Graben, where they coincide with normal faulting seismicity along the graben faults (Kreemer et al. 2020, Hinzen et al. 2021). Vertical velocities eventually decrease downstream from the Eifel area to near-zero and very low negative rates (Fig. 2-1c).

Conclusion: Because of the Mittelrhein decoupling downstream from upstream effects, future control of the downstream sections on the Aare or Hochrhein incision is not expected.

Summary of tectonic control on baselevel of the Aare – Hochrhein system

The analysis of deformation histories within the different morphotectonic domains traversed by the Rhine River suggests that the main influence on the incision of the Hochrhein is caused by low rates of North Alpine Foreland rock uplift and low, potentially episodic subsidence within the Upper Rhine Graben. These trends have in turn been balanced by continuous uplift of the Rhenish Massif and Eifel areas, leading to a relative baselevel rise with sediment accumulation and flattening of the Oberrhein profile. Accordingly, the Hochrhein is presently buffered against fluvial processes approximately downstream of the city of Bingen (Fig. 2-8).

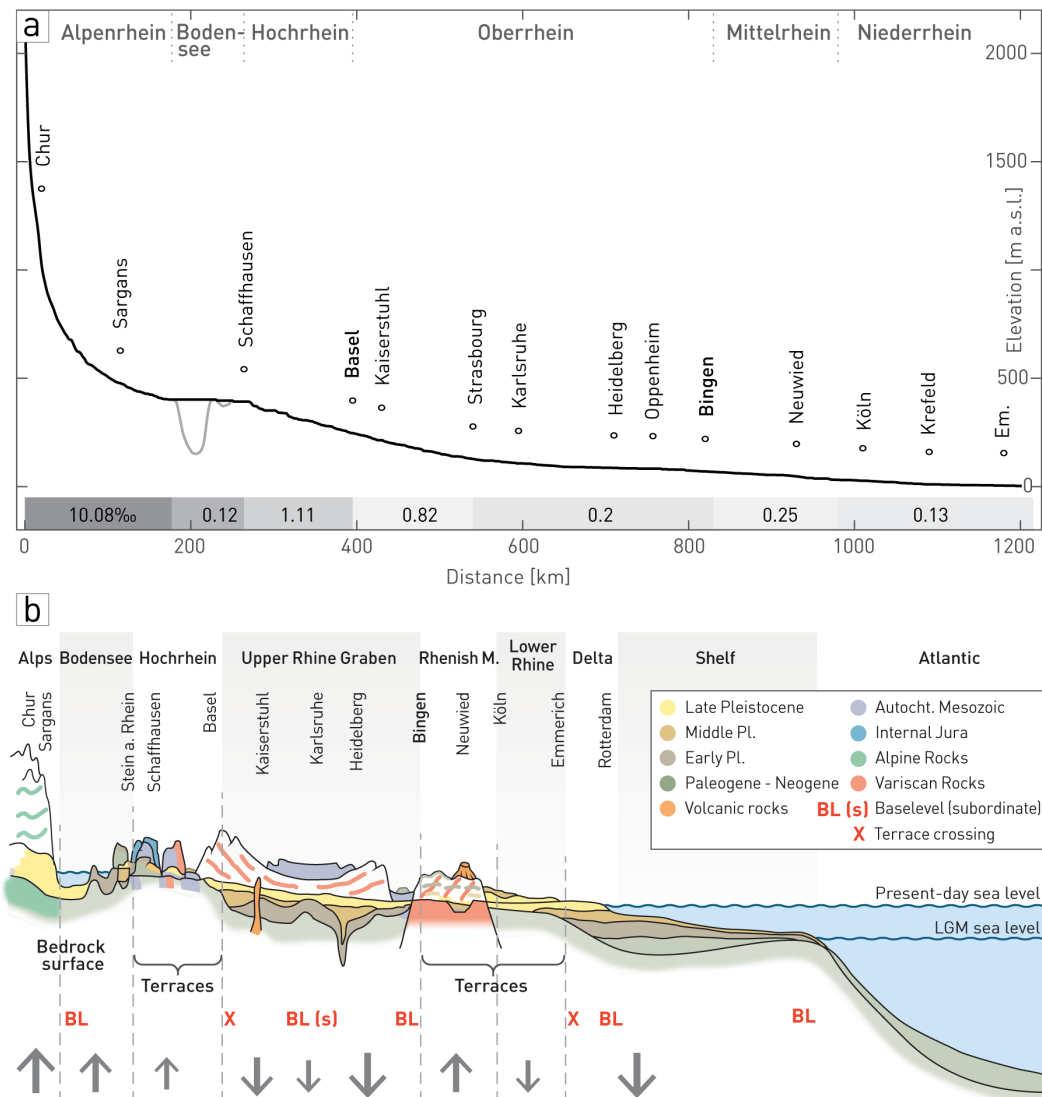


Fig. 2-8: Profile and sketch of Rhine River course with domains of Quaternary vertical motion

Rhine River profile from source to sink. (a) Longitudinal river profile crossing tectonic and geomorphic provinces. Shown are calculated gradients for significant provinces or observed changes along the profile (in per mill). Knickpoints are present near Schaffhausen and (subtle) near Bingen. Em.: Emmerich. (b) Schematic river profile with simplified general geology, occurrence of fluvial terraces and location of terrace crossings, (intermediate) baselevels of regional relevance, and variability of sea level. Not to scale, after Neeb in (Hinderer 2005) and based on Hinderer (2003, 2012). Tectonically driven rock uplift and basin subsidence (qualitatively marked by the grey arrows) influence the pattern and rates of fluvial incision in Northern Switzerland. The area is mainly influenced by North Alpine Foreland rock uplift, subsidence of the Upper Rhine Graben, and far-field rock uplift of the Rhenish Massif. Em.: Emmerich, M.: Massif. Figure draft courtesy of H.A. Kemna.

2.2 Quaternary climate as driver for erosion

Climate has an essential influence on Quaternary landscape evolution and especially on erosion in Northern Switzerland (Nagra 2024b). Climate affects the hydrological cycle and, together with topography and temperature variations, it determines precipitation patterns and amounts, the build-up and removal of glacial ice, and thus fluvial discharge, erosion processes and sediment transport in the entire Alpine realm and its foreland. Alpine glacier dynamics are largely driven by climate oscillations (Florineth & Schlüchter 2000, Cohen et al. 2018), which in turn control the extent and distribution of glacial erosion (Nagra 2024b). As shown for modern glaciated catchments in the Alps, glacier cover and fluvial sediment yield are positively correlated (Hinderer et al. 2013). In addition, extensive foreland glaciations may generate overdeepened valleys and change sediment routing and drainage pathways (Preusser et al. 2010, Häuselmann et al. 2007, Nagra 2024b). Climate and topography influence erosion also via many other processes and feedback mechanisms, including chemical weathering, ice loading and unloading, freeze-thaw cycles, formation and degradation of permafrost, vegetation cover, and frequency and magnitude of mass movements. Additionally, as indicated in Section 2.1, climate-driven erosion might contribute to uplift of the Central Alps and could therefore point to a large-scale feedback mechanism between climate, erosion, and topography (Champagnac et al. 2009, Sternai et al. 2019, Fox et al. 2015a).

Climate depends largely on insolation, i.e. the solar radiation that reaches the Earth. The amount and distribution of insolation depends on the Earth's orbit and tilt; their combined recurring effect on Earth's climate has been explained by astronomic forcing or Milankovitch cycles (Milankovitch 1941). These cycles can be divided into orbital eccentricity (ca. 400- and 100-kyr periodicity), obliquity (axial tilt, ca. 41-kyr periodicity) and axial precession (ca. 23- and 19-kyr periodicity; Fig. 2-9; Nagra 2024b). Based on Milankovitch's theory, these cycles influence the Earth's climate by modulating the summer insolation at high latitudes in the northern hemisphere (Feng & Bailer-Jones 2015). The energy received from the sun is further redistributed in the Earth system via atmospheric and oceanic circulation to equalise thermal gradients. The Milankovitch cycles determine the Earth's climate together with key characteristic of the Earth system such as atmosphere composition (e.g. CO₂ content), land-ocean distribution, thermohaline oceanic circulation, continental ice cover, glaciological mechanisms, and related feedback mechanisms. Next to the orbitally controlled parameters, CO₂ content is a key factor in climate oscillations and mirrors the general climate evolution during the Quaternary (e.g. temperature difference in Fig. 2-9; Lüthi et al. 2008, Ganopolski et al. 2016), culminating in the anthropogenically enhanced atmospheric CO₂ levels (e.g. Le Quéré et al. 2018).

The Quaternary climate cycles consist of interglacials, warm intervals with low global ice volume, and glacials, cold intervals with high global ice volume (Fig. 2-9). The climate during these periods was highly variable, characterised by cooling and warming occurring on timescales ranging from decades to millennia. Glacials are further subdivided into warmer intervals, which are known as interstadials, and the colder stadials. During glacial stages, the cold climate led to the growth of ice sheets and glaciers and the build-up of these ice-masses on land subsequently lowered sea levels. Climate cycles can be traced by palaeoclimatic proxies such as $\delta^{18}\text{O}$ records measured in the calcite shells of benthic foraminifera in deep-sea sediments, which largely reflect the growth and decay of continental ice sheets (Fig. 2-9; Lisiecki & Raymo 2005). Based on these records, the Quaternary glacial/interglacial cycles have been subdivided into marine isotope stages (MIS; Fig. 2-9). In the Early Pleistocene, the glacial/interglacial cycles were dominated by a 41-kyr periodicity. With the Mid-Pleistocene Transition (ca. 1.25 – 0.7 Ma, Clark et al. 2006), this changed to a 100-kyr periodicity (see also Nagra 2024b).

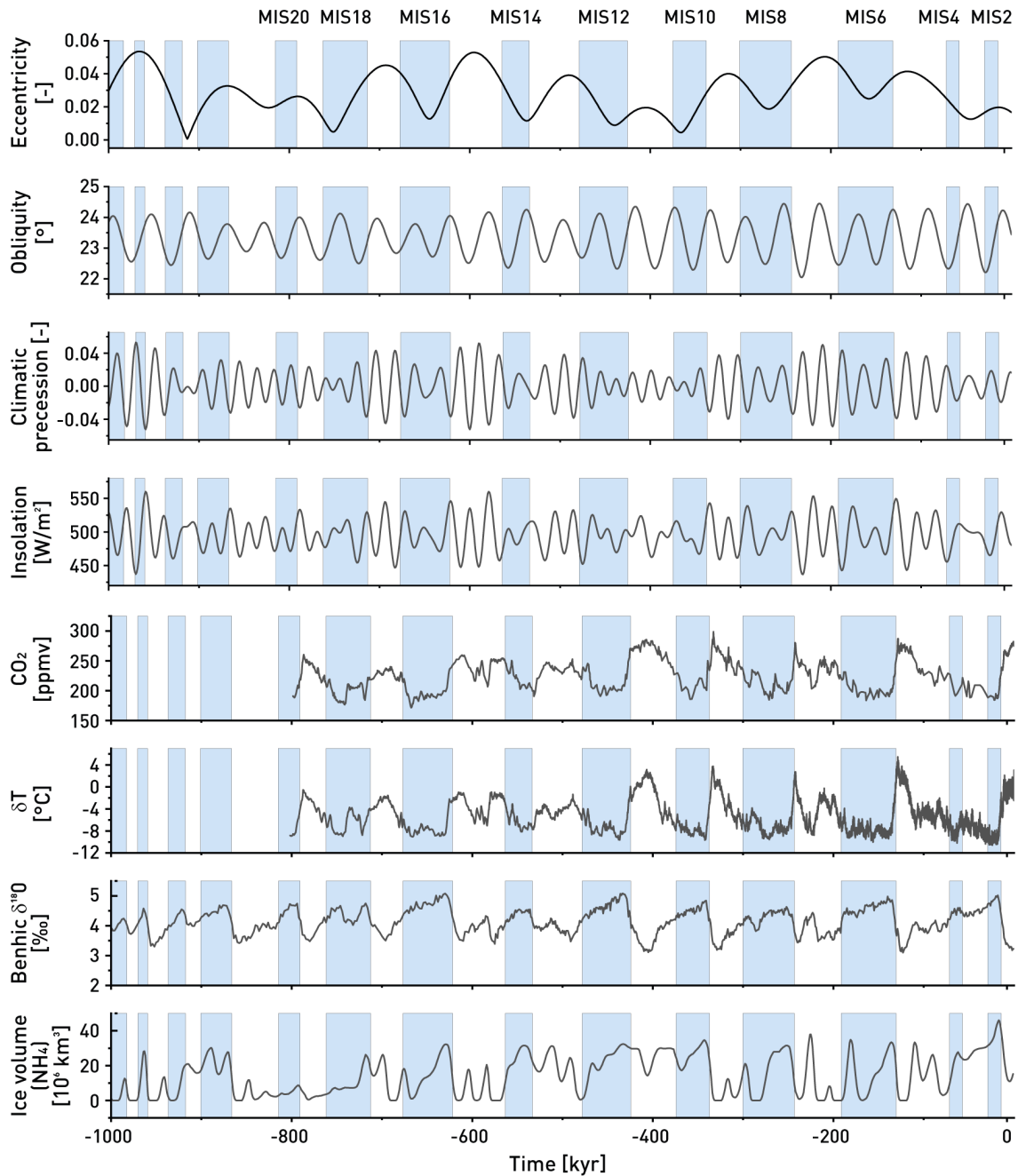


Fig. 2-9: Overview of orbital parameters and resulting insolation, CO₂ concentration, temperature difference, benthic $\delta^{18}\text{O}$ and estimated ice volume

Insolation and orbital parameters (eccentricity, obliquity and precession) are shown for 65°N (July) according to Berger & Loutre (1991). The CO₂ content follows the determinations from Lüthi et al. (2008). Global temperature differences are given as 100-yr averaged values compared to today and are taken from Jouzel et al. (2007). Data are based on measurements from the EPICA ice core in Antarctica. $\delta^{18}\text{O}$ was determined from sediment cores and can also be used to calculate an approximate estimate of the ice volume in the northern hemisphere (Lisiecki & Raymo 2005).

In conclusion, oscillating climates during the Quaternary had global repercussions regarding eustasy and base-level conditions for fluvial systems, impacting the efficiency of erosive and depositional processes. Coeval regional and local effects in the northern Alpine realm included changes in precipitation, runoff, and ice cover and thus form and rate of geomorphic processes.

2.3 Drainage reorganisation as driver for erosion

Drainage reorganisation constitutes an often disruptive process in the rather gradual evolution of river systems. The process is observed on different spatial scales and may have far-reaching effects on drainage networks and catchment configurations, on the local baselevel and subsequent erosion (Scherler & Schwanghart 2020).

Drainage reorganisation occurs due to shifts of divides between neighbouring catchments. The process is classified into three main groups (Bishop 1995) and include (1) river beheading, (2) river capture, and (3) river diversion. River beheading is a process in which drainage basins gradually grow headward at the expense of other drainage basins due to cross-divide differences of erosion rates in response to tectonic tilting or spatial trends of precipitation. River capture (or piracy) denotes the process where a river breaches a drainage divide by headward erosion, thus capturing an entire basin. This process occurs if two rivers have different baselevels or variable erodibilities. River diversion, in contrast, denotes the process where a river changes its course due to aggradation and lateral erosion, eventually overtopping and spilling over a low-lying divide into another catchment (Bishop 1995). Diversion can cause dramatic changes and lead to the formation of breakthrough channels. In the context of Northern Switzerland with repeated Quaternary foreland glaciations, breakthrough channels may develop temporarily or permanently in response to stream blockage by moraines, aggradation, landslides, or ice (see Section 3.4.1), causing channels to reroute.

Drainage reorganisation can lead to a significant pulse of erosion even in tectonically quiescent regions (Prince et al. 2011), affecting both regions upstream and downstream of the capture or breakthrough channel. The integrated catchment will suddenly need to adjust to a new baselevel, thus experiencing a wave of headward incision (Berlin & Anderson 2007, Schwanghart & Scherler 2020). Due to the drainage area increase downstream of the captured site, more runoff and sediment will also lead to a transient response characterised by a period of increased erosion and/or aggradation.

Such events have significantly influenced the evolution of the Aare – Rhine system in Northern Switzerland. For example, before the Rhine assumed its present course, there were significant changes in the catchments, which are summarised in Section 2.4.2. Today, the Aare and Rhine Rivers determine the baselevel for the siting regions. Accordingly, it is important to understand whether similar regional drainage reorganisations should be expected in the future and whether erosion in Northern Switzerland is still influenced by the past significant catchment changes.

2.4 Pliocene to Quaternary landscape evolution

2.4.1 Principles and database for reconstruction of landscape evolution

2.4.1.1 Introduction

There are three key aspects for the reconstruction of the Quaternary landscape evolution and erosion in Northern Switzerland based on Quaternary sedimentary archives. Firstly, the elevation of sedimentary key horizons or 3D bodies must be investigated. Secondly, the chronology of depositional and erosive events of the various phases must be explored. The chronology can best be investigated by determining the depositional age of sedimentary archives. Thirdly, to ensure a robust interpretation of the different archives, the sedimentary units must be interpreted in the context of environmental conditions and depositional processes. The three aspects, along with their challenges and uncertainties, are briefly discussed below.

2.4.1.2 Spatial analysis of key horizons

Knowledge of the elevation of 3D sedimentary bodies and of key horizons, such as the base Quaternary (top of bedrock below Quaternary sediments), has been considerably improved during Stage 3 of the Sectoral Plan (Chapter 1). The digital model of top bedrock / base Quaternary of Pietsch & Jordan (2014) was updated (Fig. 2-10). This updated database integrates results from more than 700 boreholes including 11 Quaternary boreholes drilled between 2018 and 2022 (Table 2-2 in Nagra 2024e). Additionally, 26 high-resolution datasets from shallow 2D reflection seismic profiles and analysis of seismic surface waves in glacial overdeepenings and fluvial valleys (Nagra 2018, 2021, Buechi et al. 2024) were incorporated. Finally, areas with Deckenschotter deposits were mapped in the field and using high-resolution digital terrain models (e.g. Hammer et al. 2019, Thew et al. 2024).

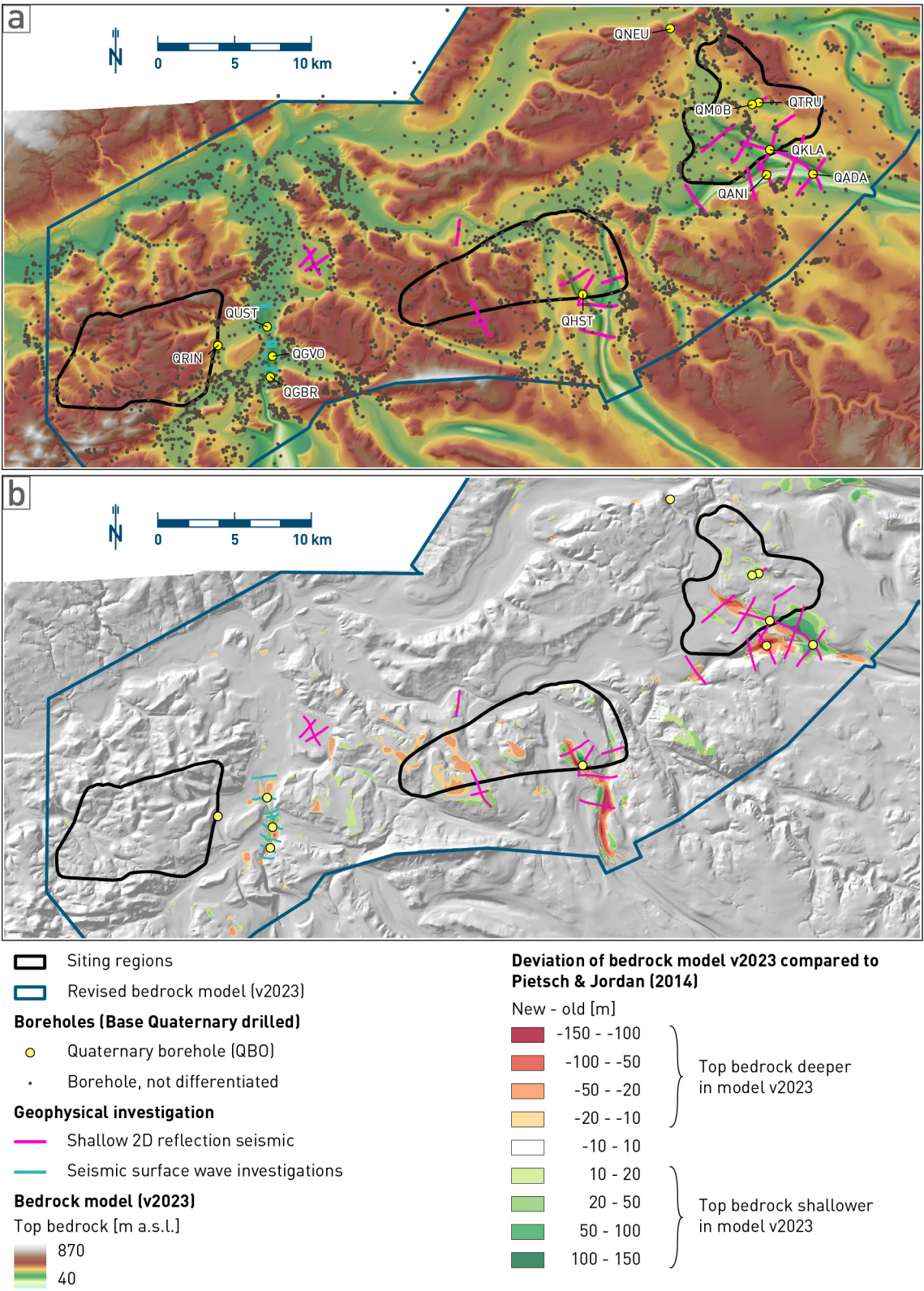


Fig. 2-10: Revised map of top bedrock (Version v2023) in the area of the siting regions (a) and difference compared to the bedrock model of Pietsch & Jordan (2014) (b)

Abbreviations of Quaternary boreholes (QBO); QRIN = Riniken, QUST: Untersiggenthal, QGBR: Gebenstorf-Brüel, QGVO: Gebenstorf-Vogelsang, QHST: Hochfelden-Strassberg, QMOB: Marthalen-Oobist, QNEU: Neuhausen, QTRU: Trüllikon-Rudolfingen, QKLA: Kleinandelfingen-Laubhau, QANI: Andelfingen-Niederfeld, QADA: Adlikon-Dätwil. Note that boreholes not reaching the bedrock surface are not displayed to enhance visibility of further map content at the given scale. However, they were also used as constraints for model interpolation. Additionally, high-resolution profiles from shallow 2D reflection seismic and seismic surface wave investigations (Nagra 2018, Gegg et al. 2021) were considered.

Form and depth of glacial overdeepenings

One key product derived from the updated top bedrock model (Fig. 2-10) are statistical analyses of the form and depth of observed glacial overdeepenings (Fig. 2-11 and Fig. 2-12 for depth and Fig. 2-32 or Fig. 3-44 for indications of form). From these data, it is learned for instance that the glacial overdeepenings in Northern Switzerland usually show cross-sectional forms that are between the endmembers of typical fluvial V-shaped valleys and glacial U-shaped valleys.

It is also an important observation that the depth of overdeepenings can be highly variable along their longitudinal axis (Fig. 2-11b). For instance, the QADA (Adlikon-Dätwil) borehole drilled the deepest part of the longitudinal depth distribution of the Marthalen Trough, but the majority of the overdeepening, including the area within the siting region of ZNO, is shallower. Nevertheless, overdeepening depth can be high in the distal part of the troughs. The boxplots shown in Fig. 2-12a highlight the depth distribution of the deepest points along the axes for a number of glacial overdeepenings in Northern Switzerland and the Bodensee Trough.

Base Deckenschotter

While the respective surfaces of the unconsolidated Deckenschotter deposits (see sections below) might be highly altered by erosion, their base (i.e. the bedrock – sediment contact) is expected to be sheltered and preserved. Base Deckenschotter is thus a valuable marker for past fluvial base-levels. On the other hand, the original bedrock surface prior to filling with Quaternary sediment was an erosional landform with a palaeotopography that is only partially observed. Furthermore, although major efforts were undertaken to establish and update the bedrock model (Fig. 2-10), uncertainty resulting from mapping the base Deckenschotter using digital terrain models and field investigations and from interpolation between data points remains. This uncertainty was described in Hammer et al. (2019) and was considered for additional Deckenschotter outcrops during the update of the bedrock model.

Based on these data, fluvial incision was calculated along selected valleys in Northern Switzerland, which are straddled by Deckenschotter deposits (e.g. Limmat Valley). The associated profiles and distribution of incision rates are shown in Appendix A.2.

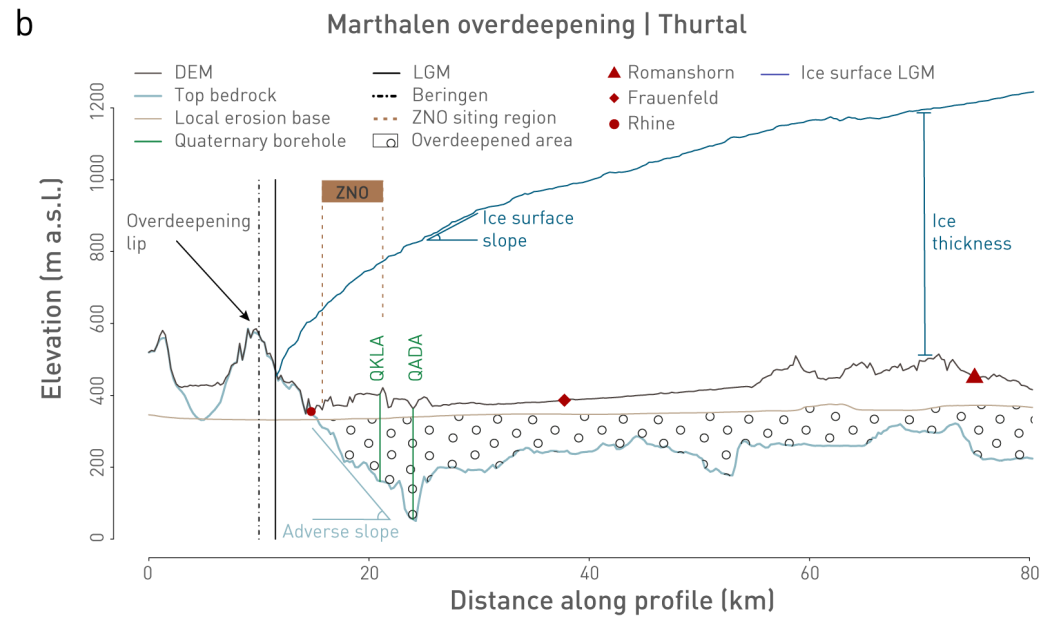
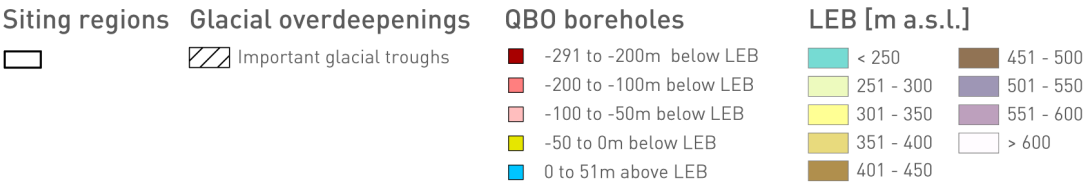
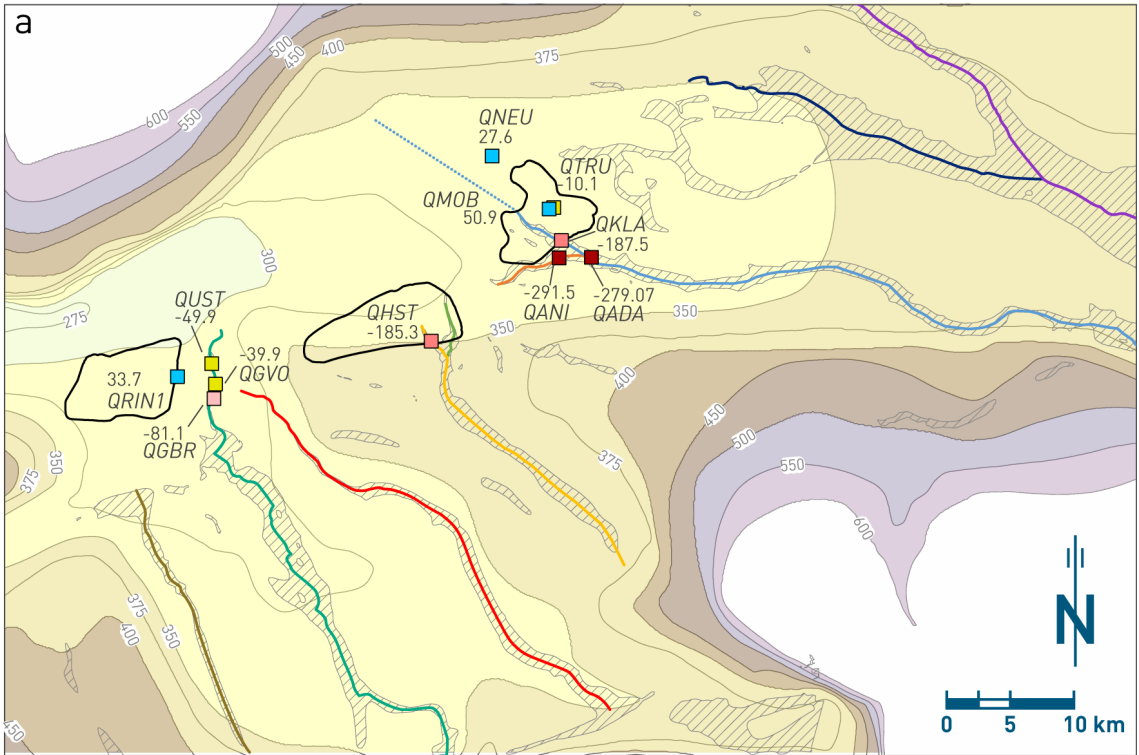


Fig. 2-11: Map of local erosion base and glacial overdeepenings and selected longitudinal profile

(a) Map showing glacial overdeepenings with the local erosion base in the background. The dashed polygons indicate areas of Quaternary deposits in glacial overdeepenings with thickness greater than 25 m. Also shown are borehole depths from the Quaternary drilling programme (QBO). Negative values indicate depth below the local erosion base, positive values above LEB. Profile centre lines through the overdeepening are shown as used for the statistical analysis in Fig. 2-12a. The profile line of the Marthalen Trough (light blue) is artificially extended across significant past ice extents (dotted part, to be used in (b)) and does not show the real continuation, which might have drained through the Neuhausen breakthrough channel or the Klettgau. (b) Example of longitudinal profile through the Marthalen Trough from a location at the Rhine River to a position near the city of Romanshorn, highlighting the important elements (local erosion base, glacial overdeepening, ice extents, modelled ice thickness, and two Quaternary boreholes).

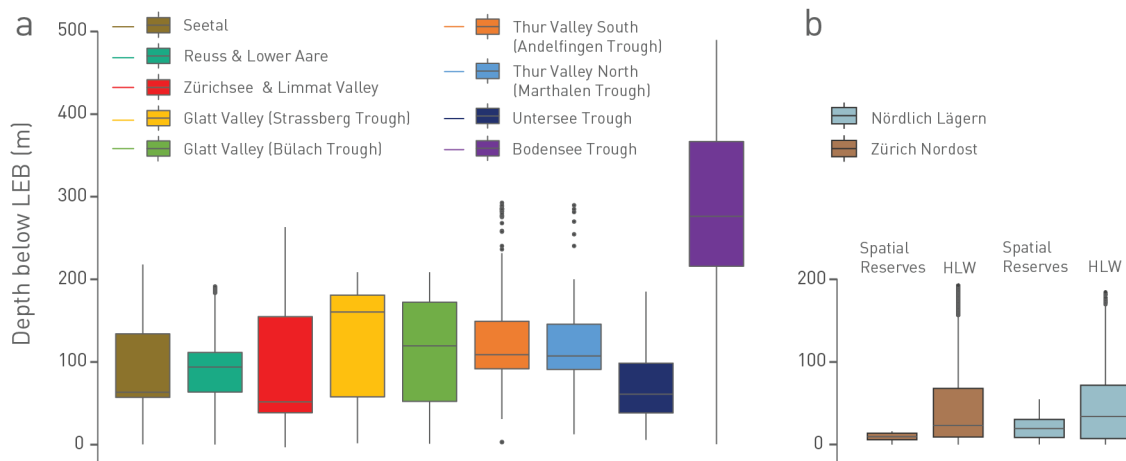


Fig. 2-12: Statistics on overdeepening depth in Northern Switzerland

(a) Boxplot statistics of the overdeepenings drawn in (Fig. 2-11a) using the same colour code. Note the significantly deeper overdeepened Bodensee Trough. In examples where glacial overdeepenings bifurcate, both corresponding boxplots include the upstream part with the exception of the Untersee. For these boxplots, depth values were recorded along the axis of the overdeepenings, thus likely reflecting the deepest points. (b) The data show the distribution of overdeepening depth within the HLW siting regions and spatial reserves (compare Section 3.1 and Fig. 4-3) of NL and ZNO and are based on the bedrock model v2023. In contrast to (a), these depth values were calculated across the entire respective areas of HLW and spatial reserves. Accordingly, they may have a tendency to lower values in their median and are not directly comparable to the depths in (a). Note that overdeepening is absent within the JO siting region and the provisional disposal areas (except minor overdeepening in ZNO provisional disposal area; see Section 3.1 and Fig. 3-2).

2.4.1.3 Chronology

Determination of the depositional ages of the Quaternary sediments in Northern Switzerland is a challenge, especially for deposits older than the Late Pleistocene (Fig. 2-13). In the following, selected key dating methods and their application for Quaternary sediments of Northern Switzerland are briefly presented. Key literature references are given for a more detailed discussion of

the methods. General information about additional dating methods (e.g. radiocarbon dating) can be found elsewhere, for example in Walker (2005). An overview of the locations with presented age information is shown in Fig. 2-14, while more details about the locations (coordinates, references) can be found in Tab. 2-1 and in Appendix A.1. The locations are grouped into the four morphostratigraphic units of Northern Switzerland, from topographically higher to lower positions, and thus from older to younger depositional ages. These principal units are the Höhere Deckenschotter (HDS, "higher cover gravels"), Tiefere Deckenschotter (TDS, "lower cover gravels"), as well as sediments of the Hochterrasse (HT, "High terrace") and Niederterrasse (NT, "Low terrace") (Graf & Burkhalter 2016 and references therein). Overdeepenings (OD in Appendix A.1) form an additional stratigraphic landscape element. The focus is on locations with dated ages from the Early and Middle Pleistocene, while for younger sediments only a few representative locations are given.

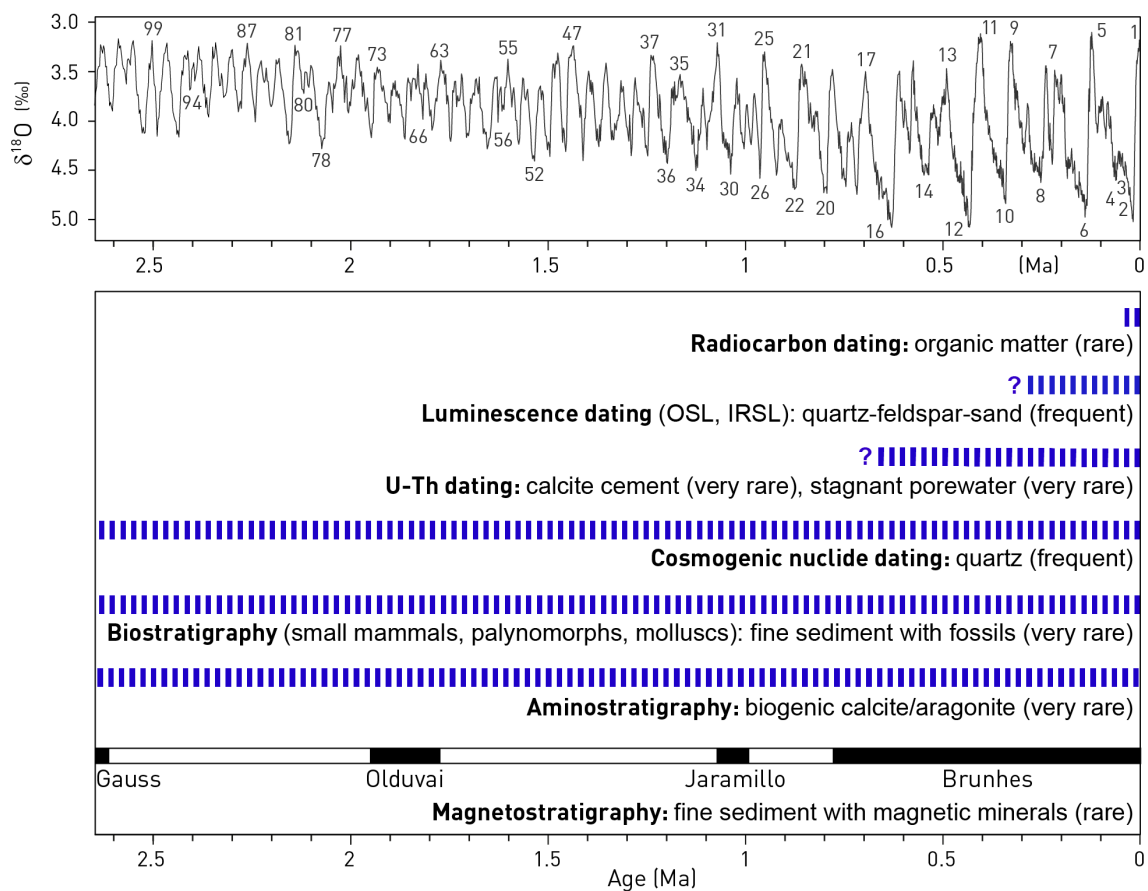


Fig. 2-13: Dating methods and their temporal applicability for Quaternary sediments in Northern Switzerland

$\delta^{18}\text{O}$ record with marine isotope stage (MIS) numbering from Lisiecki & Raymo (2005). For the different dating methods, the sample material and its frequency of occurrence in Quaternary sediments of Northern Switzerland are indicated. OSL: Optically stimulated luminescence, IRSL: Infrared stimulated luminescence.

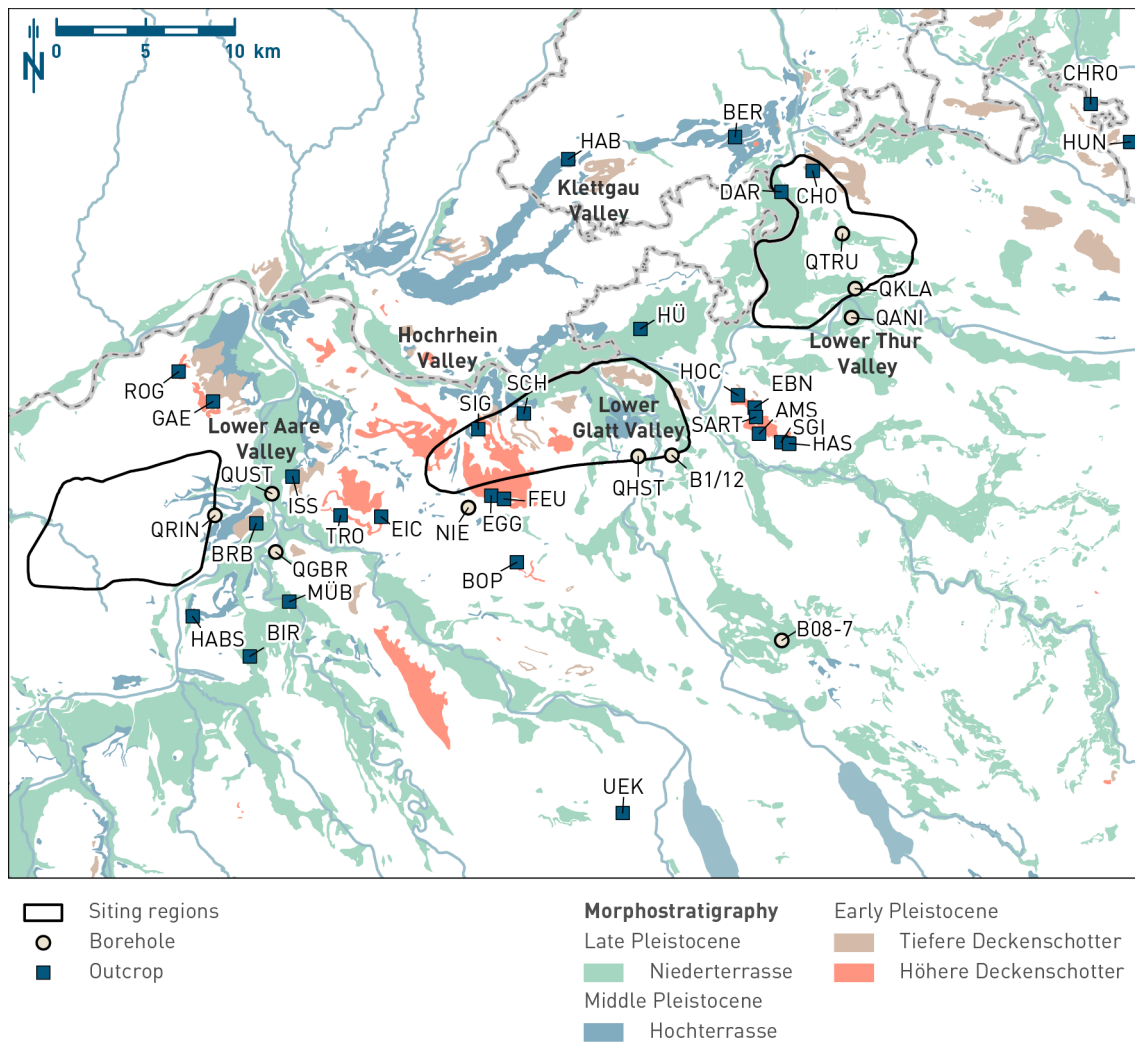


Fig. 2-14: Overview of sampling locations with presented age information

See Tab. 2-1 for names and associated landscape elements of the listed acronyms. More details about the locations (coordinates, references) can be found in Appendix A.1.

Tab. 2-1: Sampling locations with presented age information

More details about the locations (coordinates, references) can be found in Appendix A.1.
QBO: Quaternary borehole.

Acronym	Name of sampling location	Landscape element
ROG	Roggenfeld, Mettauertal	Höhere Deckenschotter (HDS) ¹
TRO	Tromsberg, Obersiggenthal	Höhere Deckenschotter (HDS)
EIC	Eichbrunnen, Freienwil	Höhere Deckenschotter (HDS)
SIG	Siglistorf	Höhere Deckenschotter (HDS)
EGG	Egghalden, Schleinikon	Höhere Deckenschotter (HDS)
FEU	Feusi, Oberweningen	Höhere Deckenschotter (HDS)
BOP	Cholholz, Boppelsen	Höhere Deckenschotter (HDS)
UEK	Uetliberg Kulm, Zürich	Höhere Deckenschotter (HDS)
BUE	Albishorn-Bürglen-2, Hausen am Albis	Höhere Deckenschotter (HDS)
HOC	Irchel Hochwacht, Berg am Irchel	Höhere Deckenschotter (HDS)
EBN	Irchel Ebni, Berg am Irchel	Höhere Deckenschotter (HDS)
SART	Irchel Schartenflue, Freienstein-Teufen	Höhere Deckenschotter (HDS)
AMS	Irchel Amselboden, Freienstein-Teufen	Höhere Deckenschotter (HDS)
SGI	Irchel Steig, Buch am Irchel	Höhere Deckenschotter (HDS)
HAS	Irchel Hasli, Dättlikon	Höhere Deckenschotter (HDS)
CHRO	Chroobach, Rielasingen-Worblingen (DE)	Höhere Deckenschotter (HDS)
GAE	Ängi, Leuggern-Mandach	Tiefere Deckenschotter (TDS)
BRB	Bruggerberg, Brugg	Tiefere Deckenschotter (TDS)
ISS	Iberig, Würenlingen	Tiefere Deckenschotter (TDS)
SCH	Schwändihalden, Weiach	Tiefere Deckenschotter (TDS)
CHO	Cholfirst, Chirchhölzli, Laufen-Uhwiesen	Tiefere Deckenschotter (TDS)
HUN	Hungerbol, Öhningen (DE)	Tiefere Deckenschotter (TDS)
HABS	Götschel, Lupfig, near Schinznach-Bad	Hochterrasse (HT)
QRIN	QBO Riniken	Hochterrasse (HT)
NIE	Niederweningen	Hochterrasse (HT)
HAB	Hablützel, Wilchingen-Hallau	Hochterrasse (HT)
QHST	QBO Hochfelden-Strassberg	Hochterrasse (HT)
BER	Beringen	Hochterrasse (HT)
DAR	Dachsen	Hochterrasse (HT) ²

¹ ROG was mapped as HDS by Diebold et al. (2006) but might be part of the HT (refer to Section 2.4.1.4).

² In the locality DAR the Schaffhausen Schotter is outcropping, which was deposited at the end of the Beringen Glacial after an incision on a level that corresponds to the Niederterrasse.

Tab. 2-1: (continued)

Acronym	Name of sampling location	Landscape element
BIR	Joriacher, Birr	Niederterrasse (NT)
MÜB	Mülligen	Niederterrasse (NT)
HÜ	Hüntwangen, Rafzerfeld	Niederterrasse (NT)
QUST	QBO Untersiggenthal	Glacially overdeepened trough
QGBR	QBO Gebenstorf-Brüel	Glacially overdeepened trough
NIE	Niederweningen	Glacially overdeepened trough
QHST	QBO Hochfelden-Strassberg	Glacially overdeepened trough
B1/12	Bülach-Hochfelden	Glacially overdeepened trough
B08-7	Hardwald, Bassersdorf	Glacially overdeepened trough
QTRU	QBO Trüllikon-Rudolfingen	Glacially overdeepened trough
QANI	QBO Andelfingen-Niederfeld	Glacially overdeepened trough
QKLA	QBO Kleinandelfingen-Laubhau	Glacially overdeepened trough

Luminescence dating

Ionising radiation resulting from the decay of natural radioactive isotopes (^{40}K , ^{232}Th , ^{238}U , ^{235}U) induces trapping of charge (electrons) at crystal-lattice defects of minerals such as quartz and feldspar (Aitken 1998). This charge can be measured as a semi-stable signal with stimulation by an appropriate wavelength of light. When the minerals are exposed to daylight, they lose their charge. The basic assumption of the method is that the grains had no charge when they were deposited. During burial, the grains are shielded from sunlight and a trapped signal will build up and increase with burial time. By measuring the trapped signal and the environmental dose rate (ionising radiation rate calculated mainly based on concentrations of K, Th and U), the duration of burial after deposition can be calculated. Luminescence dating of mostly glacially sourced Quaternary sediments of Northern Switzerland faces different challenges: the resetting of the luminescence signal prior to deposition, i.e. partial bleaching, the components contributing to the signal (e.g. presence of a prominent medium or slow component in the initial part of some signals), saturation of signal, low signal intensity, and signal stability or anomalous fading (Mueller et al. 2024). The luminescence methodology was further explored to improve the dating of these sediments in the North Alpine Foreland (e.g. Bickel et al. 2015, Klasen et al. 2016, Trauerstein et al. 2017, Lowick & Valla 2018, Rades et al. 2018, Mueller & Preusser 2022, Mueller et al. 2024). Generally, in Northern Switzerland optically stimulated luminescence (OSL) using blue light on quartz and infrared stimulated luminescence on feldspar (IRSL) were the most promising luminescence methods (Fig. 2-15). OSL on quartz has the advantage that quartz bleaches more easily. However, the signal saturates earlier, the signal intensity might be low, a part of the signal might be lost, and other components might contribute to the signal. IRSL on feldspar has the advantage that the signal saturates later, allowing potentially older sediments to be dated. However, the signal is known to bleach slower, making partial bleaching a more common artefact, and has to be corrected for fading, i.e. signal loss over time.

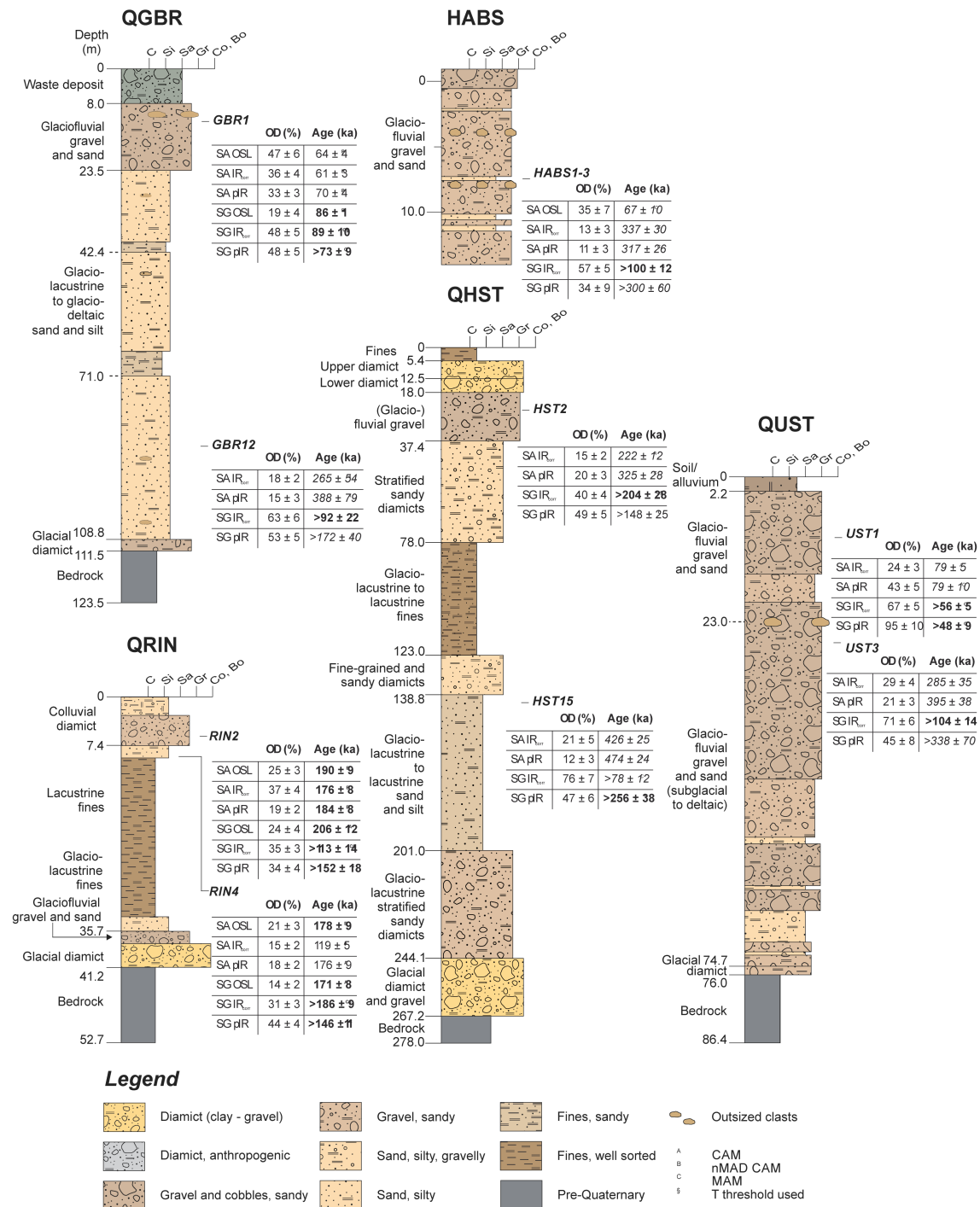


Fig. 2-15: Ages derived using different luminescence signals in QBO drill cores and outcrops

Ages indicated in bold are those considered to be most reliable. Ages marked with ">" represent absolute minimum ages. For further information see Mueller et al. (2024). SA: Single aliquot, SG: Single grain, OSL: Optically stimulated luminescence, IR_{cor}: corrected infrared stimulated luminescence, pIR: post-infrared infrared stimulated luminescence, CAM: Central age model, nMAD: normalised median absolute deviation, MAM: Minimum age model, T: Test dose. Modified after Mueller et al. (2024).

Legend:

- Siting region
- Luminescence ages (general)
- Luminescence ages (overdeepening)

Morphostratigraphy

- Late Pleistocene
- Niederterrasse
- Middle Pleistocene
- Hochterrasse

Early Pleistocene

- Tieferer Deckenschotter
- Höherer Deckenschotter

Age Ranges:

- MIS 4 = 0.071 – 0.057 Ma
- MIS 6 = 0.191 – 0.130 Ma

Ages (in Ma or as MIS) are colour-coded according to the morphostratigraphic code of the respective deposits and integrated from Kamleitner et al. (2023), Buechi et al. (2017), Anselmetti et al. (2010), Dehnert et al. (2012), Lowick et al. (2015), Gaar et al. (2019), Buechi et al. (2018, 2024), Mueller et al. (2020, 2024), Gegg (2021) and Gegg et al. (2023). For locations see Fig. 2-14 and Appendix A.1.

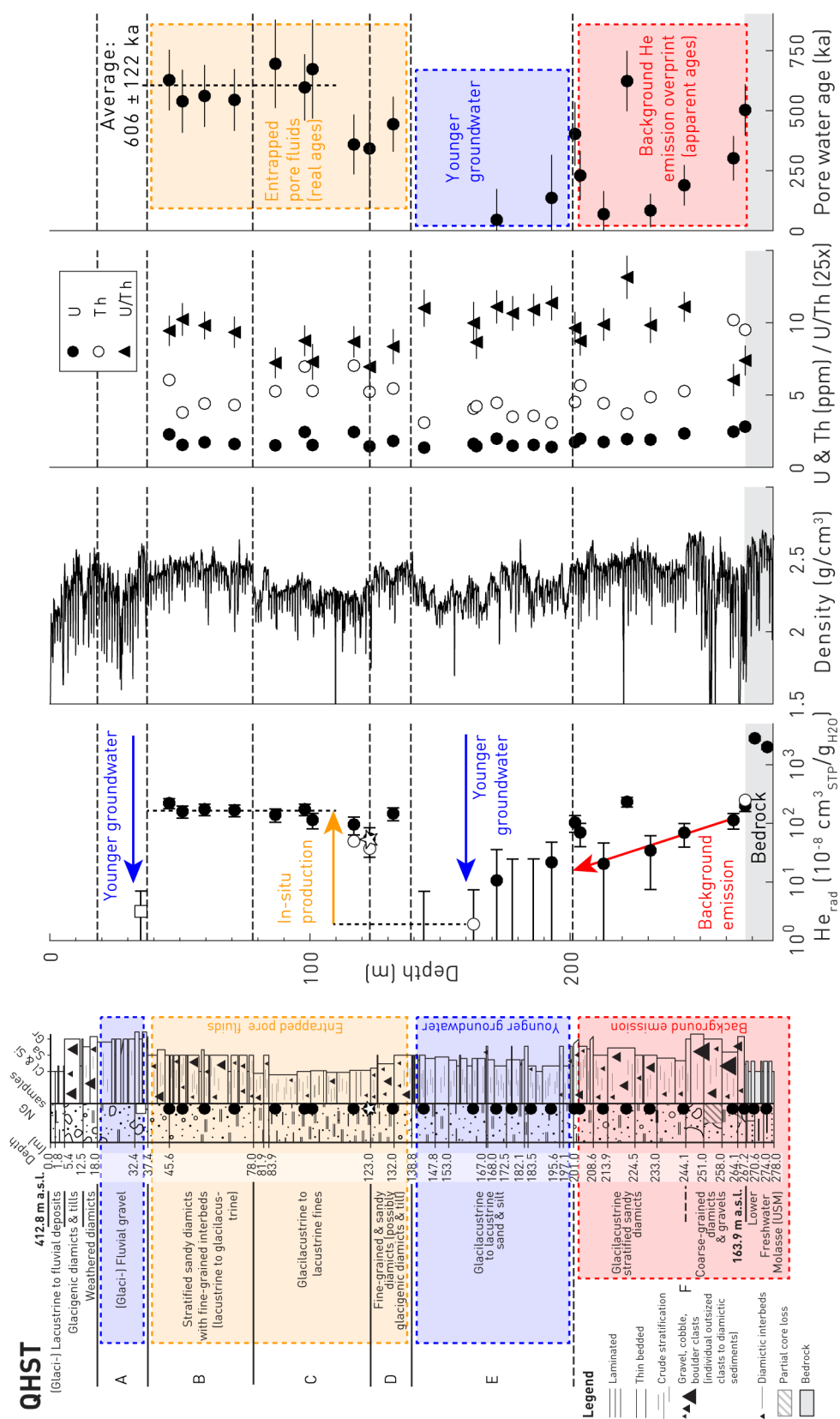


Fig. 2-17: $^4\text{He}/\text{U-Th}$ dating of porewater in QBO Hochfelden-Strassberg (QHST)

Lithology, noble-gas sample depths, radiogenic He concentrations (He_{rad}), sediment bulk density, U and Th concentrations, U/Th ratios, and porewater ages of samples from Nagra borehole QHST (modified after Tomonaga et al. 2024). Horizontal lines and letters (A – F) are lithological units relevant for the interpretation of the fluid transport in the pore space. He_{rad} determined with new method (Headspace Method, portable mass spectrometry, filled black circles) can be compared with measurements by static mass spectrometry: groundwater sample (Conventional Method, empty square), porewater samples (empty circles), and porewater sample (star). STP: standard temperature and pressure, Cl & Si: clay and silt, Sa: sand, Gr: gravel. Modified after Tomonaga et al. (2024).

$^4\text{He}/\text{U-Th}$ dating of porewater

The local decay of U and Th in the sediment matrix produces radiogenic ^4He , which can accumulate in the porewater. The porewater residence time in the sediment can be calculated by measuring the accumulated radiogenic ^4He and by constraining the ^4He production/accumulation rates (Torgersen & Clarke 1985). The porewater residence time provides a minimum age for the deposition of sediments. However, to obtain meaningful porewater residence times, the radiogenic ^4He must be retained within the sediment. Additionally, no or only minor amounts of ^4He should migrate into the sediment, e.g. from the underlying bedrock. Ideally, the radiogenic ^4He in the sediment is the result of in-situ ^4He production under closed-system conditions.

In Northern Switzerland, the $^4\text{He}/\text{U-Th}$ dating method has been successfully applied on porewater from the Quaternary borehole Hochfelden-Strassberg (QHST) by Tomonaga et al. (2024; Fig. 2-17). There, the radiogenic ^4He was entrapped in glaciolacustrine and diamictic sediments and practically isolated from external ^4He input by overlying and underlying aquifers. The derived average age of 606 ± 122 ka provides a minimum age for the Quaternary glacial valley fill in the overdeepened Strassberg Trough at QHST. Despite the promising results, the application of $^4\text{He}/\text{U-Th}$ dating remains difficult as suitable closed-system conditions seem to be very rare, and the analytical requirements render the methodology challenging.

Cosmogenic nuclide dating

Cosmic rays entering the atmosphere mainly comprise protons and alpha particles. Interactions with target nuclei in the atmosphere generate a cascade of secondary particles. The secondary particles cause reactions in the lattices of minerals in rocks and sediments exposed at the Earth's surface. The most important reactions for the production of cosmogenic nuclides are spallation and muon reactions (Ivy-Ochs & Kober 2008). The main terrestrial cosmogenic nuclide (TCN) products in use are ^{10}Be and ^{26}Al in quartz-rich rocks, which can be quantified using accelerator mass spectrometry and from which either exposure ages (e.g. of moraine blocks or landslide blocks) or denudation rates (by analysing thousands of river sand grains) can be calculated.

Exposure age dating was used for dating glacial moraine boulders to the last glacial maximum (LGM) in the northern Swiss Alpine Foreland (between 26 and 22 ka; Kamleitner et al. 2023, Reber et al. 2014).

Another application is the dating of Quaternary gravel terrace deposits by using either: a) the depth-profile dating method (Hidy et al. 2010), b) simple or isochron-burial dating method (Granger et al. 1997, Balco & Rovey 2008, Erlanger et al. 2012; Fig. 2-18), or c) the P-PINI approach (a modelling approach for interpreting burial data; Knudsen et al. 2020, Nørgaard et al. 2023). The development of the latter approach resulted from the failure or fallibility of approaches a) and b) for fluvio-glacial gravel deposits.

While the depth-profile method has failed in fluvio-glacial deposits of Northern Switzerland in the case of prevailing high erosion rates after deposition and has been disregarded for the dating of fluvio-glacial terraces (Knudsen et al. 2020), the cosmogenic burial dating methods rely on the assumption that the investigated sediment sources underwent steady rates of erosion and continuous exposure to cosmic rays in their source areas. However, sediment derived from high Alpine glaciated regions are unlikely to fulfil such steady history but are rather characterised by transient processes (e.g. glacial quarrying and/or landsliding, or temporary cosmic ray shielding beneath glaciers and/or sediment). As a consequence, such samples commonly yield low nuclide abundances and variable $^{26}\text{Al}/^{10}\text{Be}$ ratios that violate burial-age method assumptions. Further, once any sediment has been transported to a sink area (a terrace body in the foreland), post-burial nuclide production may be relatively high in the case of long exposure and/or shallow burial (Granger & Muzikar 2001). P-PINI is a newly developed dating tool for $^{26}\text{Al}/^{10}\text{Be}$ ratios where inverse Monte Carlo modelling of TCN abundances is applied to estimate the burial age and history of sedimentary deposits (Knudsen et al. 2020). This approach can be furthermore applied to sediments sourced from landscapes that are characterised by abrupt, non-steady erosion, discontinuous exposure, and catchments with elevation-dependent $^{26}\text{Al}/^{10}\text{Be}$ production ratios (Nørgaard et al. 2023).

TCN dating of fluvio-glacial terraces in Switzerland has received much attention and has been performed in various studies to decipher past geomorphic processes (e.g. Akçar et al. 2014, 2017, Claude et al. 2017, 2019, Dieleman et al. 2022a, 2022b, Grischott et al. 2020, Knudsen et al. 2020). Unfortunately, most of the sampled clasts have only low inherited concentrations, which makes age determination using the conventional isochron burial method problematic and can lead to misleading results. For the Swiss Alpine Foreland, only a limited number of samples fulfil the requirements for P-PINI analysis (Knudsen et al. 2020, Nørgaard et al. 2023); these preferentially have higher concentrations and lower $^{26}\text{Al}/^{10}\text{Be}$ ratios. All data were rigorously checked for physically valid production ratios (< 8.3) and the ratio of inherited to post-depositional cosmogenic nuclide contribution, where the latter should not dominate. As a result, several previously published cosmogenic nuclide-derived isochron-burial ages did not fulfil those P-PINI criteria and are not used in the interpretation. Considering the various challenges, most ages cluster around the Mid-Pleistocene Transition, while some individual sites yielding ages also younger or older. The distinction based on classical morphostratigraphy (HDS, TDS) tends to be supported by the cosmogenic nuclide ages, where the HDS ages are older than the TDS ages. An exception here are the gravel deposits of Hungerbol (HUN in Tab. 2-1 and Fig. 2-14; Graf 2009a), which seem to have a rather old depositional age compared to the other TDS ages (Fig. 2-19).

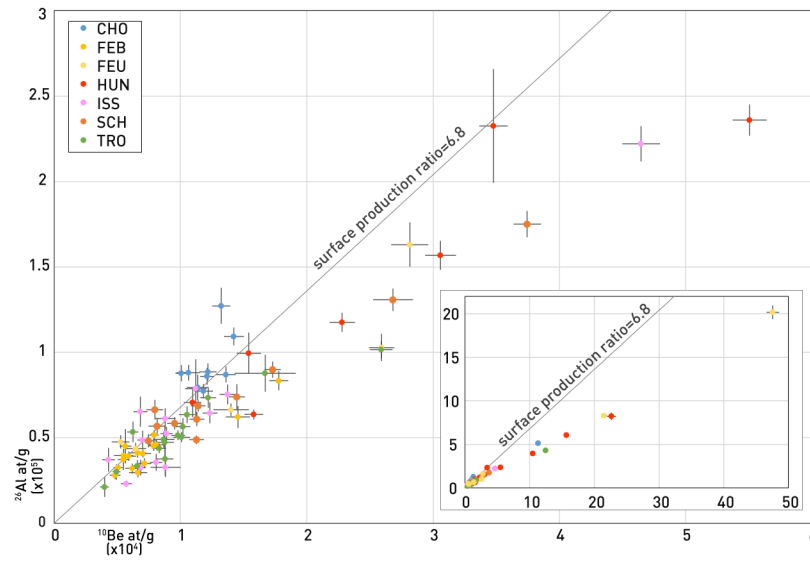


Fig. 2-18: Measured ^{10}Be and ^{26}Al concentrations for all samples analysed from the seven studied Deckenschotter units

Accelerator mass spectrometer (AMS)-measured ^{10}Be and ^{26}Al concentrations of the samples analysed in newly investigated Deckenschotter units (Nagra, courtesy of E. Bros), which were then used for P-PINI age determinations (together with recalculated published concentrations, see Fig. 2-19).

The measurement uncertainties are at the 1-sigma level. The $^{26}\text{Al}/^{10}\text{Be}$ surface production ratio of 6.8, which is generally assumed to represent the initial cosmogenic production ratio, is shown by a solid line (note that this surface production ratio may vary slightly in the P-PINI code due to the glacial environment, for details see Knudsen et al. 2020). Simplified, the age can be derived from the clockwise deviation of the samples from the line representing the initial (e.g. surface) production ratio. The inset plot shows the complete dataset with a square indicating the extent of the main plot. It reveals that most of the sampled clasts have only low inherited concentrations, which makes age determination using the conventional isochron burial method problematic and can lead to misleading results. For the Swiss Foreland, only a limited number of samples fulfil the requirements for P-PINI analysis; these preferably have higher concentrations and plot with higher concentrations in the inset. The P-PINI method has been additionally applied to published samples. See Tab. 2-1 and Fig. 2-14 for the sample site codes. FEU and FEB: samples from same locality FEU (Fig. 2-14) but taken from an upper and lower level, respectively.

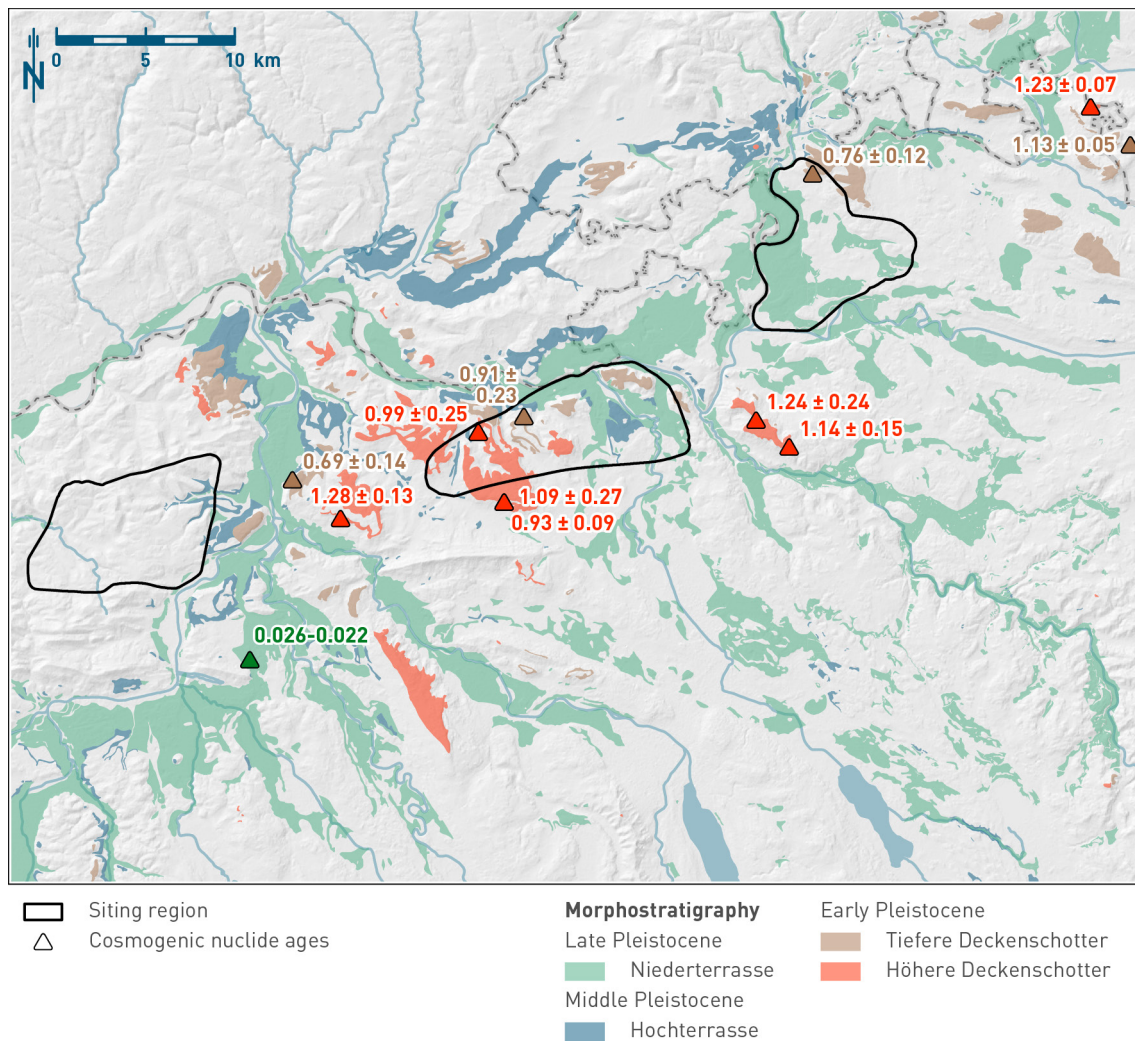


Fig. 2-19: Selected age information from cosmogenic nuclide dating in Pleistocene sediments of Northern Switzerland

Ages in Ma are colour-coded according to the morphostratigraphic code of the respective deposits and integrated from Reber et al. (2014), Akçar et al. (2017), Claude et al. (2019), Grischott et al. (2020), Dieleman et al. (2022a), Kamleitner et al. (2023) and Nagra (courtesy of E. Bros). Previously published data (with the exception of age 0.026 – 0.022 Ma at BIR) have been recalculated using the P-PINI method (Knudsen et al. 2020). Other cosmogenic nuclide data points measured with the isochron-burial method did not pass the P-PINI criteria and are not shown here. For locations see Fig. 2-14 and Appendix A.1.

Biostratigraphy and aminostratigraphy

Biostratigraphy is based on correlating and assigning relative ages of sediments by analysing the fossil assemblages contained in sedimentary strata. Special environmental conditions must be met to preserve fossil biological assemblages, which are in general rare in the case of the terrestrial, partially preserved sedimentary archives of Northern Switzerland. For biostratigraphy to be effective, the fossil assemblages must show an evolution in their morphometrical characteristics within the required timescale, which can be traced by biostratigraphic marker species or evolutionary lineages. Additionally, numerically dated sites are needed for the chronological classification of a given biostratigraphic assemblage.

In continental Europe, the remains of rodents have proven to be essential for the establishment of a reliable Quaternary biochronology (Fejfar et al. 1998). The different genera/taxa of the small mammals can be classified by a detailed analysis of their molars. In the Pleistocene sediments of Northern Switzerland, small mammal fossil assemblages are very rare. One exception is the arvicoline rodent assemblage deposited in interglacial overbank deposits at the Irchel Hasli (HAS, Fig. 2-20). This Early Pleistocene fossil assemblage was described by Bolliger et al. (1996) and was reanalysed by Cuenca-Bescós (2015) and Thew et al. (2024) resulting in an age estimate of 2.1 – 1.8 Ma (Fig. 2-23).

Terrestrial and aquatic molluscs are much more common than mammal assemblages but are still rare in Pleistocene sediments from Northern Switzerland. So far, they have mostly been used as proxies for depositional environments, ecology and past climates. However, Thew (2024) investigated the potential of terrestrial and aquatic molluscs for the temporal analysis of these sediments (malacostratigraphy). In this study more than 75'000 mollusc fragments from around 120 terrestrial and aquatic species were identified just from the Irchel Plateau and many thousands more from a number of Middle and Late Pleistocene sites, from over 150 species. Several molluscan biostratigraphic marker species were found during this study. Some of these biostratigraphic marker fossils have long been extinct, such as *Clausilia stranzendorfensis* and *Cochlostoma salomoni*, which have so far only been found in Central Europe in sediments older than 1.8 Ma (Thew et al. 2024; Fig. 2-21). Several locations were identified in which malacostratigraphic information can give an age estimation (Fig. 2-23) to improve the Quaternary stratigraphy.

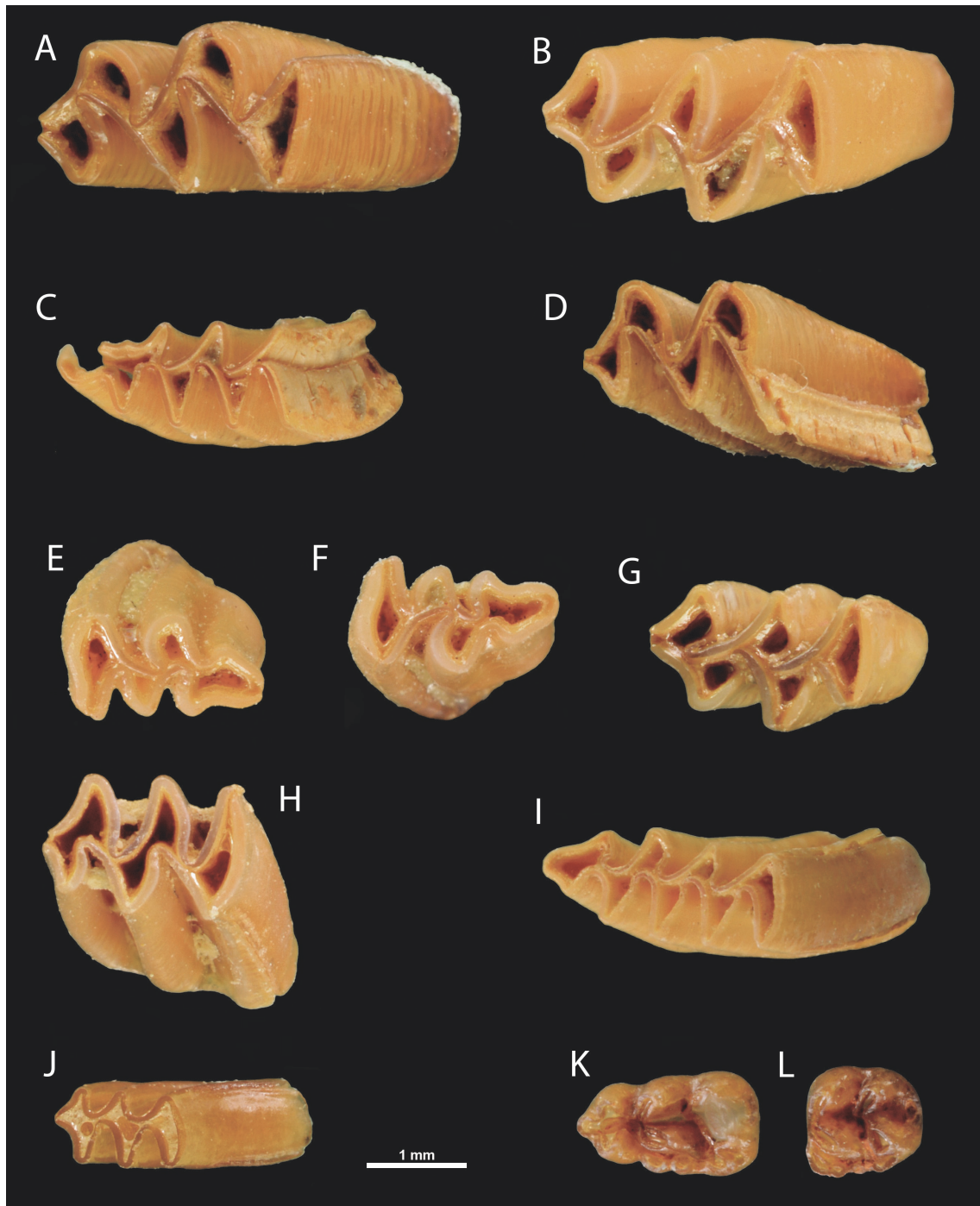


Fig. 2-20: The most important rodent molars from the Hasli Formation at the type locality Irchel Hasli (HAS, Layer 12 in occlusal view)

A. *Mimomys pliocaenicus* (M¹ sin.), B. *Mimomys pliocaenicus* (M¹ dext.), C. *Mimomys reidi/tigliensis* (m₁ dext.), D. *Mimomys* sp. (half upper molar), E. *Mimomys reidi* (M³ sin.), F. *Mimomys reidi* (M³ dext.), G. *Mimomys reidi* (M¹ sin.), H. *Borsodia* sp. (M¹ dext.), I. *Pliomys episcopalis/simplicior* (m₁ dext.), J. *Clethrionomys* cf. *kretzoi* (m₃ dext.), K. *Apodemus* cf. *sylvaticus*, (m₁ sin.), L. *Apodemus* cf. *sylvaticus*, (m₂ sin.) (Thew et al. 2024).

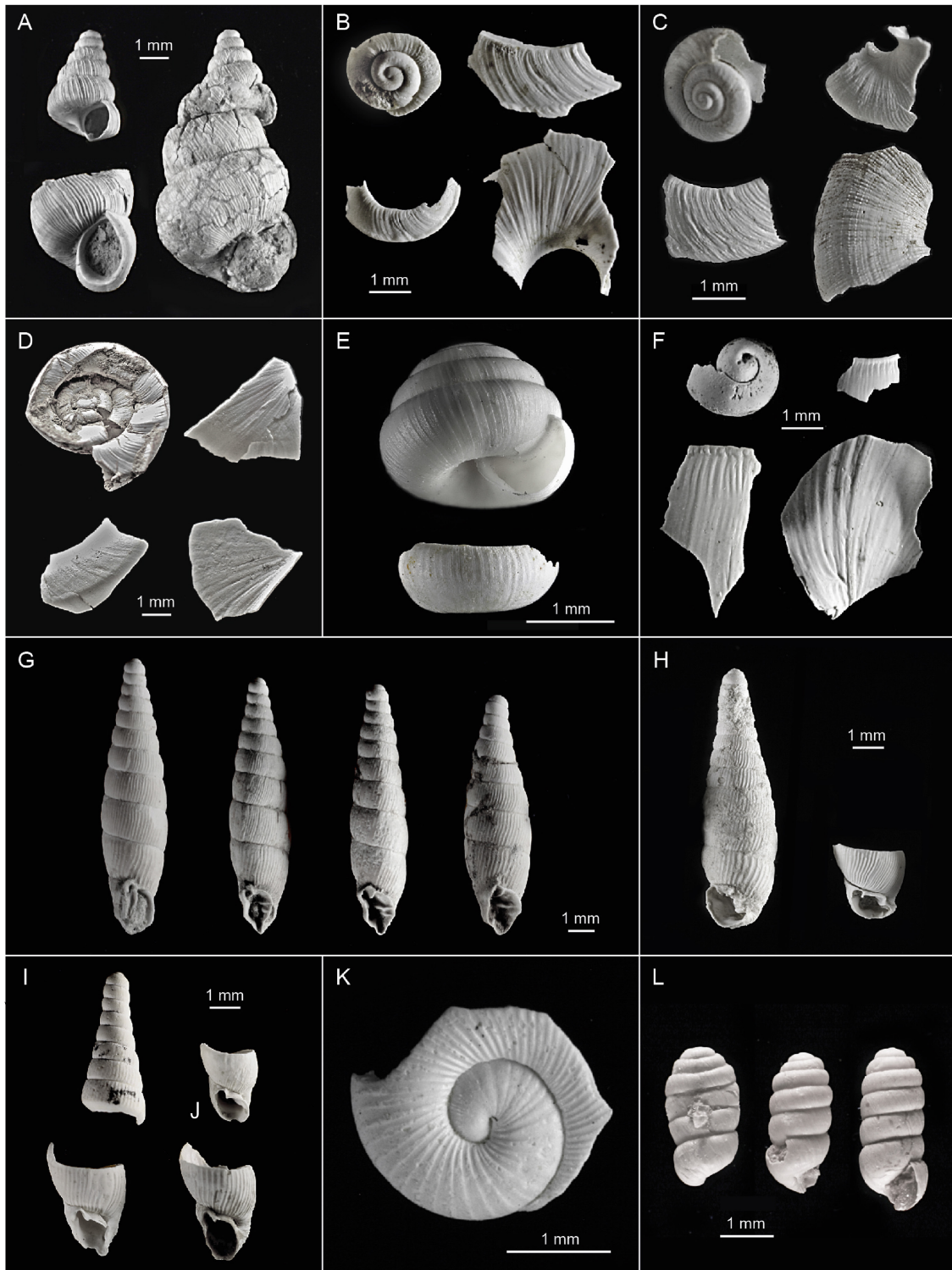


Fig. 2-21: Important molluscan species from the Irchel Plateau sites

A. *Cochlostoma salomoni*, B. *Archaeogopis acutus*, C. *Retinella elephantium*, D. *Monachoides vicinus*, E. *Spermodea lamellata**, F. *Poirertia dilatata dilatata*, G. *Clausilia stranzendorfensis*, H. *Macrogastra sessenheimensis*, I. *Neostyriaca corynodes ornatula*^{oo} + J. (top right) *Neostyriaca dehmi*, K. *Oxychilus steiningeri*, L. *Columella columella*. *Holocene reference shell from UK; *N. corynodes ornatula* shells from Hungerbol (HUN, TDS) (Thew et al. 2024).

Aminostratigraphy (also amino acid geochronology or amino-acid racemisation) is a chemostratigraphic approach that enables relative age determination in shells or other skeletal materials from marine and terrestrial fossils (Bada et al. 1970, Kaufman & Miller 1992). The approach uses reverse-phase high pressure liquid chromatography to analyse the degrading amino acid isomers after the death of an organism (Penkman et al. 2011). Most amino acids (the building blocks of proteins) can occur as two isomers that are chemically identical, but optically different: laevo-rotary and dextro-rotary isomers, termed L and D. Proteins are synthesised from L-amino acids, but, after death of an organism, the L-amino acid isomers can convert to the D-amino acids in a process termed racemisation (AAR = amino acid racemisation). The D/L proportion provides an estimate for the extent of degradation and can thus be used as an age estimate under certain circumstances. The IcPD (intra-crystalline protein decomposition) approach is based on a closed-system assumption in which other factors (water content, concentrations of cations, pH) are constant. The extent of racemisation is therefore only a function of time and temperature, so D/L data can be compared between different sites that have the same integrated temperature histories. An amino acid stratigraphy with temporal constraints can be established if several sections can be compiled and compared with independent age control (Penkman et al. 2011). Penkman et al. (2024) have created the first aminostratigraphy for the Quaternary of the Swiss Plateau using a set of biominerals (Fig. 2-22). In Northern Switzerland, the aminostratigraphy helps to define the relative age of sedimentary deposits at different locations (Fig. 2-23). For example, it shows the differentiation between the Early Pleistocene samples from the Irchel Plateau and samples from the Middle Pleistocene, Late Pleistocene and Holocene (Fig. 2-22).

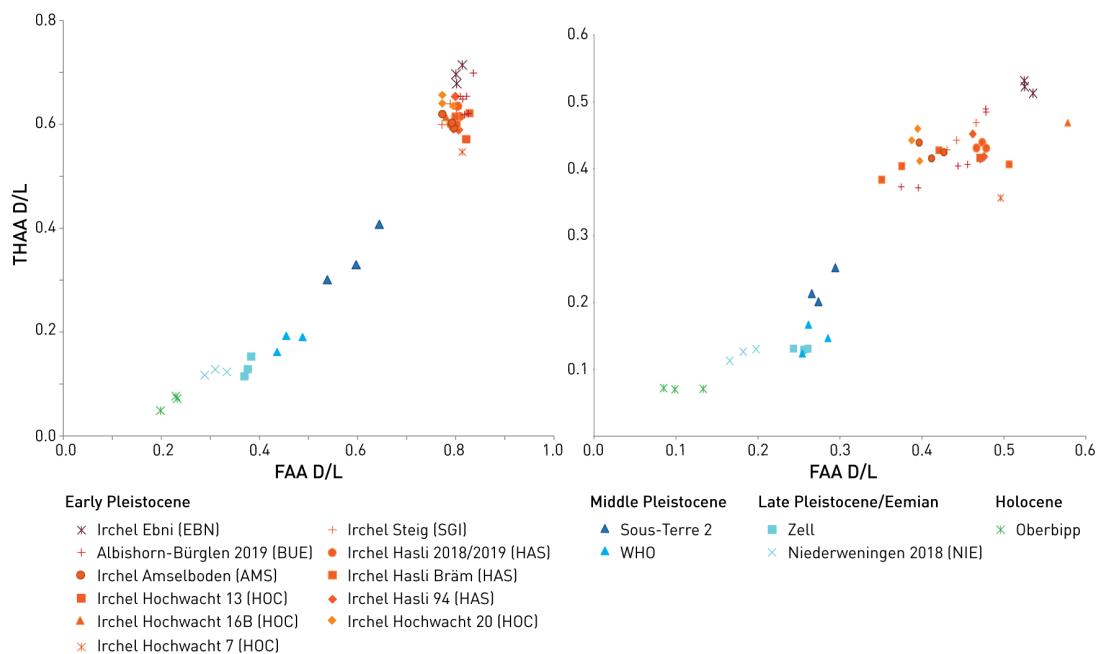


Fig. 2-22: Relative age sequence of Quaternary sites across the Swiss Plateau based on aminostratigraphy of slug plates

FAA D/L versus THAA D/L for Ala (left) and the slower racemising amino acid Glx (right). FAA = Free amino-acid fraction, THAA = Total hydrolysable amino acid fraction, Ala = Alanine, Glx = Glutamine + glutamic acid, D/L: These isomers are designated as either D (dextro-rotary) or L (laevo-rotary) depending upon whether they rotate plane polarised light to the right or left, respectively. In living organisms, the amino acids in protein are almost exclusively L, and the D/L value approaches zero. For locations see Fig. 2-14, Appendix A.1 and Penkman et al. (2024). Modified after Thew et al. (2024).

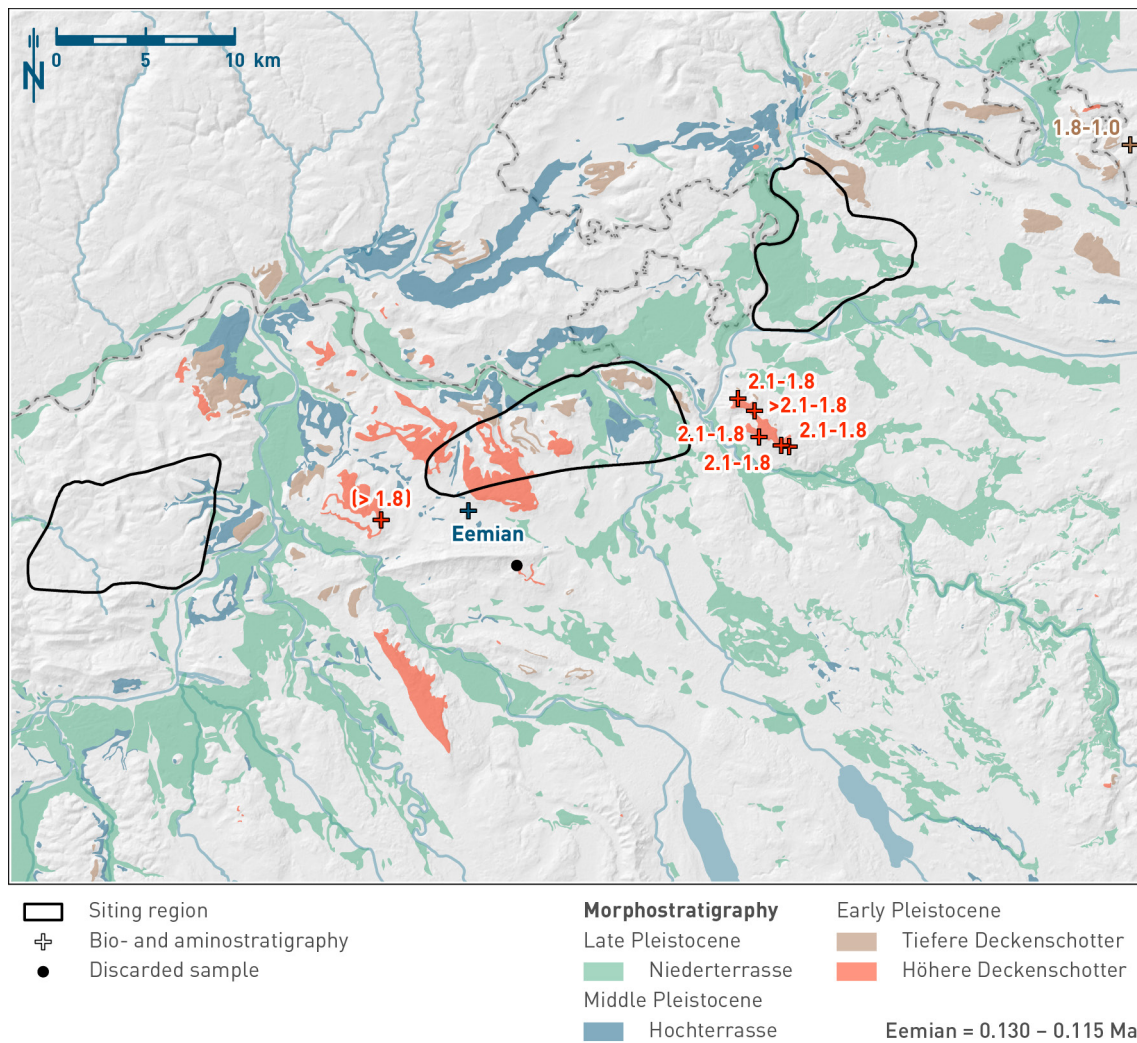


Fig. 2-23: Selected results of biostratigraphy and aminostratigraphy in Pleistocene sediments of Northern Switzerland

Ages are colour-coded according to the morphostratigraphic code of the respective deposits and integrated from Frei (1912), Penkman et al. (2024), Thew et al. (2024), and references therein. For locations see Fig. 2-14 and Appendix A.1.

Magnetic polarity stratigraphy

The Earth's geomagnetic field is a dipole field that changes its direction in periods between 10^2 and 10^7 years. These polarity changes (reversals) of the past are well dated using radiometric methods and are documented in the geomagnetic polarity timescale (e.g. Ogg 2020). Within dating accuracy, magnetic reversals are synchronous globally and can therefore be used for correlations over long distances. In particular, sediment successions may provide continuous records covering several palaeomagnetic reversals. A basic requirement is that the sediments contain certain magnetic minerals (e.g. magnetite) in a very small grain size to attain a magnetisation, also known as the natural detrital remanent magnetism (DRM). This magnetisation thus reflects the Earth's geomagnetic field at the time of deposition (Walker 2005). In Northern Switzerland,

it is challenging to find suitable fine-grained material with a strong palaeomagnetic signal that has not been altered after its deposition. Additionally, the application of magnetic polarity stratigraphy is hampered by the relictic nature of the archives.

Palaeomagnetic studies in Northern Switzerland were carried out by Bolliger et al. (1996) and Graf (1993, 2019). More recently, an extensive palaeomagnetic investigation was carried out by Scheidt et al. 2023) on the Deckenschotter deposits close to the siting regions (Fig. 2-24, Fig. 2-25). The HDS revealed reversed and normal magnetisations, indicating deposition > 0.773 Ma (> Brunhes Chron, Fig. 2-13). The TDS showed only normal directions. Palaeomagnetic reversals (inverse to normal) were found in the locations Uetliberg Kulm (UEK), Irchel Hasli (HAS) and Hochwacht (HOC). The results are especially valuable in combination with geochronological results from other dating methods (Appendix A.1).

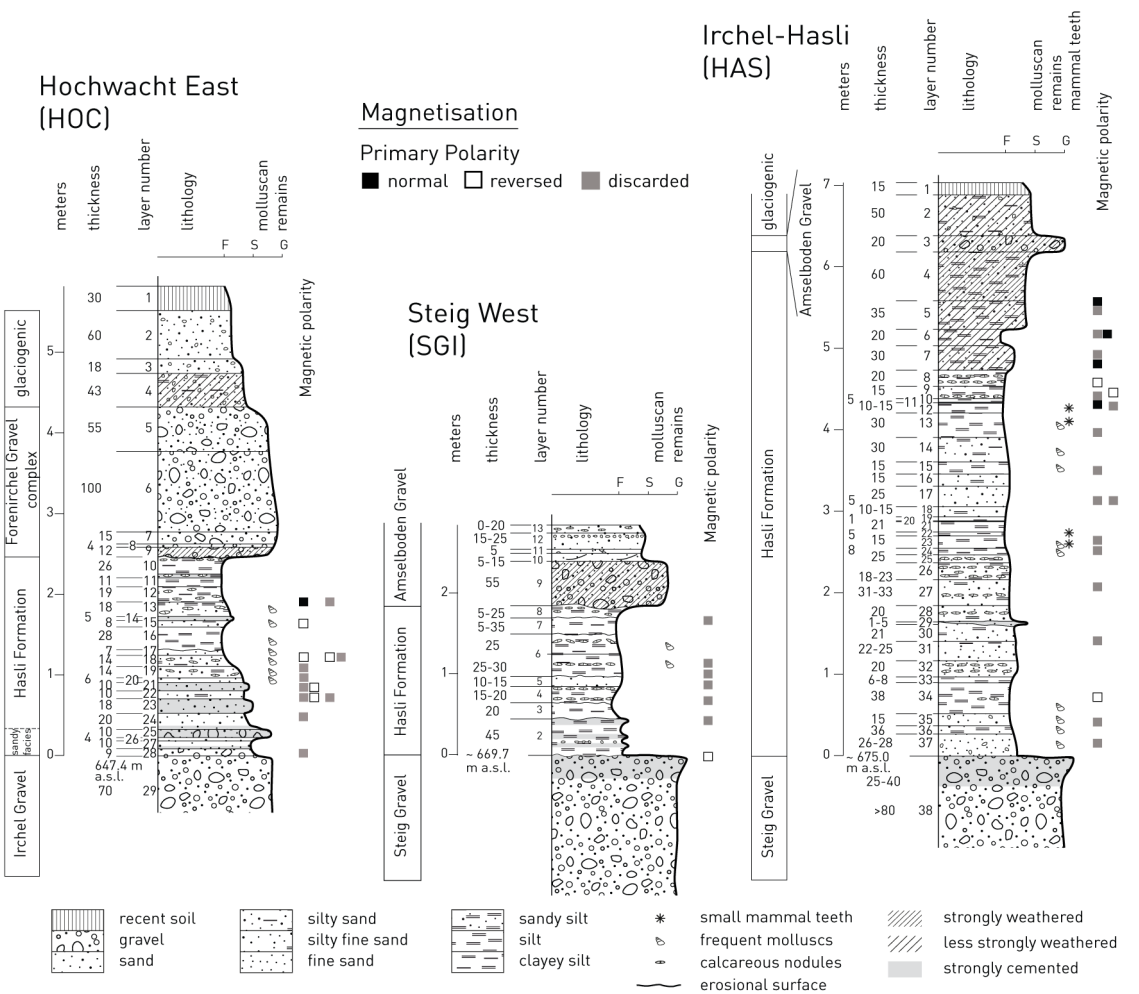


Fig. 2-24: Results of the palaeomagnetic analysis from the Irchel Plateau plotted against the study profiles

For locations see Fig. 2-14 and Appendix A.1. Modified after Thew et al. (2024).

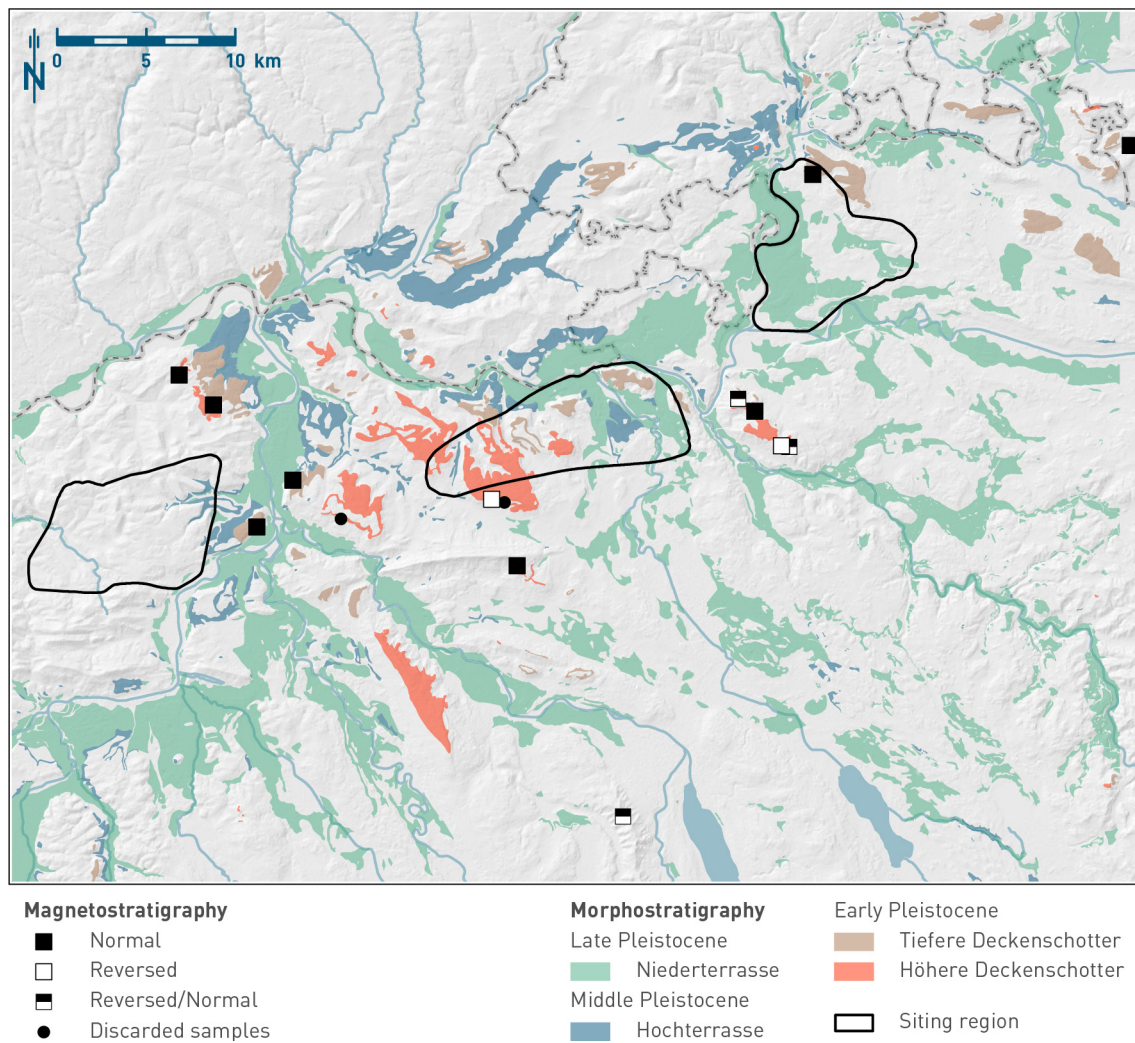


Fig. 2-25: Reconstructed polarity of the Earth's magnetic field during deposition of different Pleistocene sediments at selected sites in Northern Switzerland

Sampling locations are colour-coded according to palaeomagnetisation from Graf (1993, 2019) and Scheidt et al. (2023). For locations see Fig. 2-14 and Appendix A.1.

2.4.1.4 Genesis and processes

Introduction

With a view to a long-term reconstruction of regional fluvial incision (e.g. incision of the Aare – Rhine fluvial system), sedimentary archives are needed that trace the evolution of the fluvial baselevel. Traditionally, the division into four morphostratigraphic units in Northern Switzerland has been interpreted to reflect the stepwise lowering of the local³ baselevel (from top to bottom): HDS, TDS, HT and NT (Graf & Burkhalter 2016 and references therein; Fig. 2-29 and Fig. 2-34).

³ Local baselevel in this report is related to Northern Switzerland, thus constitutes the Hochrhein and larger tributaries (e.g. Aare, Limmat, Reuss, Thur Rivers) – related to this scale is the term local erosion base, which is introduced in Section 3.1

However, for every sedimentary archive the depositional context has to be taken into account or needs to be interpreted to infer its relationship to the former local baselevel. Caution is required, for example, if potential ice-marginal and intra-morainic deposition or fluvial transience were involved in these processes; this caveat is discussed below and illustrated with regional examples from Northern Switzerland and surrounding areas.

Ice-marginal and intra-morainic deposition

In sedimentary archives representing ice-marginal or intra-morainic deposition, it might be not possible to infer the local baselevel. During glaciations, the drainage system is modified, and a previous interglacial fluvial system might be overprinted. Following the classical theory of the "Glaziale Serie" proposed by Penck (1882), fluvioglacial gravel bodies can be accumulated starting at the terminus of the glacier to a level that is several decametres higher than the previous interglacial baselevel. Following glaciation, the river will incise into these gravels again and adjust itself to the local baselevel. Examples for such gravel bodies can be found in the Late Pleistocene gravels (Niederterrasse) of the Rafzerfeld or near Wettingen. Fluvioglacial gravels may also be deposited along the side of a glacier in an ice-marginal drainage system (e.g. kame terrace). The resulting fluvioglacial gravels may therefore occur at an elevated position along the flanks of the valley at a position that is decoupled from the local baselevel (Keller & Krayss 2000). Potentially, the two sedimentary bodies – morphostratigraphically defined as HDS – at Cholholz near Boppelsen at the southern flank of the Lägern (Scheidt et al. 2023, Thew 2024) and the sediments at Üetliberg Kulm near Zürich (Graf 2019), represent such ice-marginal deposits. In addition, lacustrine sediments may be deposited in an elevated position relative to the local baselevel due to river impoundment by ice (lake of Klettgau; Lowick et al. 2015, moraines (Ossingen-Riegel and lake of Weiningen; Müller 2013) and landslide deposits (Rinikerfeld palaeolake; Gegg et al. 2023).

Finally, gravels near Roggenfeld (ROG in Fig. 2-14, Appendix A.1) in the Mettauertal exemplify the challenge in identifying the depositional context of a gravel body of limited extent. Originally, these gravels were interpreted morphostratigraphically to be part of the HDS with an Early Pleistocene age (Diebold et al. 2006). New findings, however, indicate that these gravels could have alternatively formed in an ice-marginal, perhaps deltaic depositional environment during the Middle Pleistocene, which was much higher than the non-glacial local baselevel at that time (Scheidt et al. 2023).

Fluvial transience

The river profiles in Northern Switzerland are influenced by several drivers, such as uplift and subsidence, climate, and drainage reorganisation (Sections 2.1 to 2.3). This leads to transient states of river gradients and depositional sequences. Fluvial terraces have to be analysed as geomorphic markers in the context of their position relative to a migrating water divide and to the local baselevel.

For example, during the Late Pleistocene, after ca. 18 ka, the Rhine system captured the Wutach drainage from the Danube system. This led to a rapid incision of up to 150 m within only ca. 6 kyr, producing a deep gorge (Einsele & Ricken 1995, Hebestreit 1995). For the gravels at Irchel (Höhere Deckenschotter), a scenario was proposed where the Rhine River captured the Alpine sources of the Danube (Claude et al. 2019). Care should also be taken with sedimentary archives that represent a tributary system with a relatively small catchment and a steep gradient towards the main channel. The upstream migration of knickpoints (e.g. due to lithological contrasts, drainage diversion, tectonics) is yet another aspect that may lead to spatially and temporarily increased incision rates at local scales. A prominent knickpoint in Northern Switzerland are the

Rhine Falls with a drop height of ca. 20 m. Therefore, changes in the local baselevel and corresponding local incision rates can fluctuate temporally, but they may not be representative of the incision rate of the principal river.

2.4.2 Past landscape evolution

2.4.2.1 Pliocene

In the Pliocene to Quaternary, major drainage reorganisations with respect to the principal rivers that drained the palaeolandscape of Northern Switzerland, the Aare, Rhine, and Danube, occurred (Ziegler & Fraefel 2009, Heuberger et al. 2014). For the Pliocene, absolute age control hardly exists and therefore the age of the changes in the drainage system can only be approximated.

From the latest Miocene to the Pliocene (ca. 7.3 – 4.2 Ma), the North Alpine Foreland drained via the Aare – Danube River system in a northeastward direction to the Black Sea (Fig. 2-26; Heuberger et al. 2014, Ziegler & Fraefel 2009). This assessment is supported by relict deposits of quartzite-rich Alpine gravel at high elevation between Blumberg and Ulm (Villinger 2003), up to ~ 270 m above the present course of the Danube (refer to approximate extent of reddish-brown dots in Fig. 2-34). This is supported by the lack of heavy-mineral assemblages with Alpine provenance in the deposits of the Upper Rhine Graben during this period (e.g. Hagedorn & Boenigk 2008).

About 4.2 million years ago, the Doubs incised progressively eastwards, which was coupled with a successive eastward migration of the drainage divide between Doubs and Danube. Finally, the Doubs captured the headwaters of the Aare-Danube in the area of Waldshut (Fig. 2-26). This capture was associated with a baselevel drop of at least 200 m (Schlunegger & Mosar 2011), as indicated by the massive Sundgau Gravel in the southernmost Upper Rhine Graben (e.g. Liniger 1966, 1967, Giamboni et al. 2004; purple dots in Fig. 2-34) and Forêt de Chaux Gravel in the northeastern Bresse Graben (e.g. Petit et al. 1996, Madritsch et al. 2010). Based on biostratigraphy, the Forêt de Chaux Gravel was deposited between ca. 4.2 and 2.9 Ma (Petit et al. 1996, Fejfar et al. 1998, Madritsch et al. 2010). The Sundgau Gravel has been inferred to be of the same age. The migration of the divide between Doubs and Danube was potentially related to the sea-level drop in the Mediterranean Sea during the Messinian salinity crisis resulting from isolation from the Atlantic Ocean (ca. 5.5 Ma; Hsü et al. 1973, Clauzon et al. 1996, Krijgsman et al. 1999) and retrogressive incision of the palaeo-Rhone and Doubs systems. Alternatively, the headward erosion along the proto-Doubs River and the following fluvial reorganisation could also have been caused by strong subsidence in the Bresse Graben (Schlunegger & Mosar 2011).

Towards the end of the Pliocene, at the transition to the Quaternary, the Aare-Doubs was captured by the proto-Rhine River in the region north of Basel (Giamboni et al. 2004, Heuberger et al. 2014). Since then, the Aare – Rhine system has drained through the Upper Rhine Graben into the North Sea (Fig. 2-26). Alpine provenance within the Upper and Lower Rhine Graben deposits provides evidence for this change in the drainage at ca. 2.8 – 2.6 Ma (Hagedorn & Boenigk 2008, Westerhoff et al. 2008). This river reorganisation might have been caused by either tectonic activity in the southernmost Upper Rhine Graben or gravel accumulation in the area between Basel and the Sundgau with increased sediment discharge from the Molasse Basin. This capture was again associated with a baselevel drop, although of lower amount. The total baselevel drop from the Aare – Donau system to the Aare – Rhine system was ca. 300 – 500 m (Fig. 2-26).

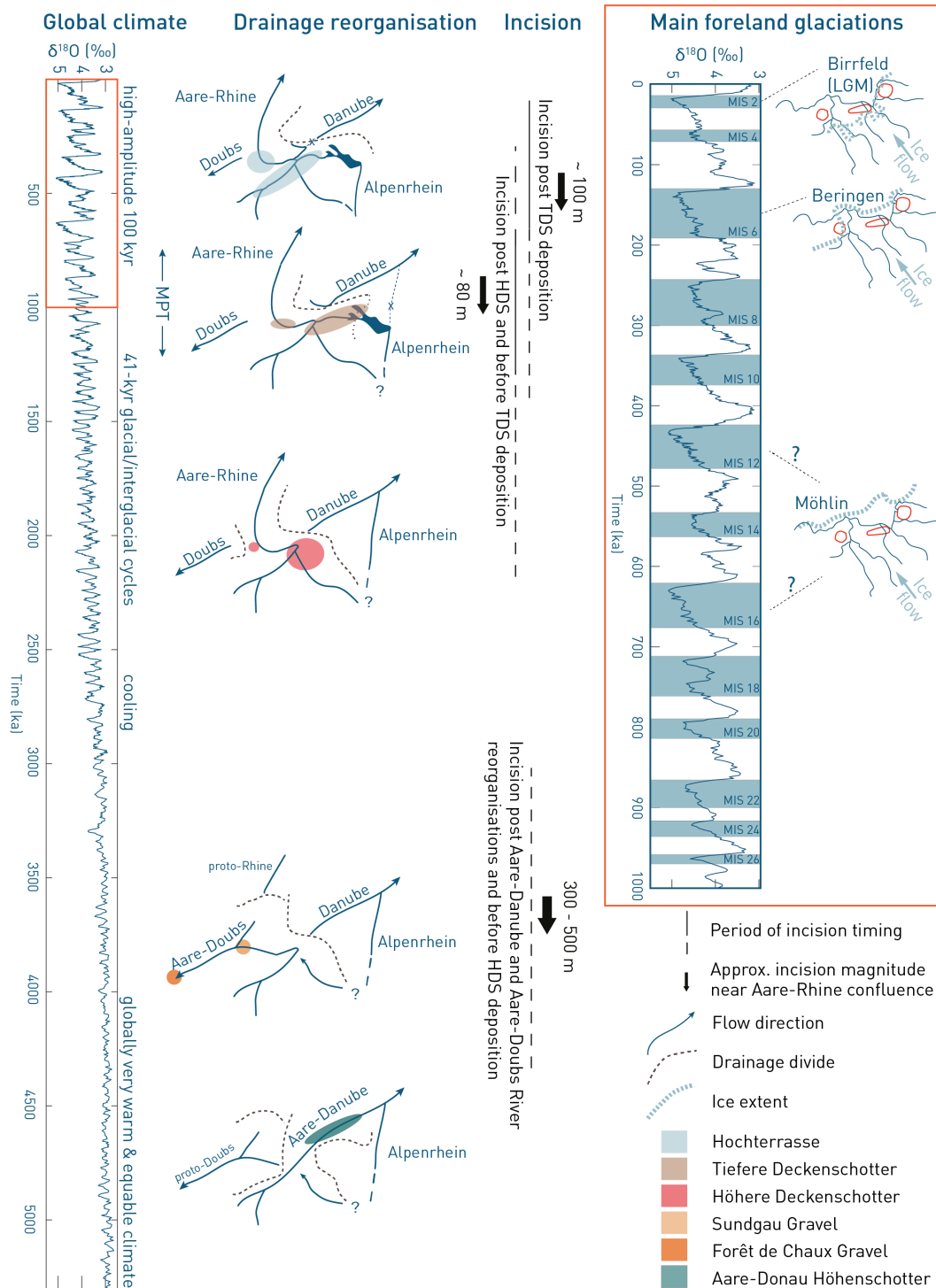


Fig. 2-26: Main landscape-forming events in Northern Switzerland during the past 5 Myr

Compilation shows global and regional events linked to the Pliocene-Quaternary landscape evolution in Northern Switzerland. $\delta^{18}\text{O}$ record with marine isotope stages (MIS) based on Lisiecki & Raymo (2005). Incision magnitudes of major drainage reorganisations are based on visual inspection of the Aare – Rhine confluence area based on Heuberger et al. (2014) and yield ~ 300 – 500 m for the Aare-Doubs incision, ~ 80 m between HDS (Höhere Deckenschotter) and TDS (Tiefere Deckenschotter) deposition, and ~ 100 m after TDS deposition. Maps with ice extents show the present-day river network for orientation.

2.4.2.2 Early Pleistocene

Between 2.6 and 0.77 Ma, glaciers advanced into the North Alpine Foreland for the first time (Fig. 2-27). Downstream of the glaciers, extensive plains with fluvioglacial gravels were formed, characterised by wider-than-today channel systems (Graf 1993, 2009a). In the Early Pleistocene Northern Switzerland had a less pronounced relief than today. This is indicated by the generally rather wide and flat valleys in which these fluvioglacial gravels (Deckenschotter) were deposited (Heuberger et al. 2014; Fig. 2-27). Glacial and interglacial periods succeeded each other, dominated by a 41-kyr periodicity (Section 2.2). With the Mid-Pleistocene Transition (ca. 1.25 – 0.7 Ma; Clark et al. 2006), the 100-kyr periodicity started to dominate the glacial cycles, leading also to more extensive Alpine glaciations. Large glaciations have reached the foreland since the Early Pleistocene, but probably had their largest erosion potential only within the last one million years, when cyclicity changed to a high-amplitude 100-kyr glacial/ interglacial cycle (e.g. Preusser et al. 2011).

During the Early Pleistocene, the fluvial baselevel was lowered, leading to incision into accumulated gravels, but also into bedrock, mainly represented by Late Oligocene to Miocene sedimentary rocks of the Molasse (Graf 1993, Kuhlemann et al. 2013). The Rhine River system evolved during several phases of drainage reorganisation that resulted in significant drops in the local baselevel, particularly in the eastern part of the study area (Fig. 2-26). The extent and timing of this incision varied spatially depending on the distance to the drainage divide or to the baselevel of the capturing river. The drainage divide between the Rhine and the Danube systems continued to shift eastwards because of the headward erosion of the Rhine River and its precursors. Beginning at these early stages, the drainage has developed in favour of the Rhine catchment and to the disadvantage of the Danube catchment (Winterberg & Willett 2019). The eastward shift of the drainage divide was probably facilitated by glacial erosion and deposition. The course of the Alpenrhein River has also changed over time and, since approximately the end of the Early Pleistocene, it no longer drains northwards into the Danube River, but towards the Hochrhein River (Heuberger et al. 2014, Ziegler & Fraefel 2009, Müller 1995, Graf 1993).

The fluvial incision led to the development of gravel terraces forming flat-topped hills in the landscape (Deckenschotter) that constitute vestiges of the old river courses (Fig. 2-28 and Fig. 2-29). Generally, during progressive fluvial incision and the formation of staircase-like terrace landscapes, older terraces are at a higher elevation, while younger terraces form at successively lower elevations (Burbank & Anderson 2012). The HDS and TDS units are further characterised by complex internal sedimentary structures (cut-and-fill sequences) in which several depositional and erosive phases can be distinguished based on clast provenance and orientation (Graf 1993, Dieleman et al. 2022a). Overall, this internal stratigraphy of the Deckenschotter is interpreted in the context of climatically controlled depositional-erosional episodes. For an internal subdivision of the Deckenschotter, the succession of (peri)glacial to interglacial evidence was analysed (Graf 1993, Graf & Burkhalter 2016). In the fluvioglacial gravels of the Deckenschotter there are limited facies indicators that could be associated with warm periods or interglacials. Deposition during interglacial periods is, for example, recorded in overbank sediments within the HDS (Hasli Formation, «Ebni silts») at Irchel (Bolliger et al. 1996, Thew et al. 2024) and above the bedrock in the TDS at Hungerbol (Graf 2009a). Additionally, relict palaeosols have been associated with warm periods (Graf 1993). Direct evidence of ice contact is also rare within the Deckenschotter. Prominent evidence can be found at Feusi above Oberweningen (FEU, Fig. 2-14), where glacialigenic diamicts are intercalated in HDS fluvioglacial gravel deposits (Grischott et al. 2020, Scheidt et al. 2023). This finding indicates that Early Pleistocene glaciers advanced into the foreland at least as far as Oberweningen, a position slightly outside of the ice extent of the Last Glacial Maximum (LGM, Fig. 2-30). Overall, in the Deckenschotter several gravel accumulation phases partly separated by warm period indicators can be distinguished that may be linked to ca. eight separate glaciations (Fig. 2-27; Graf 1993, Graf & Burkhalter 2016).

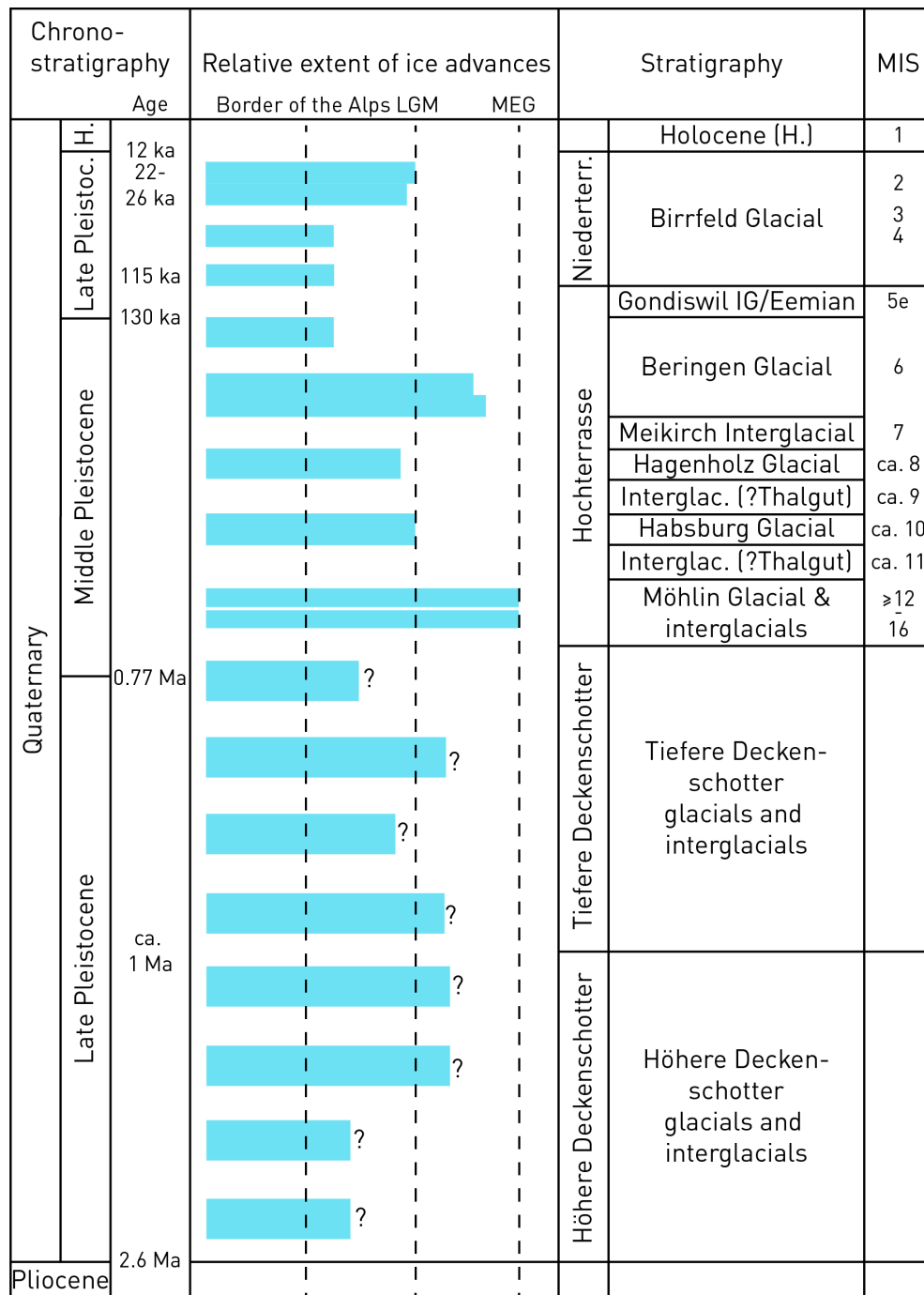


Fig. 2-27: Glaciation history of Northern Switzerland

Modified after Preusser et al. (2011), Graf & Burkhalter (2016) and Schlüchter et al. (2021).
H.: Holocene, IG/Interglac.: Interglacial, Pleistoc.: Pleistocene, Niederterr.: Niederterrasse,
MIS: Marine Isotope Stages.

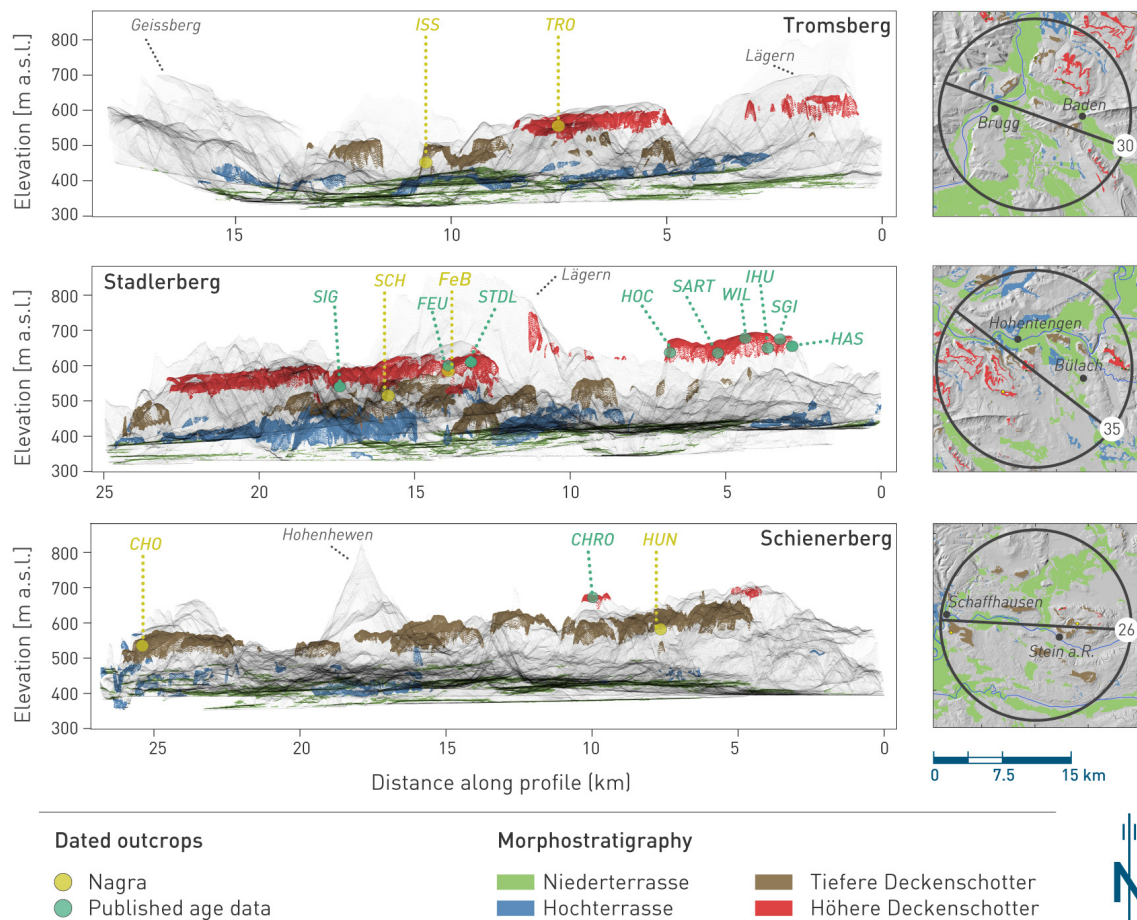


Fig. 2-28: GIS profile lines revealing terraces in the regions of the cosmogenic nuclide studies

Shown are selected profiles generated from a swath in a radial fashion and reflected by the black line in the right map views (Nagra, courtesy of Th. Pollhammer und E. Bros). The black background in the profiles corresponds to the topography in a swath along the black line of the map inset. Mapped elevations of HDS (red), TDS (brown), HT (blue) and NT (green) are colour-coded accordingly. Sampled outcrops for cosmogenic nuclides (Nagra, selected published data, Appendix A.1 and Claude et al. 2017, 2019) as indicated below.

Top: Lower Aare Valley, Middle: Central region between Aare and Irchel, Bottom: Eastern region, between W of Bodensee outlet and Schaffhausen region. Studied outcrops with cosmogenic age estimates, ISS: Iberig, TRO: Tromsberg, SIG: Siglistorf, SCH: Schwändihalden, FEU: Feusi (diamict), FeB: Feusi-Base, STDL: Stadlerberg, HOC: Irchel Hochwacht, SART: Irchel Schartenflue, WIL: Wilemer Irchel, IHU: Irchel Hütz, SGI: Irchel Steig, HAS: Irchel Hasli, CHO: Cholfirst, CHRO: Chroobach, HUN: Hungerbol.

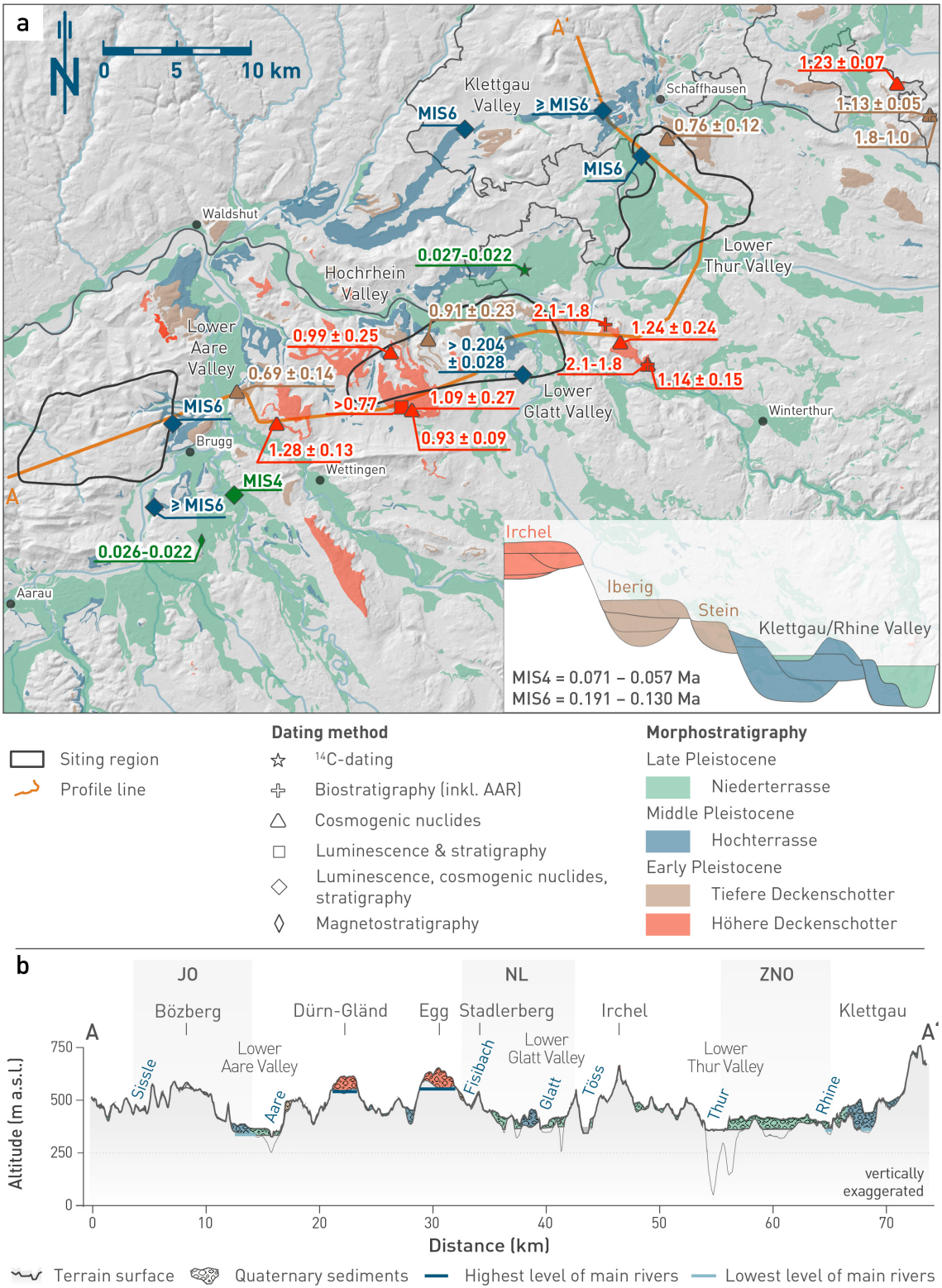


Fig. 2-29: Morphostratigraphic units of Northern Switzerland with selected age information

(a) Ages are colour-coded according to the morphostratigraphic unit of the respective deposits and listed with symbols according to dating method: biostratigraphy (Thew et al. 2024), AAR (Penkman et al. 2024), magnetostratigraphy (Scheidt et al. 2023), luminescence & stratigraphy (Mueller et al. 2024, Gegg et al. 2023, Kamleitner et al. 2023, Gegg 2021, Mueller et al. 2020, Gaar et al. 2019, Buechi et al. 2017, Lowick et al. 2015, Reber et al. 2014, Dehnert et al. 2012), cosmogenic nuclides (Dieleman et al. 2022a, Grischott et al. 2020, Knudsen et al. 2020). Inset shows sketch of morphostratigraphic relationship with complex cut-and-fill sequences within each unit (based on Graf 2009a). (b) Simplified profile (see (a) for location) through the siting regions and across several Deckenschotter deposits. Note highest and lowest fluvial levels in dark blue and light blue, respectively.

The age assignment and the morphological differentiation of the Deckenschotter into HDS and TDS has been controversially discussed. The base of these two units may locally differ by several decametres. However, the archive of the two units is relictic, and outcrops are often isolated and put into context with difficulty (e.g. Fig. 2-28). Independent dating of these Deckenschotter deposits is challenging (Section 2.4.1.3) and some ambiguities remain even after extensive efforts were made using different dating methods. For example, a prominent case of these ambiguities involves the Deckenschotter at Irchel (Fig. 2-14, Fig. 2-29; Thew et al. 2024). Small mammal and molluscan remains in fine-grained interglacial sediments of the Hasli Formation were assigned a biostratigraphic mammal stage MN 17, which corresponds to an age range between ca. 2.1 and 1.8 Ma (Fig. 2-23; Bolliger et al. 1996, Cuenca-Bescós 2015, Thew et al. 2024). However, cosmogenically derived age determinations of the gravels beneath this mammal-bearing site resulted in an age of ca. 1.14 ± 0.15 Ma (Fig. 2-19, recalculated age of Dieleman et al. 2022a using the P-PINI method, Nagra (courtesy of E. Bros), see Section 2.4.1.3).

The results of cosmogenic nuclide dating of the HDS and TDS led Dieleman et al. (2022a) to conclude, that cut-and-fill sequences accumulated in three stages during the Early Pleistocene at ca. 2.5 Ma, ca. 1.5 Ma, and ca. 1 Ma, requiring a rather stable local baselevel at that time. Consequently, the earlier differentiation of the Deckenschotter into HDS and TDS was questioned by Dieleman et al. (2022a). Yet, an evaluation and recalculation (using P-PINI) of the published ages of Akçar et al. (2017), Claude et al. (2019) and Dieleman et al. (2022a), Grischott et al. (2020) and Nagra (courtesy of E. Bros) suggest that the HDS fall into an age range between 1.5 and 0.7 Ma (Tromsberg, Irchel Schartenflue and Hasli, Chroobach, Feusi, Siglistorf), whereas most of the TDS deposits are slightly younger, defining an age range between 1.1 and 0.6 Ma (Schwändihalden, Cholfirst, Iberig) (Fig. 2-19). Note that some ages even fall into the Middle Pleistocene (Fig. 2-19). The TDS gravels at Hungerbol have an age of 1.13 ± 0.05 Ma, which is older than the other TDS dated with cosmogenic nuclides. Finally, it appears that the morphostratigraphic classification with generally older HDS and younger TDS can be maintained, but a classical, simple staircase-like lowering of the terrace levels is indeed too simple to describe the gravel deposits of Northern Switzerland. The formation of terraces and multitemporal incision and filling events may have resulted in the juxtaposition of HDS and TDS in cross-sectional channel profiles and locally the baselevel may not always have varied significantly between HDS and TDS.

Integrating all available age information and the uncertainties and caveats discussed above (Section 2.4.1.3), most HDS tend to fall into a time period between ca. 2.6 Ma and ca. 1.0 Ma (Fig. 2-27). Consequently, the TDS gravels tend to fall into a time period of ca. 1.0 Ma to ca. 0.7 Ma.

2.4.2.3 Middle Pleistocene

During the Middle Pleistocene (0.77 Ma – 130 ka), the glacial cycles were dominated by the 100-kyr periodicity, leading to the most extensive Alpine glaciations (Fig. 2-30). The onset of the Middle Pleistocene coincides with an increase in erosion and relief formation in the Alps (Muttoni et al. 2003, Valla et al. 2011). In addition, it is thought that erosion into the bedrock below the fluvial baselevel, creating major overdeepened valleys in the Swiss Plateau Molasse, has increased (Tomonaga et al. 2024). Alpine glaciers advanced and retreated in concert with the global climate cycles (Section 2.2, Fig. 2-26) and largely depended on (local) Alpine temperature and moisture availability. However, subsequent glaciations tended to modify or even obliterate the traces of earlier glaciations. Therefore, mainly the sediments of the most widespread and prominent or youngest glaciations are preserved in the North Alpine Foreland. The preserved sedimentary archives generated during these glaciations consist mostly of fluvioglacial gravel deposited in the forefield of the glaciers and downstream valleys, together forming the Hochterrasse ("High Terrace", Fig. 2-29). Interglacial sediments are rarely preserved. The earliest archives of the Middle Pleistocene might correspond to the TDS (Fig. 2-27).

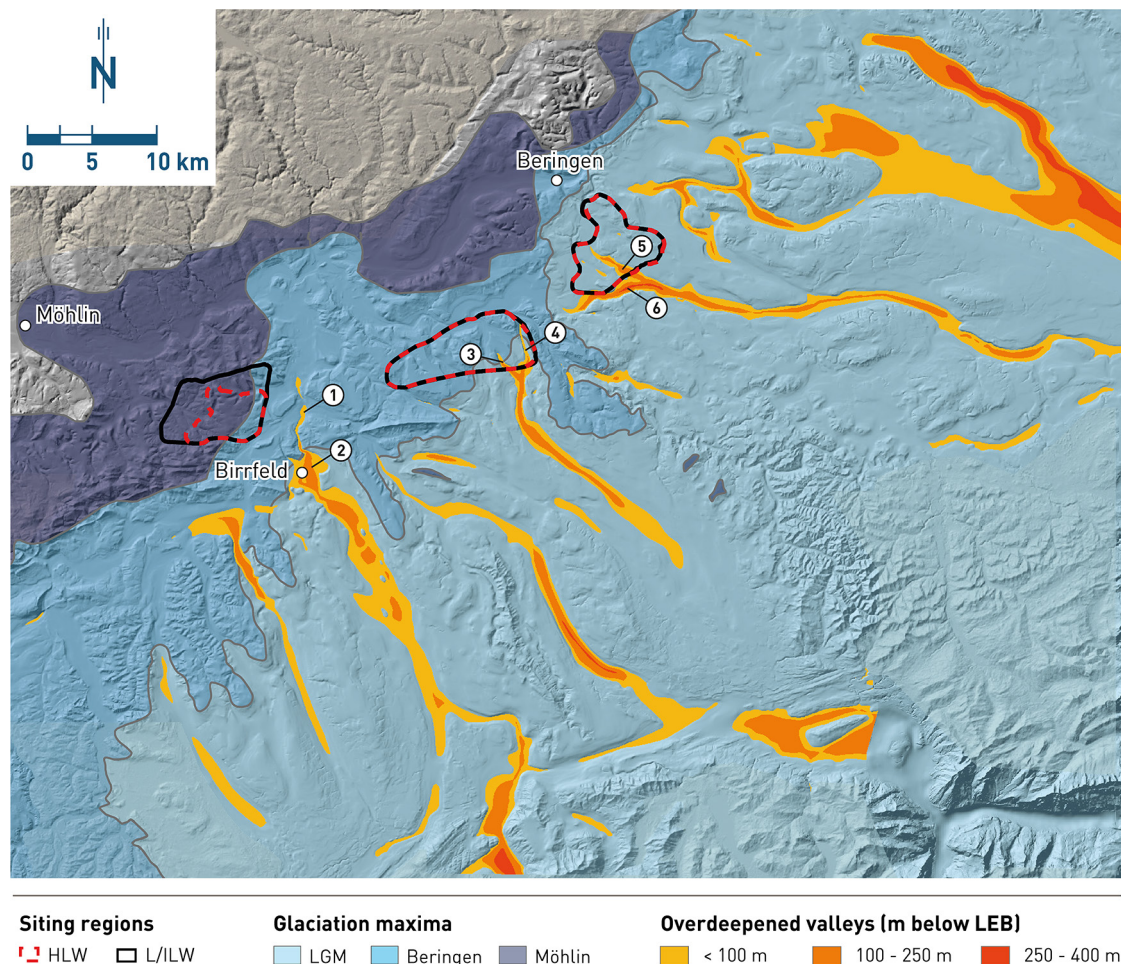


Fig. 2-30: Maximum ice extents during the Birrfeld (LGM), Beringen and Möhlin (MEG) Glacials and distribution of overdeepened valleys in Northern Switzerland

Labels in the map indicate the geographic reference locations after which the glacial extents are named. Additionally, major glacial troughs within and near the siting regions are indicated by numbers from west to east: (1) Gebenstorf – Stilli Trough, (2) Birrfeld Trough, (3) Strassberg Trough, (4) Bülach Trough, (5) Marthalen Trough, (6) Andelfingen Trough.

During the Möhlin Glacial, the Most Extensive Glaciation (MEG) of the Swiss Alpine Foreland was reached (Fig. 2-30; Graf 2009b). The glaciers crossed the Tabular Jura and reached as far as the village of Möhlin, the key region of this glaciation in Northern Switzerland. During the Möhlin Glacial, all three siting regions (JO, NL and ZNO) were covered by ice. Two ridges at Möhlin that were originally interpreted as moraine ridges were later identified as loess deposits (Gaar & Preusser 2017). However, ice-contact sediments of the Möhlin Glacial, the "Büntentill", are still interpreted to represent evidence for the MEG. Ages corresponding to the beginning of the Middle Pleistocene (Brunhes/Matuyama transition, 773 ka) and younger have been postulated for the MEG (Keller & Krayss 2010, Preusser et al. 2011). The MEG is a collective term for several glaciations that took place during that time (Schlächter et al. 2021). Based on an analysis of Quaternary aggradation of sediments in the Upper Rhine Graben, the Möhlin Glacial has been assigned to MIS 12 (478 – 424 ka) by Preusser et al. (2021). Isochron-burial dating using cosmogenic nuclides of the Büntentill also indicates a Middle Pleistocene age of 500 ± 100 ka (Dieleman et al. 2022b). However, reanalysing the dataset presented by the latter authors, Nørgaard et al. (2023) obtained a burial age of < 204 ka. Nevertheless, the strong erosional impact during the first 100-kyr cycle glaciations in the North Alpine Foreland is documented by the glacial overdeepening of the Strassberg Trough in the Lower Glatt Valley (Fig. 2-30, Fig. 2-31). In the Strassberg Trough, sedimentary evidence for several glacial advances is preserved (Fig. 2-32; Buechi et al. 2024). For the older part of the filling, a minimum age of 606 ± 122 ka was postulated applying the $^4\text{He}/\text{U-Th}$ dating method on porewaters (Fig. 2-17, Fig. 2-31; Tomonaga et al. 2024). The oldest preserved sediments of the Strassberg as well as the Marthalen Troughs have been tentatively correlated with the Möhlin Glacial (Fig. 2-33; Buechi et al. 2024). The Lower Glatt and Thur Valleys likely drained via the Neuhauserwald Channel and the Klettgau Valley to the Hochrhein Valley (Fig. 2-29; Buechi et al. 2024).

The following glaciation traced in sedimentary archives extended as far as the Habsburg area in the Lower Aare Valley and was assigned to the Habsburg Glacial and tentatively correlated with MIS 10 (374 – 337 ka, Fig. 2-27; Buechi et al. 2024). Alternatively, in light of the limited database this glaciation is older and could be part of the Möhlin Glacial or younger (MIS 8, 300 – 243 ka). During the Habsburg Glacial, the Birrfeld Trough was presumably (further) incised and parts of the Strassberg and Marthalen Troughs were reactivated, leading to the formation of inlaid glacial basins. In the Lower Aare, Glatt and Klettgau Valleys, significant gravel bodies accumulated (e.g. Habsburg Gravel, Rothenbrunnen Formation, Hardau Gravel, Fig. 2-33; Graf 2009b).

The next glaciation traced in sedimentary archives had a smaller extent and reached as far as Hagenholz near Kloten in the Lower Glatt Valley. The corresponding deposits were assigned to the Hagenholz Glacial and tentatively correlated with MIS 8 (300 – 243 ka, Fig. 2-27; Buechi et al. 2024). This glaciation left comparatively minor traces in the landscape in the distal North Alpine Foreland, but several gravel bodies in the Lower Aare, Glatt and Klettgau Valleys (e.g. Moos, Aathal and Geisslingen Gravels, Fig. 2-33). Other studies see the Hagenholz glaciation potentially as an early glaciation during the Beringen Glacial (early MIS 6, 191 – 130 ka; Keller & Krayss 2010, Preusser et al. 2011). In a more proximal position to the Alps, e.g. in the perialpine Aare Valley near Bern, there is evidence for formation of major overdeepenings during MIS 8 (Schwenk et al. 2022).

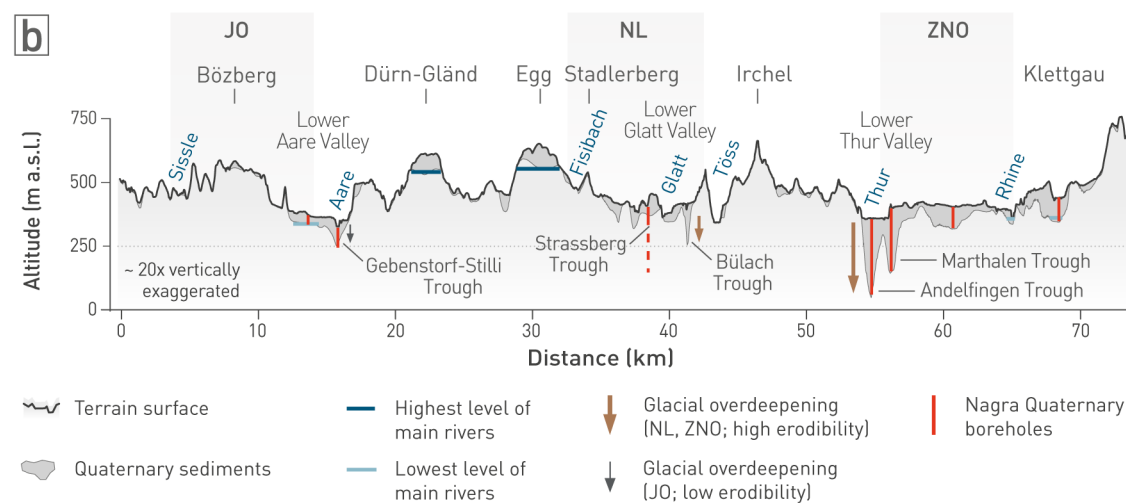
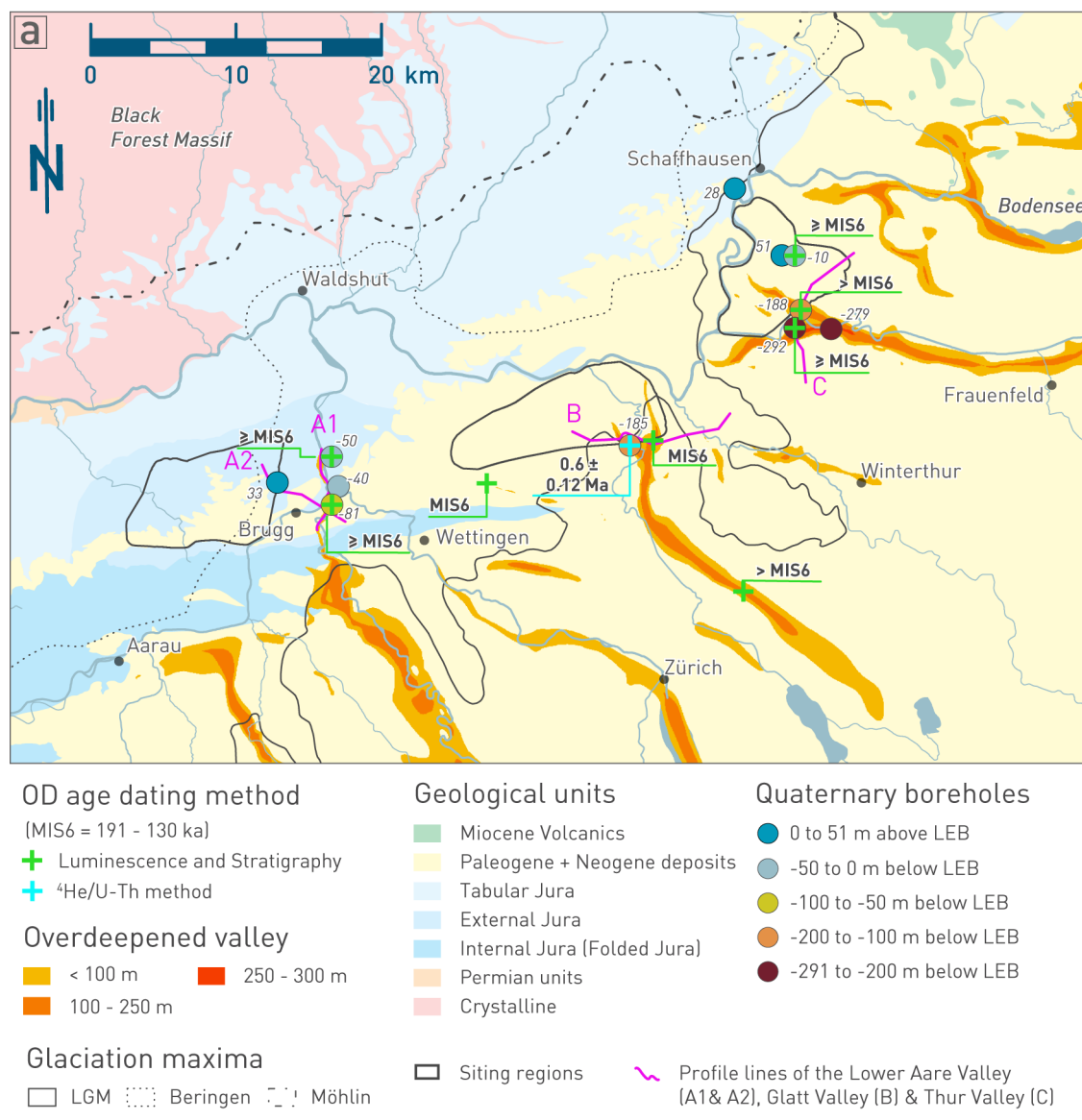


Fig. 2-31: Overview map (with accompanying profile) showing glacial overdeepenings and the underlying rock units

(a) Map showing overdeepening depth as depicted by sediment thickness below the local erosion base. The LGM (Bini et al. 2009) and the Beringen and Möhlin Glacial maximum ice extents (Preusser et al. 2011) are drawn as solid, dotted and dash dotted black lines, respectively. The locations of boreholes drilled during Nagra's Quaternary investigation programme are colour-coded and labelled with respect to their depth below the local erosion base. Selected ages reveal the timing of filling and thus postdate the deep glacial erosion event (Tomonaga et al. 2024, Mueller et al. 2024, Gegg 2021, Anselmetti et al. 2010, Dehnert et al. 2012, Buechi et al. 2018, 2024). Also shown are the locations of the four interpretative cross-sections (A) through the Lower Aare Valley, (B) across the Glatt Valley, and (C) across the Thur Valley which are shown in Fig. 3-35 (A1) and Fig. 2-32 (A2, B, C).

(b) Accompanying cross-section through the siting regions showing the present-day topography, the main palaeoriver levels (highest fluvial level in dark blue and lowest level in light blue, the latter corresponds to the LEB) and glacial overdeepenings. Quaternary boreholes drilled during Nagra's investigation programme are shown (Strassberg Trough borehole dashed because it is not located at the cross-section position, see Fig. 2-32b for comparison). Brown and black arrows indicate glacial overdeepening into (i) more easily erodible Molasse bedrock in NL and ZNO, and (ii) (amongst others) erosion-resistant Malm limestones east of the JO siting region, respectively. The location of the cross-section is the same as in Fig. 2-29.

Finally, towards the end of the Middle Pleistocene, the major glaciation after the MEG extended as far as Beringen in the Klettgau Valley and entered the Hochrhein Valley via the Lower Aare Valley (Fig. 2-30). During the Beringen Glacial, the ice advances reached only the easternmost part of JO, but covered NL and ZNO completely. The Beringen Glacial was assigned to MIS 6 (191 – 130 ka; Lowick et al. 2015). The glaciers advanced and retreated at least three times, leading to the deposition of several gravel bodies (e.g. Remigen Gravel in the Lower Aare Valley, Fig. 2-33). Unlike most of the analysed overdeepened troughs of the Swiss Plateau that were carved into the Cenozoic Molasse, the Gebenstorf – Stilli Trough in the Lower Aare Valley was incised into Mesozoic bedrock (Fig. 2-31, Fig. 2-32, Fig. 3-35; Gegg et al. 2021). In the Lower Glatt Valley, the overdeepening was focused on the Bülach Trough and the neighbouring Strassberg Trough was not reactivated (Buechi et al. 2018). In the Lower Thur Valley, the Andelfingen Trough was presumably incised and partly re-excavated in several phases (Buechi et al. 2024).

During the Middle Pleistocene, the fluvial baselevel was lowered by ca. 100 m relative to the base of the TDS at the Aare – Rhine confluence area (Fig. 2-26, Fig. 2-34; Heuberger et al. 2014). The baselevel lowering took place in several steps and was interrupted by phases of gravel and sand accumulation linked to glacial advances and retreats (Fig. 2-32, Fig. 2-33). During the Middle Pleistocene, the deepest level of bedrock incision was reached, the local erosion base (LEB) (Heuberger et al. 2014). The Rhine River flowed through the Klettgau Valley, but towards the end of the Beringen Glacial it was deflected southwards to its current course (Graf 2009b). During the Beringen Glacial, also the Aare River which originally flowed through the Habsburg – Rinikerfeld Palaeochannel, appears to have attained its current course (Gegg et al. 2023).

The major glacial overdeepenings near or within the three siting regions, i.e. Gebenstorf – Stilli, Strassberg, Bülach, Marthalen and Andelfingen Troughs, have a minimum age of MIS 6 or older (Fig. 2-31). They may have been eroded either during the Beringen Glacial (MIS 6) or the Möhlin Glacial (MIS 12 to 16). In the other glacials, glacial erosion probably led mostly to inlaid glacial basins (Buechi et al. 2024) or glaciers did not reach these distal positions of the glacial lobe.

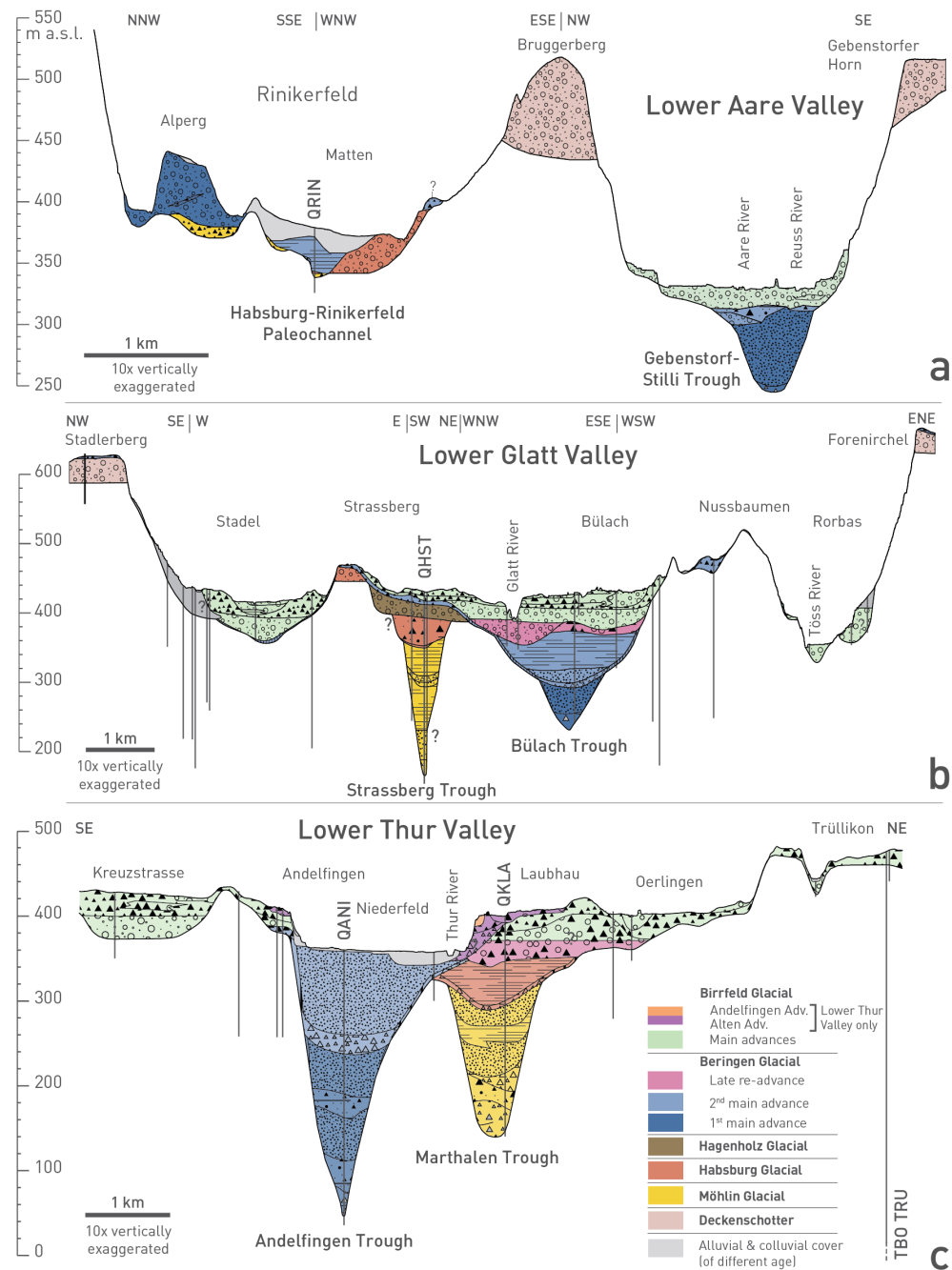


Fig. 2-32: Cross-sections highlighting the Quaternary deposits in the Lower Aare Valley, the Glatt Valley and the Thur Valley

The cross-sections represent an integrated interpretation of high-resolution 2D seismic, borehole and outcrop data (Nagra 2018, Buechi et al. 2024). The 10× vertical exaggerated profiles highlight forms, depths and filling of selected representative glacial overdeepenings. (a) Lower Aare Valley with the fluvioglacial Habsburg – Rinikerfeld Palaeochannel (QRIN) and overdeepened Gebenstorf – Stilli Trough (after Gegg et al. 2021). (b) Glatt Valley with the overdeepened Strassberg (QHST) and Bülach Troughs and (c) Lower Thur Valley with the overdeepened Andelfingen (QANI) and Marthalen (QKLA) Troughs (after Buechi et al. 2024). See Fig. 2-31 for location.

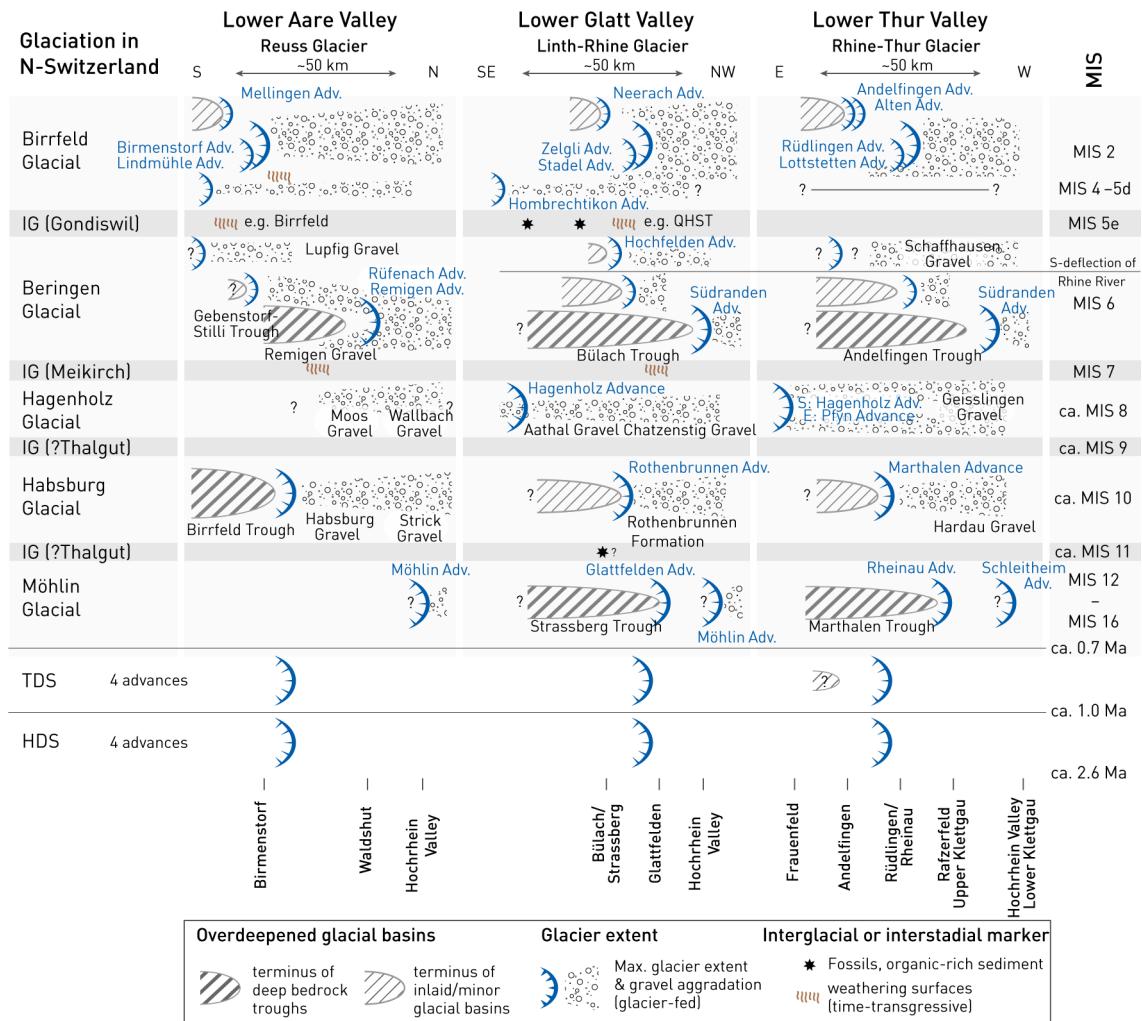


Fig. 2-33: Synthesis of landscape evolution in Northern Switzerland with focus on the formation of overdeepenings and gravel aggradation

IG: Interglacial, Adv.: Advance. Marine isotope stages (MIS) numbering from Lisiecki & Raymo (2005). Glacials and Interglacials after Graf (2009b) and Graf & Burkhalter (2016). Modified after Buechi et al. (2024).

There is evidence for a multiple reactivation of the same trough by several glaciations (e.g. Strassberg Trough, Fig. 2-32b or Middle Glattal Trough; Wyssling & Wyssling 1978, Haldimann 1978), indicating a constant positioning of trough formation over several glacials. This may indicate that also distal, meltwater-controlled overdeepenings can form relatively stable systems (Piotrowski 1994). However, in other cases, consecutive troughs can lie side by side (e.g. Strassberg and Bülach Troughs or Marthalen and Andelfingen Troughs, Fig. 2-32), indicating a temporal and lateral shift of the overdeepening especially towards their terminal zones. There are several potential controls of the shift of terminal overdeepenings (Buechi et al. 2024). Assuming similar valley morphology and glacier extents, the glacier-bed properties encountered during initial advance may play an essential role (Menzies 1989). Changes in the position of fluvial outlets by blockage and opening of new channels may also play an important part regarding subglacial drainage pathways (Cohen et al. 2023). In the Lower Thur Valley, the fluvial outlet

changed in the Middle Pleistocene from the Klettgau Valley to the recent Hochrhein Valley. This might have preconditioned the shift of the formation of overdeepening from the Marthalen Trough to the Andelfingen Trough (Fig. 2-32c).

It is conspicuous that almost exclusively the more easily erodible Molasse sediments were overdeepened (yellow areas in Fig. 2-31), pointing to a lithological influence on glacier efficiency (Section 3.5.1). An exception is the Lower Aare Valley, east of the Jura Ost siting region, where, amongst others, erosion-resistant Malm limestones were glacially overdeepened although less deeply than in the Molasse (Fig. 3-35; Gegg et al. 2020, 2021, Nagra 2021).

2.4.2.4 Late Pleistocene

In Northern Switzerland, the Late Pleistocene (130 – 12 ka) is represented by deposits of the last interglacial or Eemian/Gondiswil Interglacial, and the Birrfeld Glacial. Sediments of the Gondiswil Interglacial are rarely preserved and this period is mainly represented by weathered land surfaces (Fig. 2-33). During the last glacial period, the ice advanced and retreated several times. Up to five separate glacial advances have been identified, including a limited ice advance during MIS 4 (71 – 57 ka, Fig. 2-33). The Birrfeld Glacial is the best documented glacial period as it represents the youngest ice advance, and no later glaciations overprinted its sediments and landforms. The Rhine Glacier reached its maximum extent (LGM) between 26 and 22 ka ago (Fig. 2-29; Gaar et al. 2019, Kamleitner et al. 2023). Compared to the Beringen Glacial, the LGM ice extent was considerably reduced and reached as far as Birmenstorf/Birrfeld in the Lower Reuss Valley (Fig. 2-30). The maximum ice extent did not reach JO; glacial ice reached into the south-easternmost part of NL, and it covered ZNO completely.

During the Birrfeld Glacial, inlaid glacial basins into the previously carved sediment-filled troughs were eroded. A prominent example is the Andelfingen Trough, which might also be representative for other locations that experienced similar ice advances (Buechi et al. 2024). The LGM reached as far as Rüdlingen, near today's confluence of the Thur and Rhine Rivers. This most extensive as well as subsequent advances (Rüdlingen and Lottstetten Advances, Fig. 2-33) did not lead to significant deep glacial erosion towards the terminal zone. However, after a retreat of a few kilometres, two subsequent advances (Alten and Andelfingen Advances) caused the incision of an overdeepened inlaid basin and the formation of a reverse slope. This basin can be linked to ice-contact sediment, moraine ridges and gravel deposits. The internal position traced by moraine ridges of Alten and Andelfingen, corresponding to the 'Stein am Rhein' stadial, can be geomorphologically well correlated across the Rhine Glacier and neighbouring glacial lobes. A comparable erosion pattern after retreat from the maximum ice extent has also been proposed for Middle Pleistocene glaciations north of the Bodensee (Fiebig 2003, Ellwanger et al. 2011). This suggests greater efficiency of glacial erosion during periods of stability during an overall retreat of the glacier with more temperate conditions (Haeberli & Penz 1985, Cook & Swift 2012; Section 3.5.1).

With regard to drainage reorganisations, the capture of the Wutach River (Southern Germany) introduced in Section 2.4.1.4 represents the most prominent event of the Late Pleistocene period in Northern Switzerland and its surrounding area. It reflects a local continuation of the expansion of the Rhine catchment at the cost of the Danube catchment described for earlier periods. In this event, a former Danube tributary (Donau-Wutach) was captured by a headward eroding tributary of the Rhine River ca. 18 ka ago. This led to a significant local baselevel drop at the location of the river capture by rapid fluvial incision of up to 150 m within only ca. 6 ka (Einsele & Ricken 1995, Hebestreit 1995). As a result of this reorganisation, the Rhine basin gained an additional drainage area of ~ 400 km² at the expense of the Danube, with the lateral divide shift reaching up to ~ 15 km. The Wutach capture is described in further detail in the context of local erosion processes in Section 3.4.1.

2.4.2.5 Holocene

With the disappearance of the glaciers in the foreland around 16 – 17 ka ago, fluvial processes and hillslope processes dominated on a pre-sculpted, locally glacially overdeepened landscape. While the local baselevel (Basel, Hochrhein) controlled most of the incision of the Aare and Rhine Rivers, local streams and unstable hillslopes began adjusting to the new boundary conditions, which resulted in highly variable denudation rates (knickpoints, slope instabilities). These processes are intricately by the layered and partially folded stratigraphy of the Tabular and Folded Jura (Yanites et al. 2013, 2017, Ludwig 2018). Overall, spatially averaged landscape lowering rates have been determined to be in the order of 50 – 100 mm/kyr for low relief plateau surfaces, whereas hillslopes, connecting incising rivers and streams with the plateau surfaces, may have been characterised by rates of 100 – 200 mm/kyr (see Section 3.4.1). In extreme cases, these rates may have temporarily increased to up to 300 mm/kyr, but these are highly transient rates and illustrate the importance of recent incision processes in a deglaciated landscape (Section 3.4.1).

Today, the Aare and Rhine Rivers determine the local baselevel for the siting regions. Due to their rich history of drainage changes, it is important to understand whether similar regional drainage reorganisations could potentially occur in the future and the degree to which present-day rivers in Northern Switzerland continue to adjust to past changes in drainage basin configuration. The analysis of the digital elevation model using the geomorphometric metrics "X" of the main river system (e.g. Willett et al. 2014) may help to answer this question.

X is a proxy for elevation at any point along a river network under assumptions of the stream power model. Comparing X values of rivers on either side of a divide provides clues about potential instabilities and divide movements. The X anomaly quantifies the differences in the metric X on either side of a drainage divide. Sharp cross-divide contrasts thus highlight regions of potential future drainage reorganisation. Such contrasts can be seen at the Rhine-Danube divide north of the siting regions or at the Rhine-Danube divide southeast of Sargans (Fig. 2-34a). In turn, the drainage divide between the Rhone and Rhine Rivers across the Vosges mountains northwest of Basel seems to constitute a rather stable divide.

The closest distance between the Rhine and the Danube is found between the cities of Schaffhausen and Immendingen (labelled in Fig. 2-34a). To reach sea level from these locations is more than twice as long for the Danube than for the Rhine River. This, but also the geometry of their drainage basins, suggests that the Rhine catchment will grow at the expense of Danube catchment in the future to achieve a geometric equilibrium of the watershed (Fig. 2-34a).

Geometric imbalance is often associated with variable erosion (incision) rates, and river sections that gain watershed experience an increase in incision rates (e.g. Willett et al. 2014). However, the main pulse of increased incision from major past drainage reorganisations (Section 2.3) has largely already migrated through the Hochrhein Valley and thus the three siting regions and the river has regained an equilibrated longitudinal profile. This hypothesis is supported by the observation that the Hochrhein Valley has attained an overall along-river gradient that is comparable to the Rhine in the Upper Rhine Graben (Fig. 2-8a). Former river gradients need to be interpreted with care due to the possibility of tectonic tilting, but were still steeper than today, possibly showing a gradual adjustment since the emplacement of the HDS (Fig. 2-34b). The relatively low gradients today (see Fig. 2-8a for section-wise calculated gradients) also suggest that similar drainage changes in the future will most likely lead to minor if any increased incision. Any large events of fluvial incision in response to drainage reorganisations are therefore not to be expected within the next million years.

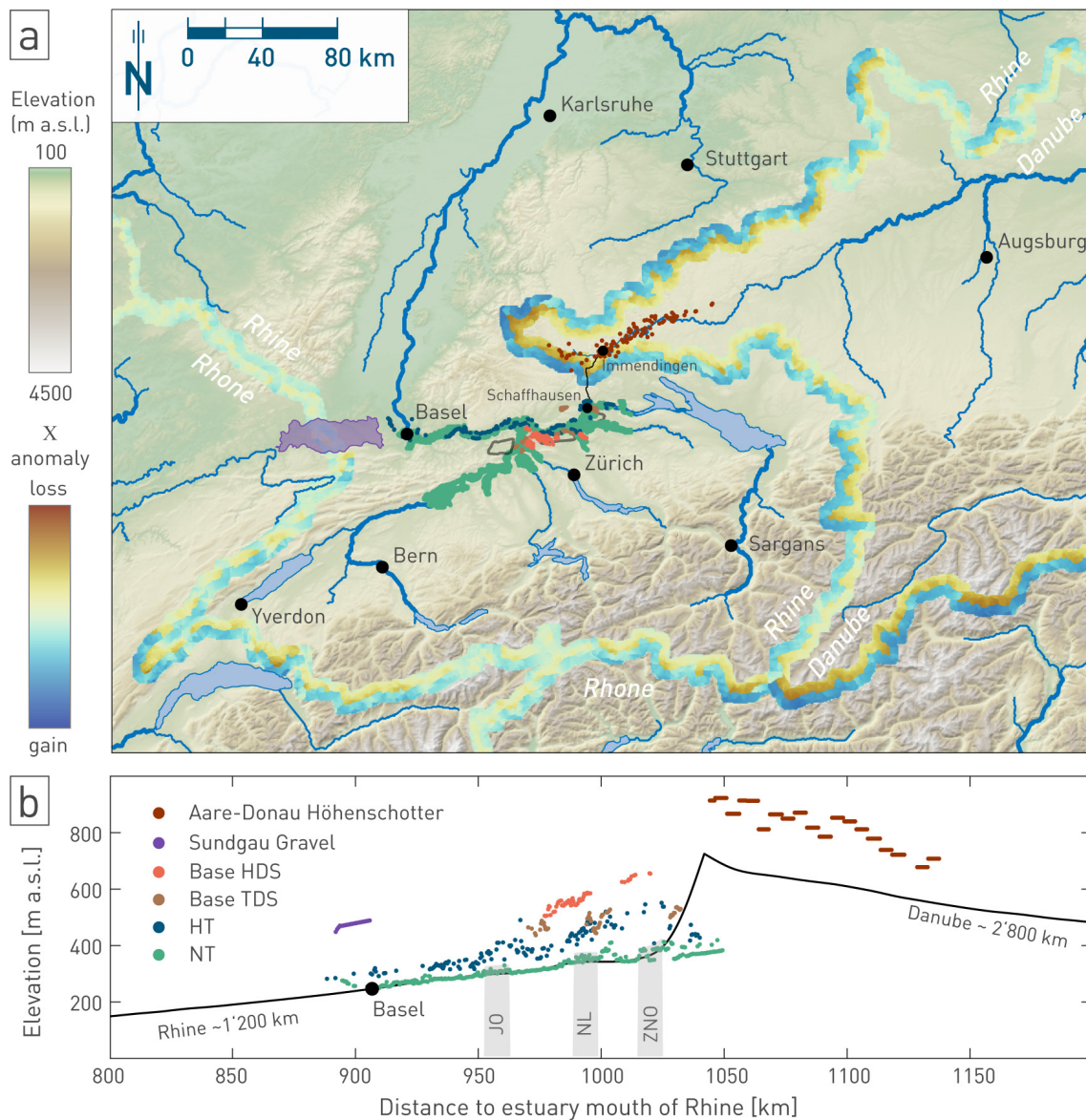


Fig. 2-34: Catchment competition between the Rhine and Danube watersheds and drainage reorganisation

Digital terrain model (a) with main river courses and their catchments. The drainage divide of the Rhine catchment is outlined by the direction of future catchment growth (parameter X, e.g. Willett et al. 2014). The X parameter is calculated for the river network and projected on the main upslope divides. Blue values indicate low X values and thus potential future gains in drainage area, whereas as red values indicate potential losses. Sharp cross-divide contrasts thus highlight regions of potential future drainage reorganisation. Conversely, similar anomaly values on both sides suggest stable divides. Shown are also locations of major fluvial to fluvioglacial deposits, see legend in (b) for colour-code. (b) Simplified comparison of partial longitudinal river profiles of the Rhine and Danube artificially connected along the black line in (a), between Schaffhausen and Immendingen. The distance along the Rhine to the North Sea is remarkably shorter than the distance of the Danube to the Black Sea, resulting in a higher altitude of the Danube system. The position of the fluvioglacial deposits associated with earlier drainages are projected to the profile. Also indicated is the approximate position of the siting regions (grey-coloured bars). The level of the Rhine is already clearly lowered at these positions.

2.5 Conclusions with respect to future erosion

- The main influence on the incision of the Hochrhein Valley are low rates of North Alpine Foreland rock uplift and low, potentially episodic, subsidence within the Upper Rhine Graben. Continuous Quaternary uplift of the Rhenish Massif and Eifel areas has caused a relative baselevel rise at these locations. This region provides an intermediate baselevel that is able to decouple downstream processes and reduce upstream migrating signals of downstream changes along the length of the longitudinal profile.
- Climate, and especially oscillating glaciations and ice retreats, have had a far-reaching influence on Quaternary erosion and landscape evolution in Northern Switzerland. Advances of Alpine glaciers controlled the extent and distribution of glacial erosion and deposition. Extensive foreland glaciations resulted in the carving of overdeepened basins and deposition of glacial sediments, which in turn promoted changes in drainage pathways, sediment routing and deposition. Glaciation of the Alps and their foreland leads to peaks in sediment yield.
- During the Pliocene to Quaternary, the Rhine fluvial system was reorganised and increased its catchment area at the expense of the Danube system. Importantly, the main pulse of increased incision from major past drainage reorganisations has largely migrated through the Hochrhein area and past the range of the siting regions. The lengths of the Danube and Rhine Rivers as well as the geometry of their drainage basins suggests that the Rhine catchment will grow at the expense of Danube catchment in the future to achieve a geometric equilibrium of the watershed. Incision processes of similar magnitude as during the Pliocene to Quaternary following hypothetical drainage reorganisation in the region are not to be expected for the next one million years.
- The landscape of the North Alpine Foreland was heavily impacted by many glacier advances and retreats in response to global glacial-interglacial cycles. The first glaciations that reached the Alpine Foreland in Northern Switzerland seem to have influenced the region during the Early Pleistocene. Although relictic and disparate in nature, various sedimentary archives and geomorphic indicators suggest that the Middle to Late Pleistocene was characterised by several extensive glaciations. In particular the last glaciation (Birrfeld Glacial), the penultimate glaciation (Beringen Glacial), and to a lesser extent the Most Extensive Glaciation (MEG, Möhlin Glacial) can be distinguished. The distinct extents of these large foreland glaciations influenced the siting regions differently; this fact combined with the relatively good knowledge about these glacials allowed them to be considered as key glacials for the assessment of future erosion processes.
- Some discrepancies and ambiguities with regard to the age and stratigraphic interpretation of the Quaternary units and events in the North Alpine Foreland remain. The interpretation of specific stratigraphic units needs a careful spatial, chronological and genetic evaluation. Overall incision and sediment accumulation represent a combination of (local and regional) tectonic processes, climate change (orbitally controlled glacial cycles) and glacial as well as fluvial dynamics such as river captures.
- Three major steps in fluvial incision can be distinguished. The first step of ca. 300 – 500 m is linked to the river reorganisations from the Aare – Donau via the Aare – Doubs to the Aare – Rhine River System. The youngest sediments of this fluvial network evolution are recorded by the gravels of the HDS. The second step, involving up to ca. 80 m, can be traced by the difference between the base of the HDS and the base of the TDS. The third step of up to ca. 100 m incision concerns the erosion between the base of the TDS down to the current local erosion base.

- Rates of incision since the Early Pleistocene may have been spatially and locally variable and were in the order of 0.1 to approximately 0.3 mm/yr (based on revised incision rates applying P-PINI derived age constraints and biostratigraphy, see Section 2.1 and Appendix A.2. The low denudation rates are largely consistent with incision rates, although averaging across a much shorter time period.
- Multiple glaciations in Northern Switzerland also caused rather localised deep erosion into the bedrock and the formation of overdeepened troughs. The majority of overdeepenings occurred in substrate characterised by Molasse rocks. Some of the troughs, such as the Strassberg Trough, record multiphase filling processes, whereas others are dominated by one phase, such as the Gebenstorf – Stilli Trough. There is evidence for a multiple reactivation of the same troughs by several glaciations (e.g. Strassberg Trough). Alternatively, these troughs may be adjacent to each other (e.g. Strassberg and Bülach Troughs), indicating a temporal and lateral shift of the overdeepening especially towards the terminal zone in the distal Alpine Foreland.
- Uncertainties in the reconstruction of the Quaternary landscape history depend heavily on the relictic and variable nature, as well as the correct identification and dating of its archive. The comprehensive scientific field investigation programme provided a better knowledge base to evaluate the Quaternary evolution of Northern Switzerland. The remaining uncertainties cannot be further reduced in due time and are therefore sufficiently accounted for in a systematic assessment of future erosion processes.

3 Assessing future erosion

After outlining the main evidence for Quaternary landscape evolution in Chapter 2, this chapter focuses on future erosion, applying the new model cascade. Firstly, the different site-specific constraints of Jura Ost, Nördlich Lägern and Zürich Nordost with respect to the three main erosion processes (fluvial incision, evolution of local topography, and deep glacial erosion) are summarised (Section 3.1). Subsequently, the evaluation outputs for forecasting erosion in the future and for assessing the remaining overburden thickness are introduced together with a brief description of the uncertainty management used in this new approach (Section 3.2). According to the relevant processes, Chapter 3 is further subdivided into Section 3.3 (fluvial incision), Section 3.4 (evolution of local topography) and Section 3.5 (deep glacial erosion). Each section starts with a brief introduction to processes and drivers, which form the basis for the workflow considering the assessment of future erosion. Subsequently, the rationale and main argumentation for scenario or model parameters are provided, followed by results and discussion.

3.1 Site-specific constraints with respect to erosion assessment

The distance of the three siting regions from the ice margins and the position within different tectonic units (e.g. position within or outside the Molasse Basin versus External Jura) has resulted in marked differences between the present-day landscapes of the three siting regions. Differences include, for instance, outcropping lithologies, the influence of principal erosion processes and repository depth. To evaluate future erosion potential at these sites, it is necessary to first assess and compare the physio-geographic conditions at the respective repository sites. In the following, the three landscape compartments and their main differences between the siting regions are briefly described and illustrated.

The main reference horizon, i.e. the first landscape compartment, for the assessment of future erosion is the *local erosion base* (LEB) – the locally deepest fluvial incision level into bedrock (Fig. 3-1). The LEB is a virtual surface that spans between the thalwegs of the main river channels of the Rhine, Aare, Limmat, and Reuss and decreases in elevation towards WNW. Some of these thalwegs are filled with sediments so that the bedrock interface defines the thalweg depth, and the LEB is lower than the present-day riverbed (see Pietsch & Jordan 2014 for more details⁴). As all other smaller streams and rivers draining Northern Switzerland are tributaries to these main rivers, no other river can significantly incise below this local baselevel. The local erosion base (LEB) separates the rocks (bedrock and unconsolidated sediments) above (here called *local topography*, Fig. 3-1c) from the sediments that fill the glacially overdeepened basins (*glacial overdeepenings*, Fig. 3-1d) below.

Fig. 3-2 shows the fractions of the different landscape elements (compartments) with respect to the depth of the Top Opalinus Clay (i.e. the repository host rock) for each siting region. These fractions are shown for (1) the entire HLW siting region, (2) the provisional disposal area (reflects the ~ 1 km² outline of the repository project for high-level waste; for location see Fig. 3-24) and (3) the provisional disposal area including spatial reserves⁵ (see Section 4.1.3 and Nagra 2024a for further information; location shown in Fig. 4-3).

⁴ The present adaptations on the bedrock model did not change the elevations of this reference.

⁵ All three siting regions have potential disposal zones in the Opalinus Clay with large lateral extents that far exceed the spatial requirements according to the current repository design, i.e. the provisional disposal area. Key factors determining the lateral extent of the potential disposal zone are (1) the avoidance of regional tectonic elements and larger tectonic faults and (2) the choice of an overall suitable minimum depth of the host rock with respect to future erosion. The spatial reserves are part of the potential disposal zones and include a multitude (i.e. for HLW 6 km²) of the lateral spatial requirement for the provisional disposal area.

The emplacement depth in the provisional disposal area (approximately 50 m below the Top Opalinus Clay; see Fig. 3-2, brown boxes) with respect to the local erosion base varies and is deepest in NL (~ 800 m at centre, > 760 m at most shallow location), followed by ZNO (~ 630 m at centre, ~ 600 m at most shallow location) and considerably less deep in JO (~ 300 m at centre, ~ 270 m at most shallow location).

The local topography in the HLW siting region decreases from the JO site in the west via NL to ZNO (Fig. 3-2). Accordingly, the local topography adds a significant overburden thickness of ~ 200 m at the JO site. Including local topography, the minimum overburden thickness above the repository in the provisional disposal area amounts to > 400 m in JO, > 750 m in NL, and > 650 m in ZNO respectively⁶.

The siting regions NL and ZNO feature glacial overdeepenings that are significantly deeper than those in the Lower Aare Valley east of Jura Ost (there is no glacial overdeepening within the JO siting region itself, see Fig. 3-2). This spatial distribution of glacial overdeepenings can be attributed to two factors. First, the glacial troughs in NL and ZNO were cut into erodible sedimentary rocks of the Molasse of Northern Switzerland. In contrast, the hard Mesozoic limestone layers in the Lower Aare Valley are likely less erodible and may have been associated with lower rates of glacial and fluvio-glacial incision. More erosion-resistant rocks such as these hard limestones can also be found at greater depths in NL and ZNO, a circumstance that will be considered for evaluation of the future depth of deep glacial erosion. Second, Pleistocene glaciations impacted the three siting regions to varying degrees (see Fig. 3-1d and Section 2.4.2). Only the most extensive glaciations reached the JO siting region, and the thickness of glacial ice was presumably lower than in NL and ZNO. During the last ice age, the ice cover in ZNO was thicker and more extensive than in NL. Of the respective HLW siting regions, about 8.5% is overdeepened (> 25 m below LEB) in ZNO and ~ 3.5% in NL (Tab. 3-9).

In summary, considering the depth dependence of the barrier properties, the initial depth of the repository, expected future rock uplift, erosion rates, and deep glacial erosion, and thus climate trends of the past and future, are important input parameters for site comparison. As outlined above and visualised in Fig. 3-1 and Fig. 3-2, the Jura Ost siting region has different original conditions from Nördlich Lägern and Zürich Nordost, which share more similar conditions. To objectively compare the three siting regions, the landscape compartmentalisation (local erosion base, local topography and glacial overdeepenings) was transformed into a model cascade (see Section 3.2) and assessed in a more holistic approach.

⁶ Note that these numbers describe the rounded minimum overburden thickness within the footprint of the provisional disposal area, found at a specific (point) location (i.e. by searching for the minimum vertical distance between the two geometric surfaces of the repository and the present-day terrain surface). This information cannot be directly extracted from Fig. 3-2, where each boxplot reflects an aggregated result from the entire footprint of the provisional disposal area. However, comparing specific percentiles (e.g. the median values from boxplots "Terrain above LEB" and "Top Opalinus Clay below LEB") and accounting for the vertical difference between the repository and Top Opalinus Clay (approximately 50 m) yields at least a similar order between the three sites (e.g. ~400 m in JO, ~800 m in NL, ~600 m in ZNO).

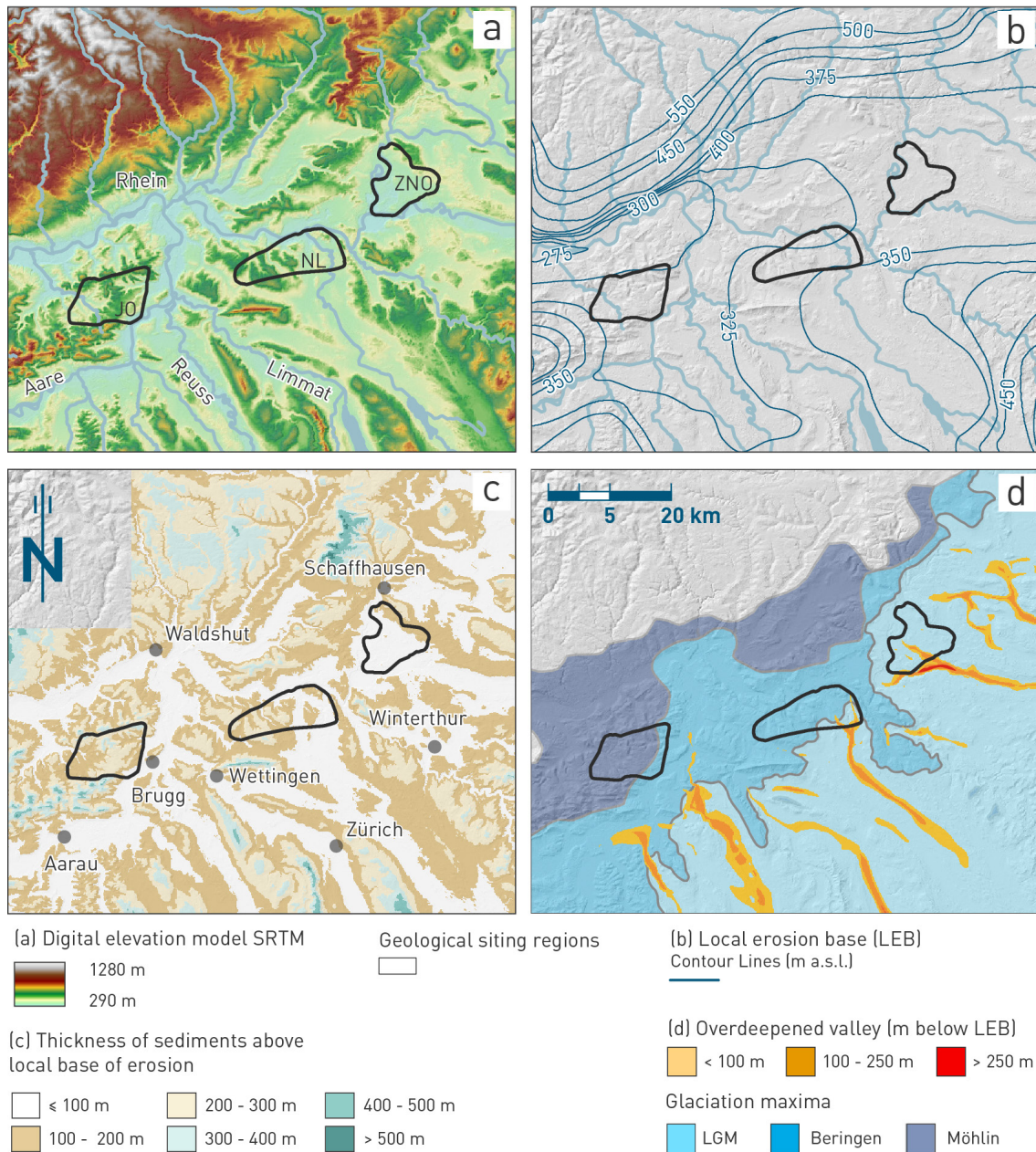


Fig. 3-1: Maps showing the main landscape compartments in the vicinity of the siting regions

(a) Digital elevation (SRTM) on a hillshade model showing the relief distribution around the siting regions. (b) Contours of local erosion base (LEB) as derived by Pietsch & Jordan (2014). (c) Colour-coded thickness of local topography, i.e. the sediment above the LEB. (d) Ice extents of the three Quaternary key glacials (blue colours) and distribution and depth of glacial overdeepenings (orange colours). The siting regions are plotted in each subfigure for comparison.

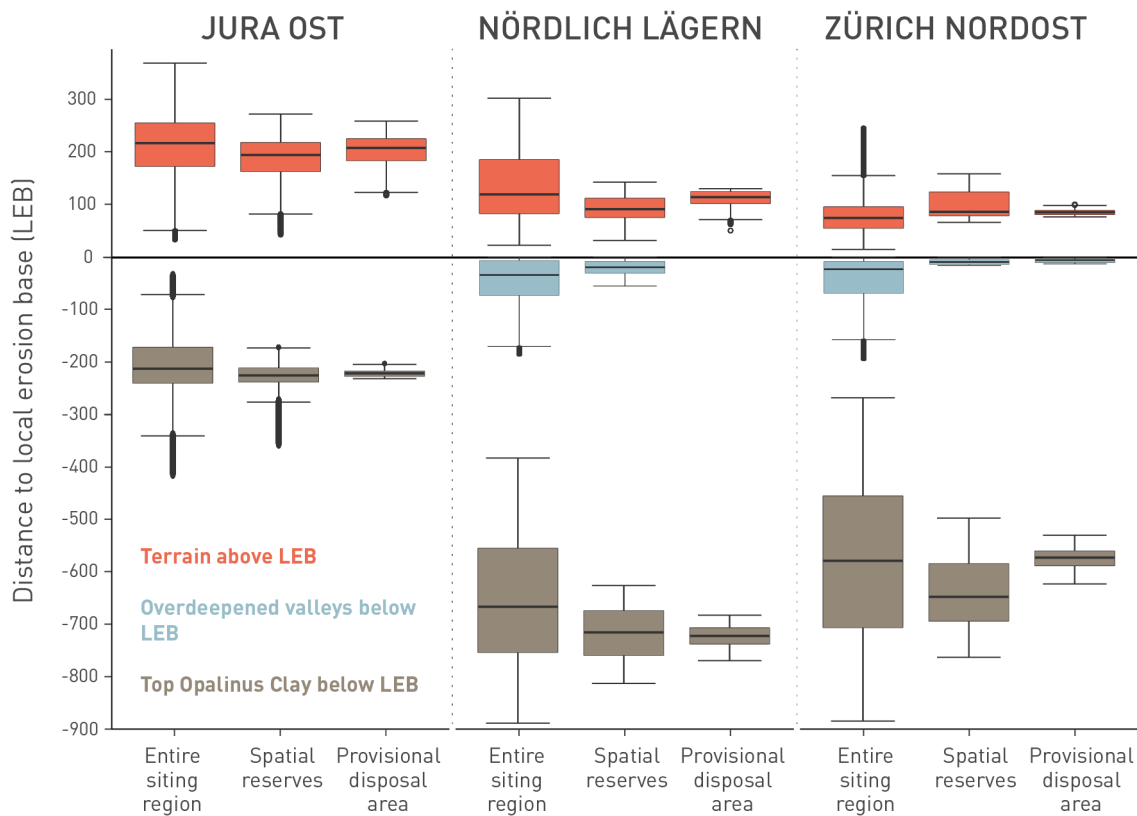


Fig. 3-2: Site-specific constraints of the landscape compartments within the siting regions

Synthetic "depth profiles" showing boxplot statistics for the main compartments with respect to the local erosion base (0 m line) and for the entire HLW siting region, the spatial reserves, and the provisional disposal areas. Boxes entail the interquartile range between 25th – 75th quantiles, the whiskers the extreme values, as derived from multiplying the interquartile by 1.5. Values outside the whiskers are considered statistical outliers. The siting regions share different portions of the landscape compartments. The depth of the provisional disposal area with respect to the local erosion base (LEB) is deepest in NL, and considerably less deep in JO. On the other hand, JO is covered by significant terrain above the LEB (local topography) and this site is less affected by glaciations. JO has no overdeepened area within the siting region.

For the calculation of the difference raster LEB – Top Opalinus Clay, raster datasets of the LEB (Pietsch & Jordan 2014) and the horizon Top Opalinus Clay (Nagra 2014a) were used. In comparison to the local model Top Opalinus Clay of 2014 (Nagra 2014a), the more recent 3D horizon model of the sites (Nagra 2024d), covering the zones of the spatial reserves and the provisional disposal areas, shows the following average differences: 18 m deeper in JO, 12 m deeper in NL, 6 m shallower in ZNO. Note that the assessment of future erosion presented in this chapter is based on the recent 3D horizon model documented in Nagra (2024d).

3.2 General approach to assessing future erosion and associated outputs

Climatic, tectonic and lithological factors and processes will determine the rates at which fluvial incision is able to remove unconsolidated sediment and rock overlying the Opalinus Clay in the future. In addition, as has been documented for Pleistocene glacial advances from the Alps, deep (sub)glacial erosion has the potential to erode up to a few hundred metres below the LEB. The timing of the formation of such overdeepenings as well as their spatial occurrence is critical to the assessment of the long-term safety of a deep repository and depends on the future climatic evolution and on the potential for new incised breakthrough channels to be formed at the sites. On long timescales involving glacial cycles, these processes interact, potentially with complex feedback mechanisms. These and associated uncertainties in the future evolution of the landscape in Northern Switzerland need to be addressed to evaluate potential future scenarios.

The new model cascade developed for this study allows a combination of numerical simulations and probabilistic calculations to incorporate uncertainties with regard to the different processes, their temporal and spatial dimensions and their rates.

The erosion model cascade combines multiple models in a hierarchical way. The cascade integrates models focused on different erosion processes that branch and merge to later provide a final assessment of the erosion while propagating uncertainties (Fig. 3-3). At the root of the modelling chain is the fluvial incision model (Model A, see Section 3.3.2). It simulates the development of the Hochrhein and the evolution of the local erosion base, thus providing information needed for an assessment of the boundary conditions for Model B which simulates the evolution of the local topography (Section 3.4) and for the calculation chain used in Model C (deep glacial erosion, Section 3.5). The separation into three different submodels and subsequent combination allow future erosion to be assessed and compared in the siting regions while considering and objectively quantifying uncertainties. Two main outputs are provided for comparison between the siting regions and subsequent safety assessments; these are the remaining vertical cover (overburden thickness, Section 3.2.1) and estimated excavation probabilities (Section 3.2.2).

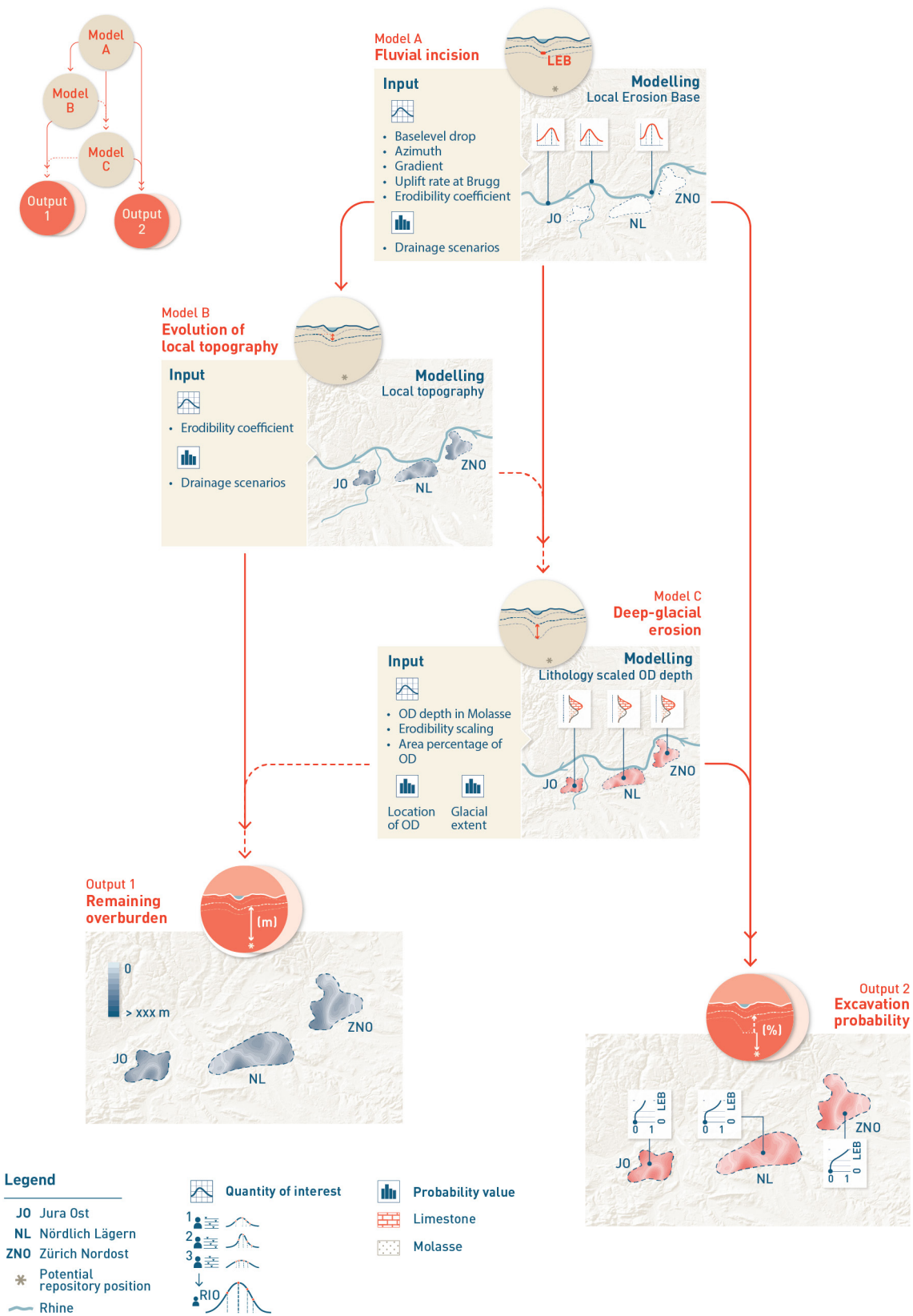


Fig. 3-3: Simplified flowchart for calculating the remaining overburden thickness and probability of excavation at time T_x

The three principal erosion processes that act in Northern Switzerland and that resulted in the previously described landscape compartments are logically combined in the model cascade. Because the local erosion base vertically separates the landscape parts above (local topography) and below (glacially overdeepened valleys), its future evolution is the first building block (Model A) of the full approach. The output provides input to Models B and C. Main outputs are remaining vertical cover (also remaining overburden thickness) and excavation probability (see Sections 3.2.1 and 3.2.2)

The "model cards" visualise the respective procedure, highlighting the input parameters on the left side and show specific characteristics in the associated maps: The results of model A for instance comprise probability density functions of future fluvial incision and are calculated with respect to the nearby downstream positions. Model B produces maps of future topography for parts of the siting regions (only for selected parameter combinations, elsewhere the procedure is similar to C), while the calculations in Model C are performed only for selected locations (e.g. the position of the provisional disposal area). The round inset pictures refer to the respective cross-section and highlight the changes in erosional depth. Two main outputs are generated: remaining overburden and excavation probability. The pale circles in the background of the output cards point to the fact that both outputs can be generated at each end of the model chain, respectively.

3.2.1 Remaining overburden thickness

The model cascade provides an assessment of the remaining overburden thickness (with respect to decompaction effects) (Chapter 1) considering changes in the local erosion base, local topography, and glacial overdeepenings. If the local erosion base is lowered at the expected rates, sufficient overburden thickness remains after one million years. In a less likely case of stronger than expected fluvial incision, local topography might still protect the repository from decompaction effects. The target performance measure is the total remaining overburden thickness above the deep geological repository at the relevant observation times of 100'000 years (for low- and intermediate-level waste, L/ILW), and 1'000'000 years (high-level waste, HLW). A probability distribution thus combines both baselevel incision and evolution of local topography⁷.

In a second step, the probability of glacial overdeepening can be similarly incorporated. For this assessment, it was decided to exclude deep glacial erosion for the 100'000-year time period. This decision was based on two arguments: firstly, modelling of future glaciations (Section 3.5.3.5, Fig. 3-46) suggests a delay of future large glaciations, especially with respect to large foreland glaciation, for scenarios incorporating anthropogenic CO₂. Although a natural scenario cannot be excluded, the extent may not be comparable to that of LGM. Secondly, in anticipation of the results with respect to the one-million-year time period, it was concluded that the outcome, i.e. remaining overburden thickness or excavation probabilities, of the shorter time period cannot be worse than that of the long time period. However, effects of such an early glaciation were considered for instance in the short-term uplift considerations.

3.2.2 Excavation probability

Deep glacial erosion is considered to be the only process to significantly erode below the local erosion base. The process produces glacial overdeepenings (Alpine periglacial lakes, e.g. Bodensee), which are loaded with ice during erosion. The ice will later be replaced by melt water and the lake basin eventually filled with sediment. Drilling campaigns into such filled overdeepenings

⁷ This step was only performed for JO (see Section 3.4.2).

show that they can be in the order of several hundred metres deep (e.g. Amschwand et al. 2020, Schuster et al. 2020, Reichenwallner et al. 2021, Buechi et al. 2022). It is expected that the filling will replace the eroded bedrock and thus diminish decompaction effects. However, to evaluate the probability of reaching the repository or even excavating it, the considered scenarios incorporate the lowering of the local erosion base, the denudation of local topography, and additional deep glacial erosion during future piedmont glaciations. The relevant quantity to be evaluated in this context is the probability of glacial and non-glacial excavation of a part of the repository at the relevant observation times.

3.2.3 Uncertainty management and expert elicitation

Managing knowledge and uncertainties in the context of assessing the long-term safety of a radioactive waste repository is challenging, in particular because of the long time span to be considered. In this section, the main reasoning and procedure used are highlighted. More details, especially on types of uncertainties and expert elicitation can be found in the App. B.

Fig. 3-4 illustrates the knowledge and uncertainty management. The first and foremost approach to assessing future erosion including uncertainties must be a rigorous evaluation of the past and present-day environmental conditions and knowledge concerning process rates. As illustrated in the introduction, Nagra has carried out a comprehensive scientific programme during the past years; this has resulted in a good database allowing characterisation of the safety-relevant aspects at the different sites (Fig. 3-4a, dark beige circles). The essence of this programme was presented in the proceeding chapter. Despite this extensive database and good process understanding on different timescales, uncertainties in the future behaviour of the system will persist and it has to be acknowledged that, in the context of the evaluation of landscape evolution within the future one million years, some uncertainties are not further reducible in due time (Fig. 3-4a, light beige circles).

In this context is important to caution that even a perfect image of the past and present-day erosion processes might not fully capture what could potentially induce future changes in form and rate. This uncertainty thus needs to be accounted for as well (Fig. 3-4a, light beige circles). Overall, data density is indeed high, but it is still sparse given the long timescales of the projected repository safety. In addition, the available data may be considered partly inconclusive or might be interpreted differently by different groups of experts. Finally, the state-of-the-art methods used to measure, evaluate and interpret erosion rates carry their own uncertainties and may not always be ideally suited for the age range or composition of the deposits to be analysed (Section 2.4.1). A prominent example of this dilemma is represented by the older Deckenschotter deposits or sedimentary sequences that were deposited in the glacial overdeepenings. Here, the geochronological and stratigraphic data and limited analytical possibilities to determine the true age of these units force interpretations that are partially ambiguous (Section 2.4.1.3).

Consequently, to enable the most independent judgement possible, it is important to consider the uncertainties in a principled, reproducible and transparent way. A probabilistic framework allows uncertainties to be included in a principled way and at the same time to "rank" them. However, considering the complexity of predicting the remaining overburden thickness above the repository after one million years, simplifying assumptions need to be made. The potential consequences of such simplifying assumptions were assessed at least qualitatively; whenever feasible, a quantitative assessment was also conducted. For the problem at hand, a hybrid approach with as much probabilistic treatment as possible in combination with individual scenarios (e.g. Sections 3.3.3.4, 3.4.3.4, and 3.5.3.3) was developed. Importantly, these scenarios were chosen to cover what was considered a critical range of possibilities. The whole process is guided by a subsequent sensitivity analysis to identify critical parameters and processes in the model cascade (Section 5.1).

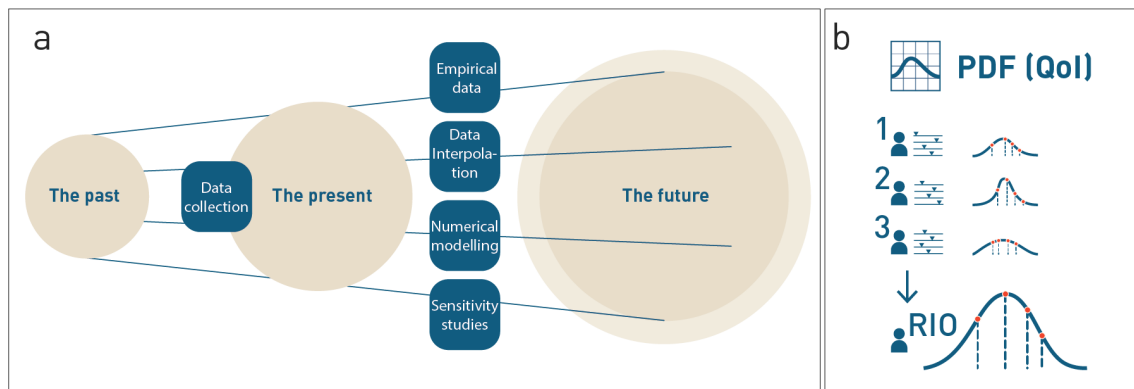


Fig. 3-4: Illustration of knowledge and uncertainty management for the erosion assessment

(a) The illustration emphasises how the knowledge about past and present conditions derived from collection and interpretation of empirical data, numerical modelling or sensitivity studies is used to evaluate potential future behaviour of the erosion system. The knowledge gained provides the majority of information (highlighted by the darker beige circle). However, remaining uncertainties (highlighted by the light beige circle) that are not easy to decipher need to be incorporated into the long-term erosion assessment. (b) Expert knowledge is incorporated in a two-step structured process from individual to group elicitations (see main text). PDF: Probability density function; QoI: Quantity of interest; RIO: Rational impartial observer.

As the model cascade combines numerical modelling, parameter uncertainties (see Appendix B.1) need to be incorporated (more details on the model are provided in the respective Sections 3.3.2, 3.4.2, and 3.5.2). In particular, epistemic components of the parameter uncertainty require special consideration. Consequently, expert knowledge (see Appendix B.2 for more detail on expert elicitation, including definitions and terminology) was included in a systematic series of workshops. The uncertainty elicitations for the parameters relevant for the erosion assessment at Nagra followed the SHELF protocol (Gosling 2018), which is a structured group elicitation framework for carrying out the elicitation of subjective probability distributions for quantities of interest (QoI) that are subject to uncertainties (see Appendix B.1). Prior to the elicitation workshops, all the experts underwent training in the fundamentals of probabilistic reasoning and the practicalities related to the elicitation of subjective probabilities (Appendix B.2).

SHELF is a two-step process (Fig. 3-4b) in which a facilitator first guides a group of experts (individually) through the elicitation of a set of quantiles for the desired distribution (red marks on distributions in Fig. 3-4b, Fig. 3-5a) using an explicit protocol. In the second step, the group acts as a single entity, deriving a single group distribution, which does not try to reach a consensus, but rather an integrated view of the uncertain quantity. To achieve this goal, the complete set of arguments raised during the group discussions is considered, as it would be seen through the eyes of a rational impartial observer (RIO). These RIO distributions thus comprise the uncertainty that is propagated into the model (see also Appendix B.2). Because the elicitations were conducted within the erosion team at Nagra with usually 3 – 5 experts per workshop, it was particularly important to include both, the expertise in local geology and geomorphology, as well as expertise in process understanding. The facilitator was supported by an external expert with a strong background on statistics and probabilities, who was trained on all SHELF aspects.

Parallel to the SHELF protocol, the elicitations were inspired by the SSHAC perspective of viewing probability distributions as centre, body and range of defensible interpretations (Appendix B.2, Fig. 3-5b; after Ake et al. 2018). Intuitively, the centre of the distributions usually incorporates the best estimate, often based on direct evidence from information gained in the

scientific programme. Part of the centre of a distribution are the statistical measures of central tendency (mode, mean, and median, see Fig. 3-5a). The body forms the shape of the distribution around the best estimate. This includes higher and lower quantile values (Q_x in Fig. 3-5a), and is based additionally on indirect evidence, for instance from numerical modelling or different interpretations of datasets. The range finally should even include as yet unseen evidence, as long as it appears physically plausible (Ake et al. 2018). Throughout the report it is often referred to *most likely* with respect to a certain value (i.e. the mode as seen in Fig. 3-5a), *most likely* with respect to a range (in this case it is referred to the "inner third" of a distribution, resembling closely the centre in Fig. 3-5b), and the *expected range*. The latter may be representative for the body of the distribution (Fig. 3-5b) and in this report refers to the 5th to 95th quantiles. Accordingly, values within this *expected range* are still called *likely*.

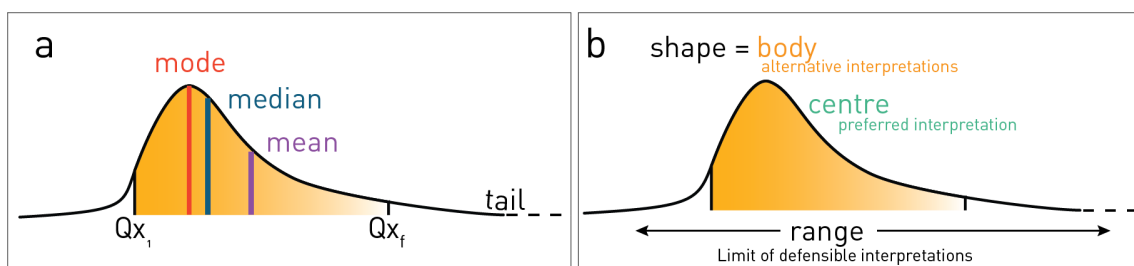


Fig. 3-5: Illustration of the properties and the interpretation of the shape of uncertainty distributions in the present context

Probability density functions (PDF) can reveal important data metrics. (a) Classic metrics are the statistic central tendency values mean, the mode, reflecting the highest probability of a value in a given distribution or the median which equals the 50th percentile. Similarly, any other quantile value (symbolised by Q_{x_1} and Q_{x_f}) can be drawn. The tails of the distribution describe the values with lowest probability (see Appendix Fig. C-1 for an introduction in cumulative distribution function [CDF]). (b) In the context of scenarios or quantification of uncertainties of parameters with poor statistics, it may be helpful to discuss the shape of the distributions in terms of interpretation of a certain evidence or quantity of interest (after Ake et al. 2018).

The elicitations are built on the evidence provided prior to the workshops which documents the factual basis for the elicitation and ensures that all experts are equally prepared and use the same factual basis. In this report, the main evidence is provided in Chapter 2, but in an updated version with respect to the time of the workshops that took place over a period of the past three years. However, all data and visualisations acquired after the elicitation workshops do not change the prior assessment.

In Sections 3.3 to 3.5, the RIO distributions are shown for the parameters used in the respective models. They are provided with a brief reasoning, which in turn is based on the evidence provided in Chapter 2, partly supported by additional global data or more process-related expertise, which might then be shown or cited individually.

In the following sections, the main processes related to erosion in Northern Switzerland are briefly described, followed by the specific workflow considering each assessment of future erosion compartments, including consideration of uncertainties. Subsequently, the rationale and main argumentation for scenario or model parameters are provided, followed by results and discussion.

3.3 Fluvial incision

3.3.1 Introduction to processes and drivers of fluvial incision

Rivers are complex and dynamic systems that evolve towards an equilibrium state between tectonic movements, sediment discharge and incision on long timescales. Rivers migrate laterally, erode and incise, and potentially capture adjacent catchments, thus reorganising the structure of entire river networks and their watersheds (e.g. Willett et al. 2014). The main processes driving fluvial incision are rock uplift/subsidence, discharge and sediment flux, modulated by the resistance of bedrock to erosion (erodibility). Over long timescales, uplift and subsidence of any landscape is mainly driven by neotectonic activity and/or isostatic compensation to mass redistribution by glacial or other erosive processes. Their effects dictate rates and location of incision and sediment accumulation and can thus cause (a) changes in the river profile (steepening, knick-points, aggradation and flattening), (b) catchment changes by fluvial piracy, and (c) variability in erosional resistance of the exposed rock.

An equilibrium longitudinal channel profile generally shows a graded, concave-upward form between a river's source and outlet. Any change in driving forces, for example patterns of tectonic movement or climate, will perturb the graded profile and prompt a river to respond by adjusting its patterns of incision and/or aggradation. Rivers thus modify their longitudinal profiles to acquire a gradient distribution that is suitable to balance the prevailing conditions and to ensure transport capacity (e.g. Hey 1979, Tucker & Slingerland 1997, Blum & Törnqvist 2000). Eventually, a river will reach its former graded state or attain a new steady-state equilibrium (Schumm 1977). However, as most controlling processes are changing before rivers adjust to their steady states (relaxation time), equilibria are usually disturbed, and rivers are therefore thought to be in transient states of adjustment for most of the time (e.g. Whipple 2001).

Changes in river gradient alter the efficiency with which water and/or sediment can be transported. Depending on the balance between sediment availability and the capacity of the river to transport it, sediment will provide the tools for eroding the bed or covering it and preventing further downcutting (Sklar & Dietrich 2001). Moreover, river captures or diversions (e.g. Wutach; other Rhine – Danube competitions, see Section 2.3) potentially cause sudden and significant changes in water and sediment flux as well as the level of the erosional base.

The rate at which rivers incise into bedrock depends on the erodibility and erosivity. Erodibility is mainly influenced by geomechanical and geochemical properties of the rock, but also by other macroscopic rock properties such as the density and orientation of fractures. Erosivity refers to the ability of exogenic forces to cause erosion. Thus, erosivity is determined by climate, and includes the magnitude of discharge and its statistical distribution and extreme events. Both erodibility and erosivity influence sediment discharge and the grain size of bedload in the channel. Erodibility and erosivity generally cause temporally fluctuating incision rates.

The Rhine, the regional erosional baselevel, connects the Alps with the North Sea, thereby traversing regions subject to tectonic uplift or subsidence (Fig. 2-1, Fig. 2-8b). The uplift of the Rhenish Massif and the volcanic Eifel (Fig. 2-1c) "decouples" downstream influences beyond the uplifting sectors, such as sea-level fluctuations, from the incision in the Hochrhein area. Consequently, an assessment of the siting regions can largely rely on the analysis of the river reach and potential forcing processes inducing incision between the Alps and the city of Bingen (Section 2.1). Today, the dominant control mechanisms influencing incision processes in the region are the uplift of the Alps, the existence of postglacial lakes (mainly the Bodensee) acting as sediment traps, and the tectonically subsiding Upper Rhine Graben (Fig. 2-6 and Fig. 2-8b). The siting regions in the Hochrhein region are located in a transition zone between the uplifting domain of the Alpine foothills and the subsiding domain of the Upper Rhine Graben; overall, the vertical crustal movements measured today are generally very small (Fig. 2-1c). Moreover, the

incision in the Hochrhein is currently reduced because the Bodensee intercepts the clastic sediments needed to erode the rocky riverbed by abrasion. Once the peri-Alpine lakes are filled with sediments, this situation may change, although the repercussions of a re-established sediment connectivity between the Alpine Rhine and the Hochrhein are difficult to project. For example, Alpine sediments that are currently trapped in the Bodensee may lead to incision upon liberation owing to an enhanced mechanical tool effect. Alternatively, the transport capacity of the Rhine may not be sufficient to transport the additional sediments, leading to aggradation. In turn, aggradation will increase the lateral mobility of the river, and a corresponding change from dominant vertical to lateral erosion.

Today, the Hochrhein is a mixed bedrock-alluvial river; such a river exhibits different channel types with fundamentally different dynamics (Turowski et al. 2008). The Hochrhein course runs across a number of bedrock sills and barriers (e.g. Mesozoic limestone of the Tabular Jura) that have no or only limited sediment cover (see Fig. 3-12). These bedrock sections alternate with transport-limited sections whose dynamics are largely governed by sediment flux and the capacity of the Rhine to transport the material (Preusser 2008). Sediment fluxes in the Northern Alpine Foreland are in turn largely controlled by the dynamics of glacial-interglacial cycles. Evidence for a sediment transport-dominated system is found in the present-day longitudinal Rhine profile, where, for example, the gradient does not change significantly at the transition into the Upper Rhine Graben (Fig. 2-8a). In such systems, incision will take place as a diffuse response to baselevel lowering and the river gradient remains largely unaltered by minor changes in uplift and local erodibility. Changes in uplift, however, potentially prompt such rivers to incise into their own deposits and the underlying bedrock.

Given the eminent role of uplift and baselevel change in the future evolution of the Hochrhein, the following assessment emphasises the quantification of uplift rates and the development of the baselevel. Moreover, as fluvial incision is strongly connected to erodibility, bedrock properties are particularly considered in the assessment. Finally, potential changes in drainage pathways and their associated catchments are analysed to study the effect of river captures and deflections.

3.3.2 Workflow to assess future fluvial incision

Several different approaches to modelling erosion and deposition exist, which can be discriminated by the degree to which they are grounded on physical rules, the types of physical rules, and by the spatial and temporal scales on which they operate. The spectrum ranges from conceptually simple statistical models with a high level of abstraction to detailed mechanistic models that make predictions of specific properties, e.g. grain size (Coulthard 2001, Nones 2020, Pazzaglia 2003). Here, the analysis of future fluvial incision and thus the evolution of the local erosion base is based on the stream power incision model (SPIM) as illustrated in Fig. 3-6.

Model selection and applicability for modelling the Hochrhein

The SPIM is among the most important models of quantitative geomorphology and is widely used to study the evolution of mountain belts and river systems at geological timescales (Lague 2014, Whipple & Tucker 1999, Kwang & Parker 2017). According to the SPIM, the rate at which rivers incise is a function of the drainage area A and channel slope $S = dz/dx$ raised to the power exponents m and n , respectively, and multiplied by a bulk parameter K which is a dimensional coefficient of erosion often simply referred to as erosional efficiency or erodibility coefficient. The SPIM is frequently used in landscape evolution models that include tectonic uplift U (e.g. Snyder et al. 2000). The main advantage is the low model complexity and its relatively straightforward numerical implementation (Braun & Willett 2013) which allows its application to long timescales ($10^5 - 10^6$ years) and thus its operation in an effort to investigate different boundary conditions.

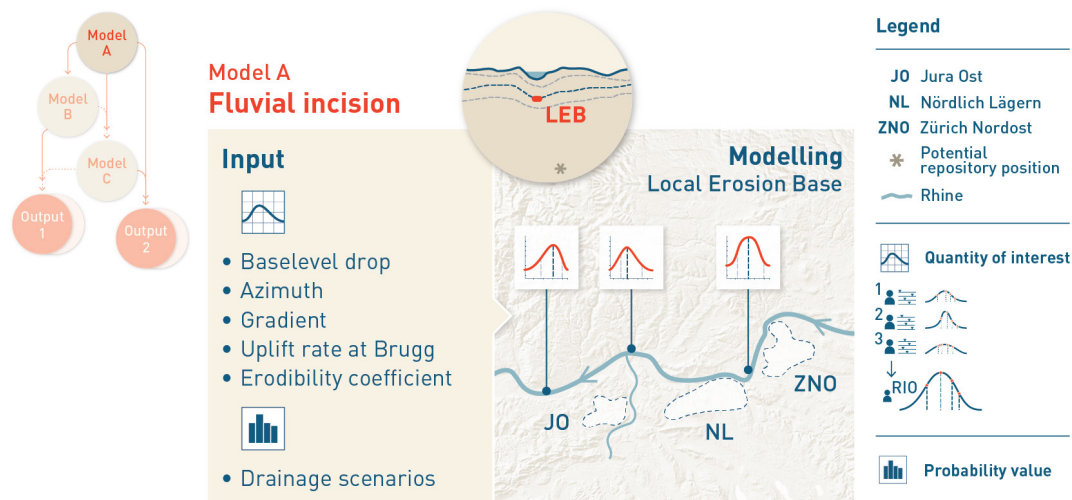


Fig. 3-6: Illustration of workflow to assess future fluvial incision using SPIM

The stream power incision model (SPIM) constitutes the main toolbox for modelling future fluvial incision (at certain time steps $T_1, \dots, T_x$) along the Hochrhein section between Stein am Rhein and Basel. The main input parameters (baselevel drop, rock uplift, changes in discharge (drainage scenarios), and erodibility (K value)) are varied and incorporated as probability distributions based on expert judgement and selected scenarios. The joint distribution of the uncertain model parameters is modelled in a Monte Carlo framework to provide probability distributions of future local baselevel in the siting regions. The respective baselevels are defined at the blue points on the map, representing the Sissle outlet for Jura Ost, the Aare outlet for Nördlich Lägern, and the Thur outlet for Zürich Nordost.

The SPIM relies on several restrictive assumptions: First, it states that the available kinetic energy of the discharge is used to incise into the bedrock, which occurs either as abrasion, cavitation or plucking. Second, the model assumes that sediments delivered from hillslopes or entrained along the river are immediately excavated. Both assumptions mean that fluvial processes related to sediment transport are not considered, thus overlooking the modulating role of sediments that may either protect the riverbed from incision or accelerate incision by the tool effect (Sklar & Dietrich 2001). Moreover, the model explicitly sets river width as a function of drainage area or discharge, thus simplifying the complex interaction between incision, aggradation, and changes in river width. Finally, the SPIM assumes that discharge can be represented by a simple function of drainage area but ignores that the stochastic variability of discharge may strongly modulate the incision behaviour of rivers.

These assumptions entail that the SPIM relies on only a few parameters (Temme et al. 2017). While a low parametrisation may not reflect the complexity of fluvial systems and thus suggest a lack of realism, there is consensus that – although against intuition – simple models perform better in prediction (Oreskes 2021). Moreover, the SPIM is based on relatively simple geometric model parameters (Lague 2014, Venditti et al. 2020) that can be obtained from the inversion of river profiles rather than measurements such as rates of sediment flux which are usually prone to large uncertainties. Moreover, such observations are usually available for observation periods that span a few years and decades and are thus much shorter than the time periods of investigation, thus limiting their applicability (Kirchner et al. 2001).

These assumptions must be cautiously applied to the Hochrhein which features numerous knick-points (e.g. Rhine Falls, Schaffhausen). Some of these knickpoints are associated with lithological variations, but others are likely mobile and reflect a transient adjustment of the profile, thus violating the assumption of a fluvial landscape in equilibrium (steady state). In addition,

lithological variations as well as spatially variable geodetically derived uplift rates suggest that K and U vary spatially. Field and laboratory information on rock types are used to address these points. In addition to lithological changes, the Hochrhein has a history of prominent reorganisations in drainage area (e.g. integration of the Donau-Wutach, see Section 2.4.2). It is unclear to what degree the present-day profile reflects these reorganisations. Finally, a major shift in fluvial dynamics is expected if the Bodensee becomes filled with sediment. Today, water and sediment fluxes of the Hochrhein largely reflect the hydromorphological situation of a non-Alpine area (Hinderer et al. 2013). The Hochrhein is transport-limited. Sediment transport rates along the Hochrhein will likely increase upon backfilling of the Bodensee. Moreover, half of the sediments mobilised in Alpine catchments currently remain in the headwater regions (Hinderer et al. 2013) due to a lack of longitudinal sediment connectivity along Alpine catchments (Heckmann & Schwanghart 2013). While the SPIM does not explicitly account for sediment transport, sediment transport can have diverse effects on river systems ranging from increased bedrock incision rates to channel bed protection and aggradation (Sklar & Dietrich 2001).

The parameters of the SPIM are usually estimated based on inversion of river profiles which are derived from digital elevation models (DEMs). The geomorphometric variables of slope and upstream drainage area are typically derived from a DEM with 30 m spatial resolution. Horizontal and vertical lithological changes along the river profiles are incorporated in the model (Fig. 3-12) based on (Pietsch & Jordan 2014). Furthermore, the Hochrhein River profile was also corrected for the anthropogenic dams.

SPIM parameter uncertainty and output variable

When addressing the future behaviour of the erosional system of Northern Switzerland, it can be assumed that parameters such as rock uplift U , erodibility coefficient K or upstream drainage area A might change. These parameters were assessed independently as subjective probabilities following the SHELF protocol (Section 3.2.3, Appendix B.2). In addition, to account for the downstream change from the tectonic domain of the Alpine Foreland into the domain of the Upper Rhine Graben, the parameter baselevel drop (BL) was introduced and elicited.

The exponent m in the SPIM largely reflects how drainage area translates into discharge. Notwithstanding, m is a bulk parameter that additionally reflects channel width, the discharge regime and the role of extreme events (Lague 2014). The parameters m ($= 0.45$ as commonly used) and n ($= 1$ for convenience and to avoid further parameter correlations) were not elicited. For these values of m and n , the units of the erodibility coefficient K are $[m^{(1-2m)} * \text{yr}^{-1}]$, which using $m = 0.45$ further translates into $[m^{0.1} / \text{yr}]$. Note that these units apply to all K values (including figures and tables) documented in this report.

Using a variety of scenarios for changes in future discharge/catchment that are ranked by their plausibility, and variants for the parameters based on expert's degree of belief, SPIM was used in a Monte Carlo approach with one million samples to generate possible realisations of future river profiles between the outflow of the Bodensee and Basel (i.e. the Hochrhein) at time T_x from all parameter combinations. For each siting region, the elevation of close downstream positions along these future Hochrhein profiles was assessed. These points that set the respective baselevels are the Sissle outlet for Jura Ost, the Aare outlet for Nördlich Lägern, and the Thur outlet for Zürich Nordost (Fig. 3-6 and Fig. 3-12). To account for the periods of concern for L/ILW and HLW, time T_x was chosen to be 100'000 yr and 1 million yr, respectively. Importantly, the results are only evaluated and discussed with respect to the final outcome at the end of these two time periods and not as a continuous evolution of temporarily varying rates. This means that, for instance, at the end of the period under consideration, short-term effects (e.g. significantly higher rock uplift rates as a response to glacial isostatic adjustment) are largely averaged out.

3.3.3 Elicited parameters: rationale and argumentation

3.3.3.1 Rock uplift

Rock uplift is one of the input parameters for SPIM modelling. An early study on parameter sensitivity revealed that uplift is indeed the most influential parameter, not only for SPIM, but also for the entire hybrid model (Sections 3.3.5.1 and 5.1). Moreover, different data and theoretical considerations suggest that uplift is not spatially homogeneous. Rather, within the wider domain of the Hochrhein and the siting regions, uplift follows a general spatial trend that mainly reflects the ongoing, low-magnitude Alpine uplift. This first-order observation can be made irrespective of the particular cause (isostasy or shortening-related) and investigated timescale. A simple model that describes these spatial uplift patterns is a linear trend function of the x and y-coordinates which, when fitted to measured uplift data, shows a decreasing trend in NW-direction, thus tracing the drainage direction of many tributaries to the Hochrhein (e.g. Limmat, Reuss, Lower Aare, and Thur etc.; see Fig. 2-5). Such a trend is also reflected by precise levelling rates (Fig. 2-5a), recent GNSS uplift rates (Fig. 2-5b) or elevations of the Deckenschotter base (Fig. 2-5c and d). Accordingly, the spatial patterns of rock uplift within Northern Switzerland are approximated by a plane, i.e. rock uplift is modelled as a linear function of easting and northing. This rock uplift plane (Fig. 3-7) is defined through the following parameters, the determination of which is subject to uncertainty and that was assessed following the SHELF protocol:

1. the rock uplift rate in Brugg (*UBR*),
2. an azimuth direction that defines the direction along which the increase in uplift is largest (assuming Alpine uplift as main cause), and
3. the slope in the azimuth direction (referred to here as gradient).

It is important to note that azimuth and gradient were elicited independent of the time period with a focus on one million years, which was assumed to be representative for both the 100'000-yr and one million-yr time spans. Uplift at Brugg was elicited separately for both of these periods (Fig. 3-8).

Based on the plane, the uplift can be evaluated for any coordinate within the domain of interest, e.g. for the locations of the siting regions (Fig. 3-9) or locations along the Hochrhein profile.

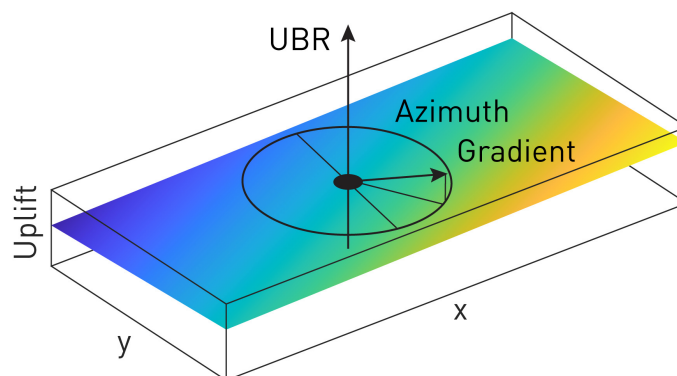


Fig. 3-7: Sketch illustrating rock uplift, gradient, as well as azimuth of the uplift plane
UBR: Uplift at Brugg.

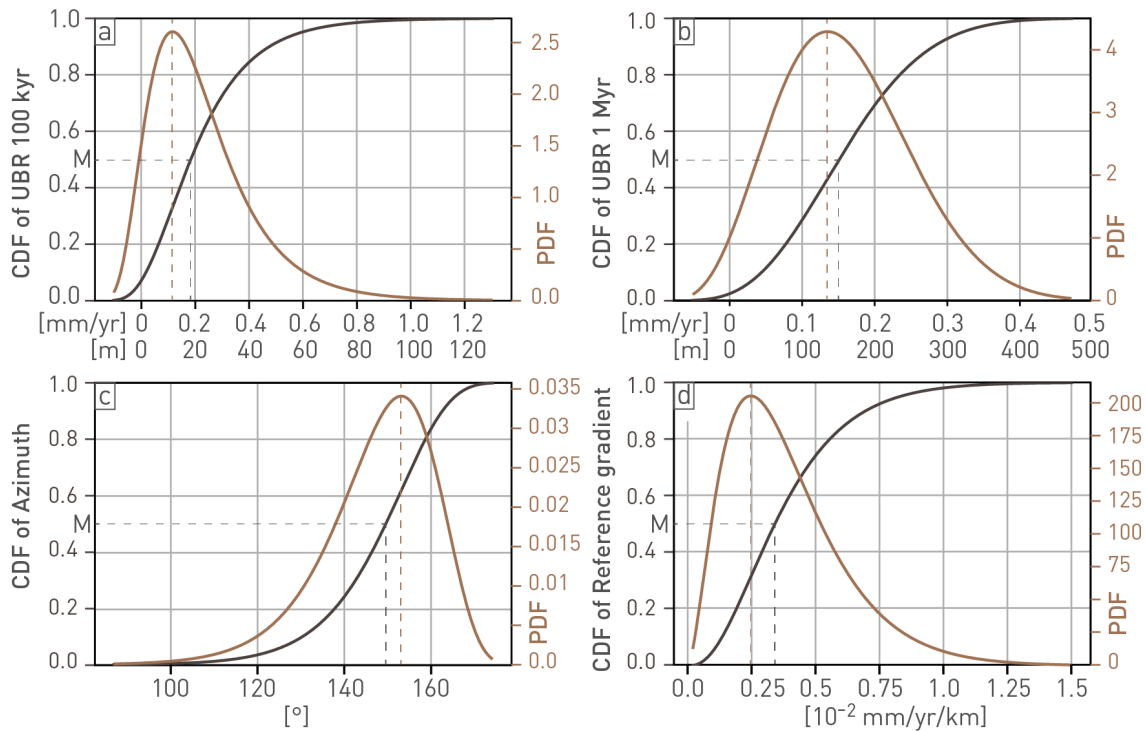


Fig. 3-8: Total assumed probabilities of rock uplift parameters

The probabilities are shown as PDF (brown, with the mode [representing the most likely value] marked as dashed line) and as cumulative distribution function (CDF, dark grey), where the value of 1.0 (y-axis) equals to 100% cumulative probability. The letter M on the y-scales refers to the median value, i.e. the 50% quantile (value 0.5). The corresponding median values can be read at the x-axes. (a) Uplift distribution for Brugg (*UBR*) for the 100'000-yr time span, represented by a log-normal distribution, shown as rate (upper) and cumulative amount (lower x-axis) (b) Uplift distribution for Brugg for the one-million-yr time span, represented by a beta distribution. Note the different scales of x-axes between (a) and (b). (c) Distribution of azimuth of steepest ascent (represented by a mirrored log-normal distribution). (d) Distribution of slope (reference gradient) along the main azimuth (represented by a gamma distribution).

Rock uplift rate at Brugg

In this report, rock uplift rate is defined as the total rate of bedrock uplift in mm/yr that is expected along the Hochrhein in response to tectonic processes and as an isostatic response to loading or unloading, including the response to loss of ice cover. The reference level for rock uplift is the geoid.

The location of Brugg as reference point for defining the rock uplift rate was chosen because it provides a central position in Northern Switzerland (Fig. 3-7) and because it is associated with rather good data coverage (GNSS levelling, Deckenschotter outcrop with exposed base and age information [Iberig]).

Both distributions, for 100'000 yr and one million yr, concentrate around uplift rates below 0.25 mm/yr (Fig. 3-8a and b).

One main piece of evidence that was used for this evaluation was Quaternary incision as reflected by the Deckenschotter geomorphic markers. It was agreed, however, that this proxy must be used with caution because the Quaternary incision history is considered to be complex and might

incorporate more drivers than uplift only (see Section 2.4). The age of terrace deposits at the Iberig outcrop using isochron burial methods was estimated to be between 0.6 and 0.8 million yr. These deposits are $\sim 100 - 120$ m higher than the deepest fluvial baselevel nearby, leading to a mean observed incision rate of ~ 0.16 mm/yr since then. It was acknowledged that incision likely ceased before 150 – 200 kyr, as was argued from dated infill of these deepest channels (Section 2.4.2.3). Thus, this period was assumed to possibly reflect pulses of bedrock incision and intervening periods of quiescence and/or sediment accumulation.

GNSS and levelling data (Fig. 2-5) were also used, although it was acknowledged that these data reflect a short-term signal and should not be overinterpreted on the long-term time windows of 100 kyr and one million yr. Furthermore, the different reference of the datasets hampers a thorough estimate. Precise levelling data result in values up to 0.08 mm/yr close to Brugg (mean error ~ 0.06 mm/yr), and up to 0.17 mm/yr near Baden (with respect to Laufenburg, see Fig. 2-5a).

The main difference between the elicitations for 100'000 and one million yr was a generally wider distribution (especially between 5 – 95%) and a slight shift in the central body towards higher rates for the shorter time span. It was accepted that higher uplift rates might be seen on the 100-kyr time span, which would be averaged with periods of low uplift rates on the million-year time span. For the upper limits of the 100 kyr, it was agreed to include the possibilities that (1) very high incision could occur within only 100 kyr and (2) that rock uplift might be the main driver. High short-term uplift pulses exceeding present-day rates were considered for the 100'000-yr period. However, high isostatically-driven uplift following deglaciation (although not excluded) was not considered very likely for the next 100 kyr, because climate models show that anthropogenically enhanced atmospheric CO₂ may delay the next glacial inception considerably (see Section 3.5.3.5) but was also not excluded.

In general, the uncertainties of age constraints and of the conceptual models for incision as well as uncertain long-term estimates based on thermochronology (Section 2.1) were incorporated for the uncertainty bounds. High long-term (incision) rates of 0.8 mm/yr, reported by Häuselmann et al. (2007) who noted a ten-fold increase in these rates ~ 900 kyr ago were discussed in the context of possible long-term uplift rates. The observations of Häuselmann et al. (2007) were conducted in the region of the Alpine front rather than the foreland, attributing the increase to climatic changes associated with the Mid-Pleistocene Transition, which suggests the influence of factors beyond uplift rates alone. Nevertheless, a distinct change in rock uplift late in the period under consideration was not ruled out entirely, as isostatic effects for instance after deglaciation might have played a role (see Section 3.5.3.4 for scenarios and timing of future large foreland glaciations). However, such a late increase in rates would only incorporate a shorter time period.

A small percentage ($< 5\%$) of the total assumed probability also allows for subsidence, as seen in the present-day levelling data west of Brugg (Fig. 2-5a). However, considering the reference for the levelling data is located at the Hochrhein in Laufenburg and not further downstream and there is no evidence for such a subsidence over geological timescales in the past, it was in general regarded as more important and more reasonable on short timescales. The probabilities for higher subsidence on timescales of 10^5 to 10^6 years were considered very low. Accordingly, a shifted log-normal distribution was chosen to best represent the RIO perspective for the 100-kyr time period, and a beta distribution was chosen for the 1 million yr time period (Fig. 3-8a and b).

Azimuth direction

The uncertain parameter 'azimuth' is defined as the direction with the strongest increase of uplift measured clockwise from north. The active Alpine orogeny is assumed to be the first-order control on future uplift rates in Northern Switzerland. Accordingly, the main azimuth direction was agreed to be oriented approximately perpendicular to the Alpine front. Geological maps as well as a visualisation of the Moho depth (Michailos et al. 2023) provided first-order evidence for the

predominant Alpine direction. Further evidence was provided by interpreting the geodetic data (Fig. 2-5 and b), as well as the isoline plots of interpolations of the base of the Deckenschotter (Fig. 2-5c and d). Reversed uplift patterns (e.g. lower uplift rates towards the Alps) are excluded. However, approaching a scenario of nearly spatially uniform uplift rates in Northern Switzerland is regarded as possible.

The elicitation of the uncertainty around the azimuth followed the SHELF protocol. Individual estimates were guided by a visualisation toolbox that allowed importing of the above-mentioned datasets, provided simple interpolations based on chosen window sizes, and allowed testing of individual (best fit) approximations of azimuth and gradient. The subfigures of Fig. 2-5 were derived from this toolbox.

The best approximations of the main azimuth (using a set of different window sizes) cluster between ~ 140 and 160° , which was accepted as a good reason for a generally rather narrow distribution around these values. The median was set to 150° , which is approximately the median value that results from the datasets available (Fig. 2-5). It was also agreed on a right-skewed distribution because a tendency to lower values was considered more likely than high values and because it was deemed unrealistic that uplift in areas such as Schaffhausen in the east would be lower than uplift at Basel in the west (which is located in the transition zone to the subsiding Upper Rhine Graben, see Section 2.1). Accordingly, only very low probability should be assigned to values larger than 165° ($\sim$ the value at which Schaffhausen would subside with respect to Basel). This consideration together with a general uncertainty allowed for the orientation of the Alpine front and the representation of a right-skewed distribution was used to limit the upper boundary to the absolute value of 190° .

The isolines of the interpolated base of the Deckenschotter deviate from a northwest to west trend in the western part of the Northern Switzerland (Fig. 2-5c and d). Although it was considered most likely that this trend reflects the baselevel influence of the Upper Rhine Graben rather than Alpine uplift, it was acknowledged that such an azimuth should still be considered with low probability. It was discussed under which conditions Alpine uplift could still be considered the dominant scenario. Local processes and possibly also glacial components (i.e. GIA effects) could cause larger deviations from the reference azimuth direction. To account for such local deviations, for instance on a spatially nearly uniform uplift pattern, the lower limit was set to 100° . A mirrored log-normal distribution provides a reasonable fit to arguments based on the observed values. This means that the natural logarithm of 190° azimuth follows a normal distribution with a mean of 3.7 and a standard deviation of 0.303 (Fig. 3-8c).

Gradient (slope in azimuth direction)

Gradient is defined here as change in uplift rate per distance in azimuth direction of the uplift plane independent on the time period with focus on one million years. It was assumed that differences in the uplift rate between the 100-kyr and one million-yr time spans would be more pronounced than that of the gradient.

The gradient was elicited only after the uncertainty distribution of the azimuth was established. The elicitation of the gradient was performed in a two-step process. First, the probability distribution for the gradient was elicited for the azimuth conditionally fixed to 150° (i.e. the median of the azimuth distribution). The elicited gradient that is conditional on an azimuth of 150° is referred to as "reference gradient". The reference gradient was elicited in mm/yr/km for the one million-year time span following the SHELF protocol. Second, a transformation function that scales the reference gradient as a function of the azimuth was agreed upon. Thus, the value of the gradient used to represent the uplift plane can be determined deterministically as a function of both the reference gradient and the azimuth.

It was agreed that, for the elicited azimuth distribution, negative values for the gradient should not be allowed. However, values close to zero could occur (i.e. almost uniform uplift). Therefore, "0" was set as a lower bound for the distribution of the reference gradient, with low probabilities for values below 0.001 mm/yr/km. The elicited scaling function ensures that the probability of small values of the gradient increases the more the value of the azimuth deviates from its median.

It was also agreed that the interpolated base of the Deckenschotter deposits (and their difference to the local erosion base) provide a better representation of the long-term gradient than the geodetic data. Because there is a time component considered within the definition of the gradient, the ages of the Deckenschotter needed to be considered in gradient estimation. Following their orientations along the previously elicited main azimuth directions and allowing ages between 1 and 2 Ma resulted in gradients between ~ 0.0018 and ~ 0.0045 mm/yr/km along the main azimuth direction. A median value of 0.0033 was accepted to reflect both time periods. However, low gradients were considered less likely than higher gradients and the curve should reflect this with a left-skewed shape.

Extreme scenarios were used to estimate the low-probability values and upper limit. For instance, an interpolation of the TDS incision, considering the lower age limit of the Iberig outcrop (Section 2.4.1.3), resulted in a high gradient of 0.008 mm/yr/km. A hypothetical scenario with a HDS age of 700 ka for the Irchel Deckenschotter incision revealed a gradient of ~ 0.016 mm/yr/km. The upper limit was accepted at 0.015 mm/yr/km as this also represents the extreme values from the geodetic data. A gamma distribution with a mean of 3.9×10^{-3} provides a reasonable fit to the arguments and elicited quantiles (Fig. 3-8d).

Correlation between uplift at Brugg and gradient

Given the simplified assumption of the linear uplift plane controlled by Alpine uplift and the city of Brugg located within this plane, a dependence between the uncertain model parameters *UBR* and the reference gradient can be expected. If, for instance, the Alpine uplift increases significantly, this signal would be noticeable at locations farther north. Accordingly, a high-gradient sample should be coupled to a rather high uplift rate sample. A certain control on such dependency can be derived from expectations for uplift around a location downslope along the azimuth. The area close to the city of Pratteln, just upstream of Basel, was used as a control. A combination of a high gradient with lower uplift rates at Brugg would unrealistically result in subsidence near Pratteln. Thus, an elicitation was performed to estimate the probability value for subsidence to occur. GNSS data and incised Deckenschotter deposits that extend into the Upper Rhine Graben were used as evidence that such subsidence cannot be excluded. Pratteln is located near the boundary of the Upper Rhine Graben and normal faults extend even beyond the graben, allowing for the less likely interpretation of a transition into a normal faulting regime. However, more subsidence would be allowed on the shorter time span combined with a lower correlation than at the one-million-year time period.

Accordingly, 25% subsidence was allowed for the location of Pratteln for the 100-kyr time span, leading to a correlation coefficient of 0.3. For the one-million-year time span, 10% subsidence was allowed and a correlation coefficient of 0.95. The dependence between the uplift in Brugg (*UBR*) and the reference gradient is then modelled using the Nataf transformation and the respective correlation coefficients (of the underlying bivariate Gaussian distribution; Nataf 1962, Ditlevsen & Madsen 1996).

Based on the distributions above, the rate of rock uplift in the siting regions was inferred from the uplift plane (Tab. 3-1 and Fig. 3-9).

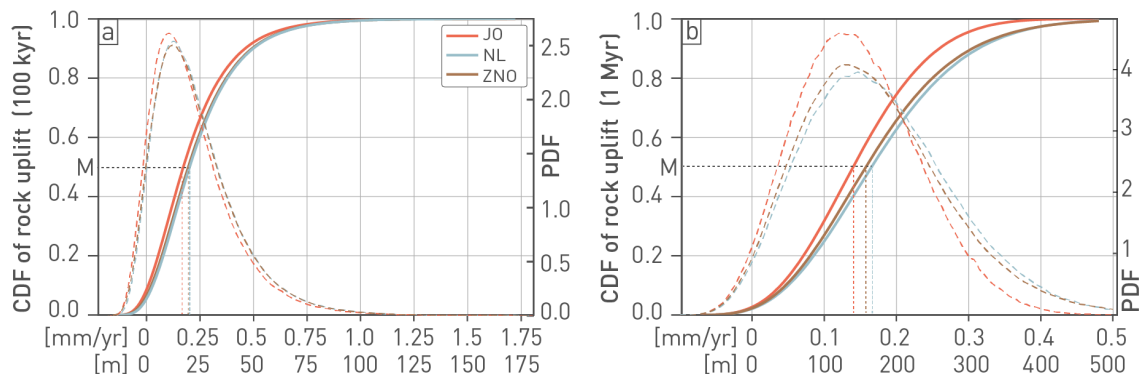


Fig. 3-9: Inferred future rock uplift rates in the three siting regions over the next 100'000 and one million years

Rock uplift rate is defined as the total rate of bedrock uplift in mm/yr that is expected as a response to tectonic processes and as an isostatic response to loading or unloading, including the response to loss of ice cover. The reference level for rock uplift is the geoid. The estimates are based on the assumption of a planar uplift pattern caused primarily by Alpine uplift. The results are calculated for the location of the provisional disposal areas and shown as probability density functions (PDF) and as cumulative distribution function (CDF, see Fig. C-1 for explanation), where the value of 1.0 (y-axis) equals to 100% cumulative probability. PDF visualisations allow to assess the most probable values, which are identified by the bulge of the curve (here close to 0 mm/yr). In contrast, lowest probabilities are reflected in the tails. The letter M at both y-scales refers to the median value, i.e. the 50% quantile (value 0.5). The corresponding median values can be read at the x-axes (a) Probability density function (PDF) and cumulative distribution function (CDF) display the estimates for the timeframe of 100'000 yr, (b) PDF and CDF showing the estimates for one million yr. Note that the shorter timeframe of 100'000 yr allows higher rates. M refers to the median values. See Tab. 3-1 for a number of quantile values.

Tab. 3-1: Inferred future rock uplift rates at the locations of the siting regions

Rates in [mm/yr] are calculated based on the uplift plane introduced above for the centroid coordinate of the provisional disposal areas in JO, NL and ZNO, respectively (same as shown in Fig. 3-9). For each site, the first value refers to the 100'000-yr and the second to the one-million-year time period.

Percentile	Uplift at JO	Uplift at NL	Uplift at ZNO
5%	$2.4 \times 10^{-2} / 1.7 \times 10^{-2}$	$4.1 \times 10^{-2} / 2.4 \times 10^{-2}$	$1.2 \times 10^{-2} / 2.2 \times 10^{-2}$
25%	$7.4 \times 10^{-2} / 8.4 \times 10^{-2}$	$9.9 \times 10^{-2} / 1.0 \times 10^{-1}$	$9.2 \times 10^{-2} / 9.5 \times 10^{-2}$
50%	$1.7 \times 10^{-1} / 1.4 \times 10^{-1}$	$2.0 \times 10^{-1} / 1.7 \times 10^{-1}$	$1.9 \times 10^{-1} / 1.6 \times 10^{-1}$
75%	$3.0 \times 10^{-1} / 2.0 \times 10^{-1}$	$3.4 \times 10^{-1} / 2.4 \times 10^{-1}$	$3.3 \times 10^{-1} / 2.3 \times 10^{-1}$
95%	$5.8 \times 10^{-1} / 3.0 \times 10^{-1}$	$6.2 \times 10^{-1} / 3.6 \times 10^{-1}$	$6.1 \times 10^{-1} / 3.6 \times 10^{-1}$

3.3.3.2 Baselevel changes (baselevel drop)

To account for the downstream change between the tectonic domain of the Alpine Foreland and the subsiding domain of the Upper Rhine Graben, the parameter baselevel drop (*BL*) was introduced and elicited. The rate of baselevel drop is defined as the lowering of the surface in mm/yr at the intermediate baselevel of the Rhine River in Basel after entering the domain of the southern Upper Rhine Graben. Basel constitutes the lowest point along the Hochrhein longitudinal profile. Similar to rock uplift, baselevel drop is estimated with respect to the geoid or its first-order representation of the sea level. Unlike rock uplift, considerations of baselevel drop are made with respect to the riverbed, including unconsolidated sediment. Thus, effects of sedimentation which will partly counterbalance subsidence are considered. Accordingly, the baselevel can be lowered by subsidence or incision into bedrock, as well as by incision into the Quaternary sediment fill. Negative rates of baselevel drop result in a net baselevel gain/rise (e.g. by sediment accumulation). For the elicitation it was considered helpful to estimate the absolute amount of lowering that a protracted rate would result in, either after 100 kyr or one million years. It is important to note that the baselevel drop evaluated for the location of Basel does not coincide with the previously considered helper variable that integrates potential subsidence near Pratteln (immediately upstream of Basel) to yield an estimate for the correlation between *UBR* and gradient (see above).

The position of the baselevel at Basel reflects downstream processes, which are impacted by subsidence of, and sediment accumulation within, the Upper Rhine Graben, and uplift of the Rhenish Massif and the Eifel near Bingen (i.e. another intermediate baselevel further downstream). Due to this decoupling of fluvial processes beyond the Rhenish Massif and Eifel, the influence of sea-level changes (ultimate baselevel) was excluded, as this would have only influenced areas downstream of the uplifting region (Section 2.1). An additional downstream buffer might be represented by the low-subsidence area of the "Karlsruher Schwelle", between the two main depo-centres of the Upper Rhine Graben (Section 2.1). Subsidence is commonly measured as sediment infill rate. Sedimentation rates are much higher in the northern Upper Rhine Graben (Ellwanger et al. 2008) than in the southern Upper Rhine Graben (Hagedorn & Boenigk 2008). The position of Basel even farther south of this "hingeline" suggests lower sedimentation rates there. Near Basel, the thickness of the Niederterrasse sediments was estimated to ~30 m above the deepest palaeochannel based on the bedrock model. The Upper Rhine River has incised up to 14 m into the Würm⁸ fill in its southern part and only up to 0.5 m in the northern part (Hillebrand & Frings 2017). Older incised terraces are rare, but do exist in the transition with the Upper Rhine Graben downstream towards the city of Sierentz (~ 20 km downstream of Basel; e.g. Claude et al. 2017). The conglomeratic gravels of these terraces comprise remnants of the pre-Rhine Sundgau Schotter, as well as Deckenschotter (HDS and TDS) and Hochterrasse sediments, the latter three covered by a thick loess-palaeosol sequence (Kock et al. 2009).

Climatic changes associated with the Mid Pleistocene Transition (MPT) have had a major influence on bedrock incision of European rivers (Gibbard & Lewin 2009), possibly controlled by higher runoff, greater supply of erosive tools, or permafrost-related weakening of underlying bedrock. There is evidence for accelerated uplift of the Rhenish Massif and the Eifel since about the MPT as well (Section 2.1, Fig. 2-7), which is related to either increased mantle activity and/or more effective erosion, potentially linked to the onset of periodic glaciations.

Because sedimentation is included in the considerations, the intermediate baselevel at Basel also reflects sediment dynamics resulting from glacial-interglacial cycles upstream. Near Basel, sediment supply most likely increased during (peak of) glaciations and resulted in aggradation (e.g. Rentzel et al. 2009, 2015 and references therein). To yield a baselevel drop near Basel, the rate of subsidence or incision must outpace the rate of sediment accumulation. Reasons for decreasing sediment delivery could be a future diversion of the Aare River into the Rhone system e.g. in the

⁸ Regional name for the last glacial cycle within the Alpine area.

Yverdon area, postglacial lakes as sediment traps (more efficient over short timescales), or a longer interglacial (less glacial erosion might lead to sediment starvation on longer timescales). Presently, the areas of the Hochrhein (downstream of Bodensee) and parts of the Oberrhein (~ to the city of Karlsruhe) are impounded and comprise a succession of weirs for power generation, which may misrepresent the present natural interglacial sediment dynamics in these reaches.

In general, the baselevel was assumed to be considerably stable with a slight tendency for base-level drop. This is reflected in the mode of the distributions for both time spans (Fig. 3-10).

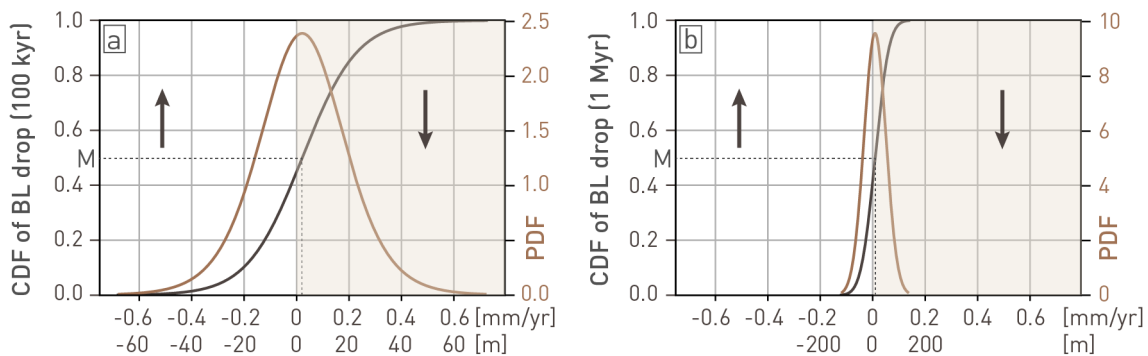


Fig. 3-10: Total assumed probabilities for baselevel drop downstream of Basel

The total assumed probabilities of the rate of baselevel drop are based on expert judgement. Probabilities are shown as cumulative distribution function (CDF) and as probability density function (PDF, brown colour). (a) 100'000-yr time period and (b) one-million-yr time period. Negative values correspond to net baselevel gain (see arrows). For both timeframes, the baselevel downstream of Basel is expected to be considerably stable with a very slight tendency towards baselevel drop (see mode of PDFs and median values of CDF). However, fluctuations of a few tens of metres in total over a one-million-year time span were considered plausible (shown for instance by the body of the PDF-distribution, compare Fig. 3-5, see text below for the numbers and arguments). The second x-axis shows the cumulative amount that protracted baselevel changes using the elicited rates would yield over the associated time periods. The axes of both plots were kept similar to highlight the higher rates that were assumed for the short-term estimate (a) with respect to the long-term estimate (b).

Considerations for 100'000 yr

For the 100-kyr time span, it is expected that the Upper Rhine Graben continues to subside as seen for the Holocene period but will also be continuously filled in with sediment. Further baselevel lowering can be expected at least until the peri-Alpine lakes are filled. For the time after that, which would correspond to approximately 20 kyr into the future, it is assumed that the lowering will either slow down or that there will be a phase of renewed aggradation. Accordingly, the net effect after 100 kyr could be both baselevel lowering or increase, which should be reflected by the total assumed probabilities (Fig. 3-10a). A low rate of baselevel drop compared to today (since the LGM) was viewed to be more likely. As Basel is located close to the source of Alpine-derived sediment, effects of reduced sediment supply should be recorded there. Between 10 and 30 m of continued incision were assumed feasible, until the sediment supply is high enough to invert the process, but the lower value was considered far more likely. Such baselevel drop could also be supported by precise levelling data (see Section 2.1 and Fig. 2-5) and assumptions about the overall gradient between Basel and Bingen (Fig. 2-8a), which would be reduced by 20% if considering a rate of 0.3 mm/yr. As an extreme upper limit, a continuous incision at the Holocene rate into the sedimentary fill (~ 0.7 mm/yr) was considered. Such a rate, leading to 70 m total

baselevel drop, would also correspond to the maximum determined subsidence/sedimentation rate (Ellwanger et al. 2008), but could only be achieved assuming subsidence as a driver and coeval sediment starvation, a condition which was not considered likely.

The chosen distribution should reflect that, in about a third of the cases, the baselevel in the 100-kyr time span could be expected to slightly rise within a few metres only and not more than 30 m (i.e. the mean thickness of the Niederterrasse sediments near Basel). A baselevel rise of 80 m (a rate of 0.8 mm/yr) was considered as an extreme value for sediment accumulation at the time period of 100 kyr. This value was based on the elevation of top Hochterrasse (near Bruderholz) with respect to the base Quaternary (~ 80 m). Although it was acknowledged that generally such pulse of sedimentation might occur within only 100 kyr, it was agreed that it would require an extensive glaciation, which is not deemed likely for the next 100 kyr (see also Section 3.5.3.5). The parametric RIO distribution is selected as a Student's t-distribution with a mean of 0.02 mm/yr, a standard deviation of 0.18 mm/yr and 8.8 degrees of freedom. The probability of a baselevel drop smaller than zero (i.e. the probability of a baselevel gain) is approximately 45% (Fig. 3-10a, i.e. a bit less than half of the PDF is located to the left of zero).

Considerations for one million years

For the long-term assessment of one million years, it was generally assumed that the effects of tectonic subsidence in the Upper Rhine Graben and sediment accumulation would largely balance out, as it is shown in the geological record (Section 2.1). This should be reflected in the rather narrow distribution by showing a higher probability for the baselevel remaining close to the present configuration (i.e. about one third between a very slight baselevel rise and only ~30 m baselevel drop). However, incision of up to ~70 m was still considered likely, which would be achieved approximately by extrapolating the present-day longitudinal profile of the free-flowing section between Bingen and Karlsruhe (e.g. Klösch et al. 2021) farther upstream. Another argument in support of such high value is reflected by the approximate height difference between the base of remnant TDS (which are younger than one million years, see Section 2.4) and the base of the Niederterrasse (approximately 22 – 26 ka old) around Basel. Part of this high incision amount is likely caused by upstream processes.

The intermediate baselevel at Bingen is presently located at an elevation of ~90 m a.s.l. and at Basel it is around 240 m a.s.l. If the longitudinal profile in the area of Bingen stays at the same elevation due to prolonged uplift of the Rhenish Massif and Eifel area (Section 2.1), the local baselevel upstream (at Basel) is not thought to be reduced by more than this difference of ~150 m, corresponding to a rate of 0.15 mm/yr for the upper limit of the one-million-yr time span.

A baselevel rise of up to 80 m was still considered likely for the one-million-yr time span, which was likewise based on the elevation difference between top Hochterrasse (near Bruderholz) and the base of Quaternary near Basel. However, the long period of one million years makes it more likely for future glaciations to occur and supply large amounts of sediment. As an extreme upper limit, a value of 150 m was chosen, which even exceeds the observed elevation difference between the HDS base and the deepest bedrock level. The parametric RIO distribution is selected as a normal distribution with mean 0.009 mm/yr and standard deviation 0.042 mm/yr. The probability of a baselevel drop smaller than zero (i.e. the probability of a baselevel gain) is approximately 41.5% (Fig. 3-10b).

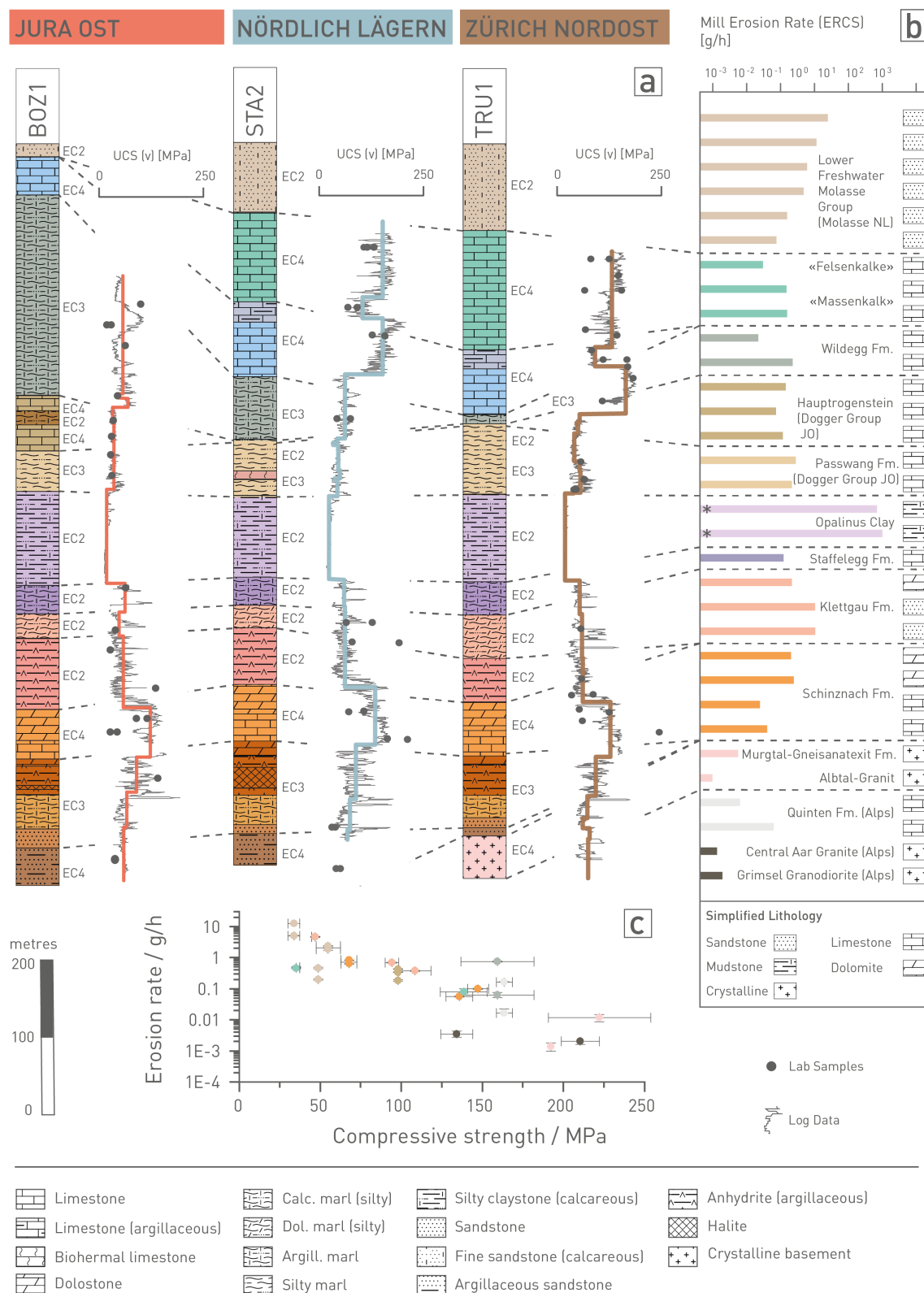


Fig. 3-11: Unconfined compressive strength and abrasion mill erosion rates as quantitative indicators of bedrock erodibility

(a) Stratigraphic columns based on the BOZ1 (Bözberg-1-1), STA2 (Stadel-2-1), and TRU1 (Trüllikon-1-1) boreholes with measured and estimated unconfined compressive strength (UCS) from these selected deep boreholes in the three siting regions. Measured data points are based on core samples tested in the laboratory (Dossier IX in Nagra (ed.) 2022a, 2022b, 2021). Continuous curves were approximated by combining the laboratory measurements with borehole log data from Nagra (2024d). Erodibility classes (EC) along the stratigraphic columns reflect the abstraction of the lithostratigraphic units as used for erodibility-related parametrisation in Model B of the erosion assessment model cascade (Fig. 3-3 and Section 3.4). Quaternary deposits (not displayed here) are assigned to EC1. (b) Relative erodibility in fluvial impact erosion as represented by erosion rates from abrasion mill experiments (Turowski et al. 2023; see further details there). Measurements were conducted on intact rock samples in the laboratory (ERCS method = erosion rate from the concentration of solids), from lithostratigraphic units as labelled. Asterisks denote results for highly erodible Opalinus Clay samples, where deviations from the standard experimental procedure were necessary. Samples were collected from surface outcrops, except for the Opalinus Clay and Alpine crystalline rock samples, which were taken from the FMT and GTS rock laboratories, respectively. Due to technical limitations, there is a sampling bias towards harder rock types. This is particularly important for more heterogeneous formations, where samples focus on the harder beds (e.g. cemented sandstone layers in the Molasse units and limestone layers within the Wildegg Fm.). Note the simplified lithological classification. (c) A strong negative correlation between mill-erosion rate and UCS is suggested by paired laboratory measurements by Turowski et al. (2023). Error bars are only discernible if larger than the data point symbols. Colours correspond to figure part (b). Note that data for the Opalinus Clay and the Passwang Formation from (b) are missing in (c) due to lack of UCS measurements on these samples.

3.3.3.3 Erodibility and K values

Generally, two main counteracting processes are important for bedrock incision: (i) the "tool effect" when bedload (tool) is available to erode the bed by impact, abrasion, or quarrying, and (ii) the "cover effect", when the riverbed is protected from erosion by a critical sediment cover (Turowski et al. 2007). Erodibility, which defines the rock resistance against erosion⁹, has an impact on sediment production and discharge, grain size, and bedload in the channel. The main influences on erodibility, in turn, are lithological contrasts, often defined by their rock compressive and tensile strengths (e.g. Fig. 3-11a, c).

The lithologies of the strata in Northern Switzerland comprise variable degrees of rock resistance (erodibility) with respect to the influence of fluvial impact erosion, as determined by abrasion mill experiments (Fig. 3-11b; Turowski et al. 2023). Resulting erosion rates derived from these experiments varied over about six orders of magnitude. Uncertainties scale with measured erosion rates.

Erosion rates measured on samples from the same lithostratigraphic unit revealed variability, most notably (for the given dataset) within the deposits of the Lower Freshwater Molasse. Nevertheless, as expected, poorly lithified sandstones or mudstones correspond to the highest erosion rates and lowest tensile or compressive strength, whereas crystalline rocks such as granite tend to show the lowest erosion rates and highest tensile or compressive strength (Fig. 3-11).

⁹ High erodibility corresponds to low rock resistance against erosion, and vice versa.

An impressive example of the effect of variable erodibility on a river's longitudinal profile is the Rhine Falls knickpoint that is located within resistant limestones. However, besides the lithological contrast, the present-day incision efficiency needed to erode this knickpoint seems to have been hampered by trapping of the Alpine sediments (i.e. the tools for efficient bedrock erosion) in large glacially overdeepened lakes, e.g. the Bodensee. This supports the notion that glacial/interglacial cyclicity (Sections 2.2 and 3.3.1) exerts a major control on the tool and cover effects and thus on the sediment transfer and fluvial dynamics of Northern Switzerland.

According to the variety of assumed resistance against erosion, the rock types of Northern Switzerland were sorted into four erodibility classes (EC). This was implemented based on lithological considerations and the lithostratigraphic units differentiated in the 3D geological model "Nagra Regionalmodell 2012" (Gmünder et al. 2013) that is used for erodibility-related parametrisation in Model B (Fig. 3-3 and Section 3.4). The classes comprise EC1 = Quaternary unconsolidated sediments, EC2 = Bedrock Class "soft", EC3 = Bedrock Class "medium», and EC4 = Bedrock Class "hard" (compare Fig. 3-11). It is generally assumed that the classification of erodibility follows the trend $EC1 > EC2 > EC3 > EC4$. This allows distributions to be assigned for the K values (erodibility coefficient, Fig. 3-13) needed for modelling fluvial incision and later local topography (Section 3.4) using the SPIM equations.

To constrain the parameter ranges, the geological map of Northern Switzerland 1:100'000 (Isler et al. 1984) was first used to map spatially variable values of erodibility coefficient K (depending on rock type) on the present-day topography, focusing on the Jura Ost siting region and surrounding areas (Graf 2022). This first-order assessment was done by extracting ranges of apparent K values from the present-day landscape based on drainage area and gradient along the channel network. Tab. 3-2 summarises the results for selected quantiles and lithological groups (as a basis for subsequent expert elicitation).

Tab. 3-2: Quantiles of (apparent) K values for different lithology groups in the Jura Ost siting region and surrounding area

Group 1: Quaternary; Group 2: Palaeogene and Neogene sedimentary rocks; Group 3: weak Mesozoic sedimentary rocks 1; Group 4: intermediate Mesozoic sedimentary rocks 2 (+ Palaeozoic sedimentary rocks); Group 5: hard Mesozoic sedimentary rocks 3; Group 6: Crystalline basement. All values in this table are $\times 10^{-6}$. See text and Graf (2022) for further details.

Percentile	G1	G2	G3	G4	G5	G6
5%	2.23	2.65	2.21	2.00	1.96	0.99
25%	3.14	3.42	2.91	2.99	2.62	2.44
50%	3.87	4.47	3.45	3.91	3.13	3.13
75%	4.79	5.09	4.03	4.85	4.25	4.31
95%	7.53	7.54	5.42	6.87	5.52	10.10

Lithological mapping along the Hochrhein (Fig. 3-12) as used for Model A is less distinct than within the 3D geological model by (Gmünder et al. 2013). Here, rock types of EC2 and EC3 were combined into one class (here ECb) of weak to intermediate resistance to erosion. Accordingly, elicited distributions of ECb were also derived by taking the average of EC2 and EC3 (Fig. 3-13).

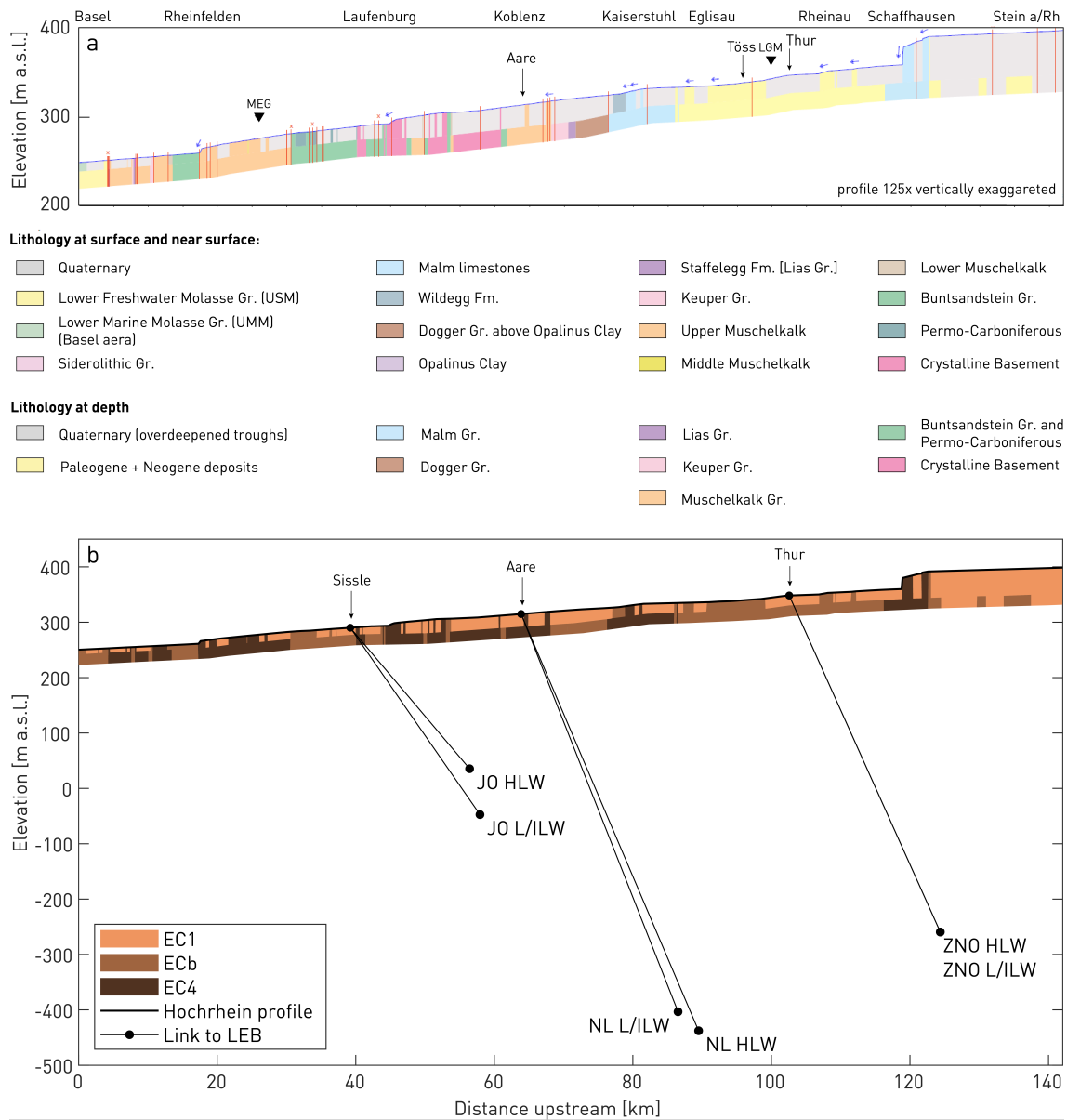


Fig. 3-12: Profile of the Hochrhein drawn with lithostratigraphic units and assumed erodibility classes (EC) as used in the SPIM (Model A)

The original longitudinal profile with lithostratigraphic units from Pietsch & Jordan (2014) is shown in (a) (modified layout). The assignment of erodibility classes is shown in (b), also indicating the position of the provisional disposal areas for both HLW and L/ILW in each siting region (projected according to their easting coordinates). Gr.: Group, Fm.: Formation.

The uncertainties of the K values assigned to the lithological units of the 3D geological model were assessed independently as subjective probabilities following the SHELF protocol (Appendix B.2). Importantly, an uplift value¹⁰ of 0.1 mm/yr was spatially uniformly applied to the area used for the K mapping presented in Tab. 3-2 (Graf 2022). Simplified, this means it was assumed that the present-day landscape has evolved under a background rock uplift rate of 0.1 mm/yr. Under the previous SPIM assumption of $n = 1$ ($m = 0.45$), uplift and K values scale linearly. Accordingly, the elicited distributions for the K values of the four erodibility classes included ranges that were scaled to include variability in the uplift value. These ranges are consistent with the previously elicited one million years RIO distribution for uplift at Brugg (UBR , see above). Although this distribution itself is elicited for the future one million years, especially the likely part of the distribution is informed by observations from the past.

Given the limited additional evidence and the inherent challenges in deriving K values, the elicitation process was approached with a practical and straightforward procedure. First, the general assumption was that the median value of each erodibility class, which was derived by the landscape inversion (Tab. 3-2) reflects the most robust K value of the respective class. The uncertainty distribution around this value should then reflect the scatter (excluding artefacts originating from low gradients at plateau areas or within flood plains). Transient effects from spatially non-uniform uplift or incision within the landscape and the uncertainty caused by the uplift assumption of 0.1 mm/yr should also be accounted for (see above).

Correlation of EC1, ECb and EC4 with UBR

As outlined above, the elicited distributions for the K values of the four erodibility classes included ranges that were scaled to account for the uncertainty regarding the past rock uplift rates (assumed to equal the past incision rates in this specific application). That is, if the past uplift rates were higher than the 0.1 mm/yr assumed above, then the K values in Tab. 3-2 (used as reference values in the elicitation process) should be increased proportionally, and vice versa.

For applying these elicited K values to the future, it is additionally assumed that the past and future uplift rates may positively correlate to a certain degree. That is, a higher uplift rate in the past is more likely to imply a higher uplift rate in the future, and vice versa. Consequently, the past uplift rate correlates positively with both the elicited K values and the future uplift rate which also implies a positive correlation between the latter two parameters.

For the sampling process used to model the future evolution of the Hochrhein in Model A (and subsequent modelling of local topography in Model B, see next paragraph), the K values of EC1, ECb and EC4 are therefore correlated with UBR . Conditional on the value of UBR , a Nataf transformation and a correlation coefficient of 0.9 are used. Even if EC1, ECb, and EC4 are independent by themselves, conditioning them on a fixed UBR value creates a strong positive dependence between them in the overall (unconditional) case. This happens because they are all linked to UBR .

Extraction of EC2 and EC3 from ECb

Model A (SPIM) utilises the three erosion classes EC1, ECb and EC4. Contrary to this, in model B (local topography) the K values of EC1, EC2, EC3 and EC4 are used, where the values of EC2 and EC3 are linked to the value of ECb. In the probabilistic model it is assumed that EC2, EC3 and ECb are fully dependent, which results in a one-to-one mapping for evaluating EC2 from ECb and for evaluating EC3 from ECb.

¹⁰ Note that in this specific application, it is assumed that uplift is equalled by incision, and vice versa.

Uncertainty distribution for erodibility class 4 (EC4)

EC4 is represented by the Groups 5 and 6 (i.e. hard Mesozoic sedimentary rocks and crystalline basement) of the landscape inversion (Tab. 3-2). Accordingly, the distribution is centred around the median of 3.1×10^{-6} .

The lower and upper bounds were chosen to respect the linear uplift scaling using the expected range of the uncertainty distribution of the parameter *UBR* (one million years), respectively (Fig. 3-8b). Because the tails of the final curve (gamma distribution) even exceed these values, it was agreed upon that the distribution using these endmembers is broad enough. Within this range, it was considered likely that the *K* values of EC4 can vary between 1×10^{-6} and $\sim 7.5 \times 10^{-6}$. This includes the average of the 5% percentiles of *K* values for group 5 and 6 (Tab. 3-2) and more likely uplift scenarios for Brugg. However, it was decided to additionally increase the width of the distribution beyond these values: the scaling should allow an uplift value of 0.25 mm/yr (i.e. 250 m uplift/incision in one million years). The most likely *K* values (inner 1/3rd of the distribution) should include the variability in inversion-derived *K* values, which is approximately expressed by the 25 – 75% inversion percentiles of Groups 5 and 6 (Tab. 3-2). It was also intended to include the median of *UBR*, which was adapted during curve fitting to be able to fix the median of the *K* value instead. The final distribution is shown in Fig. 3-13 with the fitting parameters listed in Tab. 3-3.

Uncertainty distribution for erodibility class 3 (EC3)

EC3 corresponds to a medium erodibility and includes mixed rock types (e.g. marls with limestone beds, sandstones, etc.). Accordingly, it is assumed that percentiles of the uncertainty distribution should be located between those of EC2 and EC4 with a general tendency to be closer to EC2. Thus, the distribution is centred around the median of 3.6×10^{-6} .

The large lithological heterogeneity of this class of "mixed rock types" should be reflected in a generally broader range of the distribution (than for instance for EC2). The final distribution is shown in Fig. 3-13 with the fitting parameters listed in Tab. 3-3.

Uncertainty distribution for erodibility class 2 (EC2)

EC2 is represented by groups 2 and 3 (i.e. Palaeogene and Neogene sedimentary rocks and weak Mesozoic sedimentary rocks) of the landscape inversion (Tab. 3-2). Accordingly, the distribution is centred around the median of 3.9×10^{-6} .

Analogue to the procedure in elicitation of EC4, the lower and upper bounds were chosen to respect the linear uplift scaling using the expected range of *UBR* (one million years), respectively. Because the tails of the final curve (log-normal distribution) even exceed these values, it was agreed upon that the distribution is broad enough. Within this range it was considered likely that the *K* values of EC2 can vary between 1.7×10^{-6} and $\sim 1 \times 10^{-5}$. This includes the average of the 5% percentiles of *K* values for group 2 and 3 (Tab. 3-2) and an uplift scaling to account for values of 0.25 mm/yr. The most likely *K* values (inner 1/3rd of the distribution) should include the variability, approximately expressed by the 25 – 75% percentiles of groups 2 and 3 (Tab. 3-2). It was also intended to include the median of *UBR*, which was adapted during curve fitting to enable fixing the median of the *K* value instead. The final distribution is shown in Fig. 3-13 with the fitting parameters listed in Tab. 3-3.

Uncertainty distribution for erodibility class 1 (EC1)

EC1 corresponds to unconsolidated, highly heterogeneous Quaternary sediment and is represented by Group 1 of the landscape inversion (Tab. 3-2). It is assumed that, in general, these sediments are easier to erode. Underlying bedrock in landscape sections covered by Quaternary

sediments may influence the K -inversion results. This may be especially the case within the tributaries, where the cover of unconsolidated Quaternary sediment might be thinner. In Northern Switzerland it is frequently observed that well consolidated Quaternary moraines may have a similar resistance to erosion as some variants of weathered Molasse sedimentary rocks (within EC2). For this reason, the left side of the uncertainty distribution was chosen to be close to that of EC2. In contrast, more weight to higher erodibility (than in EC2) and acknowledging the strong heterogeneity should be reflected by a wider right side of the uncertainty distribution.

To follow the general rule of $EC1 > EC2 > EC3 > EC4$ and to allow for a broader range, the distribution is centred around the median of 4.2×10^{-6} . Analogous to the procedure in elicitation of EC4 and following the above assumptions, the lower and upper bounds were chosen to respect the linear uplift scaling using the expected range of UBR (one million years). Because the tails of the final curve (log-normal distribution) even exceed these values, it was accepted that the distribution is wide enough. The final distribution is shown in Fig. 3-13 with the fitting parameters listed in Tab. 3-3.

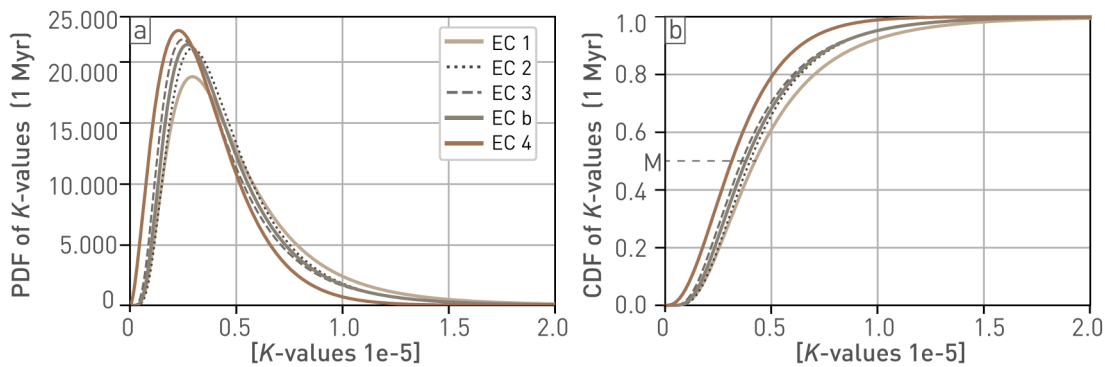


Fig. 3-13: Total assumed probabilities of K values for the erodibility classes (EC)

EC were assigned based on the stratigraphic column. Note that they are also used for Model B (Section 3.4). For Model B, EC1 and EC4 are equivalent to EC1 and EC4 in Section 3.4, respectively. ECb was deduced from the distributions of EC2 and EC3 ($ECb = 0.5 \cdot EC2 + 0.5 \cdot EC3$) (a) Probability densities of elicited K values for beforehand assigned erodibility classes (b) Cumulative distribution function of elicited K values for previously assigned erodibility classes, with M marking median values. All distributions but EC4 are represented by log-normal distributions. EC4 is characterised by a gamma distribution (Tab. 3-3).

Tab. 3-3: Distribution parameters for the elicited uncertainty of erodibility classes

EC	Distribution Type	Mean	Standard Deviation
EC4	Gamma	3.5×10^{-6}	2.1×10^{-6}
EC3	log-normal	4.3×10^{-6}	2.9×10^{-6}
EC2	log-normal	4.6×10^{-6}	2.8×10^{-6}
EC1	log-normal	5.1×10^{-6}	3.4×10^{-6}
ECb	log-normal	4.5×10^{-6}	2.8×10^{-6}

3.3.3.4 Selected scenarios for changes in drainage area

It is expected that the Aare – Rhine River system will to some degree undergo drainage divide migration. Currently, the Rhine grows at the expense of the Danube system (Section 2.4.2.5); an impressive example is the stream capture of the Wutach, which occurred in the Late Pleistocene. In the Alpenrhein and the Bodensee area, the divide migrates towards the east. In the west, the River Rhone constitutes the aggressor at the expense of the Aare – Rhine (Fig. 2-34). Based on these observations, four respective scenarios were considered, and their probabilities of occurrence elicited for the two simulation periods of 100 kyr and one million yr (Fig. 3-14): the diversion of the Aare towards Lac Léman¹¹ (Aare), the integration of the Iller River into the Rhine system (Iller), the diversion of the Rhine at Sargans (Sargans), and a continuous integration of the Upper Danube via karst sinkholes into the Rhine system (Upper Danube) (Fig. 3-14a). In general, only the integration of Iller and Upper Danube would very slightly increase incision with respect to the present-day drainage. The other two scenarios decrease incision.

All scenarios are most likely connected with ice advance. Because large glaciations are not likely within the next 100 kyr (see also Section 3.5.3.5), all but the Sargans scenario (which was assumed nearly similarly likely at both times) were considered more likely during the next one million years.

The Aare diversion scenario requires the Aare to flow through the Bielersee into Lac Léman. The threshold is considerably low. Headward erosion might capture the Upper Aare River, e.g. in the area west of Yverdon (Fig. 2-34a, see also Nagra 2014b, Dossier III). Such a scenario could cause a baselevel drop with further headward erosion because the Rhone River with ~ 800 km length is significantly shorter than the Rhine River. However, it is not expected that headward erosion would reach the Hochrhein area within the period under consideration. Instead, it is expected that in such a scenario of Aare River diversion, the Reuss and Limmat Rivers would continue to drain towards the Rhine River. In general, this scenario was not considered likely for the next 100 kyr, but similarly likely as the other scenarios during one million years.

The Iller capture and diversion through the Bodensee is another scenario that would add catchment area to the Rhine. It was considered significantly less likely during the shorter time span.

The diversion of the Alpine Rhine section through Sargans was considered most likely from these scenarios for both time spans. The discharge required to overflow the current Alpine Rhine channel has already been reached during a high-magnitude flood in AD 1480, when the Rhine River overflowed into its old channel across Walensee and Zürichsee into the Limmat (Amt für Wald Graubünden 2009). Flooding in AD 1846 has nearly missed this overflow by only about 2 m (Knäble 1970). Finally, the integration of the Upper Danube was assumed rather likely.

The RIO-type probabilities for the occurrence of the drainage scenarios represent the calculated mean of the individual elicitations (Fig. 3-14b). The scenarios are modelled as irreversible within the respective time spans and incorporated without additional time delay. That means that a drainage scenario which would be given a 25% chance to occur within one million years would be added in one out of four model runs right from the start. The scenarios are independent, and more than one scenario can occur at the same time.

¹¹ Importantly, this scenario considers that the Reuss and Limmat continue to drain into the Rhine.

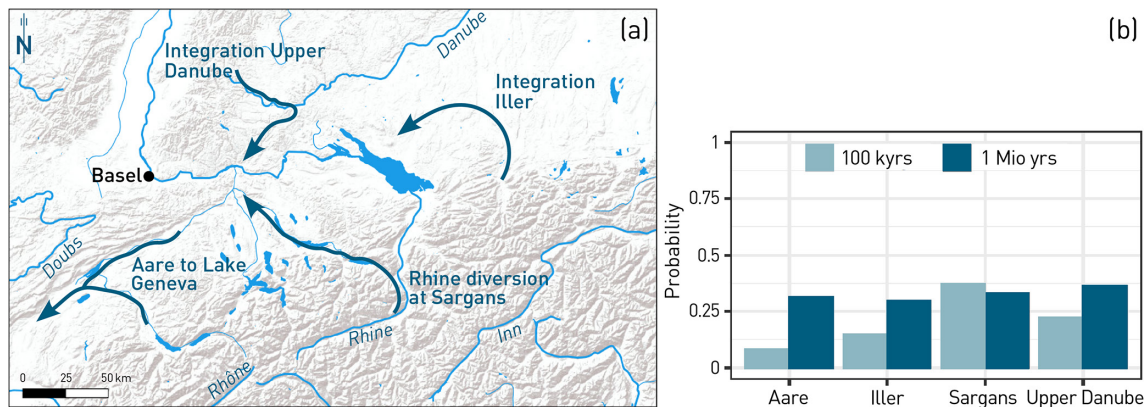


Fig. 3-14: Scenarios for drainage diversion influencing the evolution of Hochrhein incision

(a) Probabilities were elicited for the four scenarios visualised on the map: the diversion of the Aare towards Lac Léman¹² (Aare), the integration of the Iller River into the Rhine system (Iller), the diversion of the Rhine at Sargans (Sargans), and a continuous integration of the Upper Danube into the Rhine system (UpDanube). (b) Ranking of the scenario probability for the time periods of 100'000 and one million years. The RIO-probabilities represent the mean of the individual expert elicitations.

3.3.4 Future incision based on SPIM modelling results

3.3.4.1 General results

Using a range of scenarios that assess possible changes in future discharge and catchment conditions – ranked according to their plausibility – along with parameter variations based on expert knowledge, SPIM was applied in a Monte Carlo environment with one million samples to simulate potential future river profiles between the Bodensee outflow and Basel (i.e. the Hochrhein) over two timeframes: 100'000 years (Fig. 3-15a) and one million years (Fig. 3-15c).

The results reveal how future incision patterns along the Hochrhein are influenced by changes in baselevel. In scenarios where the baselevel rises, river profiles flatten; conversely, a drop in baselevel allows greater incision. The effects of these shifts are particularly pronounced closer to Jura Ost, where baselevel changes most directly impact incision.

After 100'000 years, incision into the overall uplifting rock amounts to most likely values between approximately 10 and 15 metres at the siting regions, but values up to about 55 metres may be expected for JO and NL, and up to about 30 m for ZNO (Fig. 3-15a, Tab. 3-4, see also Section 3.2.3 and Fig. 3-5 for introduction in terminology "most likely" and "expected"). After one million years, such incision may be between approximately 150 and 170 m (most likely) and may be expected to be up to approximately 300 to 390 m (Fig. 3-15c, Tab. 3-4). In general, the incision values are highest in NL.

In addition to changes in the baselevel of the river profiles, the simulations account for uncertainty in rock uplift that may affect the repository sites in the region. Based on this uncertainty, expected overburden thickness above repository sites relative to the local erosion base can be estimated (Fig. 3-15b and d for 100'000 years and one million years, respectively).

¹² Importantly, this scenario considers that the Reuss and Limmat continue to drain into the Rhine.

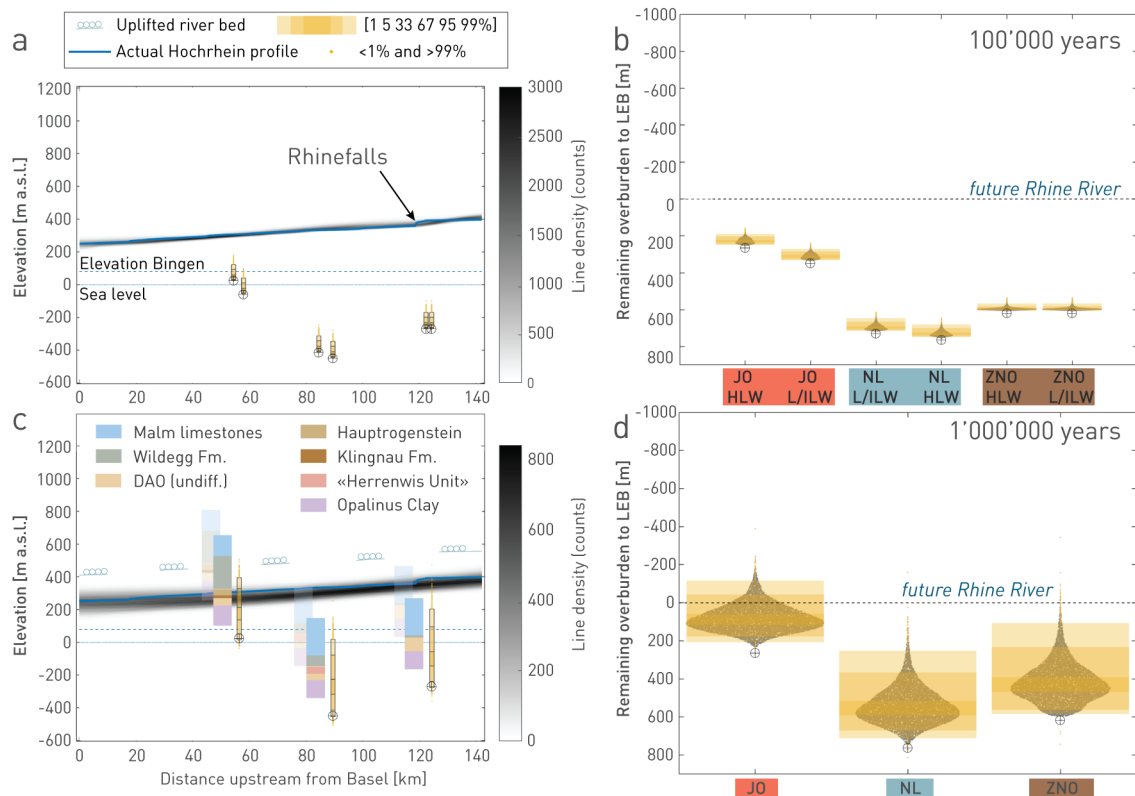


Fig. 3-15: Results of fluvial incision modelling using the Stream Power Incision Model (SPIM)

The left panel shows the realisations of future longitudinal Hochrhein profiles in grey for 100'000 years (a) and for one million years (c). The darker the lines appear the more likely such an incision depth is. Also shown are the present-day river profile (in blue) and the elevations of Bingen and sea level as reference baselevels. The circles correspond to the depth of the provisional disposal areas at the time of emplacement. The orange boxplots describe the uncertainty distributions of the rock uplift that the disposal areas might undergo in these times. As an example, the stratigraphic columns at the locations of the provisional disposal areas are uplifted according to the median and the 95th percentile (transparent) – highlighting which units might be incised after one million years, if the Hochrhein or another main river were to flow directly across the disposal area by that time. DAO: Dogger Group above Opalinus Clay. Also shown are symbolised gravel deposits assuming they were also uplifted uniformly according to the median uplift value (and subsequently incised equally along the profile). As such, they represent the present-day riverbed that becomes remnants of a future terrace surface. The remaining overburden thickness above the repository with respect to the local erosion base is shown in the right panel for 100'000 years (b) and one million years (d). Note that these distributions are calculated for downstream positions along the Hochrhein, respectively (i.e. the Sissle confluence for JO, the Aare confluence for NL, and the Thur confluence for ZNO; see Fig. 3-12). In these plots, the boxes correspond to specific quantiles (see legend) and the violins reflect the distributions, such that the range of the violins extends past the extreme data points. The bulge corresponds to the most likely values. The violin shape represents the density of data points at various values along an axis.

After 100'000 years, the remaining overburden thickness with respect to the local erosion base is most likely to be > 230 m in JO (> 200 m expected), > 730 m in NL (~ 700 m expected) and approximately 600 m in ZNO (580 m expected) (Fig. 3-15b, Tab. C-2). After one million years, the remaining overburden thickness with respect to the local erosion base (LEB) is most likely > 90 m in JO. However, it is within the expected range that the local erosion base is up to about 45 m deeper than the future elevation of the provisional disposal area (Fig. 3-15d, Tab. C-2). It is most likely that nearly 600 m of overburden with respect to the LEB remain in NL (expected ~ 370 m) and approximately 450 m (~ 230 m expected) in ZNO (Fig. 3-15d, Tab. C-2).

Notably, SPIM simulations suggest that fluvial incision alone is unlikely to remove significant overburden to compromise the barrier function in the Zürich Nordost and Nördlich Lägern regions. However, a number of scenarios indicate that the Jura Ost repository may be uplifted to, or even above, the future local erosion base (defined by the Sissle outlet elevation) after one million years. This would require Jura Ost to rely on the preservation of the overlying topography to maintain its safety margin. Further analysis of the evolution and preservation of this local topography is presented in Section 3.4.

3.3.4.2 Site-specific differences of future incision

The varying upstream distances of the siting regions along the Hochrhein profile, as well as the different depths of the disposal areas below local erosion base, influence the future evolution of the cover with respect to the local erosion base (LEB). Modelled incision varies least at the location of NL and most at ZNO.

All the siting regions are well below the local erosion base after 100'000 years of incision (Fig. 3-15a and b).

After one million years, the simulations suggest that disposal areas of all the siting regions are well below the local erosion base considering the most likely scenarios (mode). However, as visible from Fig. 3-15d, a range of scenarios (> 10%) will lift the disposal area of JO even above the LEB. After one million years, more than 50% of the simulations reveal less than 100 m remaining cover to the level of incising main rivers (Tab. C-2). Accordingly, for these cases the site JO does depend on the preservation of the additional cover given by the local topography (Section 3.4).

ZNO and NL do not depend on their cover of local topography. 95% of the simulations for one million years show more than 200 m of remaining coverage to the LEB in ZNO and more than 350 m in NL.

3.3.5 Sensitivity of incision model parameters and plausibility of the results

3.3.5.1 Sensitivity of incision model parameters

Fig. 3-16 shows the sensitivity of the elicited SPIM parameters with respect to the remaining vertical distance of the repository sites to the local erosion base (LEB). The analysis is conducted for a time span of 100'000 years (Fig. 3-16a) and one million years (Fig. 3-16b) and for different locations along the Hochrhein reflecting the three siting regions (i.e. the Sissle outlet for JO, the Aare outlet for NL and the Thur outlet for ZNO). The relative importance (or sensitivity) of the SPIM parameters depend on the simulated time span and differ for each repository site.

In tectonically active areas, rock uplift is the main driver for fluvial incision, which is reflected by the high sensitivity to parameters that control spatial patterns of uplift. In particular, over long timescales, uplift solely determines the reduction in vertical cover of a repository site, provided that uplift is compensated by denudation of the surface above.

Over time periods of 100'000 years, the modelled remaining vertical distance of the repositories to the LEB is sensitive to variations in erodibility. The parameter K , which characterises the resistance of different lithological groups (erodibility classes ECs) to erosion, controls incision in particular at the onset of simulations. The beginning of simulations involves a transient adjustment of the initial river profile during which existing knickpoints or –zones start adjusting to an equilibrium profile. This adjustment can lead to rapid changes in the river profile and, thus, the elevation of the LEB, and the rate of these changes is governed by the erodibility parameter.

Changes in baselevel constitute the downstream boundary conditions of the SPIM. Any changes in baselevel will be transmitted to the river profile at a pace controlled by bedrock erodibility. If erodibility is low, changes in baselevel will eventually not propagate as far upstream to affect upstream repository sites. It is thus understandable that JO, the repository closest to Basel, reacts most sensitively to changes in baselevel and associated steepening (or flattening) of the river profile.

Generally, scenarios of drainage reorganisation insignificantly affect simulated changes in repository depth to LEB. Scenarios that simulate the integration of neighbouring catchments or loss of drainage area lead to only minor changes in the incisional dynamics of the Hochrhein. There are two main reasons for the low sensitivity. First, the relative increase or decrease of drainage areas (and thus river discharge) is relatively low. These changes depend on the location along the Hochrhein. In Basel, the drainage area varies between -12% for the deflection of the Aare to 6% for the integration of the Iller. Second, the Hochrhein is not a steep river, so that there is little room for additional incision caused by an increase in discharge.



Fig. 3-16: Sensitivity analysis of the fluvial erosion model

Figure illustrates the relative influence of the SPIM model parameters on estimates of the depth of the repository sites relative to the local erosion base (LEB). The sensitivity analysis is conducted for simulation time spans of (a) 100'000 and (b) one million years as well as for each repository site. The sensitivity analysis is based on a linear regression with normalised parameters (zero mean, unit standard deviation) as input variables. The length of the bars, rescaled to sum to a value of 11 (the total number of parameters), displays the regression coefficients which are a measure of the relative importance of the parameters. The effects of the parameters can be both positive and negative. The figure shows the absolute values of the effects. The analysis shows that the parameter sensitivity differs for each considered site and time span. EC corresponds to the different erodibility classes.

3.3.5.2 Plausibility of the fluvial incision model results

For Nördlich Lägern, the SPIM-based estimates of fluvial incision after one million years are listed in Tab. 6-1. The mode of these estimates corresponds to $\sim 170\text{m}$ at the location of the Aare – Rhine confluence. At this position, the vertical fluvial bedrock incision between the Base Höhere Deckenschotter (HDS) and Tiefere Deckenschotter (TDS) comprises approximately 80 metres. Incision following deposition of the TDS down to the local erosion base comprises approximately 100 metres (Fig. 2-26). A mean TDS age of 0.9 Myr and standard deviation of 0.13 Myr and a mean HDS age of 1.38 Myr with standard deviation of 0.31 Myr (Fig. 2-5, Section 2.1) results in incision rates that range between ~ 110 to 170 m/Myr with respect to HDS to LEB incision and ~ 100 to 130 m/Myr regarding TDS to LEB incision at a similar location. The range of these rates include the mode of the future incision estimates but are closer to the 25th percentile of the estimate (Tab. 6-1).

On the other hand, the incision rates obtained from the Base Deckenschotter elevations estimated at the Aare confluence refer to the lowest points of the basal surface of a contiguous Deckenschotter body. This approach was chosen to best capture the level of a paleo-river flowing on a bedrock surface with a certain paleotopography (Heuberger & Naef 2014). Such an approach may underestimate the uncertainty related to capturing the values of the Base Deckenschotter, which has been characterised by an asymmetric distribution, with higher uncertainties towards lower values (Hammer et al. 2019). An alternative approach calculating Quaternary incision rates along selected valleys and including the uncertainty in Base Deckenschotter elevations are higher with values up to 300 m/Myr , although this corresponds to values at various places in Northern Switzerland. Independent of the approach for extracting the information on the height of the ancient river course, Quaternary incision rates do not directly reflect past uplift rates, as the past incision is, at least partly, the result of transient river profile adjustments to changes in large-scale drainage patterns (drainage reorganisation, Section 2.3), for which there are no indications in the next million years. To this end, it is believed that, based on the comparison with past, long-term incision rates of the Hochrhein, the estimated incision rates are plausible.

Agreeing that the modelled future incision rates (which represent the final results at either 100'000 or one million years) are plausible, allows a few temporal evolution scenarios that may be envisioned within the range of estimates to be examined. The simplest evolution which may be encompassed by the modelling results is for instance a steady low-rate uplift and subsequent incision, similar to the potential incision pathway *a* in Fig. 6-2. Another evolution may show strong early incision, e.g. associated with a prolonged interglacial period and relative sediment starvation, as depicted by incision pathway *b*. Similarly, alternating phases of increased rapid downcutting (a few tens of metres in a few tens of thousands of years) caused by short phases of increased rock uplift related to post-glacial rebound and longer phases of low-downcutting dominated by sediment accumulation and lateral erosion may be expected. After each large future foreland glaciation, the phases of downcutting may be delayed by a few tens of thousands of years until the periglacial lakes are filled and provide the bedload as erosive tools (compare Section 3.3.1). Such an incision pathway is shown in line *d* in Fig. 6-2, although for the more extreme endmember of incision magnitude.

At the high, positive end of the distribution of incision rates (i.e. 95%), values of up to $\sim 385\text{ m}$ occur. Importantly, this implies incision through an uplifting rock column and these extreme values may be connected to parameter realisations that involve additionally rapid baselevel drop. The results of baselevel drop are visible in Fig. 3-15c. The profile lines are shifted with respect to their present-day elevation at Basel at the western downstream position of the Hochrhein by the same amounts that were elicited prior to the modelling (Section 3.3.3.2, Fig. 3-10). Baselevel drop up to 70 m was included in the likely range (95%), corresponding at the maximum to a hypothetical future elevation of the Rhine River at Basel of 170 m a.s.l. (note that it may also include cases of incision into previously accumulated sediment). Such an elevation is still higher

than for instance the present elevations of the cities of Karlsruhe (~ 115 m a.s.l.) or Strassbourg (~140 m a.s.l.), which straddle the Rhine River. Such strong baselevel drop, however, requires a larger-scale equilibration and steepening of the longitudinal Rhine profile (compare Fig. 2-8a) of the Oberrhein upstream of Karlsruhe and the Hochrhein section. For such an extreme scenario to occur, very different sediment dynamics (with long-term sediment starvation) than seen for the Quaternary period is needed in addition to tectonically controlled subsidence. However, increased uplift rates and erosion in Northern Switzerland will rather lead to a sediment pulse reaching the Upper Rhine Graben and a situation with sediment starvation is hard to imagine. As uplift in Northern Switzerland and baselevel evolution are treated as independent model parameters, the positive end of the distribution and particularly the distribution tail may overestimate the total incision (see also discussion in Section 5.2, Fig. 5-4).

A baselevel elevation of 170 m a.s.l. also corresponds to the elevation of the city of Lyon (France) along the Rhone River. Although a drainage scenario in which the Aare River is diverted through the Bielersee into Lac Léman was incorporated (Section 3.3.3), it is expected that in such scenario the Reuss and Limmat Rivers would continue to drain towards the Rhine – a scenario that would inhibit increased incision of the Hochrhein. The strong baselevel drop down to an elevation of 170 m a.s.l. at Basel (and associated strong incision in the Hochrhein) would include the extreme scenario that the Rhone River captures the diverted Aare and Hochrhein Rivers and their tributaries together. The relatively large Aare catchment may bring the potential for incision downstream of Lac Léman, but such a scenario requires in addition a very strong headward incision of the Rhone River, as seems to have been the case after the Messinian Salinity Crisis (e.g. Loget et al. 2006; see also Section 2.4.2.1 about the Doubs-Aare capture). While an incision pulse may propagate quickly given the shorter longitudinal length of the Rhone River compared to the Rhine River, a plausible trigger for such a scenario within the next million years was not identified.

Finally, the SPIM model does not explicitly account for sediment transport and deposition, which may in reality lead to phases of channel bed protection and aggradation (Sklar & Dietrich 2001).

3.4 Evolution of local topography

3.4.1 Introduction to processes that drive evolution of local topography

Channel-hillslope coupling

Fluvial incision and deep glacial erosion processes determine the pace at which hillslopes respond to baselevel changes. Topographic slope stability depends for instance on the slope angle, controlled by the fluvial baselevel and associated mass movements. Thus, the nature of channel-hillslope coupling will drive denudation of the topography (Fig. 3-17). This coupling also provides the link between the landscape compartments that define the *local erosion base* and *local topography* (see Section 3.1).

Rock properties (e.g. rock strength, fracture spacing, orientation of discontinuities with respect to slope), vegetation cover, and climatic factors add to the controlling factors of hillslope stability or failure and can influence denudation rates. In light of these considerations, the distinct morphological features and different geomorphic process rates that define a landscape need to be incorporated into any future assessment of erosion.

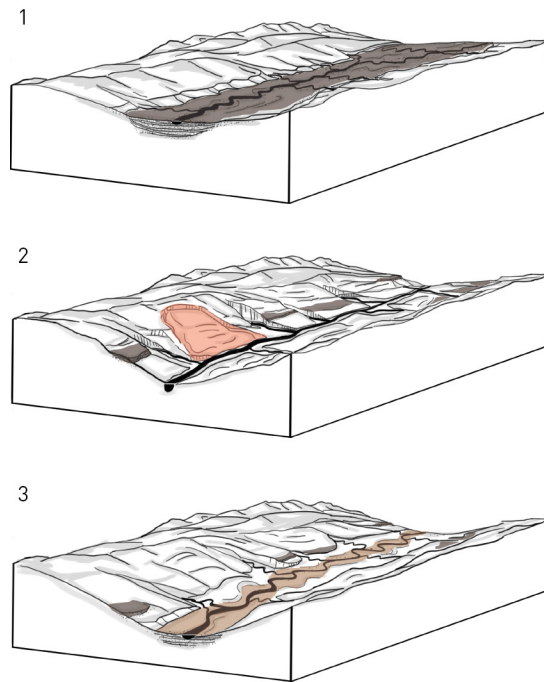


Fig. 3-17: Conceptual figures of local landscape evolution associated with local baselevel fall of a main river

The sequence starts with a mature main river valley characterised by a wide aggraded valley floor (1; dark brown). A pulse of fluvial incision of the main river through the former valley infill and into the bedrock (local baselevel fall) follows in (2). This stage includes the formation of terrace relicts above the former valley floor (dark brown), propagation of the incision signal along tributary streams and hillslope response (e.g. landslides; red). More advanced adjustment to the new local baselevel is indicated in (3), with the formation of a new main valley floor (light brown) and a general widening of the valley shape by continuing local erosion. Drawings by A. Voigtländer.

Local topography in the greater Bözberg area of Northern Switzerland – a qualitative and quantitative comparison with the Wutach area of the Swabian Alb

The majority of the terrain in Northern Switzerland is characterised by a low-relief cuesta landscape of alternating highly erodible and less erodible strata that are slightly inclined (see Fig. 3-18 and Fig. 3-20a as examples). Rather wide fluvial channels and extended terrace systems define the postglacial terrain of Northern Switzerland. While both Höhere and Tiefer Deckenschotter deposits are mostly preserved in form of isolated relicts, the Hochterrasse and Niederterrasse deposits and landforms follow a continuous and traceable fluvial network (Fig. 2-29). Thus, landscape rejuvenation times based on this observation might be in the order of a few hundred thousand years. However, deviations from these general patterns of a distinct long-lived fluvial network exist on various scales. Small-scale lateral channels (breakthrough channels) for instance are observed throughout Northern Switzerland. They include temporary features (e.g. Wangetel between Jestetten and Wilchingen or Herblingertal between Thayngen and Schaffhausen) or persistent channels (e.g. Rhine Valley between Rüdlingen and Tössegg). Most of these channels developed in association with large foreland glaciations, either by glacial and postglacial impoundment of a present drainage and subsequent overflow or by ice-marginal drainage (e.g. Keller & Krayss 2000). Subsequent formation of new waterways and cutting through bedrock relief can result in the formation of new main valleys as illustrated in Fig. 3-19. Alternatively, the

new valley formation may also be facilitated by direct (sub)glacial erosion. These processes are more likely to occur in low-relief landscapes. However, even in areas of higher local relief, corresponding scenarios cannot be excluded over long time periods. Such local changes in the drainage can have profound influence on the adjusting landscape.

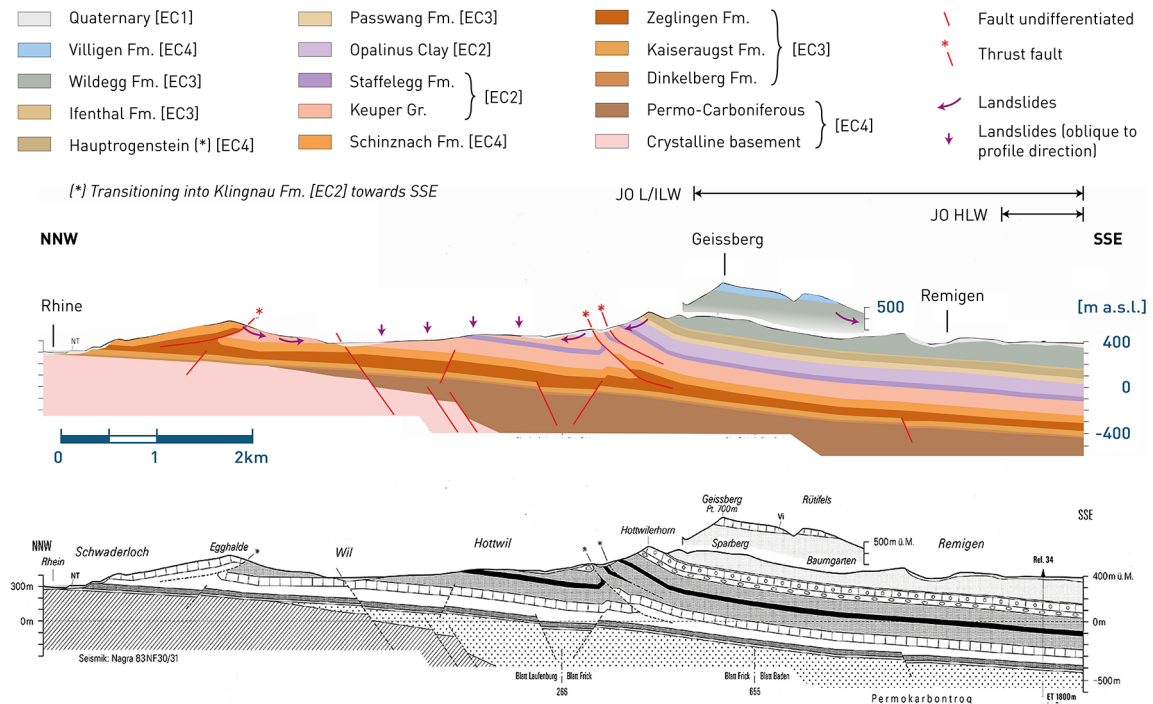


Fig. 3-18: Layered stratigraphy cross-section and spatial pattern of landsliding in Jura Ost and surrounding area

The lower figure shows the original geological cross-section by Diebold et al. (2006) with a vertical exaggeration of 1.2. The same cross-section is displayed above, with the colour scheme of lithostratigraphic units and assignment of erodibility classes (EC) according to Fig. 3-11. The profile highlights the alternation of inclined erosion-resistant (e.g. Villigen Fm.) and more easily erodible units (e.g. Opalinus Clay) of the External and Tabular Jura. The black horizontal arrows above the profile indicate the spatial intersection with the JO L/ILW and HLW siting region perimeters. The distribution of landsliding (adopted from Ludwig 2018) may be regarded as a rough proxy for the spatial variability of (hillslope) erosion rates. This setting suggests i.) different rates of fluvial incision within and across local tributary catchments, and ii.) varying susceptibility of lithological units to hillslope processes, with landsliding being typically associated with weak rock types (e.g. claystones, marls) of EC2 and partly EC3. Gr.: Group, Fm.: Formation.

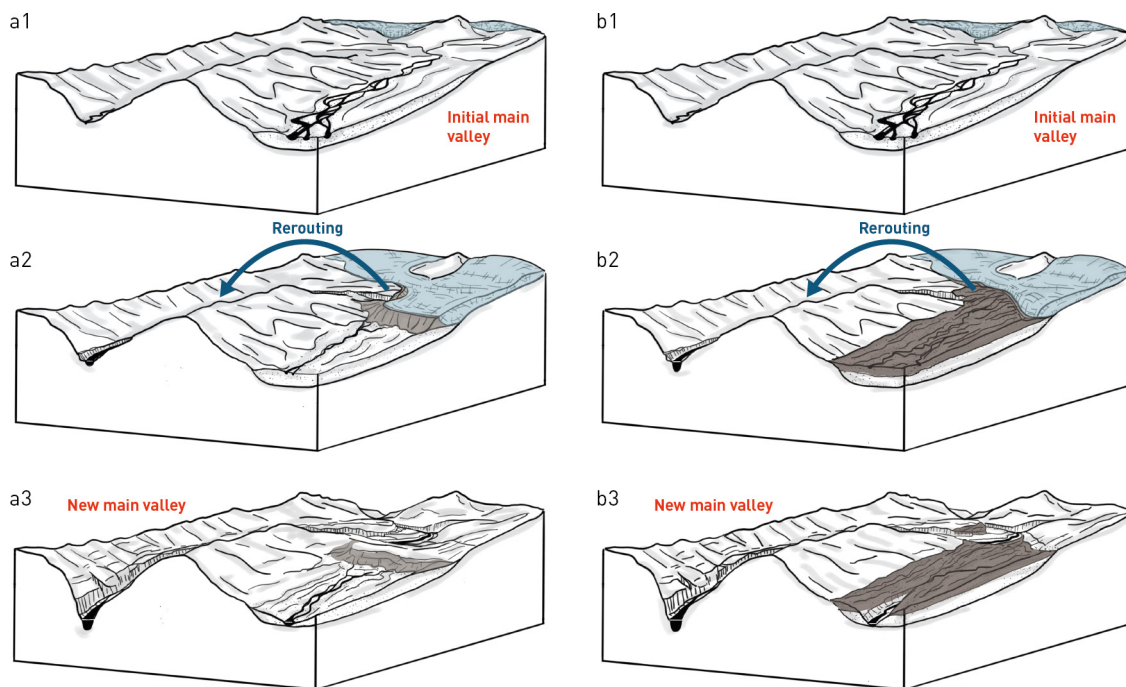


Fig. 3-19: Conceptual figures of local landscape evolution associated with the formation of a breakthrough channel in an ice-marginal setting

The figure distinguishes two alternative processes that cause the formation of a breakthrough channel: blocking of the drainage by a moraine in sequence (a), and aggradation of the valley floor in sequence (b). Both sequences start with a pro-glacial main valley (a1 and b1; glacier shown in light blue). Glacial advance in (a2) causes blocking of the former main valley by a terminal moraine (greyish brown), and subsequent rerouting of the ice-marginal main drainage into an adjacent valley. Alternatively, glacial advance in (b2) with increased pro-glacial sediment supply raises the valley floor by aggradation (dark brown), until it reaches overspilling and subsequent rerouting of the ice-marginal main drainage into an adjacent valley. For both sequences, the new drainage erodes the former ridge at the location of over-spill, thereby creating a breakthrough channel that eventually evolves into a new main river valley after glacial retreat (a3 and b3). Drawings by A. Voigtländer.

Holocene geomorphic process rates (e.g. landsliding, overall denudation processes) in the Bözberg area of the JO siting region are low based on a morphological landscape assessment (Fig. 3-20a and c; Ludwig 2018). Present-day terrain elevation within the extended study area ranges from around 700 m a.s.l. at the highest hilltops of the External Jura (Schinberg, Geissberg; see also Fig. 3-18) to 300 m a.s.l. at the outlet of the Sissle River into the Rhine. Given this comparably high local relief, hillslope processes play a relevant role in shaping the landscape of the Jura Ost siting region and its surrounding areas. For instance, historic landslides in the study area have been documented by Baltzer (1876), Hartmann (1928, 1950) and Haefeli (1940). Fig. 3-20c shows that, in general, the central part of the siting region appears inactive with respect to landsliding. A large proportion of this region consists of the "upland low-relief area" of the "Bözbergplateau" and "presumably unfailed hillslopes", associated with the lithostratigraphic domains of Palaeogene and Neogene sedimentary rocks (Molasse), Villigen Formation and Wildeggen Formation (compare Fig. 3-11). Conversely, landslides within the mapping area appear to be clustered with respect to outcrops of Palaeogene and Neogene sedimentary rocks (Molasse) along the Aare River, i.e. the southern margin of the Bözbergplateau (southeastern corner of siting region) and within various lithostratigraphic units along the Mandach Thrust north of the siting region, from Frickberg in the west to Böttstein next to the Aare in the east (Fig. 3-20c).

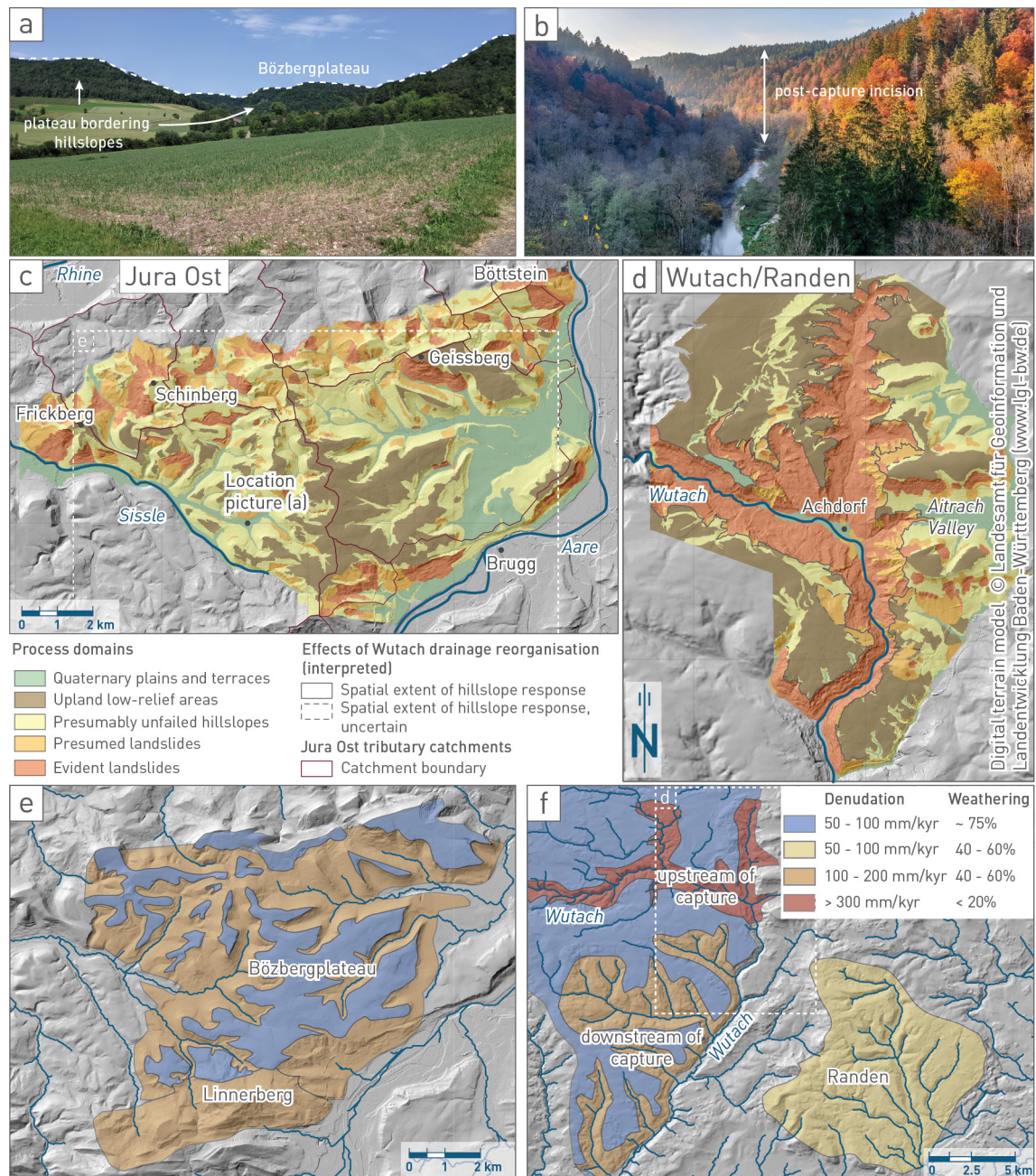


Fig. 3-20: Comparison of a geomorphically quiet (JO) with a highly active landscape (Wutach)

Comparison of the study areas. (a) Field view of a valley carved into the Bözbergplateau. Note the relatively straight mountain tops of the plateau. (b) Field view of the narrow Wutach gorge cut into Muschelkalk carbonates after the river capture event ~18'000 years ago. (c) Map of (hillslope) process domains for the JO study area (Ludwig 2018; see details there for definition and criteria of individual process classes). (d) Map of (hillslope) process domains for the Wutach catchment around Achdorf (digital elevation model: LGL 2010). (e) Simplified synthesis of interpreted denudation rates based on basin-wide and hillslope sampling of cosmogenic nuclides of the Bözberg area. (f) Interpreted denudation rates of the Wutach and Randen area. The colours also indicate the approximate percentage of weathering contributing to the total denudation rate. (Fig. 3-20a, b, e, and f are courtesy of R. Ott). See Fig. 3-21 for the database of (e) and (f).

A prominent example of a larger-scale landscape adjustment to drainage changes is represented by the Wutach gorge in the Swabian Alb (Fig. 3-20b and d), where a stream capture by a tributary of the Rhine River ca. 18 kyr ago caused a significant baselevel drop (Einsele & Ricken 1995, Hebestreit 1995) (as introduced in Sections 2.4.1.4 and 2.4.2.4). From its headwaters at Feldberg (the highest elevation of the Black Forest, 1'493 m a.s.l.), the Wutach flows east until reaching the location of the Late Pleistocene stream deflection near the village of Achdorf. Prior to drainage reorganisation, the former Donau-Wutach¹³ here continued its eastward course along the present-day Aitrach Valley (Fig. 3-20d) towards the Danube. This changed when the Donau-Wutach was captured by the Rhine basin from the south, as a result of headward erosion by a pre-existing Rhine tributary, referred to as the Rhine-Wutach.

Alternatively, coeval aggradation by the former Donau-Wutach leading to an overflow across the drainage divide has been presented as an additional explanation for the stream deflection, at least as a final trigger mechanism (Hebestreit 1995). Such aggradation may have been favoured by glacially enhanced sediment flux from the former Feldberg Glacier that occupied the Donau-Wutach headwater region during the last glaciation. As determined by the course of the former Rhine-Wutach, the present-day Wutach is directed towards the south / southwest at Achdorf (Fig. 3-20d), until finally joining the Rhine to the south at an elevation of ~315 m a.s.l. This elevation of the Rhine corresponds to a much lower local erosion base than was set by the Danube prior to drainage reorganisation.

The adjustment of the drainage system to this lower local baselevel is expressed by the remarkable incision of the Wutach stream, which reached its present-day level by an incision of up to 150 m within only 6 kyr (Einsele & Ricken 1995). Both upstream and downstream from this point, the incision of the Wutach has produced deep valleys and gorges in the present-day landscape. The adjustment of local topography to the local baselevel fall is further expressed by an attenuated, but still ongoing hillslope response by pronounced landsliding along these incised valleys (Fig. 3-20b, d). Similar to these effects caused by the Wutach capture, such increases of local incision and denudation might be associated with the formation of breakthrough channels described earlier in this section.

In the following paragraphs, the transient landscape around the Wutach gorge and the neighbouring Randen area is briefly compared with the landscape of the Bözberg (JO siting region), using denudation and chemical weathering rates based on cosmogenic radionuclides (¹⁰Be, ³⁶Cl; Section 2.4.1.3), alongside weathering rates from stream water chemistry (Fig. 3-20e, f and Fig. 3-21). Both areas share a comparable stratigraphy with alternations of slightly inclined erosion-resistant (e.g. limestones) and more easily erodible units. A total of 26 samples of ³⁶Cl and 11 samples of ¹⁰Be concentrations for catchment-average and hillslope¹⁴ denudation rates in calcite and quartz were measured, respectively.

Catchment-wide denudation rates in mixed landscapes, consisting of both soluble (carbonate rocks) and insoluble (felsic rocks) target minerals can be biased under different weathering scenarios (Ott et al. 2022). This bias can be identified and corrected using paired nuclide measurements of minerals with different solubility coupled with knowledge of either the bedrock or regolith mineralogy to derive denudation and long-term weathering rates. Similarly, single-nuclide measurements on soluble or insoluble minerals can be corrected to determine denudation rates if a weathering rate and compositional data are available. For the Wutach, Jura Ost and Randen regions, which presumably represent an increasing state of landscape maturity and hence generally lower, landscape lowering rates, this was accomplished by measuring ¹⁰Be and ³⁶Cl in

¹³ The names "Donau-Wutach" and "Rhine-Wutach" are adopted from Einsele & Ricken (1995). Donau = Danube.

¹⁴ In the Bözberg region, two ³⁶Cl samples of amalgamated colluvium from the bottom of two hillslopes were collected for comparison with nearby streams.

various catchments exposing homogeneous and inhomogeneous distributions of carbonate and felsic rock units (compare Fig. 3-21). Based on these measurements, and by considering the reliability of individual samples, a map of interpreted denudation rates is derived in Fig. 3-20e and f.

Catchment-wide denudation rates indicate the lowest rates in the low-relief areas of the Bözbergplateau and the incised carbonate valleys of the Randen area (0.05 – 0.1 mm/yr; Fig. 3-20e and f), similar to what has been concluded earlier by analogy with surrounding regions (Schnellmann et al. 2014). In these areas, weathering accounts for ~75% and 50% of the denudation, respectively (based on measurements of denudation and weathering rates outlined in paragraph above). Along the flanks of the Bözbergplateau, where hillslope channel processes are more active, these rates double and are within the range of long-term incision of the Rhine and Aare (Fig. 3-20e). Erosion and weathering each contribute ~50%. Highest rates (0.3 – 0.5 mm/yr with < 20% weathering) were measured in the rapidly incising Wutach gorge upstream of the capture point (Fig. 3-20f). Such high rates reflect adjustments to the transient process of the upper Wutach catchment capture, but will reduce with increasing maturity of the landscape, as observed in the neighbouring Randen area.

Combined, the data show that in areas of low relief (e.g. Bözbergplateau) chemical weathering dominates over physical erosion. In regions of intermediate slopes (Randen, Bözberg flanks), the amounts of erosion and weathering are similar. In contrast, the rapidly incising upper Wutach is dominated by pronounced physical erosion processes.

These new quantitative data confirm the range of previously reported intermediate-term denudation rates from the North Alpine Foreland (usually below 0.3 mm/yr) or adjacent regions, such as the Black Forest (~0.01 – 0.1 mm/yr), which are also based on catchment-wide denudation rates measured using cosmogenic nuclides (see Schnellmann et al. 2014 and Section 2.1 for a summary). It is concluded that, on average, intermediate-term denudation rates below 0.1 mm/yr can be expected in the area of the siting regions. In the case of future modulation of the drainage system with baselevel drop or if exposure of different rock types leads to increased local relief, these rates can multiply locally but for a limited period in time.

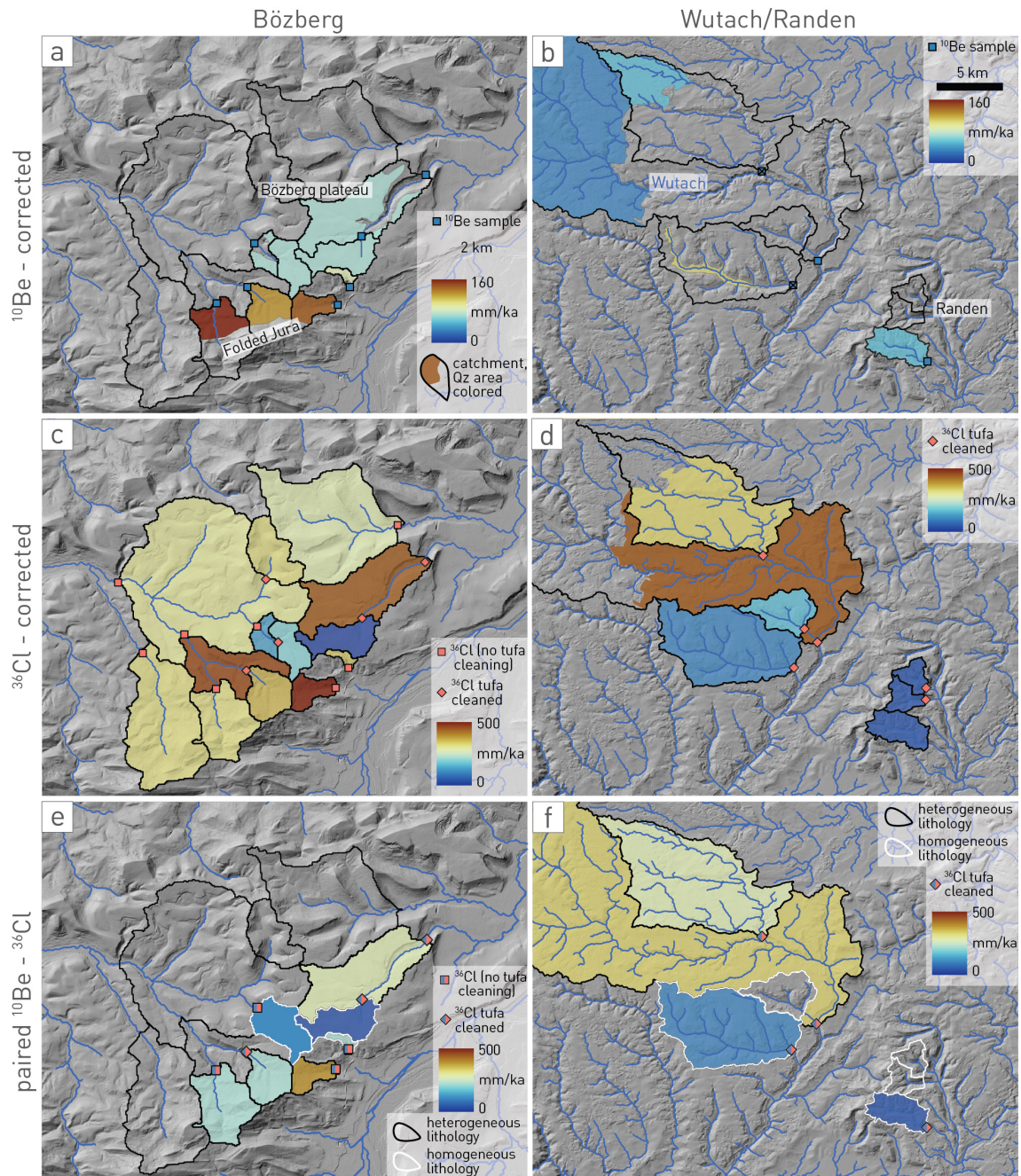


Fig. 3-21: Denudation rates from weathering-corrected single nuclide and paired nuclide approaches

^{10}Be denudation rates corrected with water chemistry weathering rates for Bözberg (a) and Wutach – Randen (b) areas. (c), (d): Same as (a), (b) for ^{36}Cl -based denudation rates. (e), (f): Denudation rates calculated from paired $^{10}\text{Be} - ^{36}\text{Cl}$ measurements (courtesy of R. Ott). The dataset presented here forms the basis for the map of interpreted denudation rates shown in Fig. 3-20e and f.

3.4.2 Workflow to assess the evolution of local topography

Model selection and landscape condition in Northern Switzerland

Local topography considers bedrock and unconsolidated sediments above the local erosion base. Because the characteristics of local topography vary between the sites, it is important to evaluate under what circumstances such additional overburden thickness can counterbalance potential disadvantages of a shallower site with respect to the local erosion base (Section 3.1).

It is likely that erosion rates will always be unevenly distributed in a cuesta landscape with alternating strata of differing susceptibility to erosion as in Northern Switzerland (Fig. 3-18), and that the landscape will remain in disequilibrium (e.g. Forte et al. 2016, Perne et al. 2017, Yanites et al. 2017, Darling et al. 2020), including for instance the continuing migration of drainage divides. The most resistant units can have both the steepest (cliff forms) and shallowest (plateau) gradients¹⁵, and, locally, transient erosion rates can vary greatly and in either direction from the controlling rate of the main channel. Such processes render a forecast of the topographic evolution difficult.

In addition to glacial processes (Section 3.5), the main components that drive topographic evolution are fluvial incision and hillslope processes. Modelling local topography requires to account for dynamic changes of the drainage networks and divides, which in turn are driven by changes in underlying rock types and baselevel changes. To be able to focus on a wide range of realisations, the choice was made against models with high degrees of complexity, as for instance landslide modules (e.g. Booth et al. 2013) or sediment transport (Section 3.3.2). Instead, the aim was to use a simple landscape evolution model that is able i.) to include the 3D lithological information of selected areas around the siting regions in Northern Switzerland, and ii.) to reproduce local relief within a plausible range as compared to the present-day landscape. To analyse the resulting differences in topographic evolution and associated distribution of overburden thickness, the model should be able to adapt scenarios of alternative main drainage networks. To do so, the Parsimonious Integrated Landscape Evolution model (MuddPILE) (Mudd et al. 2022, Graf 2022) was used, and integrated (as Model B) into the probabilistic framework (Fig. 3-22).

¹⁵ However, more erosion-resistant units tend to form steeper gradients in bedrock and slopes (e.g. Fig. 3-25). Currently, tributaries cross some lithologic boundaries without significant changes in profile steepness, and sediment in the riverbed masks the lithologic differences.

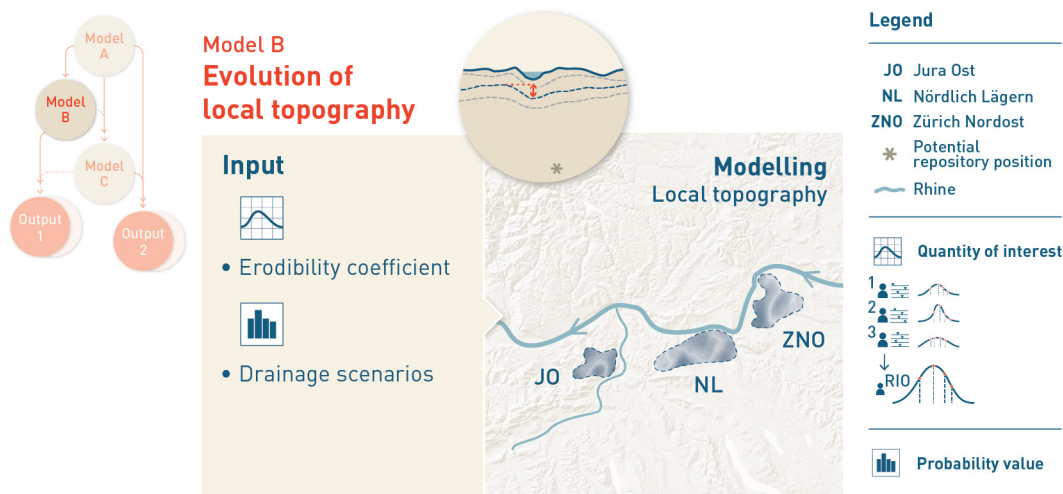


Fig. 3-22: Illustration of workflow to assess the evolution of local topography (MuddPILE)

MuddPILE (Parsimonious Integrated Landscape Evolution) model (Mudd et al. 2022, Graf 2022) was used to explore the effect of local baselevel change, heterogeneous lithology and drainage reorganisation on drainage divide migration, as well as to document their impact on the relief structure of the landscapes. The main input considers the SPIM-based modelled change in LEB according to Section 3.3 (see model cascade in Fig. 3-3, and more detailed illustration of the interface in Fig. 3-23). As further parameters, the fluvial erodibility coefficient K and different main drainage scenarios are varied based on expert judgement and incorporated as probability distributions and weighted scenarios, respectively.

For each model run of MuddPILE, local relief is calculated using steady-state solutions of the stream power incision model (for references on SPIM, see Section 3.3.2), starting from an initial drainage network (e.g. horizontal arrangement of the present-day drainage network). Hillslope relief in turn is calculated using a characteristic slope gradient (S_{CH}) on a gridded domain. On pixels that have a small drainage area (set here to 200 pixels, 25 m grid size, based on visual inspection of channel thresholds in digital elevation models) and are located away from the river network, hillslope angles were set to the characteristic slope value S_{CH} that depends on rock type (Section 3.4.3.3). The 3D geological model provided as a cube of lithological voxels (based on Gmünder et al. (2013)) was used to account for spatially variable values of erodibility coefficient K (depending on rock type; Section 3.3.3.3) and hillslope gradients. A "capping routine" avoids unrealistic landscapes: if a modelled pixel elevation exceeds that of the present-day landscape, the present-day elevation is assigned to this pixel instead. To minimise sharp breaks in relief, drainage divides are allowed to migrate. Lithologically driven discontinuities in elevation across divides lead to changes in channel-flow routing and channel relief. Such behaviour is controlled by "snapping-cycles", which work iteratively until the basin shapes stabilise (in general a few hundred snaps).

Accordingly, the uncertain parameters and input values are the (fluvial) erodibility coefficient K and hillslope angle S_{CH} . The uncertainties of the K values assigned to the lithological units of the 3D geological model (cube of lithological voxels) were assessed independently as subjective probabilities following the SHELF protocol (see Sections 3.3.3.3, 3.4.3.2 and Appendix B.2). The hillslope angles of the landscape were assessed in a previous study (Ludwig 2018). This work formed the basis for the characteristic hillslope values (Section 3.4.3.3).

In addition, for future evolution of local topography, interest lies in the response to local baselevel change. Fig. 3-23 illustrates the interface between the SPIM simulations for the Hochrhein (Model A) and MuddPILE (Model B). In Model B, the main rivers are lowered according to the

change in local erosion base (with respect to the repository), as simulated in Model A for the reference times of interest (i.e. 100 kyr, 1 Myr). Dividing the magnitude of this change by the reference time yields a net incision rate. In MuddPILE, this value is used as the erosion rate for the SPIM equation (Section 3.3) to calculate the steady-state profiles of the local tributary channels.

Consistent with Section 3.3.2, the SPIM parameters m ($= 0.45$ as commonly used) and n ($= 1$ for convenience and to avoid further parameter correlations) were not elicited.

Moreover, changes with respect to alternative main drainage networks are considered (Fig. 3-24).

Surrogate model

As outlined in Section 3.3.2, the probabilistic approach for Model A uses a Monte Carlo environment to generate possible future Hochrhein river profiles, based on one million samples. Regarding the input to Model B, this corresponds to one million samples of K values and associated changes in local erosion base. Moreover, they must be applied individually to i.) each siting region, ii.) each of the considered main drainage scenarios, iii.) and each reference time under consideration (1 Myr, 100 kyr). Because a single model run of MuddPILE is much more computationally intensive than using Model A, a surrogate model had to be generated to handle this large number of possible parameter combinations. It acts as an interpolation model that is incorporated in the probabilistic model.

Instead of running MuddPILE one million times in each configuration (siting region, drainage scenario, reference time), the number was reduced to a subset of $\sim 5'000$ model runs. The subset comprises ~ 700 representative combinations of K values. Each of these combinations is paired with 7 quantiles of local baselevel fall, derived from a thousand auxiliary model runs with Model A, using the same set of K values. For each such model run of MuddPILE (based on one specific parameter combination), a future terrain surface is calculated, enabling an estimate of the remaining overburden thickness at a given location of interest, e.g. the provisional disposal area.

The finally established "grid" of model realisations ($\sim 5'000$) is used for the interpolation model. For each of the one million Monte Carlo samples, a remaining overburden thickness at the location of interest is interpolated by taking into account the parameter combinations closest to the given sample. By integrating over all Monte Carlo samples, the final output of the surrogate model is a probabilistic distribution of remaining overburden thickness. This is calculated for the future scenarios of local topography in all three siting regions, using the present-day or alternative drainage networks (results in Section 3.4.6).

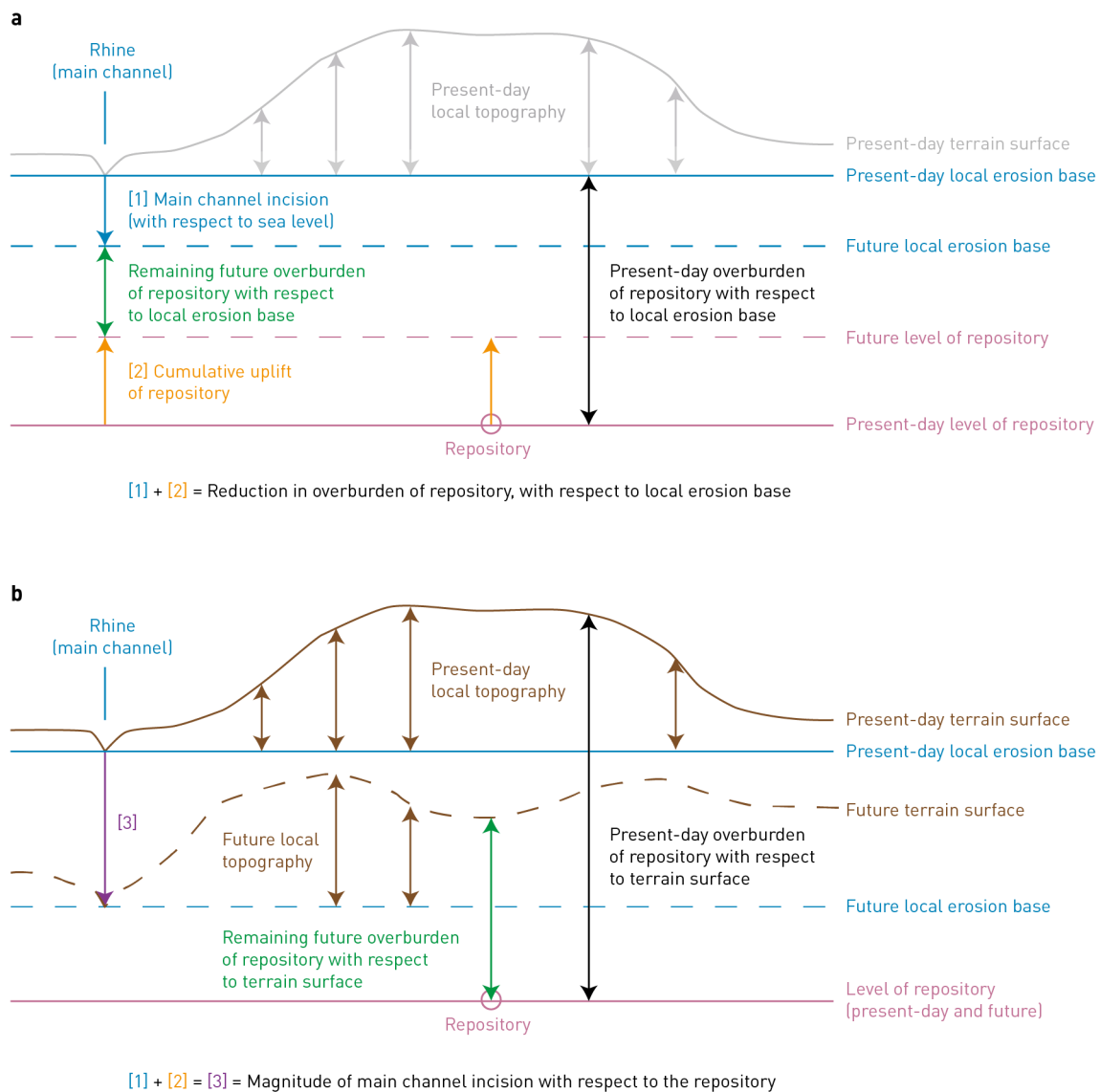


Fig. 3-23: Conceptual illustration of the interface between SPIM simulations of the Hochrhein (Model A; Section 3.3) and local topography (Model B)

Note that the figure is not to scale. (a) For the Hochrhein simulations (Model A, Section 3.3), the final quantity of interest (in green) is the remaining overburden thickness of the repository with respect to the local erosion base. b.) For the modelling of local topography (Model B, this section), it is the remaining overburden thickness of the repository with respect to the terrain surface (green). Importantly, the "reduction in overburden" (of the repository, with respect to local erosion base) from the Hochrhein simulation (Model A) can be transferred into the "magnitude of main channel incision relative to the repository" used as input for the local topography modelling (Model B). This is expressed by the two sums beneath (a) and (b). Note also that for the Hochrhein simulations in (a), all vertical changes are modelled with respect to sea level. In contrast, the modelling of local topography in (b) uses the repository as a fixed reference level, and all vertical changes (e.g. of the local erosion base) are modelled relative to this reference level.

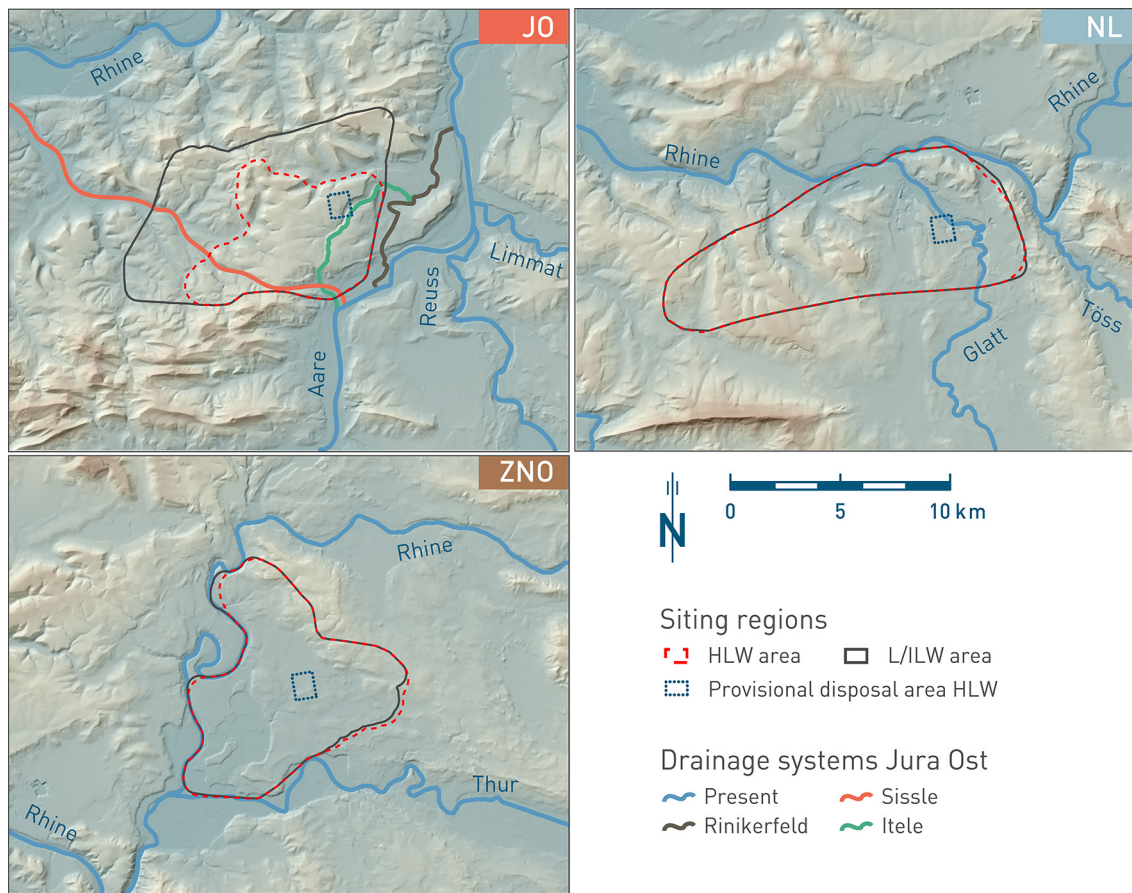


Fig. 3-24: The siting regions with present-day and alternative main drainage scenarios

Local topography scenarios (Tab. 3-6) based on present-day (all sites) and alternative drainage networks (JO) are considered for the assessment of erosion in the provisional disposal areas at one million years. Potential for lateral changes in drainage scenarios can be drawn from knowledge of fluvial or fluvio-glacial processes and denudation. Note that for NL and ZNO, the assessment regarding alternative drainage networks is directly based on the SPIM results from Section 3.3, implying that main rivers would directly flow across the location of the disposal area (see Section 3.4.3.4).

3.4.3 Elicited and additional parameters: rationale and argumentation

3.4.3.1 Fluvial incision as model input

The surrogate model for MuddPILE (Section 3.4.2) representatively applies the full distribution of fluvial incision of the Hochrhein from the SPIM modelling results in Section 3.3.4.1 as input to the local topography modelling. For each model realisation of MuddPILE constituting the surrogate model, the output – input relation shown in Fig. 3-23 is used. A list of the corresponding percentiles of fluvial incision for each siting region is displayed in Tab. 3-4.

Tab. 3-4: SPIM-derived fluvial incision as input parameter to MuddPILE

Listed values correspond to vertical changes of LEB of the Hochrhein with respect to the repository (in metres), as modelled with Model A (Section 3.3).

Metric	100'000 years			1'000'000 years		
	JO L/ILW	NL L/ILW	ZNO L/ILW	JO HLW	NL HLW	ZNO HLW
mode	16	13	11	143	163	149
mean	29	25	14	175	211	189
P05	10	8	6	79	82	46
P25	17	14	9	125	145	121
P50	26	22	12	166	197	176
P75	38	33	15	215	263	243
P95	54	55	27	300	387	375

3.4.3.2 Erodibility and K values

As described in Section 3.3.3.3, the 3D geological model (Gmünder et al. 2013) used for the local topography modelling comprises 15 lithostratigraphic units. They are assigned to four erodibility classes (EC) based on lithological considerations (e.g. Fig. 3-11, Tab. 3-5).

The parameter elicitation and the resulting parameter distributions for the fluvial erodibility coefficient K of the four erodibility classes are documented in Section 3.3.3.3.

The surrogate model for MuddPILE (Section 3.4.2) representatively applies the full distribution of K for each erodibility class as input to the local topography modelling.

3.4.3.3 Characteristic slope angles

Using the same 3D geological model and aggregation scheme of erodibility classes (EC) as for the SPIM parameter K (Sections 3.3.3.3 and 3.4.3.2), each lithostratigraphic unit is assigned one single deterministic value referred to as "characteristic slope" S_{CH} . These values are informed from lithology-specific histograms of hillslope gradient in the present-day landscape of Jura Ost and surrounding area (Ludwig 2018). Fig. 3-25 shows the two histograms for the Opalinus Clay and Villigen Formation as examples.

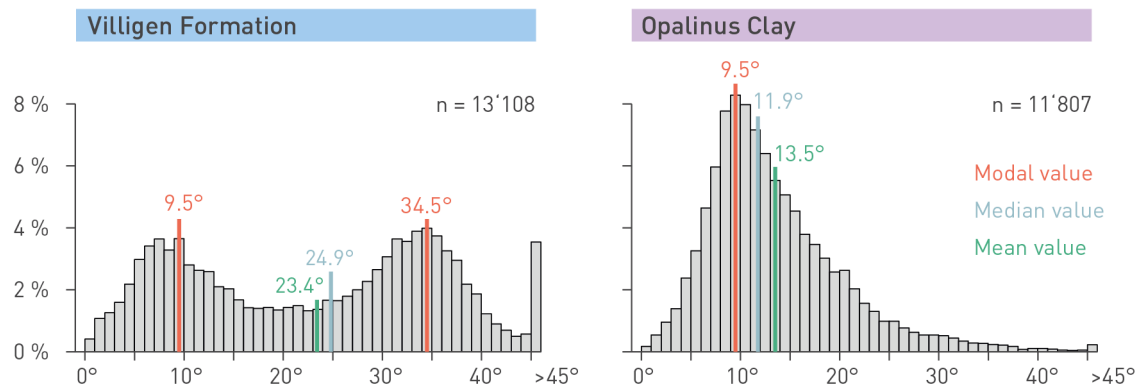


Fig. 3-25: Histogram and statistical measures of slope gradients within the hillslope domain of JO and surrounding area: two examples

The examples refer to the Villigen Formation (EC4) and the Opalinus Clay (EC2) (Ludwig 2018). Slope gradients correspond to pixel-based values from DTM-AV (2-m grid size; © swisstopo). Number 'n' indicates the number of pixels for each lithostratigraphic unit. Note that frequencies of values steeper than 45° are summarised in the histograms.

This approach aims at producing a plausible hillslope relief (as compared to the present-day landscape) in the model. It is acknowledged that, in fact, (lithology-specific) hillslope gradients are not reflected by a single value, but instead would also describe a probabilistic distribution, mostly as a function of erosion rates. However, it is deemed sufficient to focus the probabilistic consideration on the K values (Section 3.4.3.2) as (i) they may involve more uncertainty and, (ii) most importantly, they directly control the elevations of the fluvial channel nodes that (in the vast majority of cases) define the locations of smallest overburden thickness above the disposal area.

Values of characteristic slope for each EC were assigned according to Tab. 3-5. Values were determined based on the median values of associated lithostratigraphic units in Ludwig (2018), as they were considered to be a representative measure of hillslope relief observed in the present-day landscape:

- EC2 based on median values of Opalinus Clay, Molasse
- EC3 based on median values of Wildeggen Formation, Ifenthal Formation, Passwang Formation
- EC4 based on median values of Villigen Formation, Hauptrogenstein (incl. pixels of Klingnau Formation)

S_{CH} for EC1 (Quaternary) was chosen to be slightly lower than for EC2, following a similar rationale as for the K values (Section 3.3.3.3).

Tab. 3-5: Lithostratigraphic scheme and assigned values of "characteristic slope" S_{CH}

The regional 3D geological model (Gmünder et al. 2013) used for the local topography modelling comprises 15 lithostratigraphic units. They are assigned to four erodibility classes (EC), based on lithological considerations (see Section 3.3.3.3, e.g. Fig. 3-11). The table lists S_{CH} both in degrees [°] and as dimensionless gradient [-].

Lithostratigraphic Unit	EC	S_{CH} [°]	S_{CH} [-]
Quaternary units	1	10	0.18
Upper Freshwater Molasse Group (OSM)	2	12	0.21
Upper Marine Molasse Group (OMM)	2	12	0.21
Lower Freshwater Molasse Group (USM)	2	12	0.21
Upper Malm Group	4	24	0.46
Wildeggen Formation	3	15	0.27
Upper Dogger Group	3	15	0.27
Hauptrogenstein	4	24	0.46
Klingnau Formation [and eastern equivalents]	2	12	0.21
Passwang Formation [and eastern equivalents]	3	15	0.27
Opalinus Clay	2	12	0.21
Lias Group, Keuper Group	2	12	0.21
Upper Muschelkalk Group	4	24	0.46
Middle Muschelkalk Group, Lower Muschelkalk Group, Buntsandstein Group	3	15	0.27
Crystalline basement + Permo-Carboniferous	4	24	0.46

3.4.3.4 Alternative drainage scenarios

Lateral migration of the main and tributary rivers, also in connection with glacial processes, can lead to the formation of new channels by cutting through bedrock relief (Section 3.4.1, Fig. 3-19). These processes are more likely to occur in lower relief landscapes and under the influence of large foreland glaciations. However, even in areas of high local topography, corresponding scenarios cannot be excluded for long timescales and are thus considered as alternative evolutions especially in the case of JO (Fig. 3-24, Tab. 3-6). Modelled scenarios are (i.) the reactivation of the Rinikerfeld channel, an Aare shortcut using the Sissle drainage (ii.) with and (iii.) without incorporation of the Reuss and Limmat Rivers along this new route, and (iv.) a breakthrough channel following the Itele Valley. This set of alternative main drainage scenarios (ADN) is considered for the 1 Myr evolution of local topography in Jura Ost (Fig. 3-24). Alternative drainage scenarios for JO are not considered for the 100-kyr interval, because it would need at least a source of high sediment supply nearby (foreland glacier), which was considered unlikely in this short period (see Section 3.5.3.5). In JO, the higher present-day relief (as compared to the relevant parts of NL and ZNO) has more potential to influence the occurrence of alternative main river valleys. The ADN locations in JO correspond to the glacial overdeepening scenarios, which are discussed in Section 3.5.3.3. The rationale behind this is that a glacial overdeepening

constitutes the erosional part below the local erosion base. Accordingly, here the main channels (which define the local erosion base) mirror the overdeepening scenarios with two exceptions. Firstly, two parallel Aare shortcuts through the Sissle (Sissle north and Sissle south) are considered for glacial overdeepenings, which are reduced to one channel here (Sissle south). Instead of modelling a similar topographic response, the local topography model rather considers two alternative sub-scenarios, in which the Reuss and Limmat tributaries either continue to drain into the Aare (following the shortcut) or are beheaded and continue to drain towards the Rhine along the existing Lower Aare Valley (Fig. 3-24 and Tab. 3-6). Secondly, a scenario, allowing for a "random" occurrence of glacial overdeepening location was established. Because the landscape evolution model (MuddPILE; Section 3.4.2) is too computationally expensive to be able to model a significant number of random drainage locations, the percentages for the scenario "random" were distributed proportionally to the modelled alternative drainage scenarios (Rinikerfeld, Aare shortcut Sissle 1, Aare shortcut Sissle 2, Itele; Fig. 3-24 and Tab. 3-6).

Tab. 3-6: Drainage scenarios and main reasoning

The table shows the unconditional probabilities (i.e. the occurrence of glaciations is already considered and integrated out in the numbers given). See Fig. 3-24 for geographic reference.

Time [yr]	Drainage scenario	Site	Main reasoning	Probability of occurrence [%]
100'000	Present-day	All sites	No foreland glaciation	100 (not elicited)
1'000'000	Present-day	NL, ZNO	Drainage established	Not elicited
1'000'000	Present-day	JO	Drainage well established	84
1'000'000	Rinikerfeld channel	JO	Active during HT deposition, possible with high sediment accumulation (nearby glacier)	14
1'000'000	Aare shortcut Sissle 1 (Aare, Reuss, Limmat)	JO	Morphologically preconditioned narrow valley; needs large glaciation, headward erosion; shortcut lowers baselevel – reversal of Limmat and Reuss course (Sissle 1) more likely than continuing northward course to Rhine (Sissle 2)	0.7
1'000'000	Aare shortcut Sissle 2 (only Aare)	JO		0.3
1'000'000	Itele	JO	Morphologically preconditioned; more erodible lithologies; parallel to (Beringen) ice margin (Fig. 3-1), reactivation as ice-marginal drainage or localisation of subglacial stream, needs large foreland glaciation	1
1'000'000	Main channel directly above disposal area	NL, ZNO	Low relief, high randomness	Not elicited

These computational limits, but additionally the even more uncertain spatial prediction of future main drainages around the low-relief regions of ZNO and the eastern parts of NL, are also the reason why no comprehensive modelling of ADN with MuddPILE is feasible as part of the probabilistic modelling for these siting regions. Instead, the probabilistic output of the SPIM model (Section 3.3) is directly used to address the potential effects of alternative main drainage networks on remaining overburden thickness. This more pessimistic approach implicitly assumes that a main river (defining the local erosion base) will directly flow across the disposal area. However, MuddPILE is still used for NL and ZNO to assess the remaining overburden thickness using the present-day main drainage scenario.

The probability of occurrence combined with the main arguments for the different drainage scenarios has been elicited according to Tab. 3-6.

3.4.4 Modelling results

Following the rationale in Section 3.4.3.4, the remaining overburden thickness was calculated for the future scenarios of local topography in all three siting regions, using the present-day or alternative main drainage networks. Fig. 3-26 and Fig. 3-28 show examples of the corresponding results of remaining overburden thickness of Top Opalinus Clay with respect to future terrain surface for selected percentiles of main channel incision and SPIM parameter K . Fig. 3-27 and Fig. 3-29 show the corresponding (future) geological outcrop maps, based on the Regional 3D Geological Model 2012 (Gmünder et al. 2013) used within MuddPILE.

The probabilistic result of remaining overburden thickness for the three siting regions, using the present-day or alternative drainage networks, based on the surrogate model is shown in Section 3.4.6 (Fig. 3-30 and Fig. 3-31). The results from this section are discussed in the subsequent Sections 3.4.5 and 3.4.6.

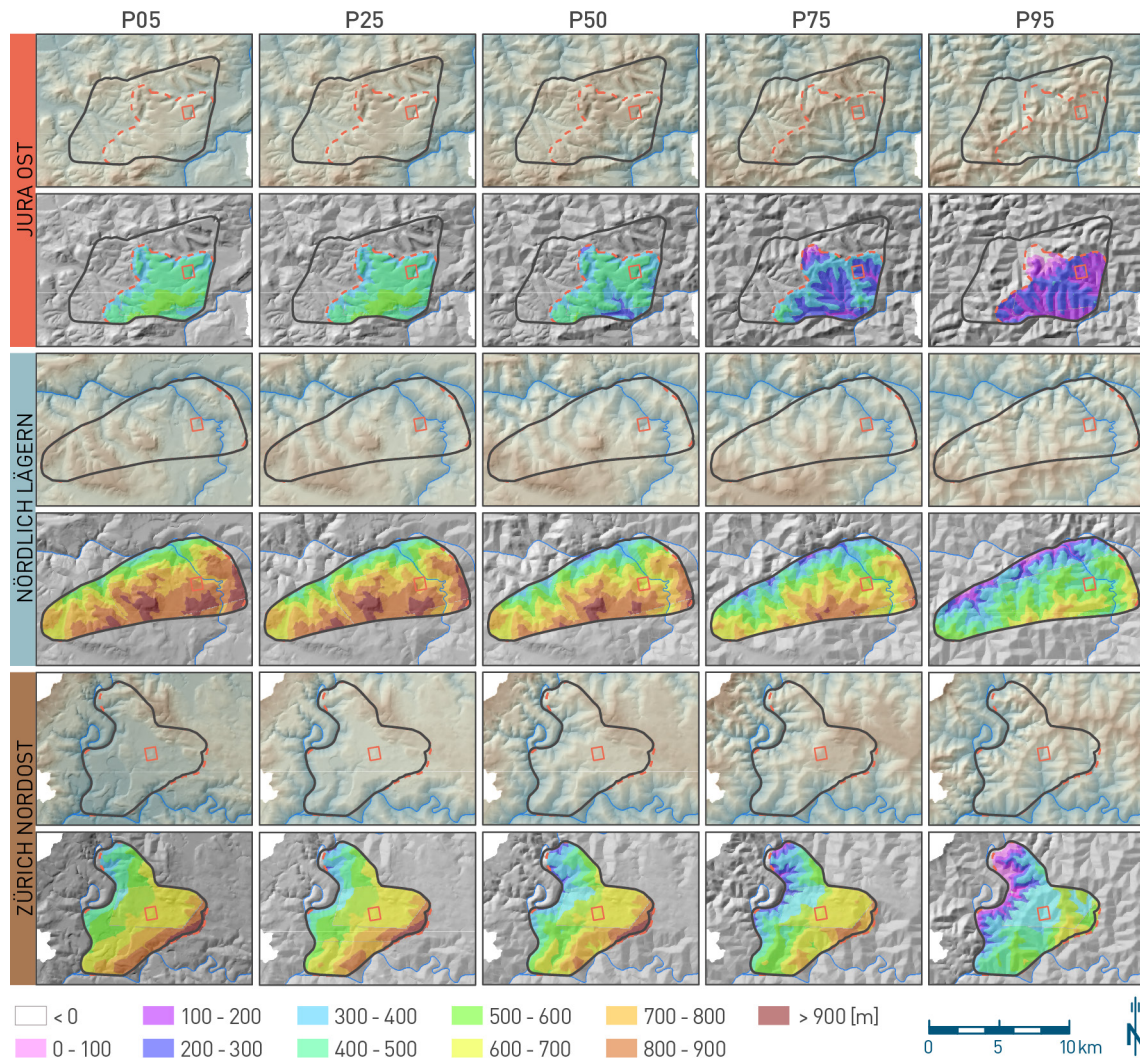


Fig. 3-26: Results of modelling the evolution of local topography at one million years using the present-day drainage network

Shown are examples of synthetic future landscapes and colour-coded remaining overburden thickness for each siting region and for certain quantiles of parameter combinations, e.g. 25th percentile of fluvial incision after one million years as derived from the SPIM modelling (Section 3.3.4), combined with the 25th percentile of the erodibility distributions (Section 3.4.3.2). Remaining overburden thickness is shown with respect to the Top Opalinus Clay (as defined by the 3D horizon model documented in Nagra 2024d, laterally complemented based on Gmünder et al. 2013). Red square shows the location of the provisional HLW disposal area.

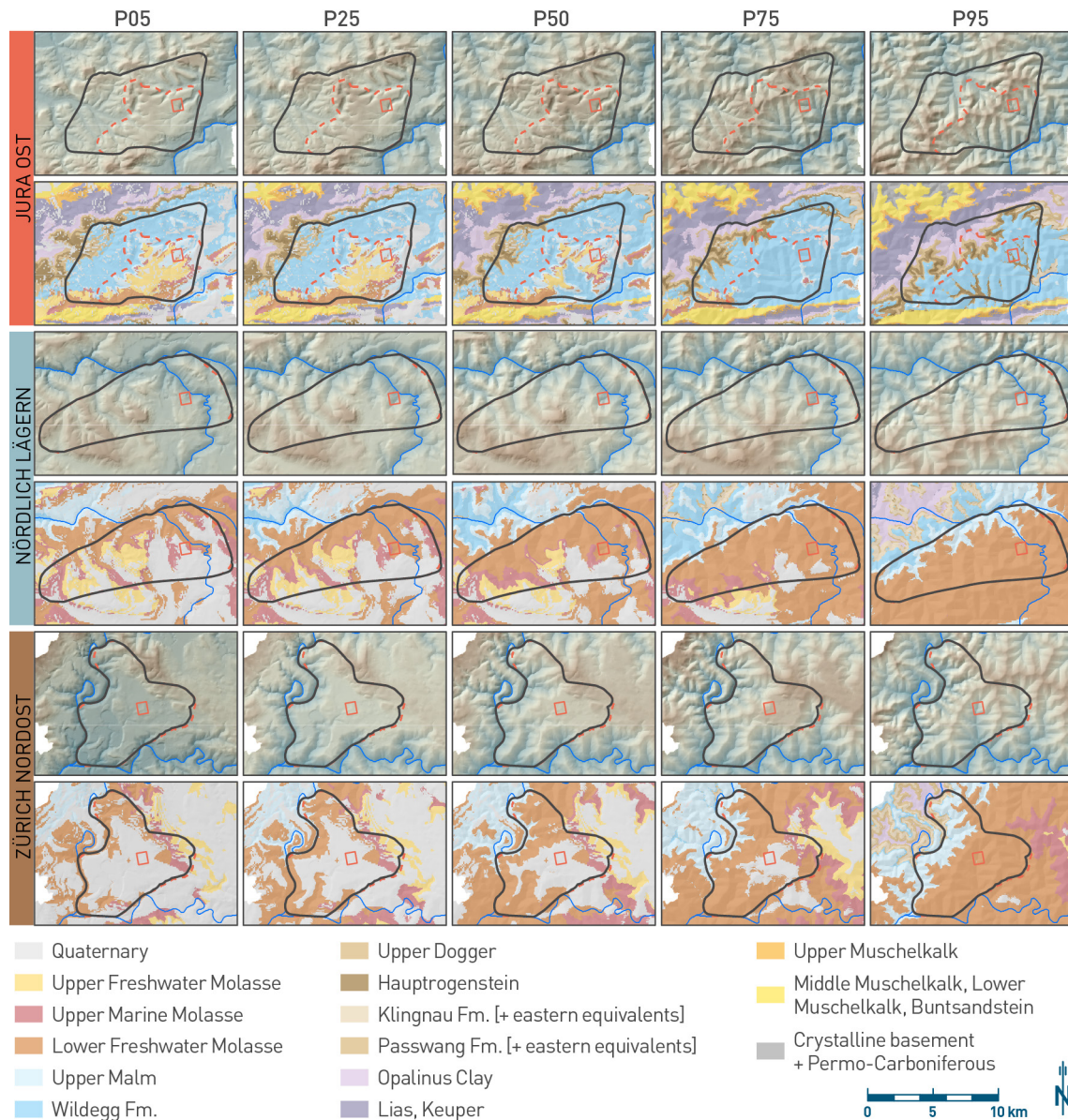


Fig. 3-27: Outcrop maps representing the evolution of local topography at one million years using the present-day drainage network

Shown are examples of synthetic future landscapes and colour-coded geological outcrop maps for each siting region and for certain quantiles of parameter combinations, e.g. 25th percentile of fluvial incision after one million years as derived from the SPIM modelling (Section 3.3.4), combined with the 25th percentile of the erodibility distributions (Section 3.4.3.2). The outcrop map is based on the Regional 3D Geological Model 2012 (Gmünder et al. 2013) used within MuddPILE. Red square shows the location of the provisional HLW disposal area.

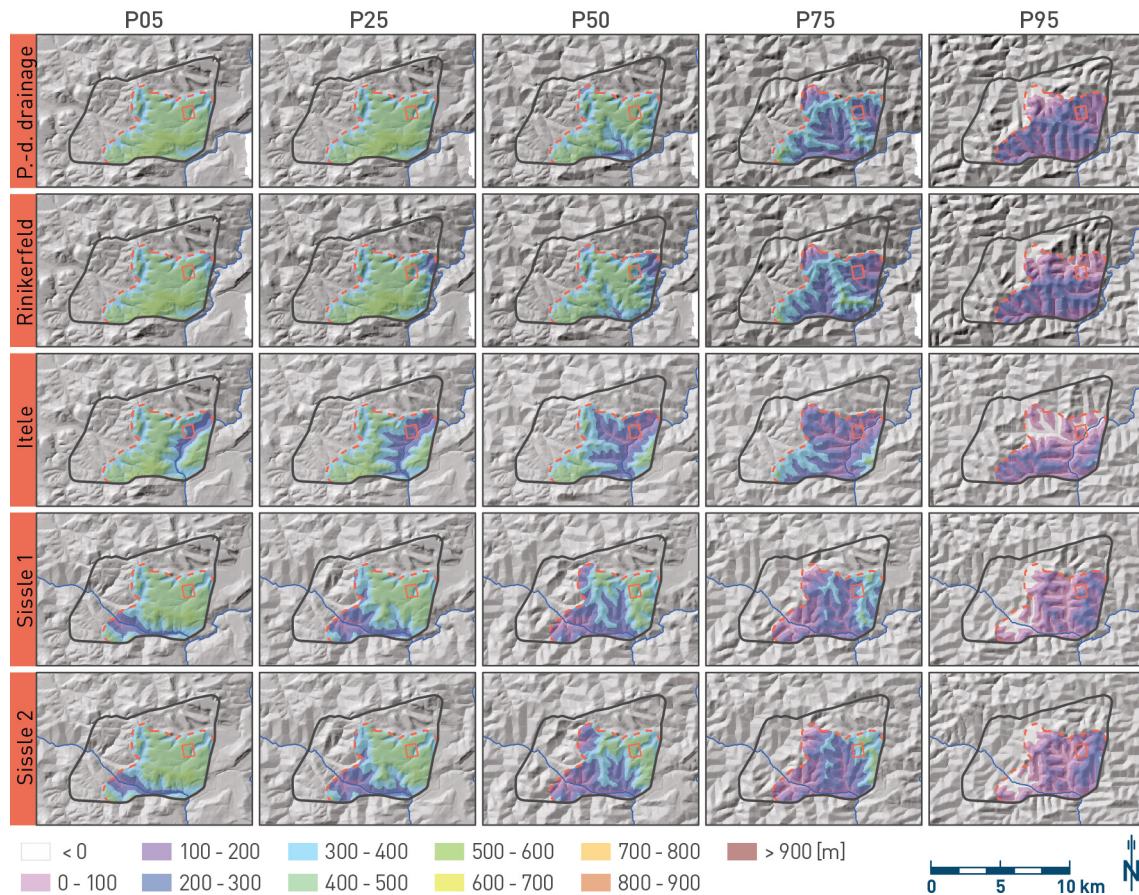


Fig. 3-28: Results of modelling the evolution of local topography at one million years using alternative drainage networks for JO

Shown are examples of synthetic future landscapes and colour-coded remaining overburden thickness for certain quantiles of parameter combinations, e.g. 25th percentile of fluvial incision after one million years as derived from the SPIM modelling (Section 3.3.4), combined with the 25th percentile of the erodibility distributions (Section 3.4.3.2) for the present-day main drainage network (upper panel) and alternative drainage networks (reactivation of the Rinikerfeld channel, breakthrough channel through the Itete Valley, a shortcut of the Aare through the Sissle Valley with [Sissle 1] and without [Sissle 2] diversion of Limmat and Reuss). Remaining overburden thickness is shown with respect to the Top Opalinus Clay (as defined by the 3D horizon model documented in Nagra 2024d, laterally complemented based on Gmünder et al. 2013). Red square shows the location of the provisional HLW disposal area.

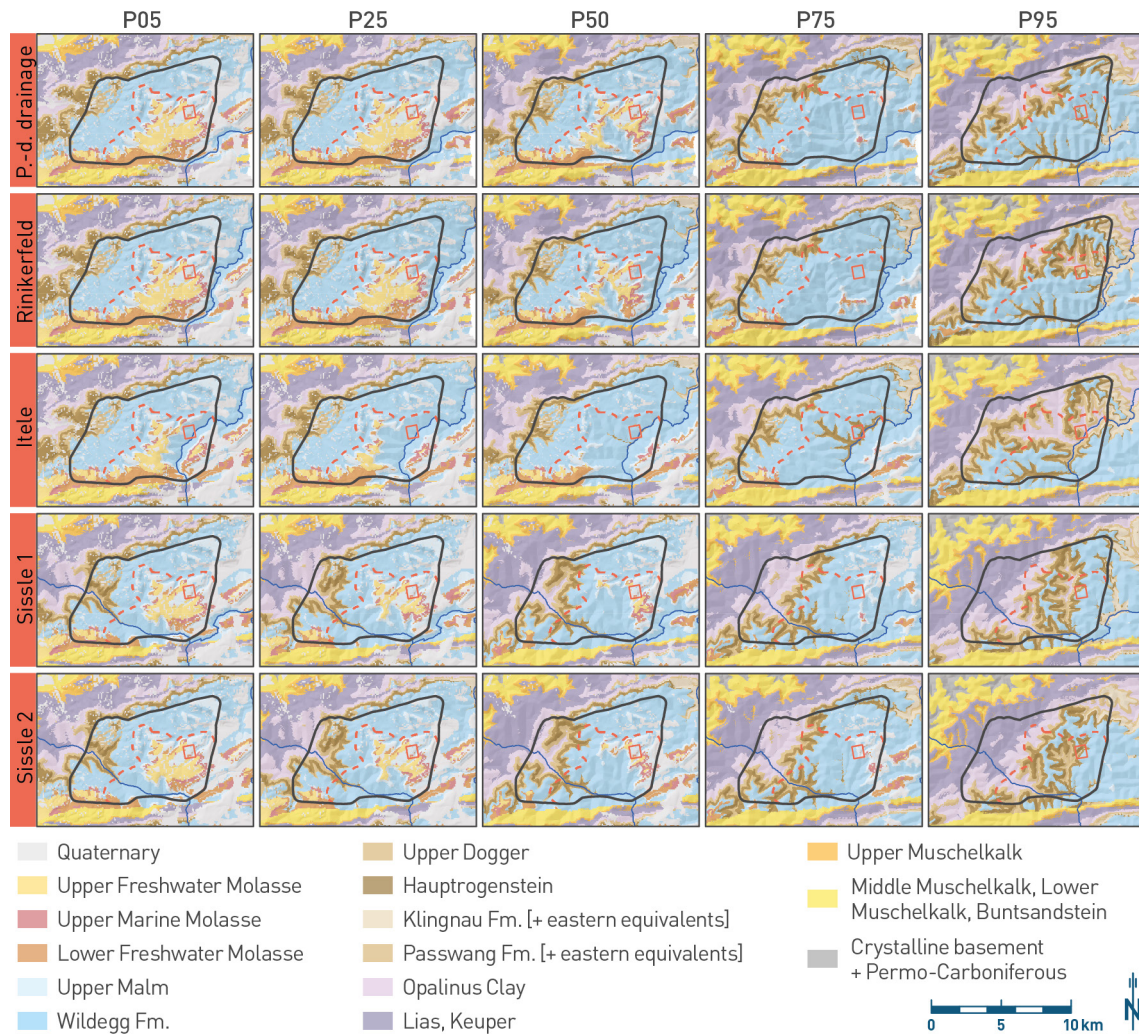


Fig. 3-29: Outcrop maps representing the evolution of local topography at one million years using alternative drainage networks for JO

Shown are examples of synthetic future landscapes overlain by colour-coded geological outcrop maps for certain quantiles of parameter combinations (e.g. 25th percentile of fluvial incision after one million years as derived from the SPIM modelling (Section 3.3.4), combined with the 25th percentile of the erodibility distributions (Section 3.4.3.2) for the present-day main drainage network (upper panel) and alternative drainage networks (reactivation of the Rinikerfeld channel, breakthrough channel through the Itete Valley, a shortcut of the Aare through the Sissle Valley with [Sissle 1] and without [Sissle 2] diversion of Limmat and Reuss). The outcrop map is based on the Regional 3D Geological Model 2012 (Gmünder et al. 2013) used within MuddPILE. Red square shows the location of the provisional HLW disposal area.

3.4.5 Plausibility

The rationale for the modelling approach was outlined in Section 3.4.2. Important requirements include the ability of the model to account for (i) different rates of local baselevel fall (main river incision; output of Section 3.3.4), (ii) present-day and alternative horizontal arrangements of the main drainage network, (iii) the 3D lithological structure reflecting the layered stratigraphy of the study area, (iv) different combinations of K values representing the probabilistic distributions estimated in Section 3.3.3.3, to account for the uncertainty in the resistance of the different rock types to fluvial erosion. In addition, a single characteristic hillslope gradient is assigned to each erodibility class to model hillslope geometry in a simplified way¹⁶. The combination of these features allows modelling of dynamic changes in the drainage networks and divides, which in turn are driven by changes in underlying rock types and baselevel changes. Moreover, the model should be able to reproduce local relief within a plausible range (as compared to the present-day landscape), with the parameter uncertainty of (i) and (iv) as well as scenario weights for (ii) being incorporated via the surrogate model. The results of this latter approach are shown in Section 3.4.6.

A first simple expectation for the model is confirmed by example results in Fig. 3-26 and Fig. 3-28 for modelled landscapes at 1 Myr for all three siting regions and considered main drainage networks. It shows how the local topography and thus the remaining overburden thickness (in this case with respect to Top Opalinus Clay) is gradually reduced, as the main river incision increases from lower (P05) to higher (P95) percentiles, indicating the corresponding response of tributary rivers and adjacent hillslopes.

Fig. 3-27 and Fig. 3-29 accordingly confirm how lower lithostratigraphic units are successively exposed in the model landscape, as the main river incision and associated reduction in overburden thickness progresses. In this scenario, the spatial pattern of exposed lithologies also continuously changes when moving from P05 to P95. In landscapes with heterogeneous lithologies and horizontal or dipping contacts, drainage divides will continuously migrate (Section 3.4.2). This behaviour is supported by Fig. 3-26 to Fig. 3-29, as well as Graf (2022), showing local drainage divides being shifted in response to increasing main river incision with associated removal of overburden thickness and changing lithological spatial patterns. In addition to the drainage divides (i.e. the spatial boundary of individual tributary catchments), also the internal horizontal arrangement of channel networks within these catchments can substantially change, as can be seen on the maps. This can have relevant implications for the remaining overburden thickness measured above a disposal area, as it may be intersected by laterally shifted tributary channels.

Importantly, the same effects attributed to the influence of vertical incision above can also result from horizontal changes of the main streams (Graf 2022)¹⁷. This can be seen when comparing results for different main drainage networks in JO (Fig. 3-28 and Fig. 3-29) for a given percentile of main river incision (e.g. P75), indicating considerable differences in the remaining overburden thickness above the provisional disposal area. In general, there is a plausible (and important) tendency towards a reduction of overburden thickness for ADNs shifting the main river closer to the repository (see also Section 3.4.6). However, this is again modulated further by the corresponding changes in the lithological spatial pattern and associated effects on local drainage divides and horizontal arrangement of the channel network within the tributary catchments, as mentioned above.

¹⁶ Regarding the plausibility of chosen hillslope gradients, as well as the justification to apply one value per erodibility class only, see Section 3.4.3.3.

¹⁷ Moreover, the two components can superimpose.

For these reasons, one specific combination of sampled K values and local baselevel drop for a given main drainage scenario will result in one specific spatial pattern of drainage divides and horizontal arrangement of tributary channel networks, which in itself involves considerable uncertainty regarding the remaining overburden thickness at the provisional disposal area. This is mitigated by the large number of combinations of K values and baselevel falls used for each drainage scenario in the surrogate model (Section 3.4.2), ensuring a mixture of model realisations with both positive and negative effects on remaining overburden thickness above the disposal area.

For the capping of too high, unrealistic model elevations (see Section 3.4.2, "capping routine"), one result is the preservation of upland low relief areas for certain (lower-magnitude) erosion scenarios (e.g. Fig. 3-26: JO and ZNO P05, P25). In fact, this model behaviour is not unlike phenomena that can also be observed in the present-day landscape, such as the nearly flat Irchel Plateau (Section 2.4) or the Geissberg Plateau (Fig. 3-18; Yanites et al. 2013). At both sites, preservation of Quaternary deposits until the present day indicates very low erosion rates on timescales of 10^5 to 10^6 years. For these capped regions, MuddPILE models zero erosion, while in a natural landscape some erosion is still expected. This is especially true for certain terrain such as the top of the Bözbergplateau, showing at least a subdued internal relief. For this reason, Fig. 3-20e indicates an interpreted erosion rate of 50 – 100 m/Myr for this area. Consequently (and by looking at Fig. 3-26, JO and ZNO P05, P25, P50), the model can be assumed to overestimate remaining overburden thickness (by an order of several decametres in 1 Myr) at the provisional disposal areas in the case of lower-magnitude erosion scenarios at JO and ZNO. Importantly, in the probabilistic model output this skews the percentile ranges of anyway thick overburden (> 200 m), an effect that is further visualised (Fig. 3-30 and Fig. 3-31) and discussed as "converging" behaviour in Section 3.4.6. However, it does not affect the percentile ranges most relevant to decompaction effects or even repository excavation. As can also be seen in Fig. 3-26 for the higher-magnitude erosion scenarios, significant erosion along tributary catchments (such as in the case of the Bözbergplateau) is still convincingly captured by the model.

The need for the capping mechanism arises from the choice of modelling with steady state assumptions. The latter is acknowledged not to cover the full complexity of the natural landscape, known to be exposed to a variety of transient perturbations. However, aside from not modelling the (potentially low) erosion rates on capped model domains (see above), the steady state approach in general is considered more pessimistic with respect to estimating remaining overburden thickness. This is because fluvial knickpoints are implicitly treated as migrating instantaneously up the entire stream longitudinal profile, thus removing any additional overburden thickness related to transient profile convexities.

Finally, while the SPIM (used to model the longitudinal profiles of tributary streams within MuddPILE) was originally developed for detachment-limited bedrock channels, many present-day tributary streams of the study area also feature sediment cover comprised of unconsolidated (Quaternary) deposits, at least in some of their reaches. Such conditions may prevail, at least temporarily, also in the future. The large range of elicited K values fed into the probabilistic assessment via the surrogate model, however, is considered to compensate for the potential uncertainty in modelled channel gradients, related to future sediment cover. This is also supported by the evidence used to estimate the K values (Section 3.3.3.3) that has been derived from the same study area and should therefore already reflect corresponding effects by the present-day sediment cover.

3.4.6 Site-specific differences

Representative results from modelling with MuddPILE for certain quantiles of parameter combinations were shown in map view for each siting region in Section 3.4.4 (Fig. 3-26 to Fig. 3-29). The actual assessment of the sites, however, uses the probabilistic distribution of remaining overburden thickness provided by the surrogate model for MuddPILE (Section 3.4.2), that is based on a Monte Carlo simulation with one million parameter combinations. For each realisation, the remaining overburden thickness of the repository is determined as the minimum local overburden thickness with respect to the modelled terrain surface within the perimeter of the provisional disposal area (represented by a 25 m grid). Results of this are shown in Fig. 3-30 for the local topography scenarios in each siting region based on a.) the present-day drainage network, and b.) including alternative drainage networks. For JO, the latter aggregates the results for the present-day and alternative drainage networks using the scenario weights (probability of occurrence) listed in Tab. 3-6. For NL and ZNO, alternative drainage networks are assessed directly based on the probabilistic output of the SPIM model (Section 3.3.4).

The result for the provisional disposal area at 1 Myr (Fig. 3-30a) using the present-day drainage network shows distinct differences between the three siting regions. A reduction of overburden thickness below 200 m, considered as the relevant overburden thickness for efficient self-sealing (Chapter 1), has a higher probability in JO than the two other sites in spite of the availability of local topography to add to the potential overburden thickness. The probability of yielding less than 200 m is low in case of both NL and ZNO, although the result indicates a slightly higher potential for NL to reach overburden thickness < 400 m than in ZNO. This can be directly linked to the proximity of the NL provisional disposal area to the present-day main drainage system, in this case represented by the Glatt River. Conversely, there is a greater distance between the provisional disposal area in ZNO and the Thur and Rhine Rivers (Fig. 3-24). As a consequence (and directly linked to the model interface between SPIM and MuddPILE according to Fig. 3-23), the distribution of remaining overburden thickness for NL from the local topography modelling using the present-day drainage network (Fig. 3-30a) is nearly identical to the SPIM output for NL shown in (Fig. 3-15d). This circumstance also explains the very different shape of the distribution for NL as opposed to ZNO and JO (again with a considerable distance between the JO provisional disposal area and the main River Aare, see Fig. 3-24). The "violin-shape" for NL closely reflects that of the SPIM result shown for the Hochrhein in Fig. 3-15d. Instead, the distributions for both ZNO and JO show only a thin distribution over most of the range of the modelled overburden thickness, and a high concentration at ~ 400 m (JO) and ~ 650 m (ZNO), also defining the median in these distributions. These values reflect the corresponding minimum present-day overburden thickness above the provisional disposal areas in the two sites. This convergence is explained by the capping effect outlined in the previous Section 3.4.5, which is put into context there (and is also recognisable in the P05 to P50 map results shown in Fig. 3-26). A median value of slightly more than 600 m remaining overburden thickness is observed for NL. This result is disadvantageously influenced by the proximity of the NL provisional disposal area to the Glatt River. However, considering the flat topography in the eastern part of NL, it is well possible that the location of the main drainage will be shifted away from the disposal area and a significant protective relief will develop in the future.

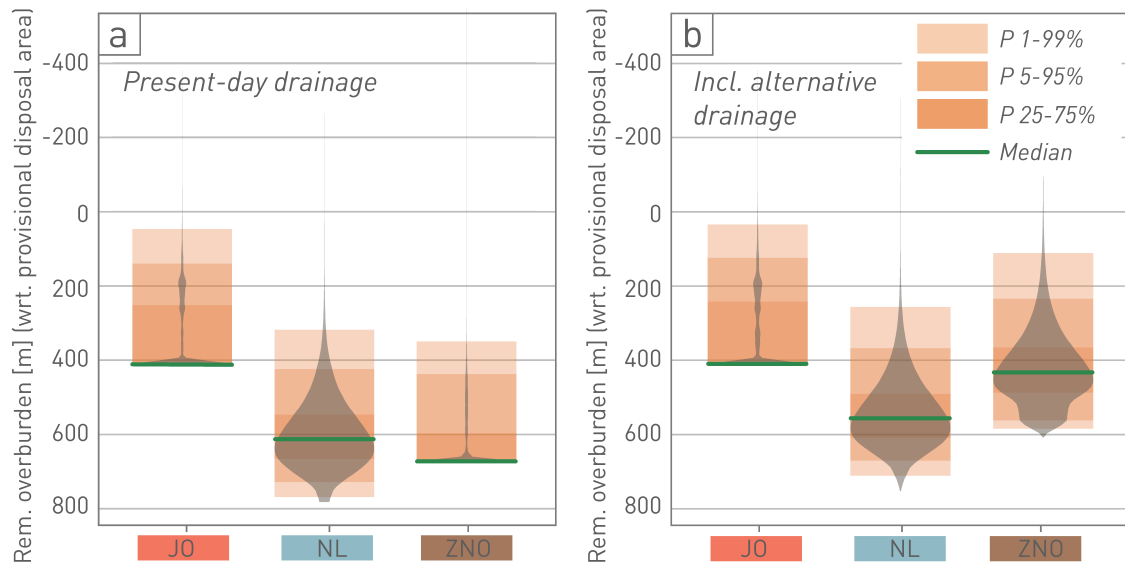


Fig. 3-30: Distributions of remaining overburden thickness after one million years including local topography for JO

Violin plots and boxes with selected percentiles of remaining overburden thickness at the location of the provisional disposal areas. (a) Distributions with respect to the evolution of local topography, assuming preservation of the present-day drainage network (i.e. main rivers remain laterally at the same position as today; see Fig. 3-24). Including local topography has a significant effect, especially for JO, if compared to the distribution shown in Fig. 3-15d. Distributions for JO and ZNO "converge" at the value of the minimum present-day overburden thickness because of the capping routine discussed in Section 3.4.5. (b) Several plausible alternative drainage scenarios were tested for JO (Fig. 3-24 and Fig. 3-31). A respective weighting by the probability of occurrence of the present-day and alternative drainage scenarios is incorporated into the uncertainty estimates for the evolution of local topography in JO. In NL and ZNO, scenarios for alternative drainage networks cannot be reasonably derived because of the low relief. Distributions are calculated using the (worst case) assumption of a main river positioned at the repository location (corresponding to direct output from Model A, i.e. Fig. 3-15d).

Distinct differences are also observed when looking separately at the results of the different alternative drainage networks in JO at 1 Myr (Fig. 3-31), as defined in Section 3.4.3.4. The least deviation from the result of present-day drainage network (as also shown in Fig. 3-30a) is indicated for the two Sissle drainage scenarios, although a minor tendency towards smaller overburden thickness is observed in these cases. This tendency involves both a higher proportion of overburden thickness < 200 m and also a less pronounced convergence at ~ 400 m associated with the capping effect mentioned above. This tendency is more pronounced in the case of the Rinikerfeld drainage scenario. This seems reasonable, given the proximity of the provisional disposal area to the westward shifted Aare in this scenario and corresponding headwater erosion as indicated in Fig. 3-28. The median value in this scenario decreases to < 350 m. Finally, the most striking reduction in overburden thickness results from the Itele drainage scenario. This is not surprising as in this scenario the Aare (as main river) would flow directly across the perimeter of the disposal area. Accordingly, and similar to the NL present-day drainage scenario described above, the distribution for the Itele drainage scenario in JO is nearly identical to the SPIM-derived distribution in Fig. 3-15d. A median value of less than 200 m is found in this case. The estimated probability of this scenario, however, is rather low and conditional on the occurrence of a large foreland glaciation (see Tab. 3-6 for probabilities that have already integrated the glaciations).

Aggregating the different JO drainage scenarios together, as weighted by their probability of occurrence according to Tab. 3-6 (Fig. 3-30b), largely resembles the distribution for the present-day drainage scenario (Fig. 3-30a), including a median value at ~ 400 m. This can be explained by the high probability of 84% assigned to the present-day drainage scenario in Tab. 3-6. Nevertheless, a certain tendency towards smaller overburden thickness is also observed here, reflecting the associated effects for the JO ADN described above, that also feed into the calculation at a combined probability of 16%.

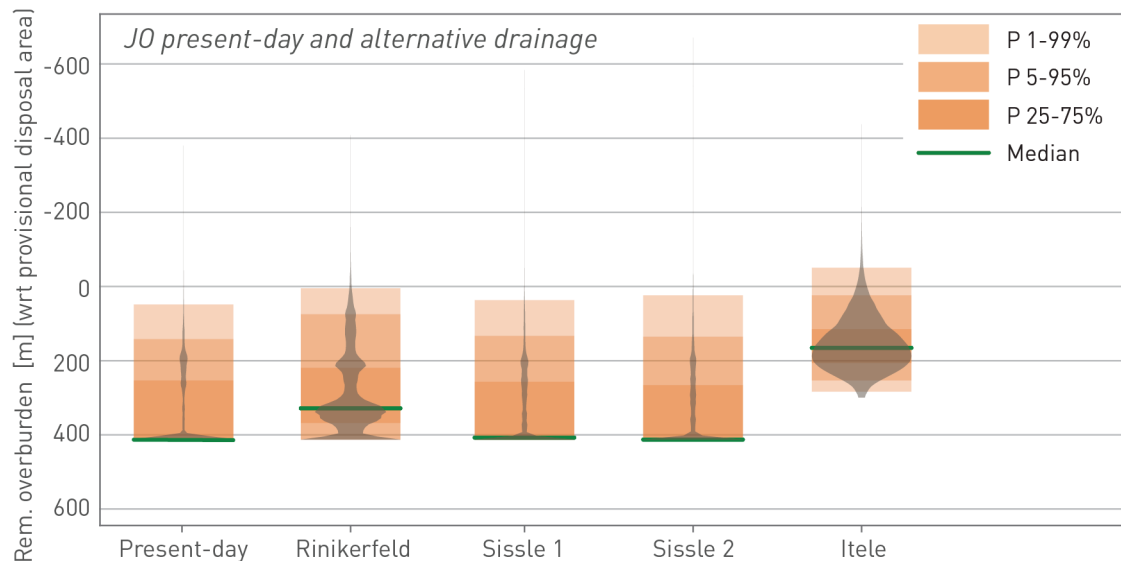


Fig. 3-31: Distributions of remaining overburden thickness after one million years including local topography for different drainage scenarios in Jura Ost

Violin plots and boxes with selected percentiles of remaining overburden thickness with respect to the evolution of local topography at the location of the provisional disposal area in Jura Ost. The figure compares the distributions in the case of preservation of the present-day main drainage network vs. alternative drainage networks (see Fig. 3-24). Distributions for JO present-day, Sissle 1 and Sissle 2 "converge" at the value of the minimum present-day overburden thickness because of the capping routine discussed in Section 3.4.5. The separate results show that especially the Itele drainage scenario, directing the Aare right across the provisional disposal area, reduces the local remaining overburden thickness considerably. A respective weighing of present-day and alternative JO drainage scenarios by the probability of occurrence (Tab. 3-6) is incorporated in Fig. 3-30b and in the uncertainty estimates for the evolution of local topography within the probabilistic model.

Overall, including local topography either based on the present-day drainage network or by including alternative drainage scenarios has a significant positive effect, i.e. larger values of remaining overburden thickness for JO as compared to accounting for the future local erosion base only (Fig. 3-15d). Nevertheless, the probability of yielding less than 200 m overburden thickness is still higher as compared to the conservative result for NL and ZNO with the future local erosion base as the reference (Fig. 3-30b). Using the present-day drainage network (Fig. 3-30a) yields the most promising result for ZNO among the three sites. However, considerable reduction in remaining overburden thickness is observed for ZNO, when accounting for the future local erosion base only (Fig. 3-30b). For ZNO, the potential range for the evolution of local topography can be considered relatively well covered by these two endmembers. For NL, in contrast, both approaches are conservative by implying a main river (Glatt River for the present-day drainage system) to flow directly above the provisional disposal area. Other main drainage networks, where a future main river would be shifted further away from the provisional disposal area, are not covered by configurations in Fig. 3-30 for NL but might add further protective relief in the future. Taking the low-relief regions of ZNO and the eastern parts of NL into account, however, a main river flowing directly above the provisional disposal area remains a relevant scenario for both sites, with NL being more robust in this respect (Fig. 3-30b).

The results shown here are readily interpreted as overburden thickness with respect to decompaction effects but are also partly used for comparison with the combined effects of non-glacial and glacial erosion (Sections 3.5 and 4.1).

3.5 Deep glacial erosion

3.5.1 Introduction to processes and drivers of deep glacial erosion

The main processes of glacial erosion are abrasion and quarrying. Abrasion involves the wearing down of rock surfaces by the grinding effect of glacier ice charged with basal debris (Fig. 3-32). Quarrying, or plucking, is the removal of well-jointed or loosened blocks of bedrock by the overriding glacier (Fig. 3-32). For abrasion and quarrying to take place, there must be relative movement between the glacier sole and the bed, which requires the basal ice to be at the pressure-melting point and sliding. Abrasion and quarrying therefore depend on the temperature and velocity of the basal ice, debris concentrations, ice thickness, and subglacial water pressure (Iverson 2002, Bennett & Glasser 2009, Burki 2009, Swift et al. 2015).

Significant wear to the glacier bed can also be caused by subglacial fluvial action. Erosion of bedrock in subglacial meltwater channels proceeds in a manner similar to the bedrock incision by subaerial streams and depends on water velocity, discharge, and turbulence as well as the quantity of suspended sediment and bedload in traction (Drewry 1986, Burki 2009, Benn & Evans 2010, Swift et al. 2015). Of specific importance in subglacial environments is that high water velocities and discharge require strong pressure gradients in relatively large, hydraulically efficient and well-connected subglacial channels. The efficiency of subglacial drainage not only affects meltwater erosion but also likely plays a predominant role in evacuating sediments produced by erosion and, thus, keeping bed surfaces accessible for further erosion.

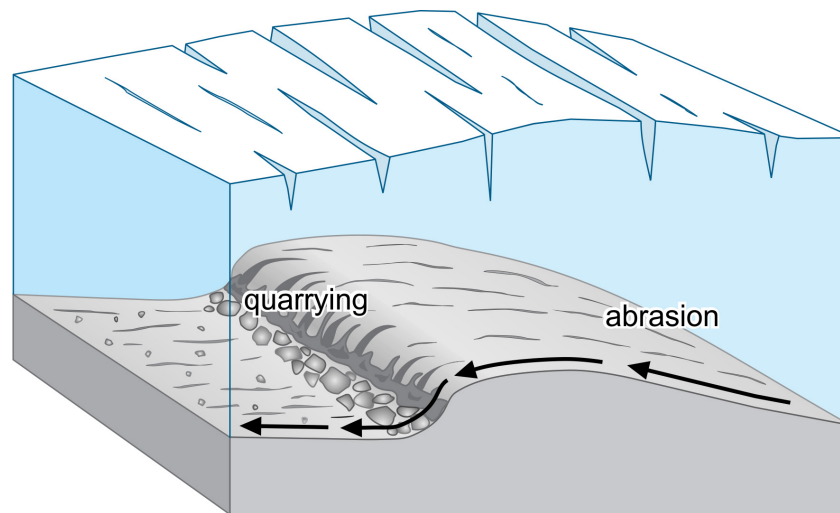


Fig. 3-32: Glacial erosion processes

Abrasion of bedrock surfaces by dragging ice-entrained basal debris across the glacier bed. Quarrying (plucking of blocks of bedrock) due to stress concentrations where ice moves over bedrock steps.

Glacial erosion occurs where ice is warm-based, which enables entrainment of debris, basal sliding and thus abrasion. Where surface meltwater is produced and reaches the bed, this promotes rapid sliding and stimulates quarrying or may directly lead to subglacial meltwater erosion. Concerning integrated effects over large areas, the processes of abrasion and quarrying may be largely responsible for the overall eroded volumes (Drewry 1986, Burki 2009). However, with respect to maximum possible erosion depths, subglacial fluvial action likely plays an important role (Fischer & Haeberli 2010).

Glacial erosion is closely linked to spatial patterns of ice flow and sediment transport and deposition (Alley et al. 1997, 2019). Maximum erosion is expected at a location close to the long-term average equilibrium line altitude (ELA) (Fig. 3-33a) where ice flux and hence basal sliding is large (Hallet et al. 1996, Boulton 1996, Anderson et al. 2006). However, without the evacuation of sediment by water from the ice-bed interface, the products of erosion may accumulate and inhibit further erosion (Alley et al. 2003) (Fig. 3-33b). Erosion may therefore be more efficient near the warm-based glacier terminus where abundant subglacial meltwater commonly flows in hydraulically efficient channels that have a high sediment-transport capacity (Alley et al. 1997) (Fig. 3-33c). Due to the ability of ice to move water and sediment along adverse bed slopes, continued erosion and deepening of the glacier bed will produce a closed topographic depression known as an overdeepening (Linton 1963, Harbor 1989, Evans 2008; Fig. 3-34). Localised bed-rock quarrying may also play a significant role in glacial overdeepening, whereby the formation of crevasses above initial perturbations caused by bed irregularities focuses the delivery of surface melt and, hence, water pressure fluctuations to the head of incipient overdeepenings (Hooke 1991, Iverson 1991; Fig. 3-33d).

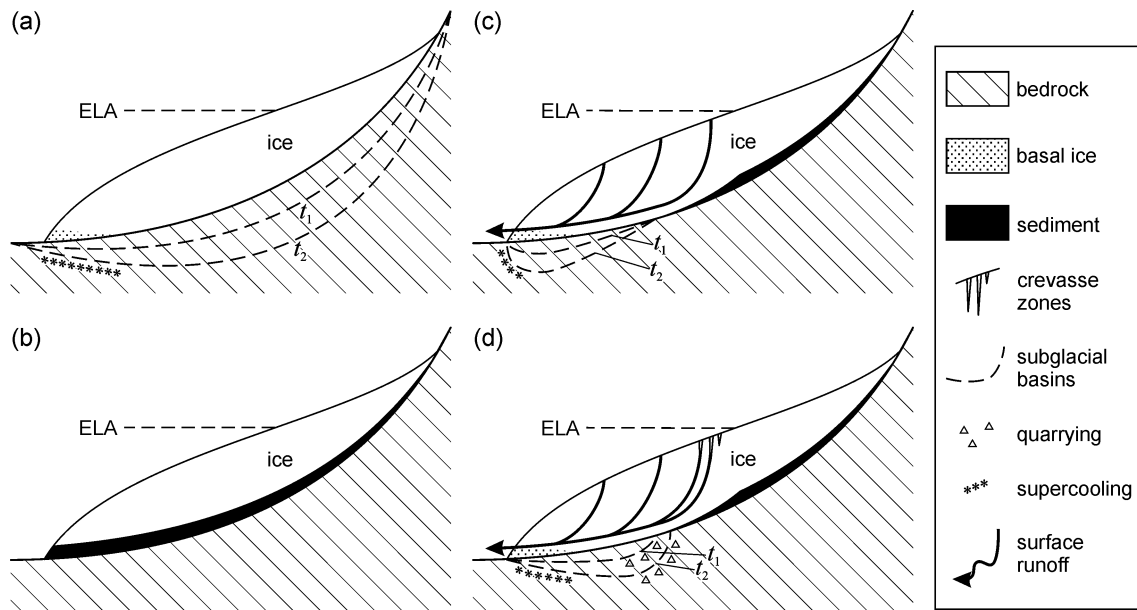


Fig. 3-33: Schematic representation of processes associated with glacial erosion

Dashed lines indicate glacier bed after erosion during two different time intervals. (a) Maximum erosion near the equilibrium line altitude (ELA) without hydrological forcing of basal sliding and sediment evacuation. (b) No erosion as a result of erosion products having accumulated to form a sediment layer that protects the bed. (c) Efficient erosion near the glacier terminus as a result of subglacial meltwater enhancing basal sliding and sediment evacuation. (d) Enhanced quarrying at the overdeepening head as a result of focused delivery of surface meltwater through crevasses (modified from Cook & Swift 2012).

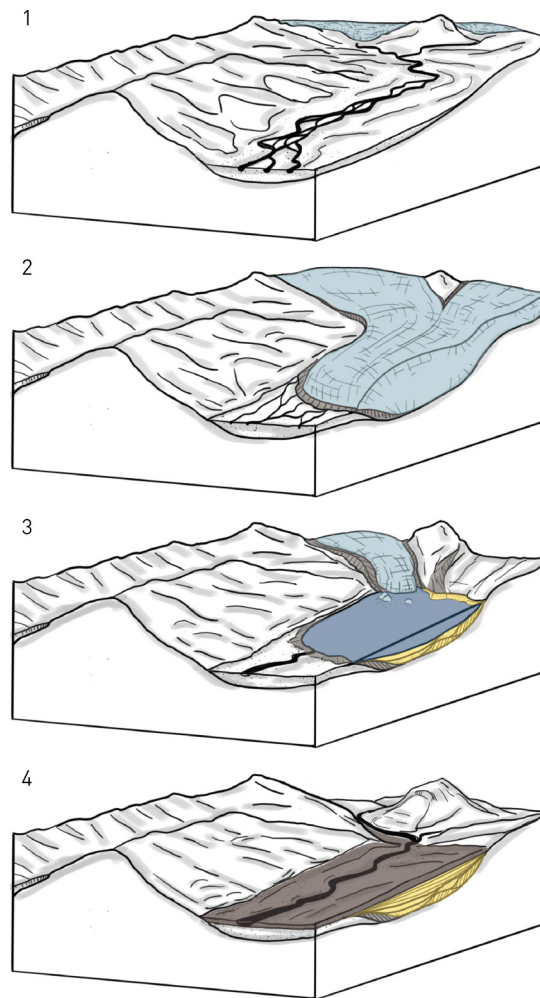


Fig. 3-34: Conceptual local landscape evolution by the formation of a glacial overdeepening

The sequence starts with a pro-glacial main valley (1; glacier shown in light blue). Glacial advance and erosion in (2) cause the bedrock surface to be lowered significantly below the fluvial baselevel, thus forming a glacial overdeepening. Glacial till (greyish brown) is deposited partly beneath the glacier and in lateral and terminal moraines. Upon glacial retreat (3), a lake evolves within the overdeepened trough that is gradually filled with lake deposits (yellow; not further differentiated). The final stage (4) illustrates the overdeepened trough (former lake) completely backfilled and the valley floor being covered by fluvial deposits (dark brown), reaching above the former fluvial baselevel. Drawings by A. Voigtländer.

The formation of glacial overdeepenings is likely the result of the combined action of ice and basal water and depends on the extent to which processes and feedbacks are able to focus erosion and maintain the evacuation of water and sediment (Cook & Swift 2012). The deeply incised troughs and overdeepened valleys in the Swiss Plateau presumably formed time-transgressively during periods of ice advance, when the steep surface slope of the ice margin provided the large hydraulic gradients necessary for a vigorous subglacial water flow and an efficient evacuation of erosion products from the ice-bed interface. They may have also formed during the retreat phase by the outburst of subglacial meltwater dammed behind a cold-based ice margin or a marginal permafrost wedge (Piotrowski 1994, Hooke & Jennings 2006, Stumm 2010, Kehew et al. 2012, van der Vegt et al. 2012). The finding that pre-LGM sediments are preserved within many overdeepenings (Dehnert et al. 2012, Buechi et al. 2018) suggests that not all glaciations were

able to excavate the postglacial sediment infill and cause renewed overdeepening. In ice-marginal areas this preservation is often the result of changes in spatial focus and magnitude of subglacial erosion between different glacial periods (Buechi et al. 2024). On the other hand, progressive deepening over successive glacial cycles may also have taken place because ice is likely steered away from hills and ridges into already existing troughs and valleys (Kessler et al. 2008). Such topographic steering causes the thicker and faster-flowing ice to generate more basal melting that will further increase basal sliding and sediment evacuation and thus result in enhanced erosion.

Notably, most of the deeply incised troughs and overdeepened valleys in the Swiss Plateau are carved into rather poorly consolidated Cenozoic marls and sandstones (Molasse substratum, see Section 2.4.2). The few overdeepenings cut into Mesozoic limestone are considerably shallower and narrower (see for instance Fig. 2-31, Fig. 2-32, and Fig. 3-35). This suggests a potential influence of bedrock lithology on the efficiency of erosion, for example, because large amounts of subglacial water may be lost into (provided there is) underlying limestone karst (a process known as underdraining) (Grust 2004, Gremaud & Goldscheider 2010, Braakhekke et al. 2017), thereby preventing the build-up of high water pressures that are favourable for active erosion, and the energetic water flow necessary for sediment evacuation.

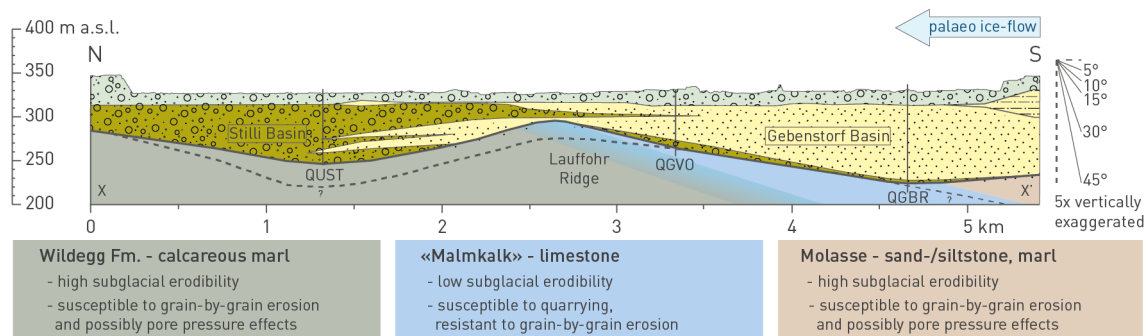


Fig. 3-35: Longitudinal profile through the Gebenstorf – Stilli Trough close to Brugg based on seismic cross-sections and boreholes (Gegg et al. 2023)

The orientation of the profile and relative amounts of erosion indicate that the Malm limestone forms a resistant zone in which the overdeepening capacity decreases relative to the Molasse to the south and calcareous marls to the north. Modified after Gegg et al. (2021).

3.5.2 Workflow to assess future potential for deep glacial erosion

Model chain and key conditions in Northern Switzerland

For future deep glacial erosion to compromise the deep geological repository, a series of events must occur. Firstly, ice during a future glaciation must advance sufficiently to cover the siting region (concept of key glacials). Secondly, the erosion must be deep enough to reach the level of the repository. This depth is a function of the future local erosion base, the size and erosional potential of the glaciation, and the incised rock type. Thirdly, the overdeepening must incise at the location of the future repository. Different spatial scenarios for overdeepening locations can be drawn from Quaternary observations and additional considerations.

Unlike the simulation of the future local erosion base and local topography, the assessment of deep glacial erosion did not include numerical landscape evolution modelling. Instead, a chain of calculations using expert knowledge-based parameter distributions (which are largely based on the information provided by Chapter 2) link the above outlined aspects of depth, location, and erodibility (Fig. 3-36). These will be summarised below.

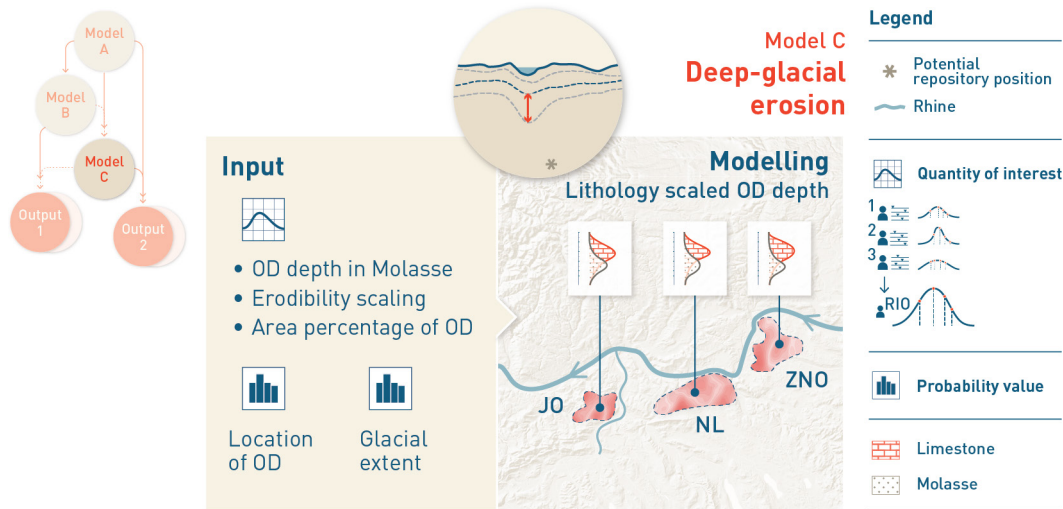


Fig. 3-36: Illustration of workflow to assess the parameters and combinations for the development of scenarios for deep glacial erosion

This approach combines several aspects to be evaluated with respect to future deeply eroded glacial troughs. The main input parameters are depth distribution (below local erosion base), erodibility scaling, the geological model (3D stratigraphy), different scenarios, and timing of the respective future glaciation, as well as the spatial occurrence and density of glacial overdeepenings within the siting region.

The statistical event that the HLW region is covered by ice is modelled. Experts can define a probability for this to occur. This implies a specific timing for which the assessment is relevant. In this case, the calculations are performed for the end of the period of concern for HLW, i.e. one million years after closure of the deep geological repository. It is important to note that the potential occurrence of a glaciation shortly before one million yr as highlighted in Fig. 3-46 was used for calculation case for one million yr.

Three key glacials with corresponding glacial extents and erosion potential associated to such events were defined based on the Quaternary observations (compare Fig. 2-30), which are complemented by the "no glacial cover" case. Accordingly, the cases are:

- none of the siting regions are covered
- only ZNO (and east part of NL) is covered (e.g. LGM-size)
- ZNO and NL (and east part of JO) are covered (e.g. MIS 6-size)
- ZNO, NL, and JO are covered (e.g. MEG-size [most extensive glaciation])

Key glacials were considered as follows: MEG-size (Möhlen) as a large glacial with significant bedrock incision, MIS 6-size (Beringen) as a mid-size glacial with significant amount of bedrock incision and erosion and filling of established overdeepenings, and the LGM as a small glacial with minor known bedrock incision within the foreland, but erosion and filling of inlaid basins within established overdeepenings (see Section 2.4.2.4).

Spatial overdeepening occurrence is assessed through a number of different overdeepening scenarios, which are modelled as a set of statistical events. Each of these events is associated with two regions:

- One region (location of overdeepenings, see Fig. 3-38 as example and Fig. 3-43 for the full maps) in which OD can potentially occur. This is fixed for each event.
- A second one (areal percentage), which is a random variable of which only the size is fixed but the location of which is assumed to be unknown within the location of overdeepenings (see Fig. 3-38).

Overdeepening depth distribution is defined for each of the key glacials and for incision into Molasse sediments. An overdeepening depth of 0 m equals the local erosion base (and is actually an oxymoron because it would, by definition, not be overdeepened). Overdeepening depth is then scaled to account for the 3D stratigraphy and different erodibility of rocks below Molasse sediments.

Influence of repeated glaciations, on overdeepening depth and spatial occurrence

Repeated glaciation events are not calculated on a one-by-one basis, which could be ultimately summed or stacked. Instead, repeated glaciations were implicitly considered within the argumentation of the depth distributions (Fig. 3-37). This implies a more direct relation between observations from the Quaternary (e.g. Fig. 2-27 and Fig. 2-33) and the estimate for the future. Considering Preusser et al. (2011) (see Fig. 2-27), there were about four larger ice advances of "key-glacial" size since Middle Pleistocene (1× Möhlen-size, 1× Beringen-size, 2× LGM-size). Earlier glaciations may have reached the foreland as is indicated by intercalated till within Deckenschotter deposits. It seems, however, that the above-mentioned four large glaciations took place only after the accumulation of the TDS and probably after (or associated with) the Mid-Pleistocene Transition (MPT), when the glacial cyclicity switched from a dominant 41-kyr to a 100-kyr cycle (Sections 2.2 and 2.4.2). The ages of the Möhlen and Beringen glaciations are not unambiguously resolved. The so far oldest dating of an overdeepening fill points to 606 ± 122 ka (Tomonaga et al. 2024; Section 2.4.2.3) and thus MIS 12 to 16. In summary, this suggests that the observed overdeepenings today are the cumulative result from the past ~ 800 kyr and more. Accordingly, the overdeepening depth could be considered an "implicit stack" (e.g. Fig. 3-37d and e). Because depth distribution is placed below the future local erosion base, the most influential overdeepening with respect to its distance to the repository level is one that occurs late (when the local erosion base is already lowered significantly), and which stems from a large glaciation with high glacial erosion potential (e.g. Fig. 3-37f).

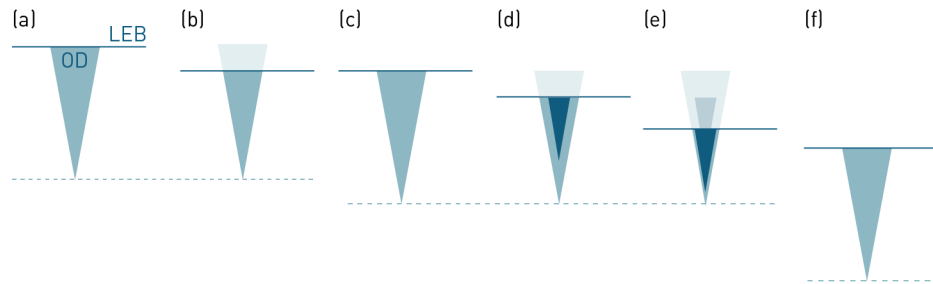


Fig. 3-37: Illustration of overdeepening formation during repeated glaciations

Glacial overdeepening (OD) is defined as glacial erosion below the (fluvial) local erosion base (LEB). The cross-sections above illustrate a potential temporal sequence of overdeepening formation. (a) Present-day situation. A glacially eroded trough reaches a certain overdeepening depth (dashed line). (b) Without a subsequent glaciation, the local erosion base will be lowered, eroding the upper sedimentary cover, reducing the overdeepening depth, while the glacial erosion base does not change. (c) A new large glaciation with significant glacial erosion occurs and incises to a deeper level (while the OD depth is similar to the present-day case). (d) and (e) show cases of repeated glaciations at later stages, where the LEB is further lowered, but the glaciations and associated overdeepenings are smaller and do not erode deeper than the previous level. (f) The most influential overdeepening with respect to its potential to erode down to the level repository level is one that occurs late (when the local erosion base is already lowered significantly), and which stems from a large glaciation with high erosion potential.

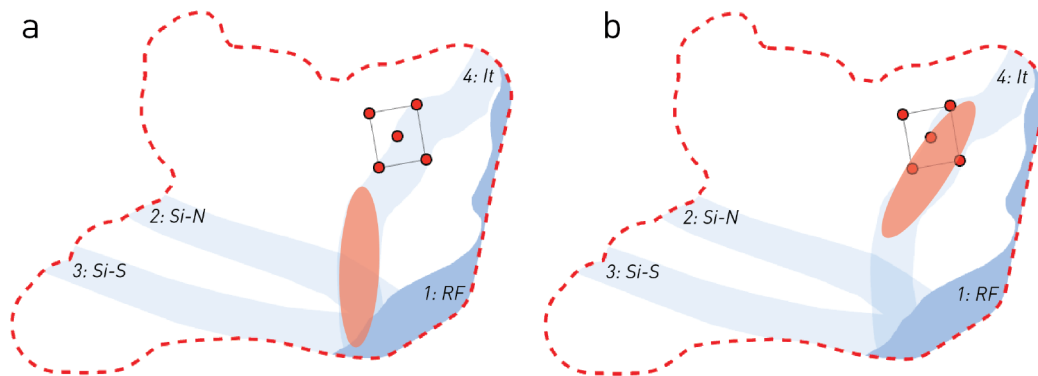


Fig. 3-38: Spatial characteristics of overdeepenings and associated areal density

The example of Jura Ost highlights four possible discrete scenarios of overdeepening occurrences in blue: RF: Rinikerfeld channel, Si-N: Northern Sissle breakthrough channel, Si-S: Southern Sissle breakthrough channel, and It: Itele breakthrough channel. Along these discrete scenarios it was considered plausible that overdeepening may occur (the darker the colour the more likely overdeepening occurrence was assumed). However, the area that may actually be overdeepened within such a location (red ellipse) may be smaller than the full size of the polygon delineating the respective overdeepening scenario, i.e. smaller for instance than the Itele breakthrough channel, whose location overlaps with the provisional disposal area (rectangle). Accordingly, the model algorithm includes the binary decision whether (b) or not (a) the domain in which an overdeepening scenario (e.g. Itele) overlaps the disposal area (rectangle, represented by five points) is actually overdeepened.

Repeated glaciations were also implicitly considered within the argumentation for spatial occurrence of glacial overdeepenings. For instance, it was argued that repeated future glaciations imply more random future occurrence (instead of reoccupation of present-day overdeepenings) and a higher potential for new future breakthrough channels to occur. The assemblage maps of the overdeepening scenarios (Fig. 3-43) should thus be understood as the cumulative map of where to expect glacial overdeepening to occur at the time under consideration, in this case one million years.

3.5.3 Elicited parameters: rationale and argumentation

3.5.3.1 Depth distributions of deep glacial erosion

Quaternary insights as a basis for future glacial erosion depth

The assessment of future glacial erosion depth is informed by Quaternary glacial history and is provided as probabilistic distribution of potential overdeepening depths below the local erosion base at the provisional disposal area. Glacial overdeepenings, which vary along their axes (Fig. 2-11b), are evaluated with particular attention to the erodible Molasse formations in Northern Switzerland, where most observed overdeepenings occur (Section 2.4, Fig. 2-31a). Accordingly, the reference depth of glacial overdeepening (i.e. with respect to Molasse rocks) is defined as the uncertainty distribution of the total overdeepening in metres below the local erosion base at the location of the provisional disposal area.

For each repository site, reference depth distributions were developed for key Quaternary glaciations, specifically the Möhlin-size, Beringen-size, and Last Glacial Maximum (LGM)-size extents (Fig. 2-30). These distributions for Molasse rock were determined through expert judgment, using data on overdeepening depths from the North Alpine Foreland and the HLW perimeters, along with insights from QBO boreholes and selected global analogues (Fig. 3-39).

Because these key glaciations are derived from past Quaternary events, the depth distributions for future erosion can be estimated independently of specific predictions about future glacial extents. However, when assessing remaining overburden thickness or excavation probability (see Section 3.2), these depth distributions are conditioned on the probability of each key glacial event's occurrence (Sections 3.5.3.4, 3.5.4, Chapter 4).

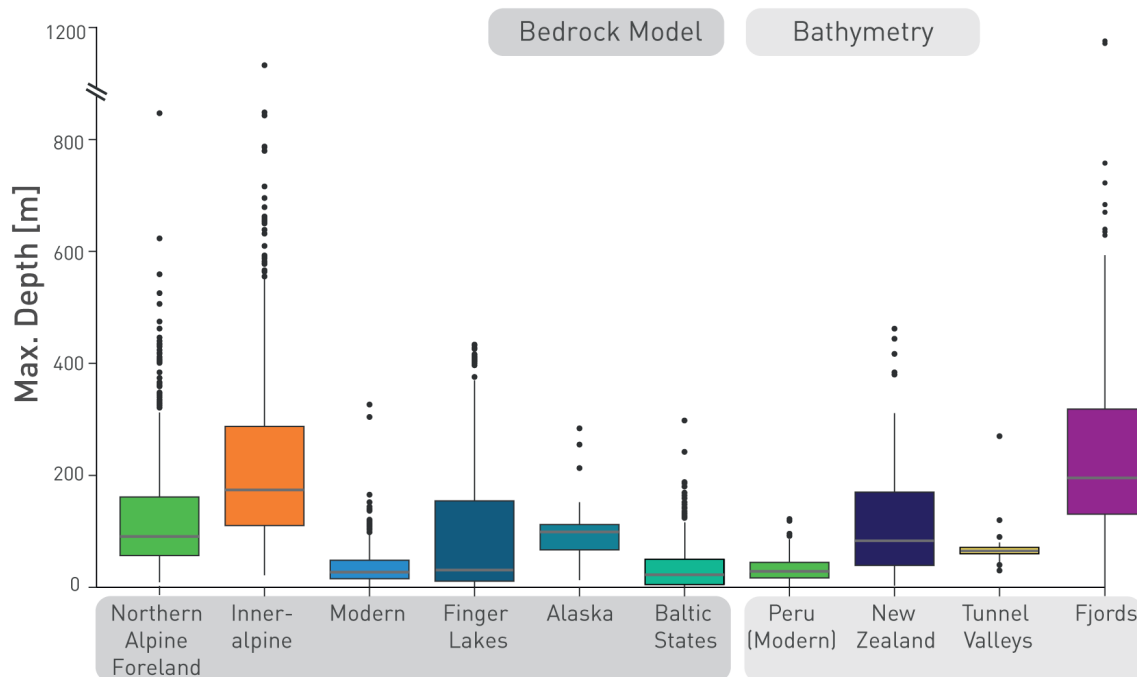


Fig. 3-39: Global analogue examples of glacial overdeepenings (depth distribution)

Boxplot of maximum overdeepening depths for various regions. Note the break on the y-axis. The box extends from the 25th to 75th percentiles and whiskers from the 10th and 90th percentiles; the median is represented by the solid grey line. Glacial overdeepenings in Northern Switzerland, Austria & Southern Germany were defined as the area between the surface of the bedrock and the level of the main rivers. Overdeepenings in other regions represent either the thickness of the Quaternary sediment infill (Finger Lakes NY, USA), parts of Alaska) or the bathymetry (with maximum depths as minimum values). Largest maximum and medium depths of overdeepenings are found in inner-Alpine Valleys and in the fjords (caused e.g. by topographic steering). In contrast, overdeepenings beneath modern glaciers in high-Alpine settings are rather shallow. Note that boxplots from glacial overdeepenings in the Finger Lakes and New Zealand (excluding fjords, and although based on bathymetry) show a comparable statistic.

Considerations for future glacial erosion potential

Two main considerations from glacial history and process understanding are applied to refine the model of future deep glacial erosion potential, balancing opposing effects on erosion rates:

Potential for increased erosion in the Rhine catchment: Ellwanger et al. (2011) suggest that, during past glaciations, the Bodensee region experienced intensified glacial erosion as the Rhine catchment expanded westward, redirecting more ice away from the Danube and towards the Rhine. This could mean that a future Rhine glacier, especially in ZNO and NL, might further intensify erosion, while JO would be less impacted, having historically been influenced by Reuss and Rhone glaciers. The observed maximum erosion in the Bodensee Trough, located east of the siting regions, is used to establish extreme upper bounds for erosion depth distributions in ZNO and NL (Fig. 3-40).

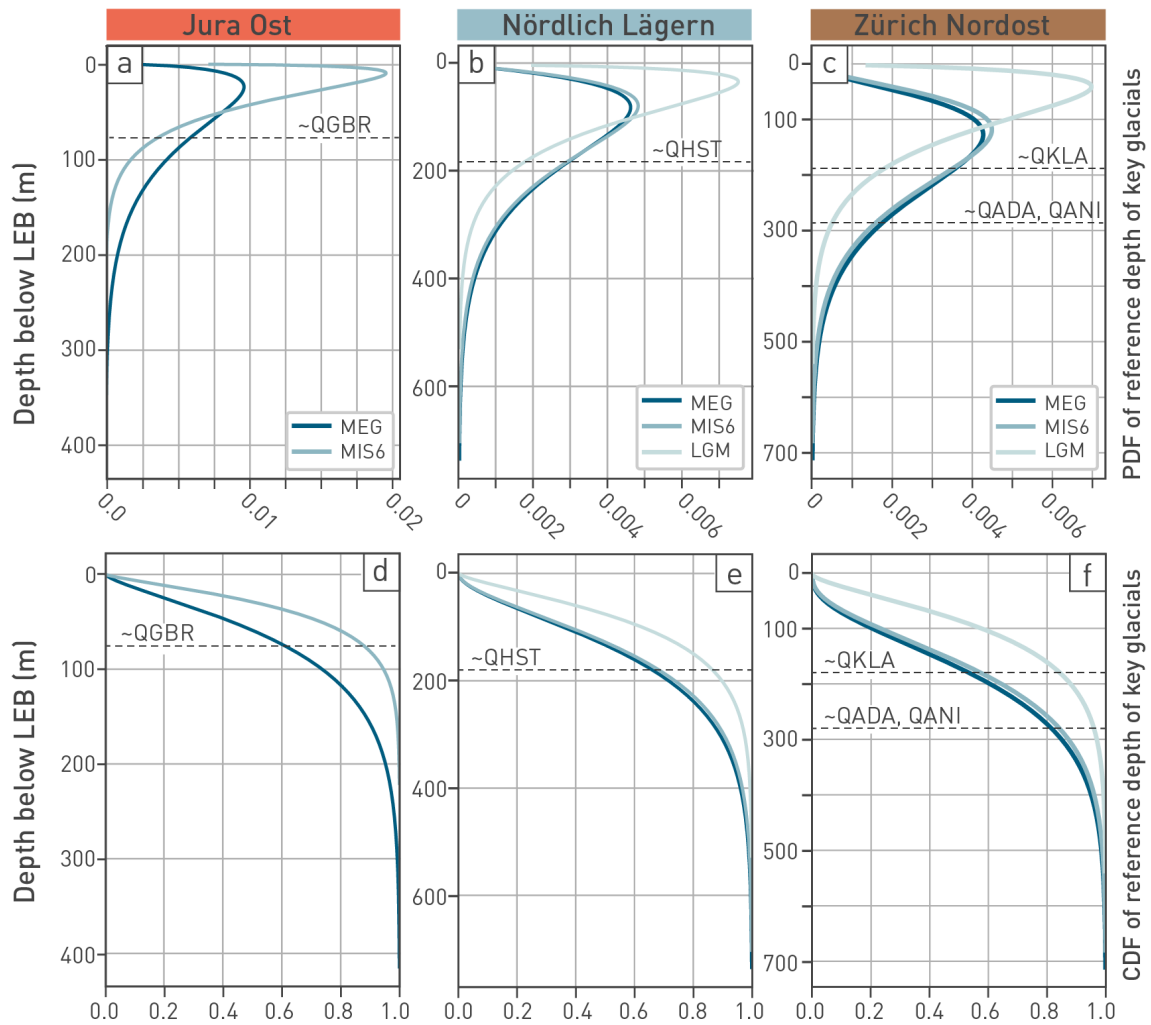


Fig. 3-40: Total assumed probabilities for reference overdeepening depth per key-size glaciation

Total assumed probabilities for reference depth distributions (Molasse) for each site and the respective key glacial. Note that MIS 6 refers to the Beringen glaciation and MEG to Möhlin. Figures in the upper panel show the probability density function (PDF), those in the lower panel show the respective cumulative distribution function (CDF) for Jura Ost (left), Nördlich Lägern (centre), and Zürich Nordost (right). Horizontal dashed lines correspond to the approximate depths (below LEB) of Quaternary boreholes that Nagra drilled within the wider areas of the siting regions (Fig. 2-11a). QGBR: Gebenstorf-Brüel borehole (Gegg et al. 2019) is located within the Lower Aare Valley, but not within the HLW area of JO. QHST: Hochfelden-Strassberg borehole (Amschwand et al. 2020) is located in the southernmost part of NL. QKLA: Kleinandelfingen-Laubhau borehole (Schuster et al. 2020) is located in the southernmost part of the ZNO-HLW area. QADA: Adlikon-Dätwil borehole (Buechi et al. 2022) is located in the south, just outside of the ZNO-HLW area. QANI: Andelfingen-Niederfeld borehole (Reichenwallner et al. 2021) is also located in the south, just outside of the ZNO-HLW area.

Possibility of reduced erosion due to downcutting: Progressive downcutting from past erosional processes could reduce the accumulation area for future glaciers, potentially limiting glacial extent and associated erosion. This effect, as suggested by Egholm (2022), implies that, even with similar climatic conditions, future glaciations might lead to less extensive glaciation and reduced erosive power.

Specific considerations for each siting region and associated key glacials

Fig. 3-40 shows the total assumed probabilities for the reference depth (Molasse) of future glacial overdeepenings of associated key glacials at each site, i.e. all three key glaciations for ZNO, Beringen and Möhlin for NL and Möhlin for JO. To account for uncertainty in the extent of the respective key glacial, depth distributions for the "next-smaller glaciation" were additionally elicited for JO (i.e. Beringen-size) and NL (i.e. LGM-size). Tab. 3-7 lists the parameters of fitted total assumed probabilities. Below, the specific considerations for each siting region and associated key glacials are briefly listed.

Tab. 3-7: Fitted distributions for reference depth of overdeepening in Molasse

Parametric distribution models for the reference depth of overdeepening in Molasse were elicited in a series of expert workshops. The elicited distributions represent the uncertainty about the depth of overdeepening (below LEB) in Molasse in each siting region and with respect to the key glacials.

Glaciation	Siting region	Distribution	Mean [m]	Standard dev. [m]
MEG	JO	gamma	75	62
	NL	gamma	159	112
	ZNO	gamma	190	110
MIS 6	JO	gamma	38	33
	NL	gamma	152	108
	ZNO	gamma	179	105
LGM	NL	gamma	96	79
	ZNO	gamma	104	81

Reference overdeepening depth distributions for Jura Ost

Today, there is no glacial overdeepening within the HLW area of Jura Ost. Three main reasons are to be considered: (1) Only the largest foreland glaciations reached this siting region (see Fig. 2-30). (2) The topographic relief provides a potential obstacle and prevents fast sliding, thus reducing erosional potential. (3) The siting region is located outside the Molasse Basin and surface lithologies in the vicinity comprise harder-to-erode rocks (see Fig. 3-35). The closest glacial overdeepenings can be found within the Lower Aare valley (Section 2.4), but even there, overdeepening depth is limited. The Nagra borehole Gebenstorf-Brüel (QGBR) reached bedrock at ~ 81 m below the local erosion base (Fig. 3-40a and b, Fig. 3-11a).

Accordingly, only a MEG-type glaciation was considered important for JO. However, because the suggested disposal area is located close to the margin of the MIS 6-type glaciation and to

allow some uncertainty in the delineation of the glacial extent, two depth distributions were elicited using the SHELF protocol (MEG- and MIS 6-type, see Fig. 3-40a and d and Tab. 3-7, see also Section 4.1).

Considerations for a Möhlin-size glaciation in Jura Ost

According to the above-mentioned arguments, the total probabilities should focus on shallow values. The lower limit was located at zero metres, which corresponds to the local erosion base and means that no glacial overdeepening is observed. The upper bound was placed at 300 m, which is a bit higher than observed overdeepenings in the Birrfeld area south of Brugg (Fig. 2-31).

If deep glacial erosion was to occur in such a future glaciation, it was considered most likely that overdeepenings could reach depths between 35 and 90 metres (considering Molasse bedrock), thus being consistent with observations from the Lower Aare valley nearby (see Fig. 3-37 and Fig. 3-40a and b), but allowing slightly larger-than-observed values because very shallow overdeepenings in the Lower Aare valley were eroded into limestones – a fact that is considered separately with the lithological scaling (Fig. 3-42).

It was considered likely that overdeepenings in JO considering a future Möhlin-size glaciation (given that there are breakthrough channels) could be between almost absent and up to ~ 150 m deep. Such higher-than-observed values might also incorporate the uncertainty about the height of the palaeo-baselevel: if the past glacial overdeepenings were older than the deepest fluvial bedrock incision that forms the LEB, they may be linked to higher fluvial palaeo-baselevels and overdeepening depth may be underestimated. The chosen gamma distribution, which best fits the arguments, overrepresents the elicited values for the 95% percentiles, allowing nearly 200 m deep overdeepenings. It was agreed to accept this higher probability of larger values instead of shifting the inner tertiles of the distribution accordingly.

Considerations for a Beringen-size glaciation in Jura Ost

The low chance of ice even reaching the site and creating an overdeepening during a Beringen-size glaciation (Section 3.5.3.5) should be reflected in the limits of the depth distribution. Accordingly, the total probability should be located only between 0 and 200 m. Nevertheless, overdeepening depths similar to those already observed in the Lower Aare Valley (e.g. in the QGBR borehole) were considered likely. The most likely third of the distribution was placed between ~ 20 and 50 m, reflecting additional depths observed presently in the Stilli Trough (see Fig. 3-40a and b).

Reference overdeepening depth distributions for Nördlich Lägern

The eastern part of Nördlich Lägern comprises two significant glacially overdeepened troughs, i.e. Strassberg and Bülach Troughs, that reach into the siting region coming from the south (Fig. 2-31 and Fig. 2-32). The sedimentary fill of the Strassberg Trough was dated using noble gas methods (see Section 2.4) to be 606 ± 122 ka (Tomonaga et al. 2024), thus likely being eroded during MIS 12 – 16. It is potentially associated with the Möhlin Glacial. Statistics from the overdeepenings within the HLW area of NL show that the majority of overdeepenings are eroded less than 100 m below the local erosion base, with a maximum of 185 m below LEB, which was observed in the QHST borehole (Fig. 2-11a, Fig. 2-12b). Further upstream, the Bülach and Strassberg Troughs converge and show general overdeepening depth up to approximately 200 m below LEB, with a higher median value in the Strassberg than in the Bülach Trough (Fig. 2-12a).

Considerations for a Möhlin-size glaciation in Nördlich Lägern

The lower limit was naturally located at zero metres, which corresponds to the local erosion base and means that no glacial overdeepening is observed. The upper bound was placed at 550 m, which is an extreme value associated with the Bodensee at the Alpine front (Fig. 2-12b). Such

high values do not seem plausible for the more distal foreland, thus providing a hard upper limit. By definition, the glacial overdeepening considers depth below local erosion base. This, however, has lowered during the past one million years. Adding at least ~ 120 m to the observed depth of the Strassberg Trough should account for this effect. Accordingly, depths up to 350 m were considered still likely for a Möhlin glaciation and are represented by the 95% quantile. On the lower part, it was acknowledged that data from the QBO campaign and other boreholes show that the majority of overdeepenings in NL are shallow with respect to the present LEB (Fig. 2-11a and Fig. 2-12). 25 m was considered to best represent the 5% quantile. In general, it was agreed that the estimates for the overdeepening depth in NL should be in the order of, but somewhat smaller than, those for a similar glaciation in ZNO. The most likely third was placed between 95 and 195 m, with a median of 135 m. All these values are considerably deeper than the statistics of the observations within the HLW perimeter of NL (Fig. 2-12b). More than 75% of observed depth values are included in the lowest third of the estimated future distribution, the median value is about 100 m higher than observed. These considerations account for the uncertainty associated with a potentially higher fluvial palaeo-baselevel than the LEB. The most likely third should also include the observed value of ~ 185 m of QHST.

The chosen gamma distribution which best fits the arguments overrepresents the elicited values for the 95% percentiles, allowing nearly 380 m deep overdeepenings. The higher tertile underrepresents the elicited value, but almost captures the QHST observation. It was agreed to accept these deviations (Fig. 3-40c and d).

Considerations for a Beringen-size glaciation in Nördlich Lägern

Because of the uncertainty in age control of the observed overdeepenings, the estimates for the Möhlin and Beringen glaciations are very close to each other. The limits were kept similar (0 – 550 m). Overdeepening depths between 25 and 340 m were considered likely, with a similar argumentation as for the Möhlin glaciation. The other RIO values for a Beringen Glaciation in NL are between 5 to 10 m lower than for the Möhlin glaciation, resulting in a very similar total probability. Accordingly, the most likely third was estimated to be between 90 and 185 m with a median at 130 m.

Considerations for an LGM-size glaciation in Nördlich Lägern

An LGM-size glaciation covers only the easternmost area of the HLW perimeter of NL. To account for some uncertainty in the glacial extent of a similarly sized future glaciation and to assess the sensitivity of including this smaller key glaciation, the LGM estimates for the ZNO site were adopted (see below) and only the quantiles defining the inner third of the distribution were reduced by 10 m.

Reference overdeepening depth distributions for Zürich Nordost

The HLW perimeter of ZNO is located within the glacial extents of all three key glacials. Because of the closer proximity to the Rhine – Linth lobe, this siting region is more often glaciated than the others. In addition, the extents of the key glacials are closer to each other near ZNO than in the vicinity of the other siting regions (Fig. 2-30). Glacial overdeepenings are prominent within the HLW perimeter of ZNO (Fig. 2-11a and Fig. 2-12b). Especially the Marthalen and Andelfingen Troughs are distinct features (Fig. 2-32c). Although the statistics seems to suggest that the majority of overdeepenings are shallower in ZNO than in NL (Fig. 2-12b), it has to be noted that more total area of the HLW perimeter is overdeepened (Fig. 3-44). The maximum depth value is observed in ZNO (QKLA borehole), but the deepest parts that were drilled are located south, just outside the HLW area (QADA, QANI).

Considerations for a Möhlin-size glaciation in Zürich Nordost

Similar to NL, the limits were considered to be between 0 and 550 m (see arguments above). It was also noted that the outer bounds might be rather similar between the two larger glaciations (Möhlin and Beringen), because the age control of the observed overdeepenings is still inconclusive (see Section 2.4.2). The values of the inner third that comprise the most likely estimates, however, were chosen to be slightly deeper for the larger glaciation of the two. Accordingly, it was considered likely that overdeepenings can reach depths between 40 to 350 m, and are most likely to be between 130 and 225 m below LEB with a median at ~ 170 m.

The chosen gamma distribution which best fits the arguments overrepresents the elicited values for the 95% percentiles, allowing nearly 400 m deep overdeepenings. It was agreed to accept these deviations (Fig. 3-40e and f).

Considerations for a Beringen-size glaciation in Zürich Nordost

Similar to NL, the limits were considered to be between 0 and 550 m (see arguments above). Values between 40 and 340 m were considered likely (5 – 95%) and it was agreed that the observations of the QADA and QANI boreholes, which are nearly 300 m deep, should be well attributed within this limit. The most likely third was estimated to be between 125 and 215 m, with a median at 165 m. This third entails the depth of QBO borehole QKLA (195 m) and the values are only slightly lower than for a Möhlin-size glaciation at ZNO, but higher than for a Beringen-size glaciation at NL (see above).

Considerations for an LGM-size glaciation in Zürich Nordost

The depth distribution for glacial overdeepenings caused by an LGM-size glaciation was strongly influenced by the lack of observations of bedrock incision from this period. This, although the observation base is generally better than for the older glaciations. On the other hand, it was argued that the proposed Hagenholz glaciation (Preusser et al. 2011; see Fig. 2-27), which is believed to have produced overdeepenings, is of similar glacial extent to the LGM. Nevertheless, the total probability that was agreed upon generally shows significantly lower values than those for the other two key glaciations. The limits were considered to be between 0 and 450 m, and still likely up to 250 m. The LGM was associated with a rather low erosion potential that produced no bedrock overdeepenings but inlaid basins within already established overdeepenings (see Section 2.4.2.4). It was agreed that this aspect should be highlighted within the central third of the total probability. Accordingly, it was considered most likely that an LGM-size glaciation could produce overdeepenings between 50 and 120 m below LEB in ZNO, with a median at 90 m (Fig. 3-40e and f).

3.5.3.2 Lithological scaling of reference depth distribution

Deep glacial erosion is expected to slow down in more erosion-resistant rocks, such as limestone. This pattern is evident in areas like the Lower Aare Valley (Section 2.4.2 and Fig. 3-35) and New York's Finger Lakes region. For instance, in the Cayuga Lake Basin, subglacial incision appears to be limited upon reaching the Onondaga limestone layer (Mullins et al. 1996, Bloom 2018, see Fig. 3-41a). In such cases, subglacial erosion typically widens the overdeepening within softer materials rather than cutting deeper into harder rock layers (Fig. 3-41a upper panel and Fig. 3-41b). However, across bedrock sills, typically consisting of more resistant bedrock, narrow canyon-like structures have been proposed (e.g. Bandou et al. 2023).

In contrast, inner-Alpine overdeepenings formed in more resistant rocks, such as those found in the Alps, can reach substantial depths (e.g. Magrani et al. 2022). However, in these Alpine regions, topographic steering along preconditioned valleys seems to play a greater role than rock type in determining erosion depth.

To account for variability in erodibility, other rock types are factored relative to the Molasse assemblage, assuming the median erodibility (M) decreases from EC1 to EC4: $M(\text{EC1}) > M(\text{EC2}) > M(\text{EC3}) > M(\text{EC4})$. For example, rocks in EC4, such as Malm limestones, are considered on average to be half as erodible as Molasse rocks. Consequently, the depth of overdeepenings in different rock types scales according to erodibility, with deeper incisions in softer rocks when topographic constraints are minimal, and shallower incisions in harder rocks.

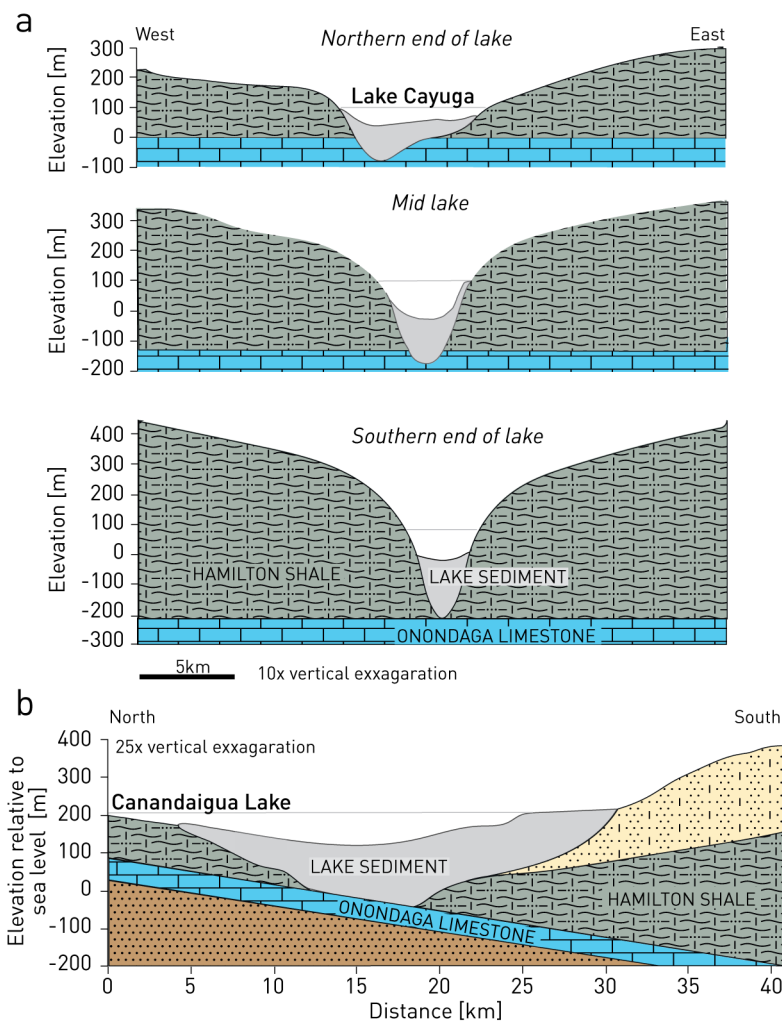


Fig. 3-41: Example of potential bedrock control on glacial overdeepening in the Finger Lakes region (NY, USA)

The figures are redrawn and simplified after Mullins et al. (1996) and show cross sections through Lake Cayuga (a) and a longitudinal section along Lake Canandaigua (b). These glacial overdeepenings incise a sequence of rock types with different resistance to erosion. It seems that incision occurs preferentially in the easier to erode Hamilton Shale, if space is available, and in down-dip direction of the bedding interface between the softer and harder formations.

However, variations within each erodibility class can occur, where specific conditions may yield different erosion resistances than the typical material properties. This is reflected in the probability overlaps shown in the total assumed probabilities (Fig. 3-42a).

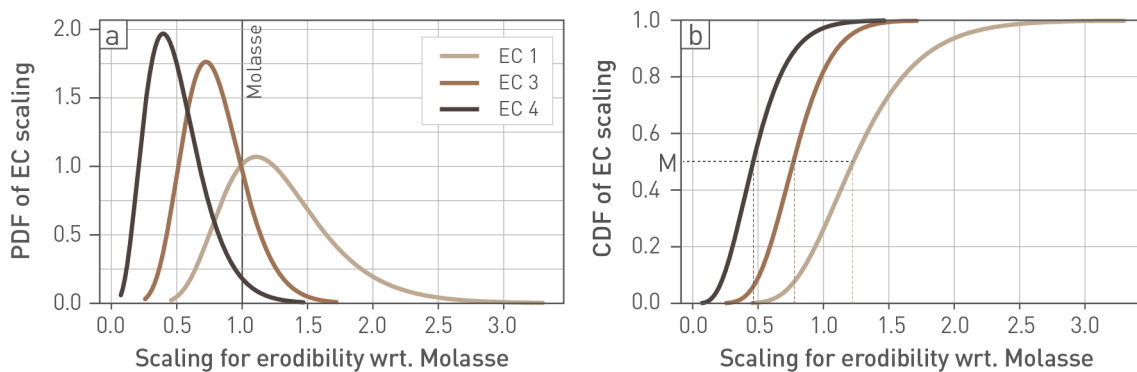


Fig. 3-42: Total assumed probabilities of erodibility scaling with respect to the Molasse of Northern Switzerland

The sediments of the Molasse assemblage of Northern Switzerland with their variety were considered as the base unit with the scaling value of 1.0 (labelled in a). Molasse sediments are within erodibility class EC2 (Fig. 3-11). Accordingly, rock types within EC1 are generally easier to erode, and rocks within EC3 and EC4 are more resistant than Molasse. Probability density function (a) and cumulative distribution function (b) are based on expert judgement under the prerequisite that the median values (M) follow the assumption of $M(EC1) > M(EC2) > M(EC3) > M(EC4)$. EC1 is best represented by a log-normal distribution, EC3 and EC4 by gamma distributions. Note that the tail of EC1 was cut for better readability of the graph. (b) Cumulative distribution function of the same estimates. All three erodibility classes are shown. Note that this factorisation is not directly comparable to the K value estimates for the same erodibility classes that incorporate more than material properties.

Considerations for EC4

The lower limit is set to 'zero'. Arguments for a zero-metre lower erosion limit focus on geological evidence and observations of minimal glacial erosion in certain areas. Studies in the Rhone Valley, near St. Maurice, where glacier thickness was significant, show stable bedrock that remained largely uneroded despite the ice cover. Similar cases are observed in regions with resistant limestone, where glacier movement either did not erode the bedrock or resulted in only very shallow erosion, around 5 metres at most. The "Giant Sieve" theory from North America (a process known as underdraining; Grust 2004, Gremaud & Goldscheider 2010, Braakhekke et al. 2017) was mentioned (Section 3.5.1). Given these examples, extreme scenarios of negligible or zero erosion are considered realistic, justifying a hard lower erosion limit of zero metres. The upper limit for limestone erosion factors is set to 2 due to limestone's potential for concentrated erosion along fractures, allowing for deeper erosion through mechanisms like plucking and quarrying, which are more effective than abrasion. Observations such as the deep canyon under the Grindelwald glacier support the possibility of significant erosion along fracture lines, although this is considered rare and conditional. Conversely, studies from the Lower Aare Valley, Finger Lakes, and rock property data indicate that limestone in EC4 is generally more resistant to erosion than Molasse, with greater compressive strength and lower abrasion rates. Most experts thus favour an upper limit of 1, with values > 1 deemed unlikely and assigned low probabilities. Although theoretical slot canyon formation is acknowledged, it is expected to be unlikely. Due to

the limited data on extreme limestone erosion under glacial conditions, an upper hard limit of 2 is chosen as a balanced estimate. This compromise allows for rare, deeper erosion scenarios, acknowledging the uncertainty in defining a definitive upper limit for such erosion-resistant materials.

Observations from Northern Switzerland, particularly in areas like the Lower Aare Valley and Klettgau, show that limestone experiences significantly less erosion than Molasse. Evidence from these regions suggests an erosion factor in limestone that is markedly lower, motivating a median close to 0.5. Experts reference local erosion depths in different bedrocks to support the median and justify a left-skewed distribution, emphasising the lower erosion rates in limestone. The mixed lithologies in EC4, with varying layers of harder and softer materials (e.g. Villigen Formation and Hauptrogenstein), create resistance to erosion that differs across the sites. This variability adds uncertainty, reflected by a widening of the range for the distribution's centre. While abrasion studies in controlled environments indicate lower erosion in limestone, the potential for processes like plucking in natural settings leads some experts to position the centre slightly higher. Experts agree to a broad central range from 0.3 to 0.63, balancing observed lower erosion rates with the variability in limestone resistance.

The lower boundary at 0.10 and the upper boundary at 0.90 offer a broad distribution that accommodates both minimal and modest erosion possibilities, with the 95% upper limit reflecting the rarity of extreme erosion factors.

Considerations for EC3

Similarly to EC4, the lower hard limit for EC3 is set to 'zero' metres, capturing the possibility, however unlikely, of no erosion under certain conditions. Since a hard lower limit of zero is selected for EC4, there is a rationale to apply the same lower limit to EC3 for consistency. EC3 is expected to be somewhat more erodible than highly resistant classes, but it was concluded that unique conditions (like the presence of resistant limestone layers in this more heterogeneous rock group) could occasionally result in no erosion at all, which cannot be ruled out. The upper hard limit for EC3 considers its higher resistance relative to EC2 but includes flexibility to capture the uncertainty in erosion variability, leading to an agreed upper limit for rare scenarios up to 2.

The reasoning for setting the median (50%) and tertile values (33% and 67%) for erodibility class EC3 takes into account the class's mixed composition and variability, such as limestone, marl, and claystone, leading to a broad range of erodibility with moderate resistance. The distribution is centred closer to the values for Molasse (EC2) but remains broad, allowing for the variability in erodibility due to the diversity of rock types within EC3. Compressive strength data and mill experiments showed EC3 is more comparable to EC2 (Molasse), leading to a shift toward slightly higher values. The chosen values are 0.70 for the 33rd percentile, 0.78 for the median, and 0.85 for the 67th percentile.

Considerations for EC1

Similarly to the other erodibility classes, the lower hard limit for EC1 is set to 'zero' metres, recognising the rare but possible cases of zero erosion due to the stability of certain Quaternary sediments (e.g. very hard, compressed till) and the dynamics of meltwater erosion, which may drain through porous Quaternary sediments without causing erosion, while simultaneously eroding harder materials, such as Molasse. Observations such as the Aathal Gravels near Uster (ZH) support this resistance, with cemented gravels that are more difficult to erode than typical Quaternary material. The upper hard limit for EC1 is set to 10, capturing the potential for extreme erosion in loose Quaternary sediments, acknowledging the potential for substantial erosion over long periods or during glacial surges. This limit considers exceptional cases where fluvio-glacial transport rapidly erodes or removes loose Quaternary sediments, although probabilities for

erosion factors this high are low. In glacial overdeepenings, the more erodible sands and silts dominate, which allows for high erosion rates. Glacial erosion rates on soft ground can be an order of magnitude higher than on hard ground (e.g. Taku Glacier in Alaska, erosion rates of 4 m per year; Stumm 2010).

Experts agree that while Quaternary sediments contain both loose and cemented materials, the overall erodibility is slightly above the Molasse reference class. The median of 1.2 reflects this moderate resistance, acknowledging that Quaternary deposits are generally less consolidated than Molasse but still contain a significant portion of harder, compacted till that balances out the softer, more erodible components. The central part is set to be between 1.0, to represent the resistant components of Quaternary sediments, and 1.4, allowing for the high erodibility of unconsolidated Quaternary materials. The lower boundary at 0.8 and the upper boundary at 2.5 balances the potential for both high and low erosion scenarios within EC1, based on observed variability of Quaternary sediments, expert judgment, and physical plausibility in the absence of comprehensive data.

By considering the rock types at a specific overdeepened location and adjusting the reference depth of incision accordingly, the depth distribution of overdeepening for that point can be assessed. This depth distribution is conditional on both the occurrence of glacial overdeepening at the location and a specific local erosion base (LEB).

3.5.3.3 Scenarios for spatial occurrence of deep glacial erosion

The history of multiple phases of backfilling in many deep glacial troughs suggests a pattern where glaciations tend to reoccupy existing overdeepenings (Section 2.4, Fig. 2-32). This reoccupation makes present-day overdeepenings a primary scenario for potential erosion pathways. Additionally, spatial buffers around current overdeepenings are considered to account for possible future expansions in length or width, allowing for lateral shifts in the erosion axis.

However, the possibility of new, glacially diverted channels and overdeepenings cannot be excluded. Thus, this scenario is included within the probabilistic model. While exact future topography remains unpredictable, certain factors inform these additional scenarios, including glacial extent, subsurface conditions (such as pervasive fracturing), rock type, and topographic relief. These elements may influence where new overdeepenings could occur in the future.

A key consideration is the topographic steering of ice flow through existing valleys, where ice is generally channeled away from elevated areas like hills and ridges and into pre-existing troughs and valleys (Kessler et al. 2008). This is an observed phenomenon and is factored into scenario development. In contrast, areas with low relief, such as Zürich Nordost and the eastern part of Nördlich Lägern, may exhibit more randomness in erosion paths due to a lack of natural features that would divert ice flow.

Topographic steering is also thought to be more influential during glacial advance phases when ice thickness is lower. At later stages, when glaciers are larger or near their maximum extents, subglacial erosion is likely more affected by the slope of the ice surface and associated hydraulic gradients (Section 3.5.1; Nagra 2024b).

Using expert judgment, a set of plausible spatial scenarios was developed for each HLW site, which informed the creation of scenario maps (Fig. 3-43). In these maps, darker colours represent areas with higher likelihoods of future erosion scenarios. The total surface areas for each scenario are detailed in Tab. 3-8. To follow the general procedure of the safety-based site comparison, a subset of the scenario maps for NL and ZNO was generated for criterion 2.2 (Fig. 3-43d and e, see Section 4.1 for more information).

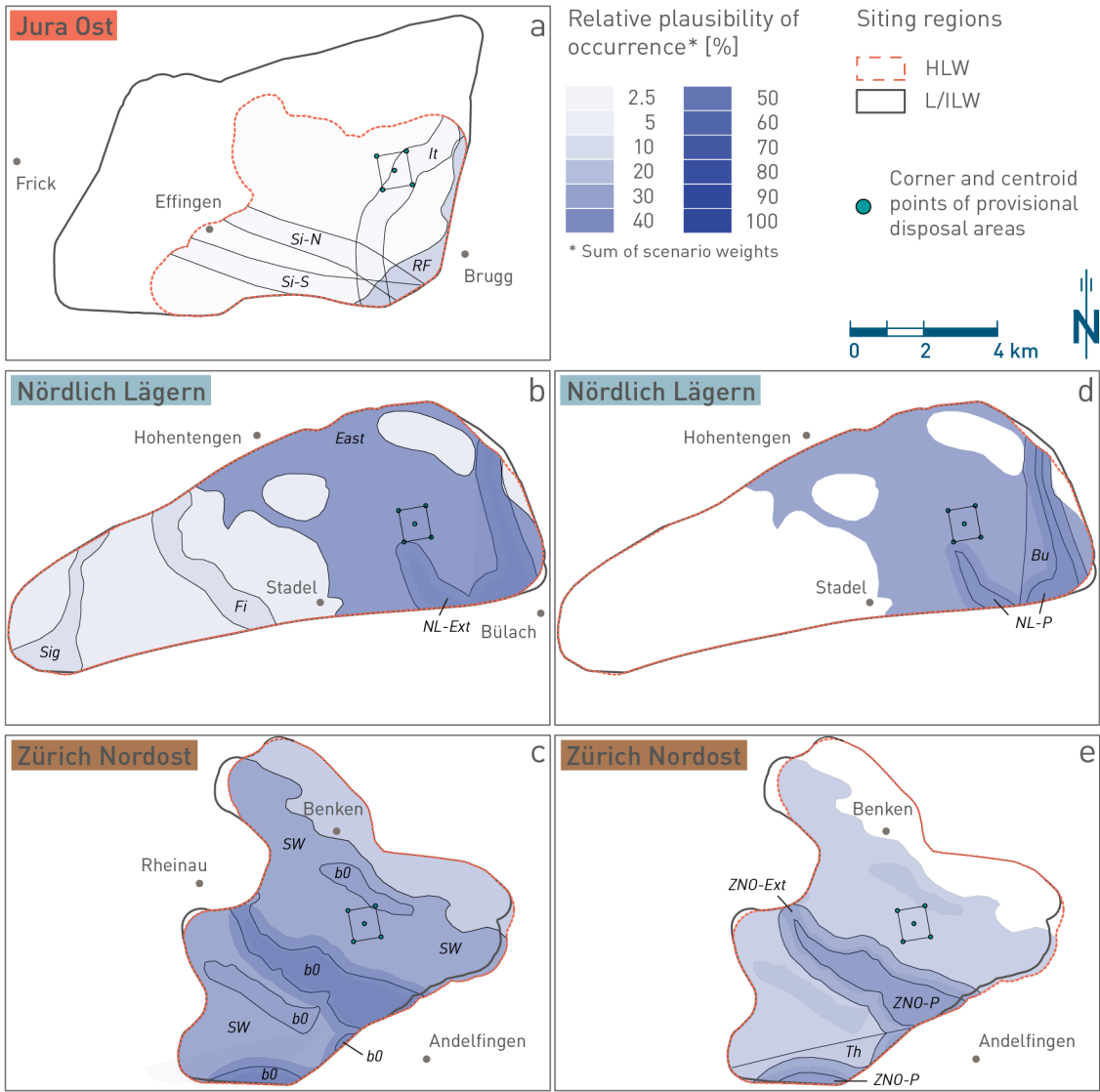


Fig. 3-43: Relative plausibility of occurrence of future overdeepenings (RIO) within the respective siting regions for one million years

Shown are colour-coded maps with darker colours reflecting a higher plausibility of occurrence of future glacial overdeepenings in the siting regions JO (a), NL (b) and ZNO (c), as used for the assessment including alternative evolutions (criterion 3.3; Section 4.1). In JO, the generally less likely occurrence of overdeepening is strongly linked to the occurrence of the (alternative) drainage scenarios (Section 3.4.3.4). Subsets considered for the expected evolution (criterion 2.2; Section 4.1) are shown for NL (d) and ZNO (e); for JO, the expected evolution assumes no overdeepening for the next one million years. The provisional disposal areas are shown with their respective corner and centroid points. Note that not the entire coloured areas would be overdeepened (see additional considerations for overdeepened area/density per siting region below).

The spatial extents of discrete scenarios of overdeepening occurrences as described in the main text are delineated with thin black lines and labelled as follows. In (a), RF: JO Rinikerfeld channel, Si-N: JO northern Sissle breakthrough channel, Si-S: JO southern Sissle breakthrough channel, It: JO Itele breakthrough channel. In (b), Sig: NL Siglistorf breakthrough channel, Fi: NL Fisibach breakthrough channel, East: NL random east, NL-Ext: NL extended. In (c), NL-P: NL present, Bu: NL Bülach Trough. In (d), SW: ZNO random SW large, b0: ZNO below zero. In (e), ZNO-P: ZNO present, ZNO-Ext: ZNO extended, Th: ZNO Thurtal Trough. Each scenario for NL and ZNO is only delineated in one of the two maps ((b) or (d); (c) or (e)) to enhance visibility. However, many of these overdeepening scenarios are considered for the relative plausibility of occurrence in both maps (e.g. NL random east, in both (b) and (d)). Moreover, for each site there is a scenario "random" where glacial overdeepenings may appear anywhere in the HLW perimeter of the siting region which is considered by the relative plausibility of occurrence in maps (a), (b) and (c).

Considerations for Jura Ost

Jura Ost presently has no glacial overdeepening (Fig. 2-11a), which was an important argument also for future considerations. It was acknowledged that the topography provides obstacles to ice flow and may inhibit significant deep glacial erosion. Because of the definition that a glacial overdeepening constitutes the part of the landscape located below the local erosion base, overdeepenings in JO can only be developed when breakthrough channels provide associated drainage pathways. Accordingly, the scenarios used for JO constitute the alternative drainage networks presented in Section 3.4.3.4, see Fig. 3-24. The following scenarios were considered and ranked for Jura Ost (Fig. 3-43a):

- JO Rinikerfeld: The Rinikerfeld in the southeast of the siting region is reactivated and glacial overdeepenings will form in this area. The Rinikerfeld channel was part of a fluvio-glacial system likely active before and during MIS 6 (Graf 2009, Mueller et al. 2020, Gegg et al. 2022). This scenario incorporates a general widening of the Lower Aare Valley.
- JO Sissle south and JO Sissle north: A breakthrough channel develops in connection with the Sissle and leads to glacial overdeepenings in the south of the siting region. Meltwater derived subglacially or proglacially along the glacier outlet might discharge across lower-level topography and allow the deep incision of a new "breakthrough channel". Subsequently, this new valley becomes overdeepened by glacial erosion. Two parallel breakthrough channel scenarios were assumed (a northern and a southern Sissle breakthrough).
- JO Itele: Similar to the Sissle breakthrough, the Itele valley was assumed to be a potential candidate for a future location of a breakthrough channel with subsequent overdeepening. The Itele valley constitutes a prominent topographic feature and is located close to the former ice margin of the Beringen glacial (MIS 6).
- JO random: In a random scenario, glacial overdeepenings can appear anywhere in the siting region. Such a random scenario considers, for instance, the uncertainty in potential future geomorphic evolution that precedes a glaciation. Potential changes might influence hydraulic conditions beneath future ice advances in these areas and thus lead to erosion of previously unaffected areas.
- JO no OD: Yet another possibility is that no glacial overdeepenings will appear in the whole siting region, despite the occurrence of a glaciation that covers the site. JO is currently the only site without present glacial overdeepenings. Furthermore, future conditions might also prevent glacial overdeepenings.

In general, the chances for larger glaciations and previous geomorphic changes are considered higher the longer time proceeds. For a one-million-year timeframe, this fact should be reflected in the scenario ranking, with a reduction of the weight for a scenario with no glacial overdeepenings for the benefit of all other scenarios.

The Rinikerfeld/Lower Aare widening scenario was considered most likely (20%). The close connection to the Aare River in the southeast, which is assumed to still be a major geomorphic agent in one million years, elevated the likelihood of this scenario with respect to the development of entirely new channels. The Sissle and Itele scenarios were considered valid potential scenarios for future breakthrough channels, but also more random options should be generally possible. As seen in Fig. 3-43a, most of the scenarios for Jura Ost were given a low plausibility. It was still considered most likely (72%) that no overdeepening would develop within the HLW perimeter of JO.

Considerations for Nördlich Lägern

The following scenarios were considered and ranked for Nördlich Lägern (Fig. 3-43b and d):

- NL present: Because observations from boreholes often reveal multiphase excavation and filling histories of overdeepenings, one scenario places future glacial overdeepenings occurrence at their present locations. Consideration was limited to overdeepenings that are deeper than 25 m below the local erosion base.
- NL extended: This scenario was complemented by an extension scenario that allows for widening/growing of the glacial overdeepenings and the possibility that the axis, and thus location of deepest erosion, shifts laterally by a limited amount.
- NL Bülach: Following the logic of multiphase reactivation from above but acknowledging that the Bülach Trough may have been active more recently than the Strassberg Trough (Section 2.4.2, see also Fig. 2-32), an additional scenario complements the above by assigning more weight to regions around younger overdeepenings. In addition, the Bülach Trough is located in a broad valley, favouring future ice and water flow.
- NL east: In a partly random scenario, glacial overdeepenings could appear anywhere within the low-relief areas of the eastern part NL. Topographic steering would support the routing of future ice flow, while higher relief would constitute efficient barriers. The eastern part of NL is a generally low-relief area, where such steering effects might be considered negligible.
- NL random: In a random scenario, glacial overdeepenings can appear anywhere in the HLW perimeter of the siting region.
- NL Siglistorf and NL Fisibach: Two breakthrough channels were assumed that may develop near the town of Siglistorf or along the Fisibach channel (Bachsertal). Meltwater derived subglacially or along the glacier margin discharges across lower-level topography and allows the formation and subsequent overdeepening of a new valley.
- NL no OD: a scenario considering that no glacial overdeepenings will appear in the entire HLW perimeter of the siting region was also assumed for NL. Although NL presently has glacial overdeepenings, large-scale geomorphic changes might result in reorganisation and rerouting of channels, such that future major ice streams and/or subglacial drainage channels bypass NL. Similarly, different climatic conditions or contrasts in the exposed rock types might prevent the formation of glacial overdeepenings.

Most agreement between individual estimates existed for the scenario NL east. This and the scenario entailing the extension of present overdeepenings (NL extended) were considered more likely than future overdeepenings at the exact location of present overdeepenings. The low relief of the eastern part of the HLW perimeter together with the more proximal position with respect to ice flow may expose this part to future deep glacial erosion.

The breakthrough of new overdeepenings within the HLW perimeter was generally regarded as possible, but less plausible than all scenarios considering the eastern part of NL.

Because of the long time span, geomorphic and subsequent erosional changes might happen, even a scenario without future glacial overdeepening was considered with low likelihood (1%).

Considerations for Zürich Nordost

The following scenarios were considered and ranked for Zürich Nordost (Fig. 3-43c and e):

ZNO present: Because observations from boreholes often reveal multiphase excavation and filling histories of overdeepenings, one scenario assumes future glacial overdeepenings occurring at their present locations. Consideration was limited to "significant" overdeepenings that are deeper than 25 m below the local erosion base.

ZNO extended: This scenario was complemented by an extension scenario that allows for widening/growing of the glacial overdeepenings and the possibility that the axis, and thus location of deepest erosion, shifts laterally by a limited amount.

ZNO below zero: Another scenario was added to account for very shallow present overdeepenings as potential future reoccupation pathways.

ZNO Thurtal: Yet another scenario was suggested to allow focused deep glacial erosion in the vicinity of the Thurtal Trough, which is a prominent geomorphic feature and might guide future ice flow.

ZNO random SW large: Glacial overdeepenings could appear anywhere within the low-relief areas of ZNO (excluding the Cholfirst area) or

ZNO random: Glacial overdeepenings could appear anywhere in the HLW perimeter of the siting region.

ZNO no OD: Finally, the scenario that no glacial overdeepenings will form in the whole siting region was assumed.

Because of the close proximity of ZNO to the ice lobes, it was regarded as likely that more glaciations may affect the area within one million yrs, potentially changing the river incision pattern and fluvial and glacial deposition. Nevertheless, the main existing overdeepenings were still considered as important for guiding ice flow in one million yrs. Accordingly, the scenarios with erosion below and around existing overdeepenings sum up to 43% (Fig. 3-43c). The "significant" present overdeepenings may still comprise a preconditioning because they are filled with less resistant material and are generally favourably oriented with respect to main ice flow. This observation is especially true for the Thurtal. The Thurtal scenario was considered to be important because the potential SSW-shift of the drainage system (Ellwanger et al. 2011) might focus the fluvial system and drainage, potentially guiding future ice flow of advancing glaciers.

The Cholfirst area was discussed to still be a significant topographic obstacle because the deposits appear very resistant and have already resisted previous glaciations, but it was agreed upon that the weight of the random scenario should be slightly higher than that of the scenario excluding the Cholfirst area. Thus, the experts agreed upon 28% for the scenario with random occurrence of future glacial overdeepenings within the HLW perimeter and 26% for the random scenario excluding the Cholfirst area.

A slight chance may persist that future glaciations would not produce glacial overdeepenings within ZNO within the one-million-year time span (3%). Accordingly, the glaciers might bypass the siting region and erode only outside the HLW perimeter.

Tab. 3-8: Characteristics of overdeepening scenarios

Overdeepening scenarios are listed with their area in km² and in percentage of the respective HLW siting regions, as described in Section 4.5.3.2 as well as the probability of occurrence and the associated main reasoning. Prob.: Probability, LAV: Lower Aare Valley, OD: Overdeepening., RF: Rinikerfeld.

OD scenario	Area [km ²]	% of HLW	Main reasoning	Prob. [%]
JO RF	2	7	Established topographic feature may steer ice flow and subglacial water flow. Scenario includes widening of LAV.	20
JO Sissle south	4	12	Breakthrough channels could be formed during a major glacial advance and overdeepened in a successive advance. Probability of occurrence for OD is very low, because of pronounced topography, resistant bedrock and orientation of scenarios oblique to ice-flow direction.	2
JO Sissle north	3	11		2
JO Itele	4	13	Breakthrough channels could be formed during a major glacial advance and overdeepened in a successive advance. Probability of occurrence for OD is very low, because of pronounced topography, and resistant bedrock. Note orientation of the Itele valley approximately parallel to Beringen ice margin.	2
JO random	29	100	Scenario covers the possibility of unexpected ODs occurring across the siting region, potential for glacial erosion to create new channels even in less expected locations.	2
NL present	2	4	Existing troughs might generally guide future OD patterns. Quaternary reoccupation observed in many areas. However, probability of occurrence moderate because erodibility contrast to bedrock is small and the topography in the eastern part of NL is moderate.	5
NL extended	6	9	Existing troughs might generally guide future OD patterns, exact pathways are unlikely to persist.	15
NL east	30	48	Exposure to glacial activity from the south, low topography facilitates glacier access and the formation of new troughs.	49
NL Siglistorf	3	4	Significant topographic features vulnerable to the formation of breakthrough channels and OD, but protected by surrounding topography, focusing too heavily on these specific locations might lead to an overly condensed likelihood on the heat map.	7.5
NL Fisibach	4	6		7.5
NL random	63	100	Scenario covers OD events at random locations in the siting region, e.g. in conjunction with the formation of breakthrough channels.	10
NL Bülach	5	8	Area between Bülach and Eglisau considered particularly vulnerable due to its low elevation and the existing OD.	5

Tab. 3-8: (continued)

OD scenario	Area [km²]	% of HLW	Main reasoning	Prob. [%]
ZNO present	4	8	Reoccupation and deepening of existing OD during successive advances has been observed in various places. Less likely that current troughs would remain unchanged. Low likelihood was retained.	8
ZNO extended	10	20	Broad patterns of existing troughs might guide future erosion, precise locations would likely shift. Includes a widening of existing ODs.	23
ZNO below 0	9	19	Extension of the "extended" scenario but with less likelihood, covers the possibility that shallow present channels guide future deep glacial erosion.	9
ZNO random SW large	40	80	Areas of low elevation in the southwest that could be prone to glacial erosion over long periods. Exact location of OD hard to predict.	26
ZNO random	49	100	Very likely scenario acknowledges the uncertainty in predicting future topography over one million years. Multiple glacial cycles and unpredictable river redirections would significantly reshape the landscape and lead to breakthrough channels across topographically elevated areas.	28
ZNO Thurtal	4	7	Thur valley is topographically low. Potential long-term shifts in the Rhine drainage system to the south could intensify glacial erosion in the Thurtal valley.	3

3.5.3.4 Considerations for overdeepened areas/density per siting region

To evaluate the overall unconditional probability of repository excavation due to glaciation, additional factors must be incorporated, including the likelihood of future glaciation events, the spatial probability of glacial overdeepening, and the extent of the overdeepened area. The exemplarily step-by-step procedure for calculating the probabilities of both fluvial and deep glacial incision at a given location is shown in Fig. 3-48.

The HLW siting regions are variably impacted by glacial overdeepenings. Considering overdeepenings greater than 25 metres below the local erosion base (LEB), less than 4% of the HLW area in Nördlich Lägern and about 9% of the HLW area in Zürich Nordost are currently affected by these features (Fig. 3-44b, Tab. 3-9). There are no existing glacial overdeepenings in Jura Ost.

In scenarios allowing for a high degree of randomness in future glacial overdeepenings (see Section 3.5.3.3), only a portion of the area within each siting region would likely be overdeepened. To account for this, the future spatial density (percentage of surface area) of overdeepenings deeper than 25 m below the LEB within each HLW region was estimated over a one-million-year time span, assuming that some overdeepening will occur. This estimation followed the SHELF protocol (Fig. 3-45).

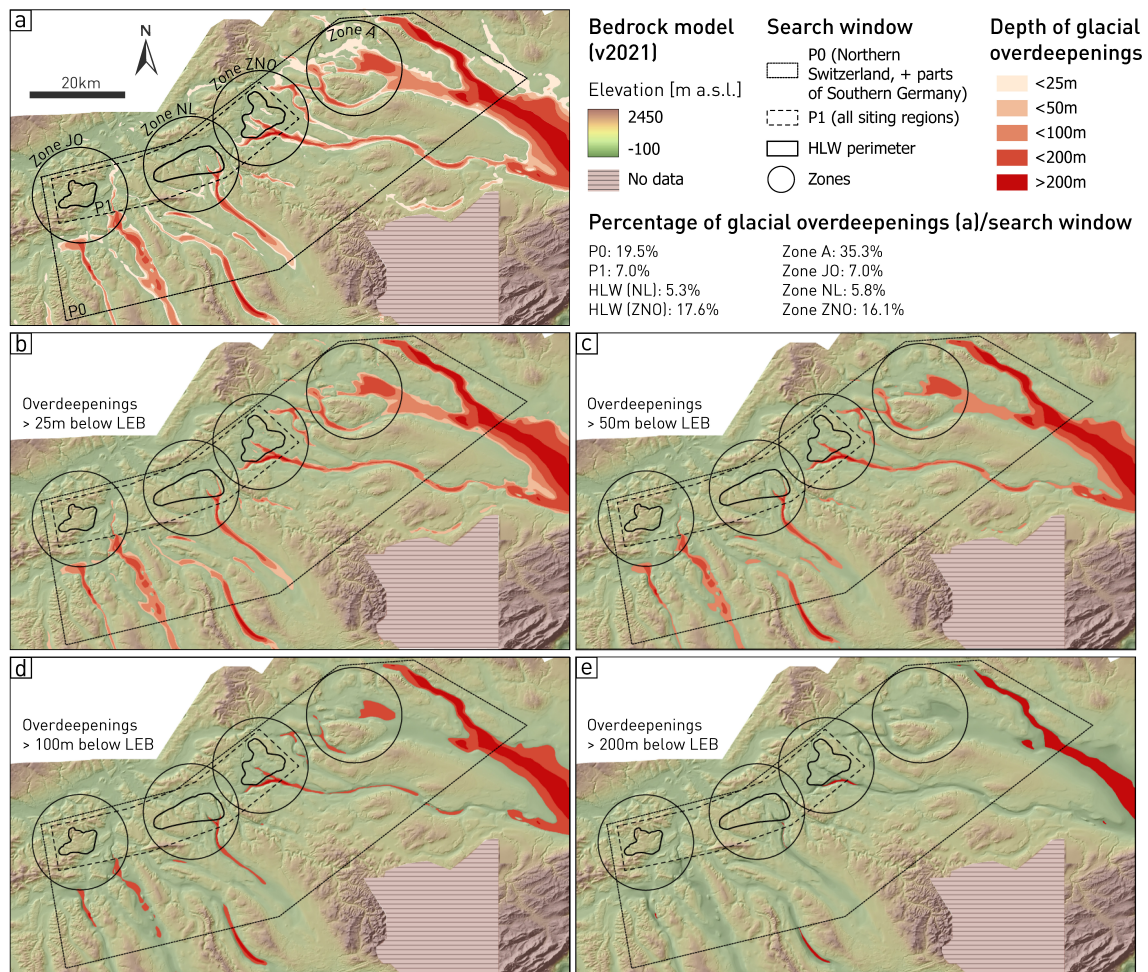


Fig. 3-44: Spatial percentage of overdeepening distributions using different search windows

Spatial density was calculated for present overdeepenings (OD) below the local erosion base. Calculations were based on the updated bedrock model (Version 2021¹⁸). (a) all ODs, (b) OD deeper than 25 m below LEB, (c) OD deeper than 50 m below LEB, (d) OD deeper than 100 m below LEB, (e) OD deeper than 200 m below LEB. The fraction of overdeepened area is calculated for different search windows and shown in km² and %, respectively (within legend only shown for calculations in (a)). The other values are shown in Tab. 3-9.

The primary evidence base for this estimation consisted of spatial overdeepening distribution statistics across Northern Switzerland, assessed within various search windows, each representing different depths and areas (Tab. 3-9). The smallest search window was defined by the HLW perimeters for each region. For broader spatial comparison, additional search windows were added: a circular area around the HLW perimeters (called Zone JO, Zone NL, Zone ZNO), and a larger area extending into Untersee (Zone A).

¹⁸ Note that the difference between version 2021 and version 2023 refers to the addition of the results of seismic investigations in the Bodensee in the later one. These calculations shown here (which were used as elicitation basis) are based on version 2021.

Tab. 3-9: Fraction of glacial overdeepening within different search windows

Shown are areas in km² and fraction in % of overdeepenings within different search windows (corresponding to Fig. 3-44). Statistics are calculated for all overdeepenings and considering selected depth below LEB (> 25 m, > 50 m, > 100 m, and > 200 m, respectively). Calculations are based on the bedrock model v. 2021.

Search window		JO HLW	NL HLW	ZNO HLW	P0	P1	Zone A	Zone JO	Zone NL	Zone ZNO
Area [km ²]		29.2	63.5	49.5	3'123.8	542.3	314.2	314.2	314.2	314.2
Area [km ²]	all OD	0	3.4	8.7	608.5	37.9	110.8	22.1	18.4	50.5
Area [km ²]	> 25 m	0	2.2	4.2	413.6	20.9	62.9	12.6	9.5	31.2
Area [km ²]	> 50 m	0	1.3	2.8	294.8	14.9	40.2	7.9	6.9	23.4
Area [km ²]	> 100 m	0	0.4	1.3	136.8	8.5	12.4	2.8	3.8	12.5
Area [km ²]	> 200 m	0	0.0	0.0	38.4	1.6	0.0	0.0	0.3	2.2
Fraction [%]	all OD	0	5.3	17.6	19.5	7.0	35.3	7.0	5.8	16.1
Fraction [%]	> 25 m	0	3.4	8.4	13.2	3.9	20.0	4.0	3.0	9.9
Fraction [%]	> 50 m	0	2.1	5.7	9.4	2.8	12.8	2.5	2.2	7.5
Fraction [%]	> 100 m	0	0.7	2.7	4.4	1.6	3.9	0.9	1.2	4.0
Fraction [%]	> 200 m	0	0.0	0.0	1.2	0.3	0.0	0.0	0.1	0.7

Two additional rectangular windows included:

- **P1:** A narrower window encompassing all three HLW regions together, and
- **P0:** A wider window covering large parts of the upstream overdeepened valleys in Northern Switzerland.

The spatial statistics – measuring the percentage of each search window coinciding with overdeepenings – were calculated at multiple depth intervals: surface level (Fig. 3-44a), 25 m depth (Fig. 3-44b), 50 m depth (Fig. 3-44c), 100 m depth (Fig. 3-44d), and 200 m depth (Fig. 3-44e). These results are summarised in Tab. 3-9. The observed decrease in overdeepened areas with depth reflects both the depth distribution of overdeepenings (not all reach 100 or 200 metres) and the narrowing of these features with increasing depth, particularly in the distal foreland areas.

Considerations for Jura Ost

Experts agreed on setting the lower hard limit at 0%, reflecting the possibility of minimal or no overdeepening in the siting region due to strong topographic barriers and the likely presence of resistant rock formations. For the 5% boundary, they collectively chose 0.25%, a value representing the lower end of potential overdeepening (OD) depth.

For the median and tertiles, experts based their values on scenarios where minor ODs occur primarily in the eastern part of the siting region, e.g. in conjunction with widening of the Lower Aare Valley (i.e. the Rinikerfeld scenario). These scenarios reflect smaller-scale OD areas, possibly aligned with existing channels. A median of 2.5% reflects the potential occurrence of overdeepening scenarios but at a level still lower than today in NL. Tertiles were set at 1.5% (33%) and 3.7% (67%), keeping values close to the median to avoid a bimodal distribution.

The upper boundaries considered potential maximum OD coverage while staying within feasible limits for the siting region's geological and topographic constraints. The 95% percentile was set to 8.5%, which experts determined based on visual assessments of similar OD densities in comparable regions, e.g. ZNO (Fig. 3-44). The upper hard limit was capped at 25%, avoiding extreme scenarios that are unlikely given the siting region's pronounced topography and rock types (e.g. limestone). This choice ruled out improbable high-OD scenarios, such as full over-deepening akin to Lake Constance, while still allowing for larger valley-like ODs in very unlikely cases (Fig. 3-44, Fig. 3-45 and Tab. 3-10, note that information in figure and table is provided in km²).

Considerations for Nördlich Lägern

It was noted that OD density in NL would likely exceed that of JO due to differences in geology and topography. It was emphasised that areas covered with Molasse are more likely to experience broader ODs, while regions underlain by limestone would likely have narrower erosional features. The east-west topographic variation within NL, with low-relief areas in the east more prone to glacial impact, further supported these conclusions. The selected values reflect a broad range of possibilities, accounting for the complexity of glacial and erosional processes over a long timescale.

The lower hard limit was set to 0%, representing the possibility of minimal or no overdeepening within the siting region, similar to prior discussions for the JO region. For the 5% boundary, chose a value of 1%, informed by current trough depths in the range of 100 – 200 m was chosen, reflecting an estimate of possible minor glacial impacts.

Experts based the median and tertiles on current OD densities, particularly in the eastern NL area. They agreed on a median of 5%, reflecting the likelihood that existing channels and valleys, such as Windlacherfeld, could broaden and deepen. The most likely third was considered to be between 3% and 6.5%.

For the upper boundary, experts discussed extreme scenarios, including the hypothetical stronger westward orientation of the Rhine Glacier (Ellwanger et al. 2011) and a potential increase in OD density in NL. The 95% percentile was set to 10%. The upper hard limit was capped at 50%, reflecting an extreme but conceivable scenario where significant portions of the eastern NL region experience OD, though not to the extent of replicating a major feature like the Bodensee (Fig. 3-45 and Tab. 3-10).

Considerations for Zürich Nordost

Experts agreed on a lower hard limit of 0%, representing a minimal probability of ODs in ZNO under certain conditions. For the 5% boundary, they selected 2%, which accommodates scenarios where the siting region is relatively unaffected by glacial overdeepening. This lower range reflects the region's resilience during smaller glacial periods, like the Last Glacial Maximum (LGM), when much of the ZNO area was minimally impacted by glacial erosion.

Experts aligned the median and tertile values with current OD densities, particularly in regions where future glacial erosion could reshape existing features like the Andelfingen and Rudolfingen Troughs. They agreed on a median of 10%, based on current densities and the likelihood of continued erosion over time. Tertiles were set at 7% (33%) and 12.5% (67%), allowing for a wider body of the distribution that reflects ZNO's flat topography and susceptibility to new erosional features due to river and glacial shifts.

For the upper boundaries, experts considered ZNO's proximity to Lake Constance and the Rhine system, which may increase its potential for extensive OD formation over time. They set the 95% percentile at 25%, reflecting a moderate scenario of substantial erosion within the siting region.

The upper hard limit was capped at 60% (similar to Zone A in Fig. 3-44), indicating a theoretical maximum in which extensive glacial activity results in significant OD density. This value, higher than NL's, was justified by ZNO's larger flat areas and its higher exposure to erosive forces from nearby major glacial and fluvial systems (Fig. 3-45 and Tab. 3-10).

Experts highlighted that ZNO's OD density distribution should be the highest of all siting regions, given its susceptibility to glacial forces (ZNO was overridden by ice during all three key glaciations (see Fig. 2-30) and the flat terrain that makes the area prone to overdeepening. Experts also acknowledged the potential impact of lithology, as increasing exposure of limestone in there northern parts of ZNO after long time periods versus Molasse layers in the southern parts might produce both narrow and broader ODs, depending on which materials are exposed over time.

Tab. 3-10: Estimates of future overdeepened area within the HLW siting regions

The area that is overdeepened was elicited as a gamma distribution, whose parameters depend on the specific HLW siting region (JO, NL, ZNO, see Fig. 3-45). The mean and standard deviation of the gamma-distributed overdeepened area are listed.

Siting region	Mean [km ²]	Standard deviation [km ²]
Jura Ost	0.95	0.85
Nördlich Lägern	3.44	2.34
Zürich Nordost	5.53	3.59

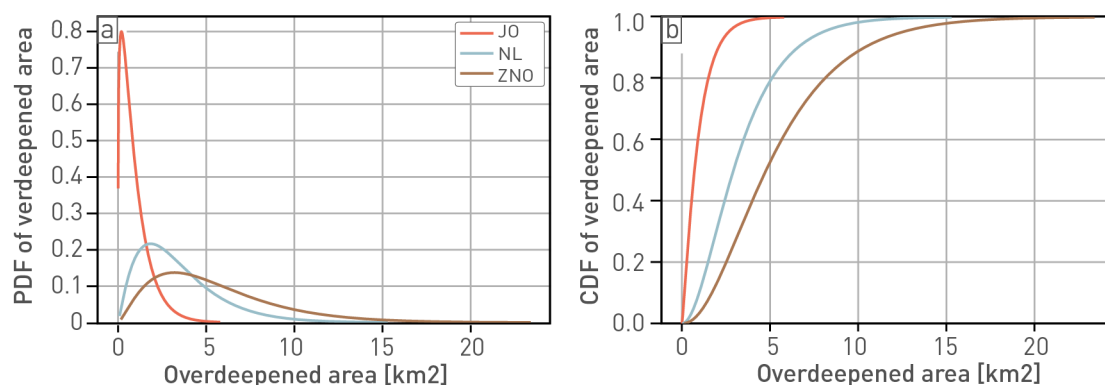


Fig. 3-45: Total assumed probabilities of portion of scenarios that can be overdeepened

Probability density function (a) and cumulative density function (b) based on expert judgement (compare Tab. 3-10).

3.5.3.5 Probabilities for occurrence of future key-size glaciation

Quaternary insights as a basis for future key-size glaciations

The specific glacial extents of the main recognised Quaternary glaciations Birrfeld (including LGM), Beringen (likely MIS 6), and Möhlin (likely MIS 12/16) with respect to the siting regions (Fig. 2-30) were used as key glacials in the future assessment of erosion. Accordingly, Jura Ost will only be covered by a future glaciation of Möhlin-size, Nördlich Lägern by Möhlin- and Beringen-size, and Zürich Nordost by all three key glaciations. For a full assessment it is also considered that a foreland glaciation might occur, but smaller than LGM, i.e. in this context too small to reach Zürich Nordost. Such a glaciation will not have any impact on the estimated erosion. The concept of key glacials was used to (1) estimate the probability of occurrence of alternative drainage scenarios (see Section 3.4.3.4) conditional on the occurrence of a key glacial, (2) to estimate the overdeepening depth for each of the key glacials, (3) to estimate the probability of occurrence of the key glacials for the future one million years. Note that it was decided to not include glaciations for the erosion assessment of the 100'000-year period for L/ILW (Section 3.2.1).

Considering Preusser et al. (2011), there were about four larger ice advances of "key-glacial" size ($1 \times$ Möhlin-size, $1 \times$ Beringen-size, $2 \times$ LGM-size (see Fig. 2-27). Earlier glaciations reached the foreland, as is indicated by intercalated till within Deckenschotter deposits. It seems, however, that the above-mentioned four large glaciations took place only after the accumulation of the TDS and probably after (or associated with) the Mid-Pleistocene Transition (MPT), when the glacial cyclicity switched from a dominant 41-kyr to a 100-kyr cycle (Section 2.1 and 2.4). The ages of the Möhlin and Beringen glaciations are not unambiguously resolved. The so far oldest dating of an overdeepening fill points to 606 ± 122 ka (Tomonaga et al. 2024; Section 2.4.1.2) and thus MIS 12 – 16. In summary, this suggests that the observed overdeepenings today are the cumulative result from the past ~ 800 kyr and more.

Scenarios of future ice volumes as proxy for future large foreland glaciations

To assess the potential for future glaciations to reach the North Alpine Foreland, the focus is on models of future ice volumes (Talentó & Ganopolski 2021, Nagra 2024b). Parameters for orbital configurations can be calculated accurately for the past and future and are for example reported in Berger (1977, 1978) and Berger & Loutre (1997). However, the future atmospheric CO₂ content cannot be accurately estimated as the additional anthropogenic emissions cannot be easily forecasted. To accommodate this uncertainty, several emission scenarios were used to cover a range of likely scenarios, see Fig. 3-46 (Talentó & Ganopolski 2021).

The following statements are based on simulations using an intermediate complexity model and anthropogenic CO₂ anomalies (Talentó & Ganopolski 2021). Paleo-reconstructions of the past 800 kyr (global mean surface temperature, atmospheric CO₂ concentration and normalised global landmass ice volume) were used as part of the training and validation set for their (predictive) model. Additional constraints of model realisations and subsets were based on threshold values and can be summarised as follows:

Accepted model solutions:

To model the impact of anthropogenic CO₂ emissions on Earth's climate, Talentó & Ganopolski (2021) used a parameter called K (a measure of the sensitivity of the critical orbital forcing to the atmospheric CO₂ concentration) that influences how the model behaves. However, different values of K can lead to vastly different outcomes. To ensure that the future predictions align with known climate dynamics, only solutions that match past climate simulations (called "Paleovalid" solutions) are considered. Out of these, only a subset meets a specific criterion for K (requiring it

to be less than -150 W/m^2), narrowing the simulation down to 29 solutions, which are called the "accepted" solutions. These "accepted" solutions are more likely to provide realistic future scenarios under different CO_2 emissions levels.

Best solution:

Among the "accepted" solutions, the "best solution" is the one that most accurately reproduces historical climate patterns over the past 800'000 years, including past ice volumes, atmospheric CO_2 levels, and surface temperature. This model was particularly effective in matching the timing and scale of glaciations. In essence, it is the most precise and reliable model for both past climate variability and for projecting future climate changes due to emissions.

The argumentation with respect to future glaciations is based primarily on the "best solution" simulation, which is shown for four different CO_2 scenarios in Fig. 3-46. A brief note is added on the deviation of other model solutions at the end of the paragraphs describing the scenario, respectively. We furthermore focus on the proxy of ice volume which performed best and which is normalised with respect to the global LGM ice volume. As learned from Quaternary glaciations in Northern Switzerland (Section 2.4.2, Fig. 2-30), LGM-size glaciations did not have the highest erosion potential with respect to glacial overdeepenings. Accordingly, the results cannot be directly transferred but offer a valuable indicator. In the model simulations the occurrence of full glacial conditions and large glaciations are defined to be at 0.5 and 0.8 of the normalised global landmass ice volume, respectively, see Fig. 3-46 for best solution scenarios. Glacial inception is the first time before reaching full glacial conditions at which ice volume is non-zero; this is exemplarily marked in Fig. 3-46b for the inception of the first modelled glaciation. With respect to the long-term geological evolution of the repository sites, we are only interested in the potential future timing of large glaciations, as only they may reach the North Alpine Foreland. This timing approximately corresponds to the timing of the peaks above the 0.8-unit ice volume threshold in Fig. 3-46.

The first case is a natural scenario with no additional anthropogenic emissions (0 petagram Carbon [PgC], Fig. 3-46a). For the period 1750 – 2017, land use and fossil-fuel cumulative emissions in carbon equivalents are estimated to 660 PgC (Le Quéré et al. 2018). Future projections for additional cumulative emissions might reach between ~ 700 PgC (incorporating fossil-fuel reserves exploitable today) and $\sim 3'000$ PgC (considering resources that might become exploitable in the future) (McGlade & Ekins 2015). Accordingly, scenarios with additional cumulative emissions of 500 (already passed today), 1'000 and 3'000 PgC were analysed (Fig. 3-46b, c, and d, respectively).

In general, the higher the carbon concentration in the atmosphere, the later the inception of the next glaciation will be (Fig. 3-46 for best solutions). In the natural scenario ($E = 0$ PgC), the next large glaciation is simulated to peak around ~ 110 ka. It will, however, not reach LGM volumes (maximum simulated ice volume is 0.84 for the first major glaciation; Fig. 3-46a). Although amongst the "accepted" simulations there is one solution with an earlier glaciation, but this one does not reach the 0.8 units threshold.

For the low emission scenario ($E = 500$ PgC), the next large glaciation is postponed by $\sim 90 - 100$ kyr compared to the natural scenario, with a first peak at around 220 ka and ice volumes of 0.8 (Fig. 3-46b). This peak also corresponds to the first large glaciation (although with slightly higher ice volumes) when adding the "accepted" model assemblage.

For the intermediate emission scenario ($E = 1'000$ PgC), the next large glaciation may only occur after $\sim 680 - 700$ kyr, with simulated ice volumes of about 0.9 (Fig. 3-46c). For this scenario, however, the "accepted" model assemblage shows two additional earlier simulations for large glaciations: at about 480 ka and about 580 ka in the future (Talento & Ganopolski 2021, their Figure 7).

Finally, with a cumulative emission of 3'000 PgC, simulations indicate mostly ice-free conditions for the next 800 kyr (Fig. 3-46d). Using the "best" solution, the next large glaciation is not expected to occur before ~ 880 ka ("accepted" solutions show a predominantly 100-kyr cycle with large glaciations from approximately 700 ka onward).

Once the first larger glaciation has occurred, all scenarios show a cyclicity of approximately 100 kyr, similar to the glacial/interglacial cycles established during the Late Pleistocene (after the MTP, see Section 2.2 and Nagra 2024b).

According to the different scenarios presented in Fig. 3-46, 3 – 5 large glaciations (normalised ice volume > 0.8) that may reach the Alpine Foreland in the next one million years seem very likely, which is quite similar to the past assumptions of large foreland glaciations based on Preusser et al. (2011) (see Fig. 2-27). Again, adding the less likely but "accepted" simulations, even 4 to 8 glaciations may be able to reach the Alpine Foreland after 200 ka for the non-natural scenarios (Talento & Ganopolski 2021).

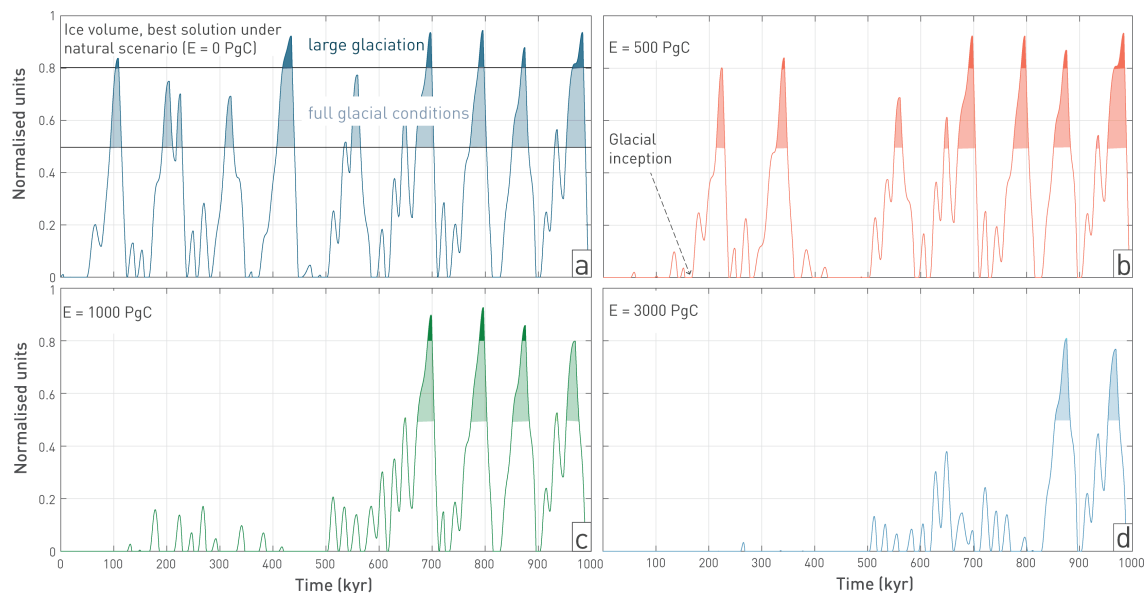


Fig. 3-46: Future climate simulations (global ice volume) based on different anthropogenic CO₂ scenarios

The plots show the "best solution" simulations and are redrawn from Talento & Ganopolski (2021), using normalised global landmass ice volume as proxy for future glaciations. A value of 1 refers approximately to the same amount of ice as determined for the peak of the LGM. Full glacial conditions are suggested to occur at 0.5 units (filled parts), while a value of 0.8 may suggest a large glaciation (darker filled peaks). With respect to the long-term geological evolution of the repository sites, such large glaciations might reach the Alpine Foreland. (a) Scenario with zero anthropogenic CO₂ emissions. A large glaciation (potentially reaching the Foreland) occurs only after ~ 100 kyr from now. (b) Cumulative emissions of 500 PgC have already been reached in the present-day atmosphere. In such a scenario, the next glacial inception is significantly delayed, with the next large glaciation (ice volume > 0.8) occurring only after ~ 200 kyr. The first glacial inception based on these best solutions is marked by the arrow (as the first time before reaching full glacial conditions [ice volume > 0.5] at which ice volume is non-zero). (c) The 1'000 PgC scenario considers consumption of all presently exploitable fossil fuel reserves and will further delay future large glaciations. (d) 3'000 PgC represents a worst-case scenario, including fossil fuel reserves that might become exploitable in the future.

Estimated glacial extent (one million years)

In addition to the modelling of Talento & Ganopolski (2021), selected information on anthropogenic CO₂ emissions (IPCC 2015) and CO₂ longevity in the atmosphere (Lord et al. 2016), deep future climate modelling for Sweden (Brandefelt et al. 2013), and the LR04 benthic $\delta^{18}\text{O}$ stack constructed by the graphic correlation of 57 globally distributed benthic $\delta^{18}\text{O}$ records (Lisiecki & Raymo 2005) were consulted.

The expert elicitation on the future key-size glaciations was conducted for the timing of 880'000 years in the future, as this corresponds to a potential large future glaciation visible in Fig. 3-46b to d. It was later agreed upon by the experts that (considering the uncertainty) the individual estimates would not differ if extending the timeframe to one million years (Fig. 3-47). Originally, the elicitation was conducted for a certain glaciation at the respective time, i.e. 880'000 years, considering that the final large glaciation of the period under consideration may be the most important one (Fig. 3-37). However, the model adaptation uses the elicited key-size estimates as respective percentages in the sampling process.

Given the glacial extent stack below (Fig. 3-47a), Möhlin is selected as a relevant glaciation event for a one-million-year time span (i.e. all siting regions are covered by glaciers) with a probability of 18%. With a probability of 20% a Beringen-size glaciation (i.e. only NL and ZNO are covered) may occur, and 24% probability were suggested for an LGM-size glaciation (only ZNO is covered by glaciers). 38% probability thus refers to a scenario where the glaciers (even during a large glaciation) would not reach the ZNO site, which in the model is realised as "no glaciation". Accordingly, ZNO is significantly more frequently glaciated than NL and both sites more often than JO (Fig. 3-47b).

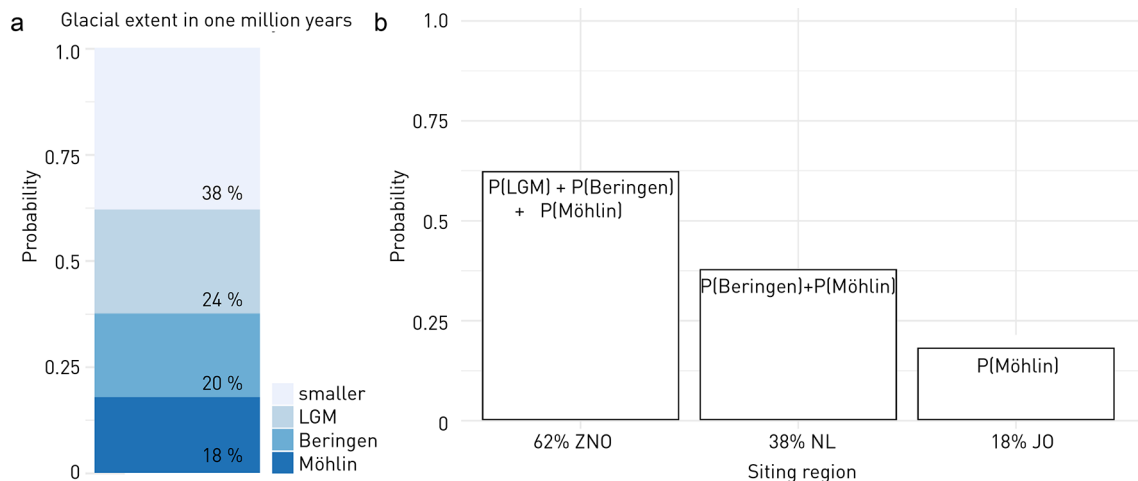


Fig. 3-47: Estimated glacial extent for glaciation (1 million years)

The stacked bar chart provides the elicited probabilities for glaciation at the one million yr timeframe to be of key glacial size or smaller.

3.5.4 Stepwise illustration of the procedure to evaluate the probabilities of fluvial and deep glacial erosion at a certain point

The distribution of the reference depth represents the deepest incision of a glacial overdeepening event in metres below the local erosion base. In the probabilistic model, the same reference depth is applied to each point within the region overdeepened by a glaciation event. Note that for evaluating the actual depth of overdeepening at a certain point within the overdeepened region, the reference depth needs to be scaled taking the actual rock types at this point into account. If Molasse is the only rock type between the repository and the LEB, then the reference depth can be used directly to evaluate the probability of repository excavation, conditional on a glacial overdeepening event directly above the repository. For evaluating the unconditional probability of repository excavation due to glaciation, additionally the probability of future glaciation events, the spatial occurrence of glacial overdeepening and the actual overdeepened area must be taken into account (Fig. 3-48).

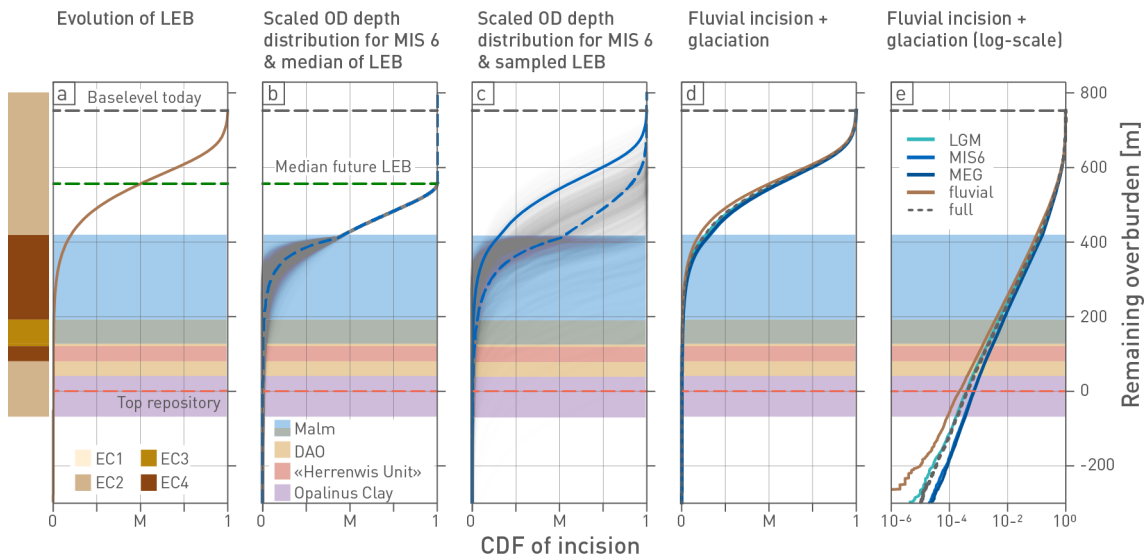


Fig. 3-48: Stepwise illustration of the procedure to evaluate the probabilities of fluvial and deep glacial erosion at a certain point

The illustration represents a stepwise visualisation of the probabilistic approach for the example of NL and the timing of one million years. The stratigraphic column below the Molasse is shown in all subfigures. The (colour-coded) associated erodibility classes are visible on the left. (a) Cumulative distribution function (CDF) of the future local erosion base (LEB, dashed green line shows median value). (b) Scaled depth distribution (CDF), shown for the example of a future MIS 6 (Beringen-size) glaciation linked to the median of the future LEB (dashed green line), derived by scaling the reference depth distribution (dashed blue CDF) with different sets of scaling factors sampled from their distributions (Fig. 3-42). The variability resulting from the scaling starts at the depth of the stratigraphic step between Molasse rocks and rocks of EC4 (grey lines). Note that this is a subset of the full samples and plotted only for illustration purposes. (c) Scaled depth distributions (grey CDF) for different Monte Carlo samples of the future LEB (blue solid CDF). The blue dashed line corresponds to the resulting CDF conditional on the glaciation and the occurrence of an overdeepening (OD) scenario at the location of the provisional disposal area. It does not involve the condition of an actual OD at the location¹⁹. This distribution is calculated conditional on the occurrence of a glaciation comparable to one during MIS 6. (d) Probability of incision (fluvial and glacial) for a certain remaining overburden thickness, including all key glacials with their probabilities of causing overdeepening within the provisional disposal area. The process termed fluvial erosion ('fluvial') describes the case of no glacial OD, e.g. for glaciations of a size smaller than during the LGM and corresponds to the distribution of the future LEB. (e) For better visibility, the incision probability (same as in d) is illustrated in log-scale. Note that in this illustration it is assumed that glacial overdeepening occurs directly above the provisional disposal area.

¹⁹ The probability for a point being actually overdeepened within an overdeepening scenario is below 100%. This is especially important for scenarios that allow random occurrence of new breaching channels within a relatively large area – not all of this area will be overdeepened.

4 Remaining overburden thickness and potential for repository excavation

Chapter 4 provides the final calculations of the erosion assessment, i.e. the remaining overburden thickness as well as glacial and non-glacial excavation probabilities. Section 4.1 provides the final calculations for the assessment indicators "remaining overburden thickness" and "excavation probability". Please note that excavation probability was calculated to show the robustness of the system. Because erosional excavation is highly unlikely, the excavation probabilities are defined by the tails of the distributions. A detailed discussion on the influence of certain parameters and combinations thereof can be found in Chapter 5.

The erosion assessment comprises the periods under consideration for L/ILW (100'000 years) and HLW (one million years), respectively. However, safety-relevant requirements for radionuclide doses are also defined for the time beyond the periods of consideration (ENSI 2023). For a discussion on erosion beyond one million years, including the hypothetical case of repository excavation, see Nagra (2024g). Based on the erosion assessment, presented here a distinction can be made between expected and alternative geological evolutions.

4.1 Expected and alternative long-term geological evolution

The main outputs shown in Fig. 3-3 result from a complex hybrid-probabilistic model cascade using many samples in the Monte Carlo approach (Chapter 3). While some intermediate results are valid for the entire area of the siting regions (e.g. the SPIM simulations of Model A), others can only be calculated for certain points or small areas associated with a limited number of points, e.g. the provisional disposal area.

Erosion as part of the long-term geological evolution needs to be considered in two of the thirteen criteria for safety and engineering feasibility of the Sectoral Plan (SFOE 2008); criterion 2.2 (Erosion) and criterion 3.3 (Predictability of long-term changes). To follow the general procedure of the safety-based site comparison, different outputs were generated for criterion 2.2 (showing the expected evolution) and criterion 3.3 (including alternative evolutionary paths). Please note that this differentiation is not to be mistaken with the considerations used for the "storyboard" in Section 6.2 (and the corresponding description of the *expected geological evolution* regarding erosion in Nagra 2024e). This storyboard narrates the anticipated erosional evolution within the next million years, defined as the 5-95% uncertainty band (also called expected range) in the full erosion assessment²⁰. It synthesises overarching conclusions with a comprehensive description, supporting the discussion on repository evolution in Nagra (2024f).

²⁰ Note that this set of results is again identical to that of criterion 3.3 (including alternative evolutionary paths) of the safety-based site comparison described above.

4.1.1 Criterion 2.2: expected evolution of the three sites

4.1.1.1 Assumptions

The future landscape assessed for criterion 2.2 (Erosion) for the non-glacial evolution strictly follows the present-day drainage network. Because it was decided to ignore glacial processes within the next 100'000 years (Section 3.2), calculations for L/ILW do not account for Model C. For one million years, only a subset of scenarios for potential future occurrence of glacial overdeepenings was used for the HLW perimeters of NL and ZNO (see Fig. 3-43) as well as the assumption that no future overdeepening will occur within the HLW perimeter of JO. In general, the 5th – 95th percentiles were assumed to represent the expected evolution for criterion 2.2.

Three key glacials of different extent and erosion potential were defined based on observations from the Quaternary that guide estimates for the future (Section 3.5.2). For criterion 2.2, the following assumption was used: JO can only be covered by a Möhlin-size glaciation, NL can only be covered by both a Möhlin-size and a Beringen-size glaciation, while ZNO can be covered by either of the three.

4.1.1.2 Results

Remaining overburden thickness after 100'000 years considering criterion 2.2

The evaluation suggests that, after 100'000 years, there is no significant change to the landscape under the assumptions used for criterion 2.2. Non-glacial excavation probability after 100'000 years is basically zero according to the results (Tab. 4-1). A detailed discussion is thus relinquished.

Tab. 4-1: Remaining overburden thickness and excavation probabilities after 100'000 years considering criterion 2.2 (abridged version)

Remaining overburden thickness and excavation probabilities are calculated with respect to the provisional disposal area of the L/ILW repository. The results are identical to those for criterion 3.3 and rounded to the nearest 5 metres.

Quantile / Overburden thickness	Jura Ost	Nördlich Lägern	Zürich Nordost
5%	490 m	835 m	670 m
25%	490 m	835 m	670 m
50%	490 m	835 m	670 m
75%	490 m	835 m	670 m
95%	490 m	835 m	670 m
Excavation probability	0	0	0

Remaining overburden thickness after one million years considering criterion 2.2

This subsection presents the combined results of the erosion assessment for the one-million-year timeframe (Tab. 4-2). Discussed are the remaining overburden thickness for the expected geological evolution as well as the probability of combined glacial and non-glacial excavation.

Tab. 4-2: Remaining overburden thickness and excavation probabilities after one million years considering criterion 2.2 (abridged version)

Remaining overburden thickness and excavation probabilities are calculated with respect to the provisional disposal area of the HLW repository. The results are rounded to the nearest 5 metres.

Quantile / Overburden thickness	Jura Ost	Nördlich Lägern	Zürich Nordost
5%	140 m	395 m	370 m
25%	250 m	535 m	565 m
50%	410 m	610 m	675 m
75%	410 m	665 m	675 m
95%	410 m	725 m	675 m
Excavation probability	0.00388	0.00028	0.00142

Fig. 4-1 shows the distributions of remaining overburden thickness above the provisional disposal areas after one million years as calculated for the expected evolution (see also Tab. 4-2). This figure is tripartite for better accessibility: the upper panel provides probability density functions (PDF), from which the most likely scenarios can be deduced (bulk of the curves, highlighted by the vertical arrows). The middle panel shows the same distributions as cumulative distribution function (CDF). The shaded areas represent the 5 – 95% probability (x-axis) and associated overburden thickness amounts (y-axis), which are considered here as "expected range" of future evolutions. Very low probabilities are not distinguished in the upper and middle panels. Therefore, the lower panel shows the CDF in log-space. The intercept with the position of the provisional disposal area (red dashed line) reveals the probability of excavation.

Remaining overburden thickness is calculated in a Monte-Carlo approach using one million samples. The respective probabilities to fall below a specific vertical datum (e.g. the level of the provisional disposal area) are deduced from the Monte-Carlo samples. These distributions were calculated for the four corner- and one mid-point of the provisional disposal areas and reflect the "worst" of them (with respect to the failure criterion). In the case of Model B (modelling of local topography), however, the remaining overburden thickness is extracted from a raster (25 m grid size) using the minimum value within the entire perimeter of the provisional disposal area.

Considering the assumptions for criterion 2.2, at least 400 m of overburden thickness in JO and even more than 600 m in NL and ZNO are most likely to remain after one million years (see arrows and thick dotted line in Fig. 4-1a). Note that there is only negligible present-day local topography above the provisional disposal area in NL, as the Glatt river is located close to this location. This is reflected by the similarity between the curves "Fluvial incision" (modelled LEB according to Model A) and "Local topography". Incorporating local topography has a positive effect with respect to remaining overburden thickness in JO and ZNO (compare Section 3.4.6). The range (5th – 95th percentile) of the expected remaining overburden thickness is highlighted by the shaded areas and vertical arrows in Fig. 4-1b. These ranges are ~ 140 to 410 m in JO, ~ 395 to 725 m in NL, and ~ 370 to 675 m in ZNO, at the location of the provisional disposal areas, respectively (see also Tab. 4-2). Using this range, the Hauptrogenstein in JO and Malm limestones in NL, respectively, may just become exposed at the surface at the location of the provisional disposal areas (Fig. 4-1b).

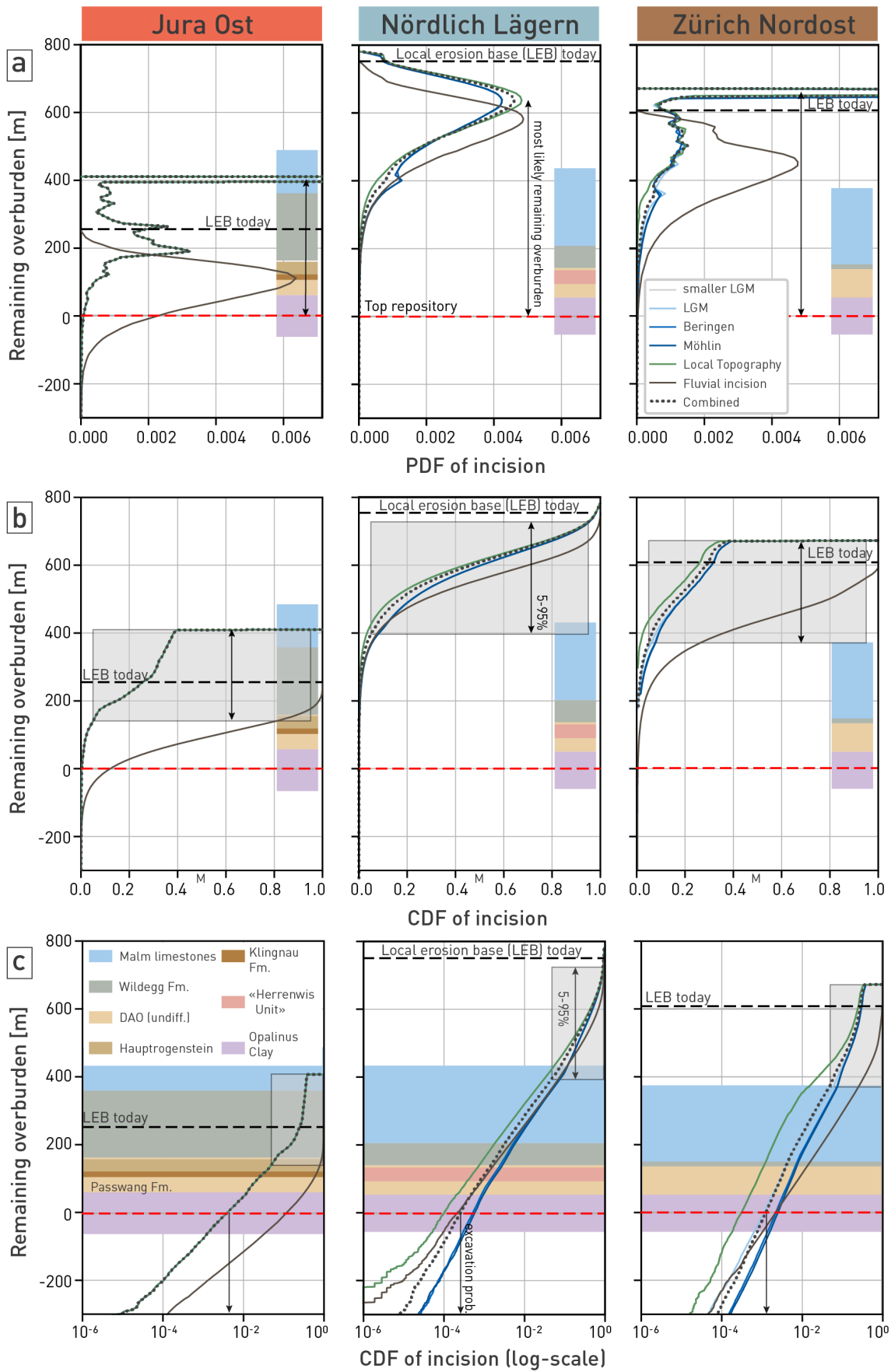


Fig. 4-1: Estimated probabilities of remaining overburden thickness after one million years as calculated for the expected evolutionary paths

The same outputs are displayed in three different formats for better visualisation (a) as probability density function (PDF), (b) as cumulative distribution function (CDF), and (c) as CDF transferred onto a logarithmic scale (see Appendix, Fig. C-1 for supporting information on CDFs). The left, central, and right panels correspond to the results for JO, NL, and ZNO, respectively. Results are shown with respect to the depth of the provisional disposal area (red dashed line). The horizontal black dashed line corresponds to the local erosion base (LEB) today. The thick black dotted line represents the final, combined, probability of remaining overburden thickness. It integrates over i.) both glacial and non-glacial processes and ii.) the occurrence of all considered key glacials. The lines "smaller LGM", "LGM", "Beringen" and "Möhlen" integrate over both glacial and non-glacial processes, but are restricted to the occurrence of the specific key glacial. The line "Local Topography" integrates over all considered key glacials but is restricted to non-glacial processes (output Model B, i.e. no glacial overdeepening). The line "Fluvial incision" describes the modelled LEB from Model A.

For JO, the final combined curve directly corresponds to the evolution of local topography (line "Local Topography") assuming preservation of the present-day main drainage network; deep glacial erosion is not part of the expected evolution in JO. For NL and ZNO, the final, combined probability is a combination of the fluvial incision by main rivers (line "Fluvial incision"), evolution of local topography (assuming preservation of the present-day main drainage network) and deep glacial erosion (output Model C; conditional on the occurrence of key glacials). The vertical arrows in (a) highlight the most likely amount of remaining overburden thickness at the location of the provisional disposal area. In (b) the 5 – 95% range is shown and highlighted by the grey shaded rectangle, see M for position of median values, and in (c) The probability of excavation is represented by the intersection of the probability curve with the depth of the provisional disposal area, which is significantly lower in NL. Note for comparison the 5 – 95% range (grey shaded box), see Section 5.2 for a discussion on the extreme end of the probabilities. Simplified lithostratigraphic columns are provided with these images, allowing visualisation of the incision depth. DAO: Dogger Group above Opalinus Clay.

An excavation of the repository within one million years was estimated to be highly unlikely in all three siting regions. This finding is best illustrated in the log-scale visualisation of the distribution (vertical arrows in Fig. 4-1c). This likelihood is about an order of magnitude lower at the provisional disposal area in Nördlich Lägern than in JO and ZNO.

4.1.2 Criterion 3.3: including alternative evolution of the three sites

4.1.2.1 Assumptions

To emphasise the possibility of different paths of future landscape evolution, evaluation of criterion 3.3 (Predictability of long-term changes) is, in addition to the scenario with the present-day drainage, also supported by a set of alternative drainage networks (ADN) for JO (Fig. 3-24). Modelled scenarios are (i.) the reactivation of the Rinikerfeld channel, an Aare shortcut using the Sissle drainage (ii.) with and (iii.) without incorporation of the Rivers Reuss and Limmat along this new route, and (iv.) a breakthrough channel following the Itele Valley. Because the relief (local topography) within the relevant parts of NL and within ZNO is too low to provide meaningful alternative scenarios for modelling of future landscape evolution, a (worst-case) scenario is added that assumes a main drainage developing just above the disposal area. In this case, the remaining overburden thickness is directly derived from Model A.

For 100'000 years (period of consideration for the L/ILW repository), no ADN was calculated based on the assumption that, within the present landscape, a glaciation would be needed for significant changes in the drainage network. Hence, only one distribution of remaining overburden thickness exists for 100'000 years, and the 95th percentile was suggested to represent criterion 3.3.

Because of the proximity of the disposal areas in NL and JO to the next-smaller of the three key glaciations, these were additionally considered for criterion 3.3 considering one million years (i.e. Beringen-size for JO and LGM-size for NL).

4.1.2.2 Results

Remaining overburden thickness after 100'000 years considering criterion 3.3 (full model)

The results of the remaining overburden thickness after 100'000 years are identical to those for criterion 2.2 (see Section 4.1.1.2 and Tab. 4-1) and thus not further discussed.

Remaining overburden thickness after one million years considering criterion 3.3 (full model)

This subsection shows the combined results of the erosion assessment for the one-million-year timeframe (Tab. 4-3). Discussed are the remaining overburden thickness as well as the probability of combined glacial and non-glacial excavation, if including alternative evolutions.

The most likely scenarios at the location of the disposal areas of all sites show at least 400 m of remaining overburden thickness above the repository after one million years (vertical arrows in Fig. 4-2a). In NL, nearly 600 m of overburden thickness are most likely. In the 5 – 95% probability range, between ~ 125 and 410 m of overburden thickness might be expected in JO, between ~ 350 and 670 m in NL and between ~ 220 and 560 m in ZNO (shaded area and vertical arrows in Fig. 4-2b). Considering the 200 m respect overburden thickness for efficient self-sealing (Chapter 1; Nagra 2024e), the Jura Ost site shows more limitations within this range than the other sites at the position of the provisional disposal area. In this scenario range, the Malm limestones in NL and ZNO and Hauptrogenstein in JO are likely to be exposed at the surface at the location of the provisional disposal area. However, suitable conditions in JO are available in the immediate surrounding of the provisional disposal area (Section 4.1.3).

An excavation of the repository within one million years was estimated to be highly unlikely in all three siting regions. This fact is best seen in the log-scale visualisation of the distribution (vertical arrows in Fig. 4-2c, see grey box showing the 5 – 95% range for comparison). This likelihood is about an order of magnitude lower for the provisional disposal area in Nördlich Lägern than in JO and ZNO.

Fig. 4-1 and Fig. 4-2 also show the main contributors to the distributions of remaining overburden thickness. Although deep glacial erosion can incise intensely into the overburden thickness, its probability at the position of the repository is strongly reduced. The main contributors consequently are fluvial incision and associated evolution of local topography (the latter for JO only) (see also Chapter 5 for a discussion on model parameter sensitivities and plausibility of the results). The present-day low relief in ZNO and in the east of NL, thus the area of the repository location, provides a good reasoning to ignore local topography and compare with the local erosion base (see Section 3.4). However, it cannot be excluded that the location of the main drainage will be shifted away from the disposal area and a significant protective relief will develop in the future. Accordingly, the remaining overburden thickness might be further increased, and the probability of excavation further reduced in NL and ZNO. In general, the results reveal considerable differences in the remaining overburden of the top of the Opalinus Clay between the three siting regions, which translate into the probability for glacial excavation when considering the depth of glacial erosion.

Special consideration should be given to the local topography when addressing the potential for glacial excavation at JO. Because the local erosion base constitutes the main reference plane for glacial overdeepenings, the strong lowering affects JO more because of the shallower original depth of the host rock below local erosion base. Local topography has strongly influenced the argumentation for the relative plausibility of overdeepening scenarios in JO, and both overdeepening occurrence (Section 3.5.3.3) and spatial densities (Section 3.5.3.4) were considered significantly lower than in NL and ZNO, but in the end (Fig. 4-2) this effect does not seem to provide an overall advantage for this siting region.

Consequently, considering the three erosion processes, NL provides the best protection against future erosion.

Tab. 4-3: Remaining overburden thickness and excavation probabilities after one million years considering criterion 3.3 (abridged version)

Remaining overburden thickness and excavation probabilities are calculated with respect to the provisional disposal area of the HLW repository. The results are rounded to the nearest 5 metres.

Quantile / Overburden thickness	Jura Ost	Nördlich Lägern	Zürich Nordost
5%	125 m	350 m	220 m
25%	240 m	480 m	355 m
50%	410 m	550 m	425 m
75%	410 m	605 m	480 m
95%	410 m	670 m	560 m
Excavation probability	0.00552	0.00043	0.00387

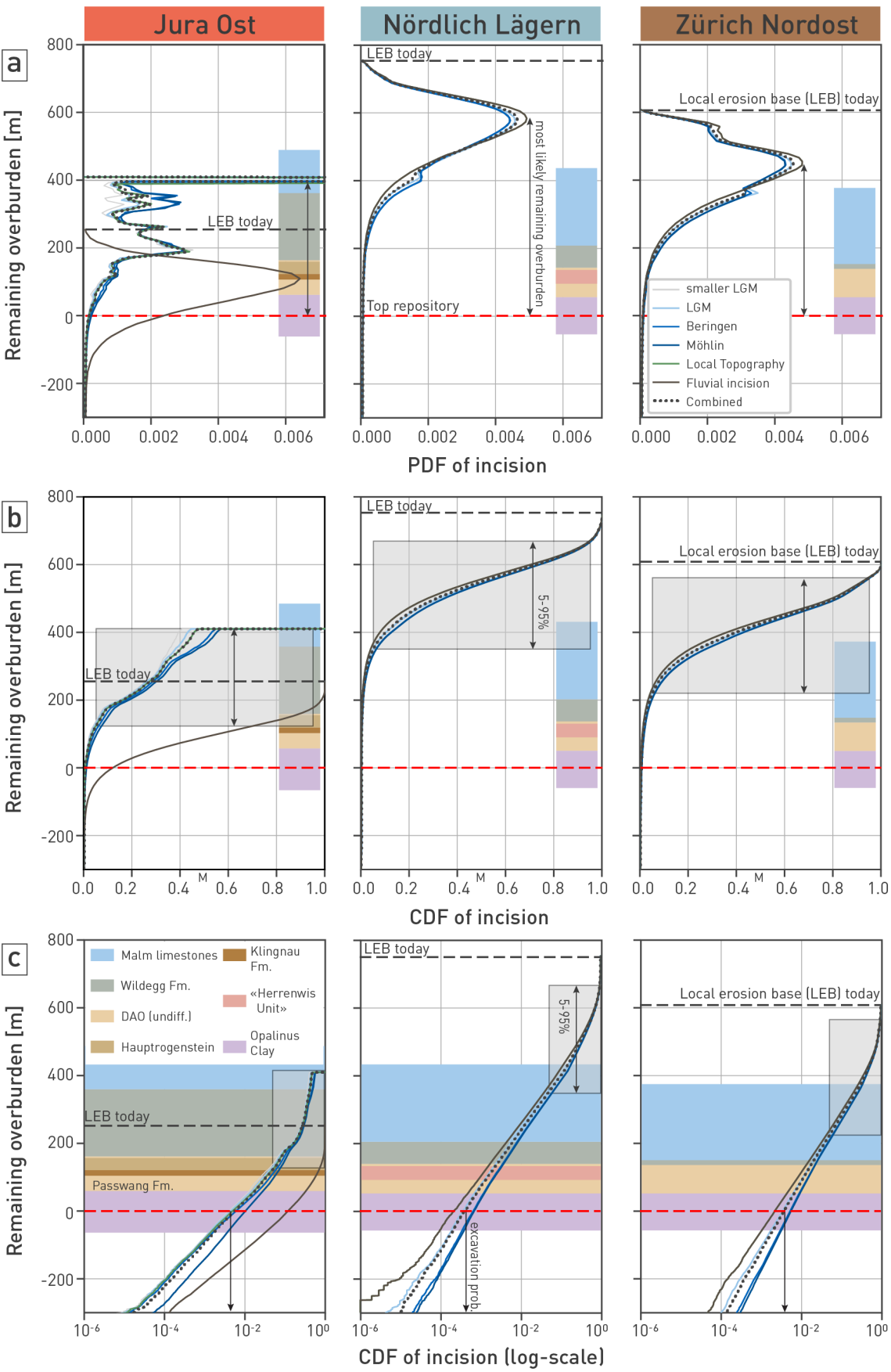


Fig. 4-2: Estimated probabilities of remaining overburden thickness after one million years as calculated with inclusion of alternative evolutionary paths

The same outputs are displayed in three different formats for better visualisation (a) as probability density function (PDF), (b) as cumulative distribution function (CDF), and (c) as CDF transferred onto a logarithmic scale. The left, central, and right panels correspond to the results for JO, NL, and ZNO, respectively. Results are shown with respect to the depth of the provisional disposal area (red dashed line). The horizontal black dashed line corresponds to the local erosion base (LEB) today. The thick black dotted line represents the final, combined probability. That is, it integrates over i.) both glacial and non-glacial processes and ii.) the occurrence of all considered key glacials. The lines "smaller LGM", "LGM", "Beringen" and "Möhlen" integrate over both glacial and non-glacial processes, but are restricted to the occurrence of the specific key glacial. The line "Local Topography" integrates over all considered key glacials for determination of alternative drainage networks but is restricted to non-glacial processes (output Model B, i.e. no glacial overdeepening). The line "Fluvial incision" describes the modelled LEB from Model A.

For JO, the final combined curve is a combination of the fluvial incision by main rivers (line "Fluvial incision"), evolution of local topography (including ranked scenarios) and deep glacial erosion (output Model C; conditional on the occurrence of key glacials). Local topography was not included in the case of NL and ZNO. Here, a main river (defining the local erosion base) was assumed to be located directly above the provisional disposal area (see Section 3.4). The vertical arrows in (a) highlight the most likely amount of remaining overburden thickness at the location of the provisional disposal area. In (b) the 5 – 95% range is shown and highlighted by the grey shaded rectangle, see M for position of median values, and in (c) The probability of excavation is represented by the intersection of the probability curve with the depth of the provisional disposal area, which is significantly lower in NL. Note for comparison the 5 – 95% range (grey shaded box), see Section 5.2 for a discussion on the extreme end of the probabilities. Simplified lithostratigraphic columns are provided with these images, allowing visualisation of the incision depth. DAO: Dogger Group above Opalinus Clay.

4.1.3 Criteria 2.2 & 3.3 - expanding the calculations to spatial reserves

The main calculations and outputs shown in Sections 4.1.1 and 4.1.2 are based on the provisional disposal area – a current layout that might be subject to future repositioning within the 6 km² perimeter known as spatial reserves (see Section 3.1 for introduction and Nagra 2024a for further information). To account for such a possibility, a slightly simplified approach was used, where the remaining overburden thickness and probability of excavation are evaluated at points (point-approach). These points were selected arbitrarily to cover the range of different boundary conditions within the spatial reserves (e.g. different depth of Opalinus Clay or different scenarios for alternative drainage or occurrence of glacial overdeepenings (Fig. 4-3). Similarly to Sections 4.1.1 and 4.1.2, the expanded calculations were performed for the differentiation between criterion 2.2 and criterion 3.3. The five points comprising the corner and central point of the provisional disposal area are included in this point calculation for comparison. Note that their values may differ from those discussed in the sections above due to the adaptations made in the point-approach.

With respect to criterion 2.2, the ranges of remaining overburden thickness (resembling the 5% samples of least overburden thickness) calculated using this simplified approach comprise approximately 185 m (in the northeast) to 285 m (in the south) of the spatial reserves in JO (Fig. 4-3b), between approximately 410 m (in the centre; due to proximity to the Glatt River, see Section 4.1.1) to 575 m (in the south) of the spatial reserves in NL (Fig. 4-3e), and between about 385 m (in the northwest) and 610 m (in the east) of the spatial reserves in ZNO (Fig. 4-3h), respectively.

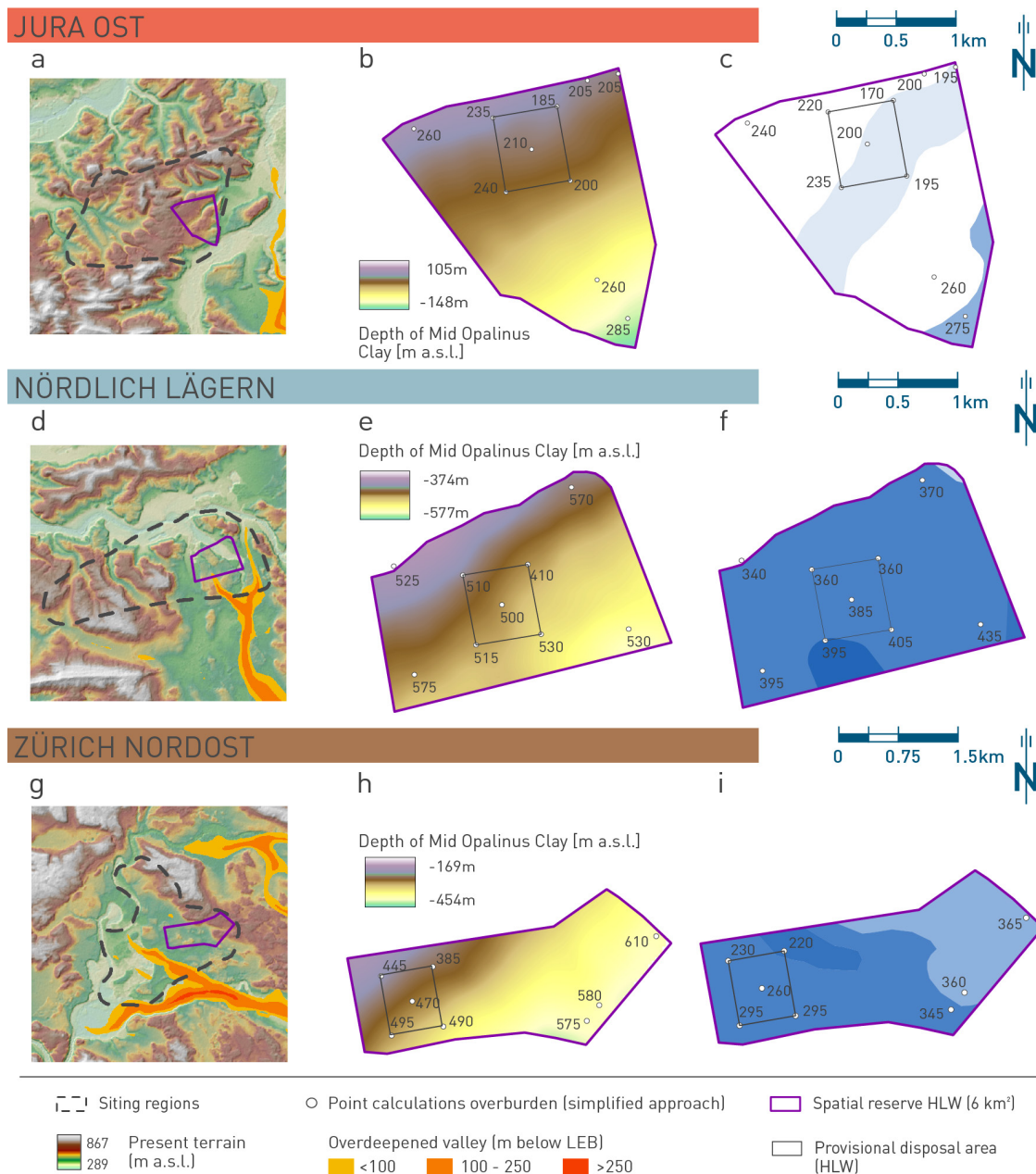


Fig. 4-3: Simplified point calculations of remaining overburden thickness within the spatial reserves

For each siting region the location of spatial reserves (purple outline) within the HLW outline (black dashed outline) is shown on the left images (a, d, g) with existing glacial overdeepenings, if present. The middle and right images are cropped to the respective extent of these spatial reserves with the mid depth of the Opalinus Clay (in m a.s.l. shown in b, e, and h) and scenarios for potential future overdeepening occurrences (c, f, and i) in the background (see Section 3.5.3.3 for more details, note here only that the darker the colour the more likely a scenario becomes). Also shown is the provisional disposal area (black rectangle). The points for which the simplified approach was used were selected arbitrarily to cover a range of different depths and a variety of potential overdeepening scenarios. They are labelled as remaining overburden thickness (in metres) on the middle images with results for the criterion 2.2 subset (see assumptions in Section 4.1.1) and on the right images for the criterion 3.3 subsets (values represent P05). The 5 points comprising the corner and central point of the provisional disposal area are included in this point calculation for comparison.

Calculations considering assumptions for criterion 3.3 reveal between approximately 170 and 275 m remaining overburden thickness in JO (Fig. 4-3c), ~ 340 to 435 m in NL (Fig. 4-3f), and between approximately 220 and 365 m in ZNO (Fig. 4-3i), respectively.

The ranges show that locations with remaining overburden thickness of more than 200 metres, as considered reliable for self-sealing in the Opalinus Clay (Chapter 1) are present within the spatial reserves of all three siting regions.

5 General discussion and plausibility of the results

5.1 Main influence on long-term erosion at the sites

Because evaluated landscape changes within the next 100'000 years are minor with respect to the remaining overburden thickness, only sensitivities for the fluvial modelling were shown for both times (Section 3.3.4). The following discussion on the sensitivity of model parameters in the full model focuses on the one-million-year timescale. First-order Sobol indices and Tornado plots are both used in sensitivity analysis, but they differ significantly in what they measure and how they present information. Tornado plots (Section 5.1.1) provide a straightforward, visual view, useful for intuitive interpretation. Tornado plots are usually based on a scenario analysis where each input is varied within a specific range while holding other inputs constant. In contrast, 1st order Sobol indices (Section 5.1.2) offer a more mathematically precise, variance-based measure, also capturing complex relationships.

5.1.1 Sensitivity of model parameters – Tornado plots

Fig. 5-1 uses tornado plots to illustrate the sensitivity of individual model parameters while keeping all others at their baseline values. Each siting region has its own plot, based on the remaining overburden thickness above the provisional disposal area over one million years, using the model configuration from Section 4.1.2.

Tornado plots typically apply to continuous variables, where baseline values can be quantitatively defined. For continuous parameters, the baseline corresponds to the mean of the probability distribution, and the 'base' model outcome is calculated by evaluating the model with all continuous parameters at their baseline values. However, the model also includes discrete nominal parameters (e.g. 'size of glaciation') that lack a natural baseline. The 'base' outcome for these parameters is defined as the expected remaining overburden thickness, calculated by considering all discrete states and their probabilities, while continuous parameters are set to their mean values.

For continuous parameters, 'low' and 'high' values are determined by the 5th and 95th percentiles of their distributions. In contrast, black dots on the plots represent remaining overburden thickness for the discrete parameters, conditional on the potential discrete parameter states with other parameters held at baseline values.

Red bars indicate parameters that reduce remaining overburden thickness, while green bars indicate those that increase it. The uplift rate at Brugg is notably influential in decreasing overburden thickness across all sites, while in NL and ZNO it may also contribute to an increase. In JO (where modelling of local topography is included), factors like the erodibility coefficient K (ECb) and alternative drainage scenarios significantly affect overburden thickness, particularly due to the Itele drainage scenario, which would cause a rerouting of the Aare River across the provisional disposal area. Other parameters, such as azimuth, gradient, and baselevel drop, influence both increases and decreases in remaining overburden thickness.

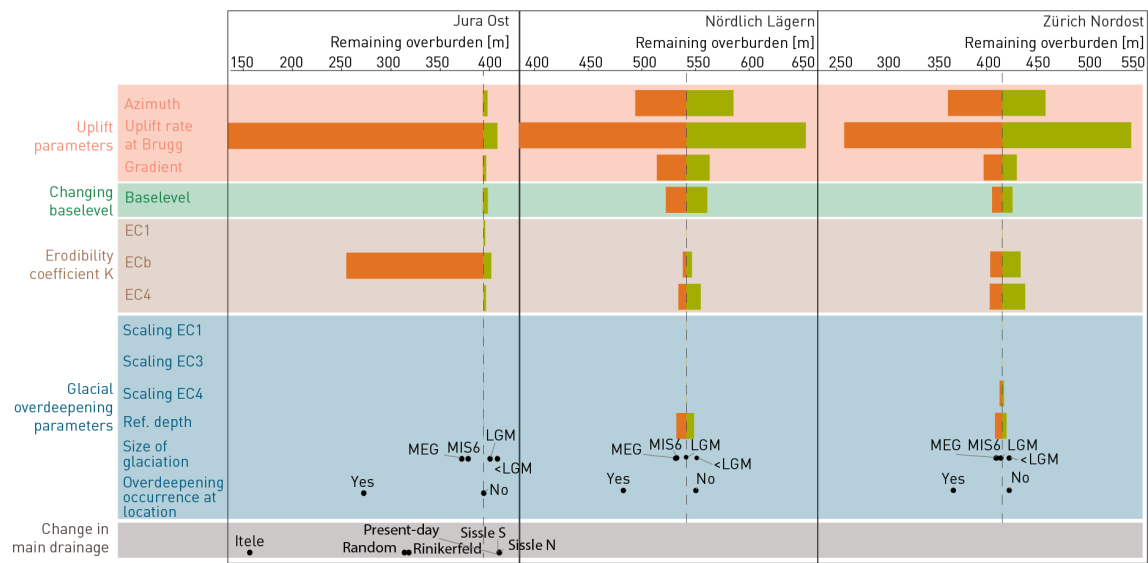


Fig. 5-1: Tornado plots highlighting the most influential model parameters of the full model

Illustration of the sensitivity of individual model parameters while keeping all others at their baseline values. Tornado plots were generated for each siting region but only for the one million year timescale. Red bars indicate parameters that reduce remaining overburden thickness, while green bars indicate those that increase it. Deterministic scenarios cannot be plotted in the same way (there are no distributions); their influence is thus marked by their relative positions (black dots). The determinator yes/no describes whether or not at the location of interest, in this case the location of the provisional disposal area, overdeepening occurs (see Section 3.5.3).

5.1.2 Sensitivity of model parameters – 1st order Sobol indices and total-effect sensitivity of the glacial parameters

1st order Sobol indices help to quantify the individual impact of each input variable on the output variance and to identify which variables are most influential in determining the variability in the model's output for a specific siting region. They identify the model parameters that on average, once fixed, would cause the largest reduction in variance (Saltelli et al. 2008). These indices are thus useful in improving complex models by identifying the most influential factors. Such a ranking of model input parameters can (1) help to enhance the model with a focus on the most influential parameters and (2) assist in the decision for which parameters collecting additional information can have the largest impact on decreasing model uncertainty (i.e. decreasing the variance in the model output). However, to achieve the latter, only epistemic uncertainties should be considered for the sensitivity analysis, as aleatory uncertainties cannot be reduced by collecting additional information. For the probabilistic model at hand, the uncertainty about most parameters is composed of both epistemic and aleatory uncertainties. The contribution of the discrete model parameters (drainage scenarios, glaciation, topo-scenario, overdeepening scenario and occurrence of overdeepening at the repository location (YES/NO)) should be viewed as purely aleatory in the current setting. This needs to be considered when interpreting the ranking. Tab. 5-1, Tab. 5-2, and Tab. 5-3 list the 1st order Sobol indices for the model parameters for each of the siting regions.

Additional to the 1st order Sobol indices, the total-effect sensitivity index of the group of all glacial model parameters (including the alternative drainage network parameter in the case of JO, which are assessed conditional on glaciations) are evaluated. The total-effect sensitivity index expresses how much the respective parameters and their interactions with other parameters contribute to the variance of the model output. For JO, NL and ZNO, the total-effect sensitivity index of the glacial group is 29%, 47% and 39%, respectively. This shows that the glacial model parameters (and ADN for JO) are responsible for a significant portion of the uncertainty in the model output. Thus, even though most of the parameters of the glaciation model have a small 1st order Sobol index (and consequently their individual contribution to the variance of the model output is small), their influence on the variance of the model output cannot be neglected, due to interactions with other parameters. For example, for an above average drop in LEB, the information that a glaciation event occurs that actually causes overdeepening above the repository, will very likely result in a small remaining overburden. On the other hand, for small drops in LEB without the occurrence of a glaciation event within the considered time period, the remaining overburden is very likely above average. Information about such interactions with all other model parameters is quantified by the total-effect sensitivity index.

5.1.2.1 Jura Ost

Considering the 1st order Sobol indices, the most influential uncertain model parameters in Jura Ost are uplift at Brugg and the parameters related to the gradient of the uplift plane. The erodibility coefficients (*K* values) of ECb, EC1 and EC4 also have a notable contribution (Tab. 5-1). Apart from the reference depth of overdeepening in the Molasse, all other uncertain model parameters have a small/negligible influence.

Tab. 5-1: 1st order Sobol indices calculated for Jura Ost

The columns are structured as follows: (1) The name of the uncertain model parameter is given. (2) The value of the 1st order Sobol index (estimated based on one million samples) is given. Note that gradient refers to the gradient in direction of the median azimuth, while Ugrad is the transformed gradient to account for any azimuth from the total assumed probability. High values of 1st order Sobol indices indicate a large influence on the output variance.

Uncertain model input parameter	Sobol 1 st order
Azimuth	0.29%
Uplift at Brugg	33.70%
Gradient	32.80%
Ugrad	32.62%
Uplift at Basel (for correlation)	0.99%
Baselevel	0.31%
Erodibility coefficient (EC1)	20.72%
Erodibility coefficient (ECb)	23.07%
Erodibility coefficient (EC4)	16.67%
Drainage scenario Iller	0.00%
Drainage scenario Aare	0.00%
Drainage scenario Rhine	0.00%
Drainage scenario Upper Danube	0.00%
Lithological scaling EC1 (JO)	0.00%
Lithological scaling EC3 (JO)	0.00%
Lithological scaling EC4 (JO)	0.00%
Glaciation event (key glacial)	0.09%
Topo scenario	0.41%
Overdeepening scenario	0.32%
Overdeepening at repository (YES/NO)	0.03%
Reference depth of overdeepenings (Molasse)	9.32%

5.1.2.2 Nördlich Lägern

Considering the 1st order Sobol indices, the most influential uncertain model parameter is the reference depth of overdeepening in the Molasse. However, a significant part of the input uncertainty about this parameter is of the aleatory type. The second most influential quantities are the parameters related to the gradient. Uplift upstream of Basel (*UB*, approximately at the location of Pratteln) is the third most influential parameter. However, note that *UB* is only a derived parameter that is computed deterministically from uplift at Brugg (*UBR*), the azimuth and the gradient of the uplift plane and is used to induce a correlation between *UBR* and the gradient. Uplift at Brugg, baselevel drop and the *K* values of EC1, ECb and EC4 also have a notable contribution (Tab. 5-2). All other uncertain model parameters have a small/negligible influence.

Tab. 5-2: 1st order Sobol indices calculated for Nördlich Lägern

The columns are structured as follows: (1) The name of the uncertain model parameter is given. (2) The value of the 1st order Sobol index (estimated based on one million samples) is given. Note that gradient refers to the gradient in direction of the median azimuth, while Ugrad is the transformed gradient to account for any azimuth from the total assumed probability. High values of 1st order Sobol indices indicate a large influence on the output variance.

Uncertain model input parameter	Sobol 1 st order
Azimuth	0.12%
Uplift at Brugg	5.17%
Gradient	9.86%
Ugrad	9.34%
Uplift at Basel (for correlation)	7.69%
Baselevel	2.52%
Erodibility coefficient (EC1)	3.10%
Erodibility coefficient (ECb)	2.98%
Erodibility coefficient (EC4)	2.98%
Drainage scenario Iller	0.00%
Drainage scenario Aare	0.00%
Drainage scenario Rhine	0.00%
Drainage scenario Upper Danube	0.00%
Lithological scaling EC1 (JO)	0.00%
Lithological scaling EC3 (JO)	0.00%
Lithological scaling EC4 (JO)	0.01%
Glaciation event (key glacial)	0.01%
Overdeepening scenario	0.01%
Overdeepening at repository (YES/NO)	0.03%
Reference depth of overdeepenings (Molasse)	10.68%

5.1.2.3 Zürich Nordost

Considering the 1st order Sobol indices, the most influential uncertain model parameters are the parameters related to the gradient and uplift at Brugg. The reference depth of overdeepening (in the Molasse) and the K values of ECb, EC1 and EC4 also have a notable contribution. Uplift upstream of Basel (Pratteln), baselevel and the azimuth of the uplift plane have a small contribution. All other uncertain model parameters have a negligible influence (Tab. 5-3).

Tab. 5-3: 1st order Sobol indices calculated for Zürich Nordost

The columns are structured as follows: (1) The name of the uncertain model parameter is given. (2) The value of the 1st order Sobol index (estimated based on one million samples) is given. Note that gradient refers to the gradient in direction of the median azimuth, while Ugrad is the transformed gradient to account for any azimuth from the total assumed probability. High values of 1st order Sobol indices indicate a large influence on the output variance.

Uncertain model input parameter	Sobol 1 st order
Azimuth	0.83%
Uplift at Brugg	12.51%
Gradient	14.87%
Ugrad	13.77%
Uplift at Basel (for correlation)	1.80%
Baselevel	0.66%
Erodibility coefficient (EC1)	7.31%
Erodibility coefficient (ECb)	7.86%
Erodibility coefficient (EC4)	7.99%
Drainage scenario Iller	0.00%
Drainage scenario Aare	0.00%
Drainage scenario Rhine	0.00%
Drainage scenario Upper Danube	0.00%
Lithological scaling EC1 (JO)	0.00%
Lithological scaling EC3 (JO)	0.00%
Lithological scaling EC4 (JO)	0.01%
Glaciation event (key glacial)	0.06%
Overdeepening scenario	0.05%
Overdeepening at repository (YES/NO)	0.18%
Reference depth of overdeepenings (Molasse)	10.04%

5.2 Main influence on the extreme endmember scenarios (distribution tails)

The most extreme overburden thickness reduction, culminating in the erosional excavation probabilities after one million years require a combination of parameters that are extremely unlikely to occur simultaneously. Therefore, it is important to discuss these outcomes, especially if using a rather complex model where many parameter combinations are included. In the following, for the "worst" 10'000 and 1'000 out of one million Monte Carlo samples, i.e. the 10'000 (or 1'000) samples that yielded the lowest remaining overburden thickness after one million years, parameters are analysed with respect to their influence on the results (Fig. 5-2 and Fig. 5-3).

Fig. 5-2 compares the cumulative distribution functions of a series of parameters (1 million years, NL) of all Monte Carlo samples (one million), as well as 10'000 (1%) and 1'000 (0.1%) samples. A parameter is influential on the extreme overburden thickness reduction if the curves comprising all samples (green) differ significantly from the others (orange and light blue). A large contribution is visible with respect to the gradient, i.e. the gradient in the direction of the reference azimuth (Fig. 5-2b), at uplift at Brugg (Fig. 5-2d), the erodibility coefficient K for EC4 (Fig. 5-2h), and the reference depth distribution of overdeepenings in the Molasse (Fig. 5-2i). Still influential are azimuth (Fig. 5-2a), the gradient that is active at the preferred azimuth (U Grad, Fig. 5-2c) as well as the other erodibility coefficients (K) for EC1 (Fig. 5-2f) and ECb (Fig. 5-2g). The rock-type scalings of the overdeepenings (Fig. 5-2j – l) seem not to play an important role for the extreme distributions.

Fig. 5-3 shows the different distributions of these influential parameters as a violin plot. The range of susceptible values that influence the extreme overburden thickness reduction can be deduced from these plots. For the 1'000 worst samples, the critical range is highlighted with the red bar at the respective right edge of the subplots. Accordingly, an azimuth range between $\sim 120 - 140^\circ$ (Fig. 5-3a), a rather steep gradient (Fig. 5-3b), uplift rates in Brugg above 0.4 mm/yr (Fig. 5-3c), reference depths of overdeepenings between ~ 200 and 450 m, low baselevels below the present elevation of Basel (Fig. 5-3e) and rather high erodibility coefficients (K) in all three classes (Fig. 5-3f – h) play a role.

While the critical azimuth samples ($120 - 140^\circ$) stem from the expected range (5 – 95%) of the main distributions and thus could be considered rather reliable (Fig. 5-3a), the very high gradients seem to correspond to the tail of the main distribution (Fig. 5-3b and c). Similar to azimuth, the unfavourable Monte Carlo samples from baselevel and depth distribution correlate with the expected range (Fig. 5-3e and d, respectively), while the erodibility coefficients (K) again resemble samples from the tails of their main distributions (Fig. 5-3g and h). Because the erodibility coefficients (K) are strongly correlated with uplift in Brugg, their strong influence may be an artifact stemming from this correlation.

However, more typically it is not the unfavourable range of a single parameter, but a combination of them that triggers extreme behaviour. The joint influence of a large number of uncertain model parameters, however, is difficult to quantify. However, this was nevertheless calculated for parameter pairs and an example is shown in Fig. 5-4 for the parameters uplift in Brugg and baselevel.

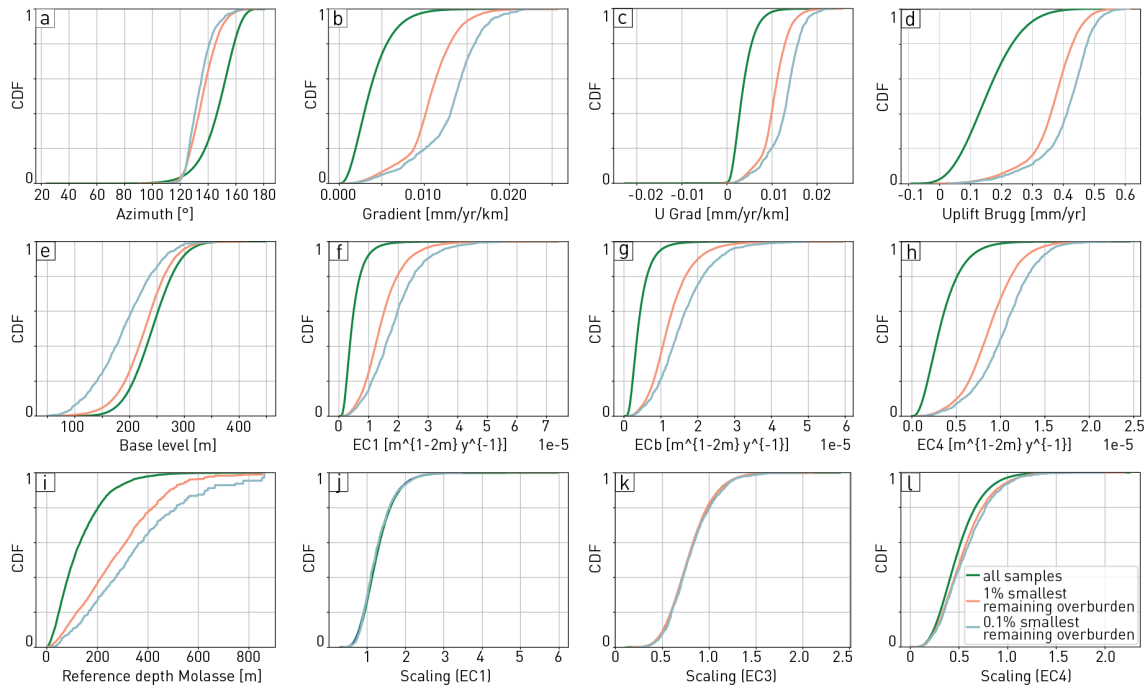


Fig. 5-2: Influence of model parameters on extreme overburden thickness reduction (one million yr, NL)

The plots show cumulative distribution functions of a number of parameters used for calculating remaining overburden thickness after one million years in Nördlich Lägern. Each plot contains the distributions for all samples, as well as worst (least overburden thickness) 10'000 (1%) and 1'000 (0.1%) samples. The parameter is influential on the extreme values if the green curve (all samples) differs from the orange and blue curves. Such differences are visible in all but the three erodibility scalings of the Molasse depth distribution for glacial overdeepening (j – l). Note that "gradient" refers to the elicited gradient in the direction of the reference (median) azimuth, while "U grad" is based on the gradient modified as function of the actual azimuth used in SPIM.

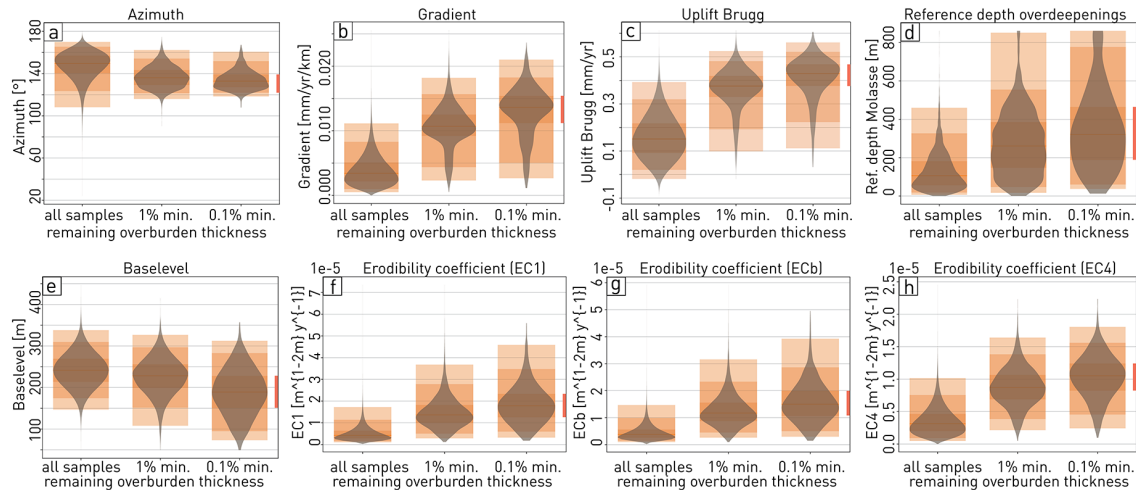


Fig. 5-3: Range of values leading to extreme distributions

Violin plots show the distributions and the same comparison as in Fig. 5-2, all samples, 1%, and 0.1%, respectively. The violin shape represents the density of data points at various values along an axis. (a) Azimuth values between ~ 120 and 140° seem to have an influence on extreme overburden thickness reduction, as well as high gradients (b), a high uplift at Brugg > 0.4 mm/yr (c), a reference depth of overdeepenings between ~ 200 and 450 m (d), baselevel lowering to $\sim 150 - 230$ m (f, presently at 250 m), and rather higher erodibility coefficients (K) in all three classes (f to h). The small red bars to the right of each figure highlight these ranges.

As an example, the visualisations in Fig. 5-4 show the parameters uplift in Brugg (UBR , Fig. 5-4a) and baselevel (BL , Fig. 5-4b) and their joint influence on remaining overburden thickness (Fig. 5-4c). In both plots (Fig. 5-4a and b), the lines deviate considerably from the mean of the parameter (the dashed bold red line) and even intersect with the 99%-quantile value of the parameter (dotted red line). Consequently, larger values of UBR and small values of BL cause smaller remaining overburden thickness (note that baselevel here corresponds to the absolute value calculated from the present-day elevation of Basel [250 m a.s.l.] and the rate of baselevel change multiplied by one million years [time period for which the rate was elicited]). Comparing the spread of the bluish points with the 1%- and 99%-quantile values shows that realisations with small remaining overburden are associated with a comparatively large spread. Thus, each of these parameters (here UBR and BL) alone does not fully explain the small remaining overburden thickness values.

An example of the plot for the interaction of UBR and BL is shown in Fig. 5-4c. There is a trend between increasing UBR and decreasing BL and a decrease in the remaining overburden thickness above the repository – this can be best seen by the coloured line of the moving average with a window of 1'000 samples. Also seen is a clear dependence between BL and UBR . The worst samples, not surprisingly, seem to require both a large uplift in Brugg and a low baselevel. This, however, is a rather implausible combination, because increased uplift rates are expected to cause enhanced erosion in Northern Switzerland, which will lead to a sediment pulse reaching the Upper Rhine Graben and a situation where sediment is expected to balance tectonic subsidence and baselevel drop.

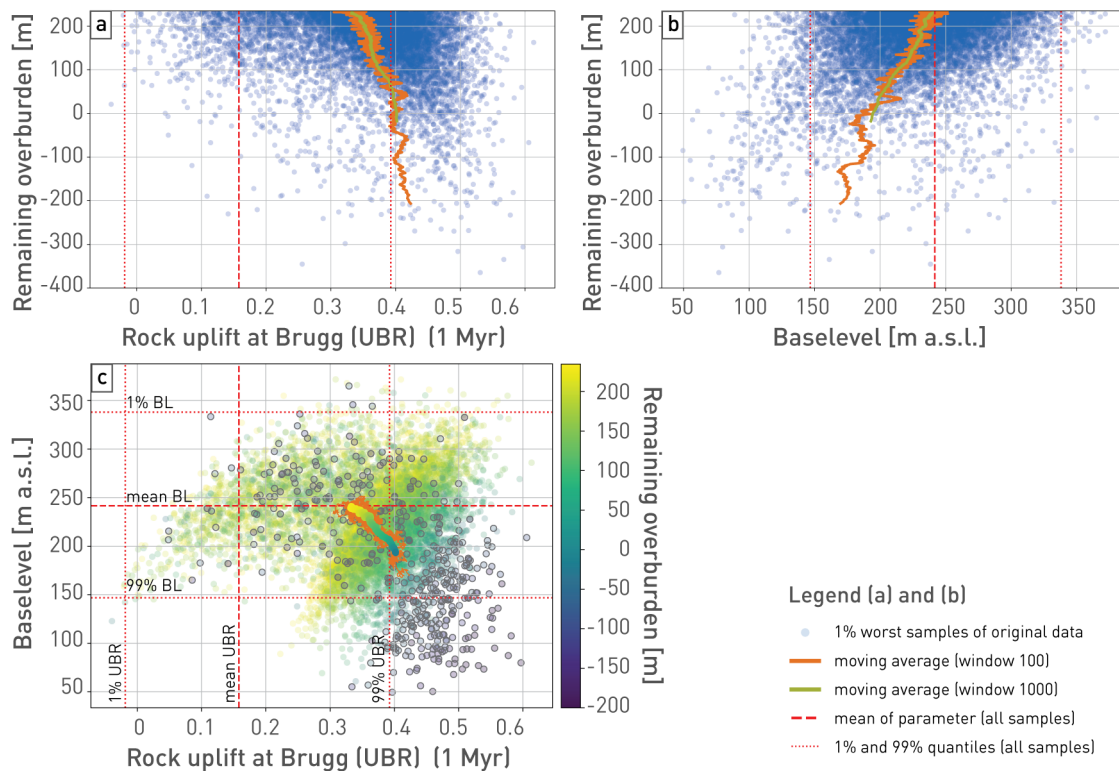


Fig. 5-4: Influence of (unreasonable?) parameter combinations

Shown are the parameters uplift in Brugg (*UBR*, a) and baselevel (*BL*, b) and their joint influence on remaining overburden thickness (c). The values of the parameters in (a) and (b) are plotted on the x-axis. On the vertical axis, the remaining overburden thickness is shown. Only the subset of samples associated with the 1% smallest remaining overburden thickness are included in the illustration (the bluish points). The orange and green lines represent estimates for the expected parameter value conditional on the value of the remaining cover for two different sampling window sizes. (c) The 1% samples with the smallest remaining overburden thickness are shown as points. The colour of the points is linked to the remaining overburden thickness above the provisional disposal area. The points with a gray edge are samples that have a cover less than zero, i.e. for these samples the repository is exposed. The orange lines in the plot show the moving average of the samples with a window size of 100. The coloured points inside the orange line represent the moving average with a window size of 1000; the colour represents the average remaining overburden thickness associated with the point. The vertical and horizontal continuous lines represent the mean of *UBR* and *BL*, and the vertical and horizontal dashed lines represent the 1% and 99% quantiles, respectively.

5.3 Conservative assumptions

Throughout the model construction, attention was paid to a realistic workflow. However, there are a few circumstances that are worth mentioning, where the approach is conservative with respect to a certain outcome.

Alternative drainage networks vs. "worst case" scenario

One prominent example of conservative assumptions is the treatment of Model B for alternative drainage scenarios and the different approaches for JO versus NL and ZNO. The present-day low relief in ZNO and in the east of NL, thus the area of the repository location, provides a good reasoning to ignore local topography and compare with the local erosion base (see Section 3.4).

However, considering the flat topography in ZNO and in the eastern part of NL, it is quite possible that the location of the main drainage will be shifted away from the disposal area (as is currently also the case in ZNO, Section 3.4.6) and a significant protective relief will develop in the future. Accordingly, the remaining overburden thickness might be further increased and the probability of excavation further reduced in NL and ZNO.

Downstream position of local erosion base calculations

For all sites, the future local erosion base is modelled with respect to a downstream position along the Rhine River (Fig. 3-6 and Fig. 3-12). The LEB represents a virtual surface that is spanned between the thalwegs of the main river channels and declines towards the WNW at present. This means that the virtual surface of the LEB is higher at the provisional disposal areas (located further to the southeast in JO and NL, and east in ZNO) than the downstream positions along the Hochrhein. For the present LEB as interpolated by Pietsch & Jordan (2014), this difference amounts to ~40 m in JO and NL, and ~10 m in ZNO. Therefore, results of remaining overburden thickness with respect to the future LEB (e.g. Fig. 3-15b, or curves "Fluvial incision" in Fig. 4-1 and Fig. 4-2) also tend to be underestimates, e.g. by a few decametres.

Given that the glacial overdeepenings in Model C are defined as depth below the LEB, the remaining overburden thickness will be similarly underestimated where future glacial erosion is involved.

For the modelling of local topography in Model B (Section 3.4.2), the main rivers are lowered according to the change in LEB simulated in Model A. This lowering step is executed instantaneously along the entire main river profiles. For the natural landscape, however, the longer the distance from the location of incision is, the longer a signal may take to arrive at the main river location nearest to the repository – a fact that is not incorporated in the modelling. Based on the present-day main river system (Fig. 3-6 and Fig. 3-24), the distances between the provisional disposal areas and the respective downstream positions may have the largest effect on JO (including a significant segment along the Aare River), followed by NL, with the smallest effect on ZNO.

Steady-state assumption in Model B

Model B uses a "snap-to-steady" algorithm, i.e. the model runs until the newly developed drainage divides are stable (Section 3.4.2), irrespective of the time required to reach such solution. It also implies that the longitudinal profiles of all tributary rivers have reached a steady state.

Equilibrium profiles calculated by the SPIM show a graded, concave upward form between a river's source and outlet (Section 3.3.1). Model B only allows a deviation from this form as a result of differently erodible stratigraphic layers, i.e. lower K values associated with a lithostratigraphic unit produce steeper channel segments, and vice versa. In contrast to the fluvial transience in natural landscapes (Section 2.4.1.4), any knickpoints in river profiles arising from transient perturbations (e.g. lowering of the main rivers) are modelled as if instantaneously migrating up the entire profile. In reality, the approximation of a steady state may only be acquired late in the period of consideration or even not at all. This model behaviour is considered conservative, as it removes any additional overburden thickness related to transient profile convexities.

Depth-depending morphology of glacial overdeepenings

The overdeepening occurrence scenarios (Section 3.5.3.3, see Fig. 3-43) together with the consideration of the effective overdeepened area (Section 3.5.3.4 and Fig. 3-45) focus on surface area of the glacial overdeepenings. The depth distribution of the overdeepening is derived along the axes of overdeepened valleys. This depth distribution is then applied to the overdeepened area

within an overdeepening scenario. Such an approach implies an extreme U-shape of the overdeepenings (with a horizontal valley floor and vertical flanks). From boreholes and seismic investigations, it is known, however, that the troughs in Northern Switzerland are often V-shaped and decrease their area towards depth (see Fig. 3-44). Because the excavation probability is targeted to the provisional disposal area and considers “the moment” when the depth of the repository is reached, the probability of “locating” the disposal area is likely overestimated.

On the other hand, the surface area of the overdeepenings that are seen today may similarly be underestimated, if they were to be associated with a higher fluvial paleo erosion base, e.g. TDS level. In this case, the overdeepenings were probably not only deeper, but also wider.

Parametric model parameter distributions vs. truncation

In the expert elicitation, the experts had to estimate lower and upper bounds. However, in the probabilistic model, except for hard (physical/mechanical) bounds (e.g. for parameters where negative values are not allowed), the fitted probability distributions are not truncated at the elicited bounds. Instead, the value of the corresponding parameter is allowed to exceed the elicited bounds, although with a relatively small uncertainty (see Section 3.2.3). In general, truncating distributions should be avoided in a probabilistic model, unless hard physical/mechanical bounds exist. In many cases it is extremely unlikely that a large value is exceeded, but not impossible. Modelling this effect with a probability distribution that has a decreasing tail is preferred compared to truncating the distribution. This states that the next larger value is less likely than the value before, whereas for truncated distributions the point of truncation exhibits a discontinuity in probability density – which can typically not be observed in nature. Nevertheless, applications exist where hard limits are used to avoid unreasonable large parameter values. One prominent example is the maximum possible magnitude in seismic hazard assessments.

The probability of exceeding the bounds was presented to, and discussed with, the experts in the RIO phase of the elicitation workshop. At the end of the workshop, the parametric distribution was accepted as RIO distribution, but with the requirement to test the influence of truncation versus fixing the elicited quantile values and bounds (see below).

Thus, the effect of truncating the tails of parametric distributions was assessed in a sensitivity study. The model parameters were truncated at the bounds and the resulting probabilistic model was compared to the “standard/main probabilistic model”.

For some model parameters, a truncation of the domain does not have a measurable impact on the probability of excavation (this includes the azimuth, as well as the erodibility scaling for EC3 and EC4). For other parameters, the truncation results in smaller excavation probabilities (including gradient, as well as K values for ECb and EC4). The parameter “ K values for EC1” is the only uncertain model parameter for which the exposure probability increases – however, the resulting increase is small. For the “reference depth of overdeepening in Molasse”, the exposure probability decreases for the siting regions NL and ZNO but does not change measurably for JO. Truncating the distributions of all model parameters at the lower and upper bounds simultaneously, the exposure probability for all siting regions decreases significantly (for NL, the decrease is more than one order of magnitude). The associated probabilities of excavation after one million years are JO: 0.00099, NL: 0.000039, and ZNO: 0.00087 (see Tab. 4-3 for comparison with the standard/main probabilistic model).

This significant decrease in excavation probabilities shows that parametric distributions indeed introduce a serious conservatism. However, estimating the plausibility of extreme realisations (i.e. of values in the vicinity of the bounds or exceeding the bounds and even the bounds themselves) is even more challenging than quantifying the uncertainty in the main body of the distribution, especially if data and observations are rare.

5.4 Plausibility based on the geological past

The results presented in Fig. 4-1 show that fluvial incision is the main geomorphic process that ultimately determines the remaining overburden thickness in the case of NL, where the Glatt, representing a main river, is located in direct vicinity of the provisional disposal area (see comparison of grey curve and thick dotted curve). The same applies to NL and ZNO in Fig. 4-2, where the effect of local topography is ignored. For ZNO in Fig. 4-1 and JO in Fig. 4-1 and Fig. 4-2, the evolution of local topography, especially considering alternative drainage networks, is also important (green curve).

The distribution of future fluvial incision that was modelled based on elicited parameters (see Section 3.3) for the one-million-year timeframe reveals median values of ~ 165 m in JO, ~ 195 m in NL and ~ 175 m in ZNO. Including 95% of the samples, future fluvial incision may reach up to approximately 300 m in JO, ~ 385 m in NL, and ~ 375 m in ZNO. Comparing these values with the Deckenschotter incision, provided in Fig. 2-5e and Appendix A.2, there is in general good agreement (see discussion in Section 3.3.5). The 95% range seems higher than the distributions of Deckenschotter incision may imply but is comparable to the present-day uplift proxies derived from geodetic data (Fig. 2-5e).

Denudation rates measured today are not directly comparable to the results of modelling the evolution of local topography. Present (catchment-wide) denudation rates which average across a few thousands of years are below 0.05 mm/yr south of the siting regions and below 0.1 mm/yr north of them (Fig. 2-20), or were estimated in JO to be below 0.1 mm/yr on the more stable parts on top of the Bözberg and below 0.2 mm/yr on the flanks (Fig. 3-20). These rates are comparable to or lower than the Quaternary incision rates and would add to approximately 200 m after one million years. The modelled reduction in overburden thickness by the combined effect of fluvial incision and evolution of local topography in JO may be in the order of up to 300 m (95%) at the position of the provisional disposal area (Fig. 3-30). This is to a considerable degree due to the potential scenarios of a new drainage developing through the Itele valley or the reactivation of the Rinikerfeld channel (Fig. 3-28 and Fig. 3-31). Such events may lead to a significant temporary increase of local incision and denudation at the provisional disposal area, compared to the present-day observations from the Wutach area (Fig. 3-20). Moreover, the cuesta landscape in JO with a layering of lithologies of variable erodibility may lead to temporarily higher denudation rates in the future. Keeping these aspects and a potentially conservative effect of the steady-state assumption (Sections 3.4.5 and 5.3) in mind, the range of rates for the evolution of local topography is considered plausible.

The effect of deep glacial erosion on remaining overburden thickness and excavation probability is small compared to fluvial incision (compare blue curves and grey curve in Fig. 4-2). This is although glacial overdeepenings can reach several hundreds of metres below the local erosion base and have shaped the landscape of Northern Switzerland significantly (Sections 2.4.2 and 3.5). The distributions of deep glacial erosion that were elicited for future key glaciations closely resemble the Quaternary equivalents (compare Fig. 2-12 with Fig. 3-40) and are thus reasonable. Nevertheless, deep glacial erosion is spatially confined and Northern Switzerland or the siting regions comprise significantly more area without or with less deep overdeepenings than the area with very deep incision (see for instance Fig. 2-11, Fig. 2-12, Fig. 2-30 and Fig. 3-44). This likelihood plays a significant role in the current effect of deep glacial erosion in the model at the respective provisional disposal areas (see for instance the tornado plots in Fig. 5-1, "Overdeepening occurrence at location – yes").

6 Conclusion and comparison of the siting regions with respect to the erosion assessment

6.1 General conclusions

The erosion assessment aimed at ensuring uncompromised barrier properties during the period under consideration. Some of these properties are "depth-dependent" and persist as long as a minimum residual cover of 200 m for efficient self-sealing is maintained above the Opalinus Clay, i.e. above the main barrier of the deep geological repository (Chapter 1). Accordingly, the remaining overburden thickness after the periods under consideration (i.e. 100'000 years for L/ILW and one million years for HLW) was identified as the main target output of the erosion assessment.

The overall knowledge about the timescales, form and rate of erosion processes, and the erosion history of Northern Switzerland have been improved considerably by detailed field and laboratory work during recent years (Chapter 2). The effects of recurring climate-driven erosion processes superposed on the low uplift of Northern Switzerland have been identified and analysed with respect to landscape evolution. Nevertheless, age determination and the limited availability of dated Quaternary geomorphic markers and sediments (e.g. Deckenschotter, infills of glacially overdeepened valleys) remain a challenge. Consequently, the age uncertainties of Quaternary landforms and deposits are accounted for in a systematic assessment of future erosion.

Due to the different distance to the Alps and different topography, ice thicknesses and the duration of ice cover at the sites varied during past glacial cycles (ZNO > NL > JO). The same pattern will most likely also be seen during future glaciations. Within the next one million years, several full glacials are expected. Future erosion processes (glacial and non-glacial) are expected to be generally comparable to processes that have repeatedly influenced the area of the siting regions during the Quaternary, but they are inferred to operate at lower rates: analyses of the fluvial topography show that the catchment of the Rhine River will grow at the expense of the Danube catchment, thus no major pulse of drainage reorganisation will cause deep fluvial incision comparable to the Pliocene-Quaternary past as the major wave of retrogressive erosion of the Rhine has already passed by the area of the three siting regions.

The assessment of future erosion was done using a robust, structured hybrid-probabilistic approach, based on available field data, rate determinations, laboratory tests and modelling, to evaluate the behaviour of the erosional system in Northern Switzerland for different future scenarios (Chapter 3). The following parameters were identified to be most important with respect to their influence of erosion on the remaining overburden in the three siting regions:

1. uplift parameters,
2. reference depth of glacial overdeepening,
3. scaling factors for glacial overdeepening in EC3 and EC4 at the NL and ZNO sites.

The representation of local topography is very relevant for JO.

A remaining overburden thickness of 200 m as required for robust and efficient self-sealing of the Opalinus Clay can be shown for the most likely scenarios in all the siting regions for the period under consideration (Chapter 4). Using the 5 – 95% range as reference, the JO site shows more limitations for such a 200 m criterion at the location of the provisional disposal area. However, suitable conditions in JO are available in the immediate surroundings of this provisional disposal area, such that JO can also reliably provide sufficient overburden thickness to secure the barrier function over the period under consideration. Nevertheless, because of the comparably small distance between the repository depth and the local erosion base, JO is dependent on the

preservation of the local topography and is thus more sensitive to changes in the fluvial network compared to the other sites (Fig. 6-1). The NL site is a well-suited location for the development of a repository because it is protected by the large emplacement depth against future erosion from all of the principal erosion processes capable of causing deep incision.

Fluvial incision, driven mainly by rock uplift, dominates the modelled overburden thickness assessment on a timescale of one million years (Chapter 5). Although deep glacial erosion can greatly increase the degree of incision, it is less likely at the sites and especially at the location of the provisional disposal areas, for which the assessment was performed. This is because of the low probability of glacial overdeepenings occurring at the location of the provisional disposal area, the more resistant rocks below the Molasse strata, and the longevity of elevated atmospheric CO₂ concentrations that are predicted to delay future foreland glaciations. Because of continuous erosional downcutting in the Alps and subsequent decrease of the accumulation area of future glaciers, foreland glaciations might be smaller even with similar climate forcing.

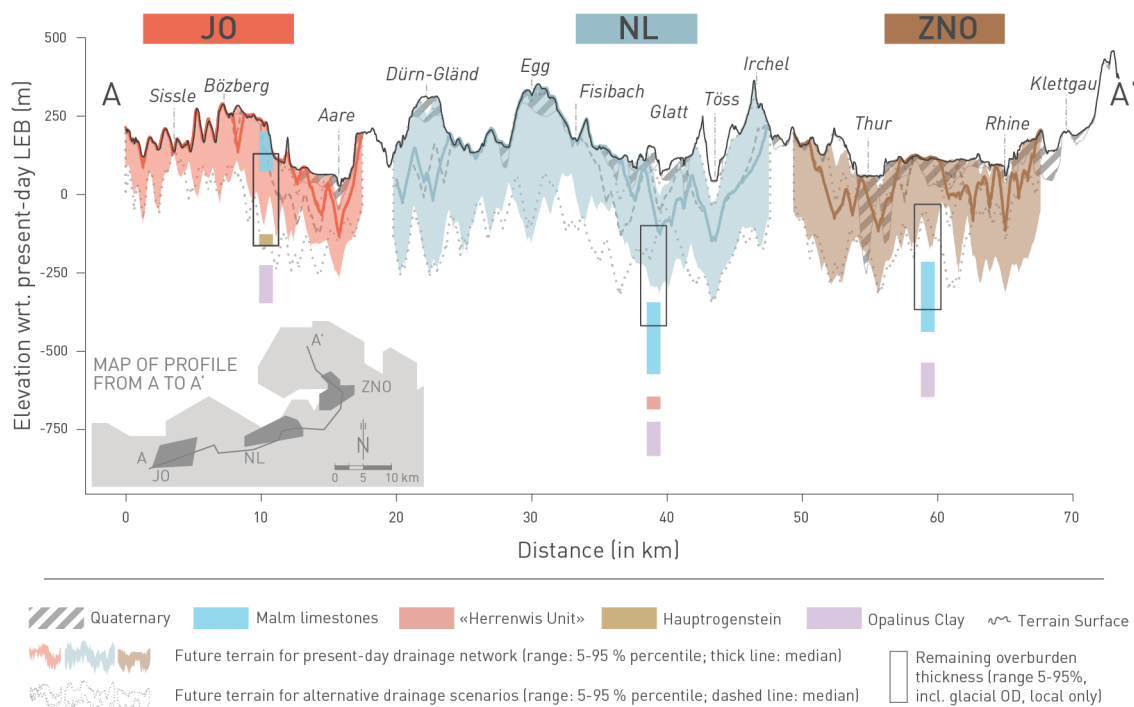


Fig. 6-1: Synthesis profiles to visualise future erosion in all the siting regions

The profile shows ranges of future evolution of the local erosion base and local topography for selected scenarios (see corresponding map representations in Fig. 3-26 for the present-day drainage network scenarios and Fig. 3-28 for the JO alternative drainage scenario Rinikerfeld). The profile location is the same as in Fig. 2-29. Note that the formation of glacial overdeepenings is only represented in the boxes showing the remaining overburden thickness at the locations of the provisional disposal areas, which are comparable to the shaded boxes in Fig. 4-2b. Vertical exaggeration is ~ 20-fold. For the provisional disposal area at JO (lateral position corresponds to the location of lithological column in the plot), the remaining overburden thickness with respect to the coloured and dashed bands of future terrain shown here contributes to the green curve ("Local Topography"; output Model B) in Fig. 4-2.

Excavation of the repository by erosion within the next one million years can be considered extremely unlikely for all the provisional disposal areas but is nearly an order of magnitude lower in NL (Chapter 4). These erosional excavation probabilities were calculated to show the robustness of the system. They are represented as extreme endmember scenarios in the tails of the overburden distribution. However, the most extreme overburden thickness reduction (leading to excavation) after one million years includes combinations of parameters that are highly unlikely to occur simultaneously (Chapter 5). The final numbers are likely overestimated and have to be treated with caution.

6.2 Storyboard for the anticipated long-term erosion of Nördlich Lägern

Within this storyboard, the anticipated future erosional development of the Nördlich Lägern site is envisioned. The story evolves around the 5th to 95th percentiles of certain parameters and the 5th to 95th percentile range of the remaining overburden thickness output as calculated for the criterion 3.3 case (Section 4.1.2). The data are shown for the provisional disposal area of the HLW repository.

The full model requires too many parameter estimates for a continuous future evolution. Accordingly, this approach was focused on the two times under consideration, i.e. 100'000 yr for L/ILW and one million years for HLW, respectively. As highlighted earlier and shown in the different curves in Fig. 4-1 and Fig. 4-2, fluvial incision is most important for the erosion at the provisional disposal area. The SPIM modelling (see Section 3.3), however, has the advantage of low computational costs. Thus, three additional fluvial model time steps (350'000 yr, 700'000 yr, and 880'000 yr) are added arbitrarily as support points for the storyboard. We use the elicitations conducted for the 100'000-year time period to model the 350'000-yr time, and the elicitations conducted for the one-million-year time period for the 700'000- and 880'000-yr time steps, respectively (Tab. 6-1).

Tab. 6-1: Fluvial incision including additional incision steps

Columns represent (1) the modelled time using SPIM, (2) the reference time for which elicitations were performed, (3) the quantile or respective metric (e.g. P05 represents the 5th quantile or the lowest 5% of the incision samples), and (4) the incision depth derived for Nördlich Lägern at the Aare – Rhine confluence. These incision depths form the basis for the boxes visualising future fluvial incision in Fig. 6-2.

Modelled time [yr]	Reference time [yr]	Quantile/Metric	Depth of incision
100'000	100'000	Mode	14 m
100'000	100'000	Mean	26 m
100'000	100'000	P05	8 m
100'000	100'000	P25	14 m
100'000	100'000	P50	22 m
100'000	100'000	P75	33 m
100'000	100'000	P95	54 m
350'000	100'000	Mode	84 m
350'000	100'000	Mean	98 m
350'000	100'000	P05	30 m
350'000	100'000	P25	62 m
350'000	100'000	P50	90 m
350'000	100'000	P75	124 m
350'000	100'000	P95	193 m
700'000	1'000'000	Mode	131 m
700'000	1'000'000	Mean	154 m
700'000	1'000'000	P05	54 m
700'000	1'000'000	P25	108 m
700'000	1'000'000	P50	146 m
700'000	1'000'000	P75	192 m
700'000	1'000'000	P95	278 m
880'000	1'000'000	Mode	162 m
880'000	1'000'000	Mean	191 m
880'000	1'000'000	P05	74 m
880'000	1'000'000	P25	131 m
880'000	1'000'000	P50	180 m
880'000	1'000'000	P75	238 m
880'000	1'000'000	P95	347 m
1'000'000	1'000'000	Mode	171 m
1'000'000	1'000'000	Mean	211 m
1'000'000	1'000'000	P05	81 m
1'000'000	1'000'000	P25	145 m
1'000'000	1'000'000	P50	196 m
1'000'000	1'000'000	P75	263 m
1'000'000	1'000'000	P95	386 m

Long-term rock uplift, fluvial incision and remaining overburden thickness

The Rhine will grow at the expense of the Danube, with headward erosion rather than through major river capture events as seen in the Pliocene and Quaternary periods. Instead, the evolution of the river profile is largely controlled by rock uplift and sediment input. The bedrock erodibility is only of secondary importance for vertical fluvial incision but might have an influence on lateral erosion.

The spatial pattern of rock uplift in Northern Switzerland over the next million years will continue to be dominated by Alpine uplift, regardless of whether it is a matter of tectonic uplift or isostatic compensatory movements. In simplified terms, the main uplift gradient can be assumed to cause increasingly higher rates of uplift from the external foreland basin towards the Alps along a direction that is approximately perpendicular to the Alpine chain. The long-term rates are low. Total uplift of ~ 170 m (between ~ 25 m and a maximum of ~ 360 m [5 – 95%]) over one million years are assumed for the NL siting region. Short phases of higher rates might be associated with, and postdate, deglaciation of extensive foreland glaciations.

Significantly greater incision can only be achieved by additionally lowering the baselevel. Two intermediate baselevels will be important in the future: As long as the Rhenish Massif continues to be uplifted, this location (approx. near Bingen) decouples the Upper Rhine (and thus the Hochrhein) from the sea. Furthermore, the sedimentary basin of the Upper Rhine south of Bingen is controlled by tectonic subsidence and sediment input at the transition to the Hochrhein near the city of Basel. Today, there is little disturbance of the river profile at this point; the subsidence is roughly compensated by sediment accumulation. Potential elevation gain by short-term over-supply of sediment is compensated over the long term by vertical incision into these sediments. Averaged over the next million years, however, it is considered possible that the baselevel at Basel might fluctuate by up to ± 70 m.

Under these boundary conditions, a vertical fluvial incision of ~ 165 m is most likely (between 80 and 380 m [5 – 95%]) at the confluence of the Aare and Rhine Rivers, i.e. near NL. This is in good agreement with Quaternary incision on a one-million-year timescale around the area of NL. Given the provided incision estimates, a remaining overburden thickness of ~ 550 m (350 – 670 m [5 – 95%]) can be expected. It is therefore possible that the Malm limestones of the Villigen Formation are already being incised fluvially in the area of the siting region or that the future exfiltration area will be located progressively closer to the siting region and to the position of the disposal area. Accordingly, the exfiltration path in the aquifers is shortened compared to today. Importantly, the remaining overburden thickness is sufficient to avoid any decompaction effects or to affect the self-sealing capacity of the host rock.

Short-term modulation of fluvial incision by glacial/interglacial cycles

The long-term incision will be modulated by phases of sediment accumulation with no net vertical incision. These sediment dynamics are largely controlled by glacial/interglacial cycles. During the interglacial warm period, a relatively high precipitation rate is expected and consequently a relatively high water discharge. The sediment flux on the other hand will probably be limited by a weak coupling between the glacially conditioned sediment sources and the active transport pathways as can be observed today. The lowest sediment flux is therefore expected during the interglacials. During the renewed Alpine glaciation, the reworking of periglacial deposits by the advancing glaciers and an increase in the area subject to subglacial abrasion should lead to a peak in sediment flux. During the main glacial phase, sediment flux is assumed to be larger than the interglacial rate because of increased glacial activity, sediment generation in adjacent permafrost terrains and the effect of greater climate variability. For the termination of glacial conditions, a peak is inferred in water discharge due to the warming-induced increase in atmospheric moisture and a pulse of meltwater from the decaying ice. The greater water availability induces faster

glacial sliding and hence an increase in glacial abrasion. In combination with the paraglacial adjustment of the landscape, this should result in a maximum contribution of sediment to the fluvial system. After glacial retreat, however, in many places the sediment is trapped in peri-Alpine lakes and overdeepened basins for up to several tens of thousands of years, thus reaching into the interglacial period.

The expected glacial scenario and deep glacial erosion

According to different anthropogenic CO₂ scenarios, 3 – 5 large glaciations (LGM-size or larger) are expected that might reach the Alpine Foreland in the next one million years. The next glacial inception will be significantly delayed, and the next large glaciation will only occur after the next ~ 200'000 or even 350'000 years (Fig. 3-46). During extensive future foreland glaciations, deep glacial erosion is expected in the eastern part of the NL siting region, where present-day relief is low. This part could be repeatedly glaciated if the future glacier is of similar or larger extent and erosive efficiency to the penultimate glaciation (Beringen).

Total cumulative depth in such a case is expected to reach between 30 and 360 m (most likely ~70 m) below the local erosion base. Because the baselevel has already lowered, deep glacial erosion can reach the Malm limestone. It is expected that the Malm limestone is on average only half as erodible as Molasse-type rocks. It is thus highly unlikely that the glacial overdeepenings will reach the host rock.

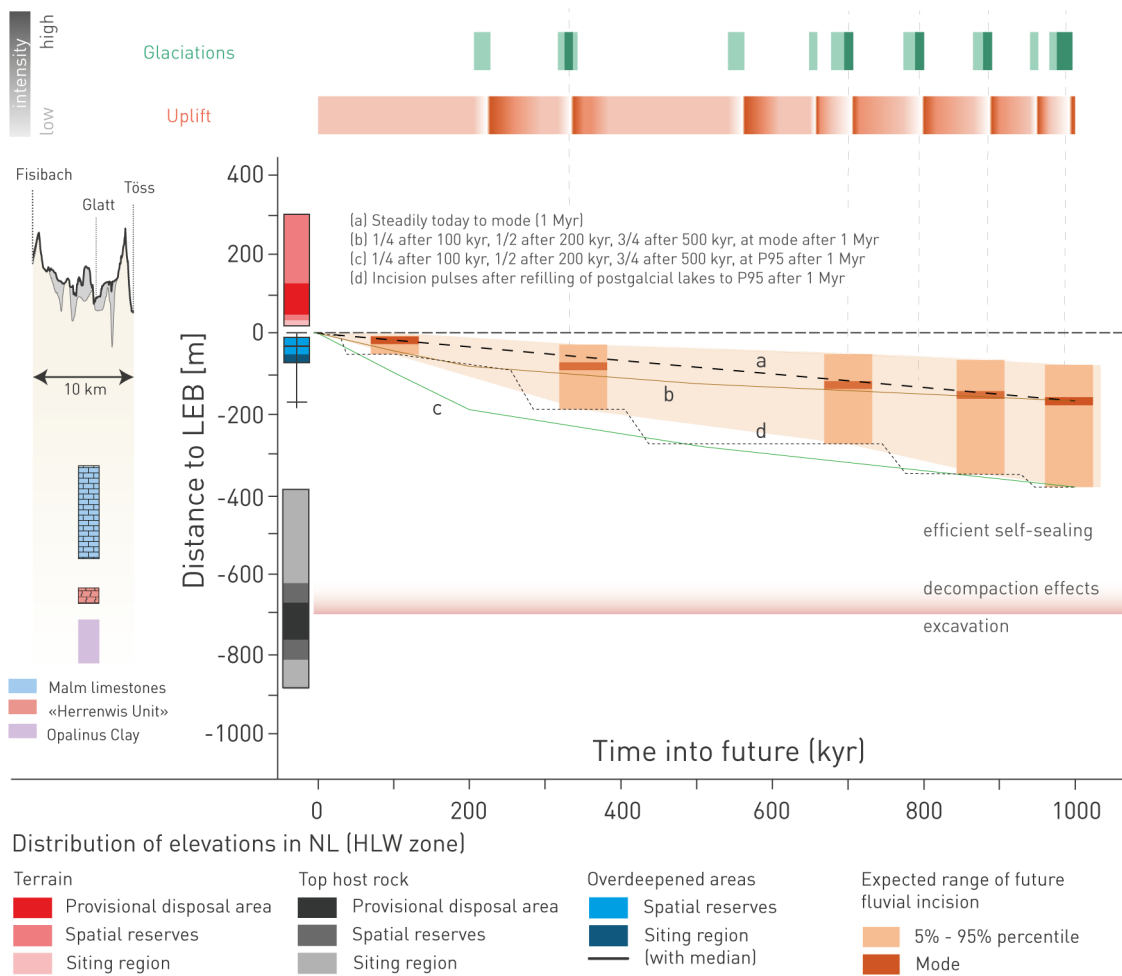


Fig. 6-2: Schematic diagram summarising the long-term geological evolution of the NL site

The diagram summarises the elements that describe the expected long-term geological evolution for the next one million years for the Nördlich Lägern site. The panel on the left-hand side shows the present-day physiogeographic conditions extracted from the cross-section in Fig. 2-29 between Fisibach and Töss and as a simplified depth profile with boxplots highlighting the distributions of host rock, local topography, and glacial overdeepenings within the siting region, and at the location of the provisional disposal area. The local erosion base (LEB) constitutes the reference for the depicted present-day elevations and the future evolution. As a result of long-term rock uplift, modified by transient climate effects, the vertical distance between the repository level and the local erosion base will be lowered by fluvial incision (orange boxes in panel on right-hand side). Potential incision pathways are possible during the one million years (e.g. a or b), others are less likely but shown nevertheless because they provide potential calculation cases to show the robustness of the site safety (e.g. c and d). Phases of glacial erosion with potential overdeepening are also highlighted in this upper panel. Such phases of glacial overdeepening are coupled to large future foreland glaciations, as depicted by the dark-green bars in the panel "Glaciations". Glacial conditions are expected to occur in semi-regular intervals. The timings delineated here correspond to the 500 PgC scenario of future climate models (Fig. 3-46). The panel "Uplift" also highlights a certain synchronicity with future glaciations, e.g. pulses of increased rock uplift after deglaciation (glacial isostatic adjustment). In general, however, the increased rock uplift is expected to balance out over the long time period.

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App. A Additional information (Quaternary landscape evolution)

A.1 Documentation of sampling locations for age determinations

Tab. A-1: Documentation of sampling locations for age determination

Morpho- strati- graphy	Acronym Sampling Location	Name Sampling Location	B: Borehole O: Outcrop	CH1903+ / LV95 (Bessel 1841)		Dezimalgrad		Age (Ma) or MIS	Dating method	Reference
				X_LV95	Y_LV95	X_WGS	Y_WGS			
HDS	ROG	Roggenfeld, Mettauertal	O	2654680	1269605	8.16535908	47.57487174	Normal Magnetisation	Magnetostrat.	Graf (1993), Scheidt et al. (2023)
HDS	TRO	Tromsberg, Obersiggenthal	O	2663770	1261525	8.28500523	47.50138651	1.28 ± 0.13	TCN	Grischott et al. (2020), Knudsen et al. (2022), P-PINI recalculated
HDS	EIC	Eichbrunnen, Freienwil	O	2666050	1261450	8.31525277	47.50048693	(> 1.8 Ma)	Biostrat., Lithostrat.	Frei (1912), Graf (1993), Thew (2024)
HDS	SIG	Siglistorf	O	2671500	1266365	8.38836895	47.54411827	0.99 ± 0.25	TCN	Akçar et al. (2017), P-PINI recalculated
HDS	EGG	Egghalden, Schleinikon	O	2672230	1262630	8.39745859	47.51044906	> 0.77	Magnetostrat.	Graf (1993), Scheidt et al. (2023)
HDS	FEU	Feusi, Oberweningen (lower level)	O	2672950	1262450	8.40698597	47.50875078	1.09 ± 0.27	TCN	Nagra, courtesy of E. Bros
HDS	FEU	Feusi, Oberweningen (upper level)	O	2672950	1262450	8.40698597	47.50875078	0.93 ± 0.09	TCN	Grischott et al. (2020), Knudsen et al. (2022), P-PINI recalculated
HDS	BOP	Cholholz, Boppelsen	O	2673665	1258893	8.4158883	47.47668178	Normal Magnetisation	Magnetostrat.	Scheidt et al. (2023)
HDS	UEK	Uetliberg Kulm, Zürich	O	2679605	1244805	8.49217212	47.34929198	> 0.77	Magnetostrat.	Graf (2019), Scheidt et al. (2023)
HDS	BUE	Albihorn-Bürglen-2, Hausen am Albis	O	2683325	1234575	8.53949955	47.2568282	> 1.8	Biostrat., AAR	Penkman et al. (2024), Thew (2024)

Morpho- strati- graphy	Acronym Sampling Location	Name Sampling Location	B: Borehole O: Outcrop	CH1903+ / LV95 (Bessel 1841)		Dezimalgrad		Age (Ma) or MIS	Dating method	Reference
				X_LV95	Y_LV95	X_WGS	Y_WGS			
HDS	HOC	Irchel Hochwacht, Berg am Irchel	O	2686077	1268269	8.58234125	47.55949214	2.1 – 1.8	Biostrat., AAR, Magnetostat., Lithostat.	Graf (1993), Scheidt (2023), Penkman et al. (2024), Thew et al. (2024)
HDS	EBN	Irchel Ebni, Berg am Irchel	O	2687008	1267577	8.59457438	47.55314647	> 2.1 – 1.8	AAR, Magnetostat., Lithostat.	Graf (1993), Scheidt et al. (2023), Penkman et al. (2024), Thew (2024), Thew et al. (2024)
HDS	SART	Irchel Schartenflue, Freienstein-Teufen	O	2687093	1267050	8.5956	47.5484	1.24 ± 0.24	TCN	Dieleman et al. (2022), P-PINI recalculated
HDS	AMS	Irchel Amselboden, Freienstein-Teufen	O	2687258	1266111	8.59760763	47.53992991	2.1 – 1.8	Biostrat., AAR, Lithostat.	Graf (1993), Penkman et al. (2024), Thew et al. (2024)
HDS	SGI	Irchel Steig, Buch am Irchel	O	2688501	1265633	8.61402087	47.53546393	2.1 – 1.8	TCN, Magnetostat.	Claude et al. (2019), P-PINI recalculated, Scheidt et al. (2023)
HDS	HAS	Irchel Hasli, Dättlikon	O	2688944	1265570	8.61989003	47.53483871	2.1 – 1.8	Biostrat., AAR, Magnetostat., Lithostat.	Bolliger et al. (1996), Cuenca-Bescos (2015), Scheidt (2023), Penkman et al. (2024), Thew et al. (2024)
HDS	HAS	Irchel Hasli, Dättlikon	O	2688953	1265544	8.6200	47.5346	1.14 ± 0.15	TCN	Dieleman et al. (2022), P-PINI recalculated
HDS	CHRO	Chroobach, Rielasingen- Worblingen (DE)	O	2705873	1284637	8.8492	47.7038	1.23 ± 0.07	TCN	Dieleman et al. (2022), P-PINI recalculated
TDS	GAE	Ängi, Leuggern- Mandach	O	2656610	1267925	8.19079473	47.55959953	Normal Magnetisation	Magnetostat.	Scheidt et al. (2023)
TDS	BRB	Bruggerberg, Brugg	O	2659040	1261075	8.22217139	47.49778115	Normal Magnetisation	Magnetostat.	Scheidt et al. (2023)
TDS	ISS	Iberig, Würenlingen	O	2661070	1263700	8.24947048	47.5212034	0.69 ± 0.14	TCN, Magnetostat.	Grischott et al. (2020), Knudsen et al. (2022), P-PINI recalculated, Scheidt et al. (2023)
TDS	SCH	Schwändihalden, Weiach	O	2674060	1267250	8.42251775	47.55179349	0.91 ± 0.23	TCN	Nagra, courtesy of E. Bros
TDS	CHO	Cholfirst, Chirchhölzli, Laufen- Uhwiesen	O	2690270	1280880	8.64061256	47.67233301	0.76 ± 0.12	TCN, Magnetostat.	Nagra, courtesy of E. Bros, Scheidt et al. (2023)

Morpho- strati- graphy	Acronym Sampling Location	Name Sampling Location	B: Borehole O: Outcrop	CH1903+ / LV95 (Bessel 1841)		Dezimalgrad		Age (Ma) or MIS	Dating method	Reference
				X_LV95	Y_LV95	X_WGS	Y_WGS			
TDS	HUN	Hungerbol, Öhningen (DE)	O	2708100	1282505	8.87837641	47.6842456	1.8 – 1.0	Biostrat., AAR, Lithostrat., Magnetostat.	Graf (2009a), Thew (2024), Penkman et al. (2024), Scheidt et al. (2023)
TDS	HUN	Hungerbol, Öhningen (DE)	O	2708100	1282505	8.87837641	47.6842456	1.13 ± 0.05	TCN	Nagra, courtesy of E. Bros
HT	HABS	Götschel, Lupfig, near Schinznach-Bad	O	2655472	1255851	8.17416701	47.45111101	≥ MIS 6	Lumi., Lithostrat.	Graf (2009b), Gegg (2021), Mueller et al. (2024)
HT	QRIN	QBO Riniken	B	2656717	1261523	8.19140066	47.50201466	MIS 6	Lumi., Lithostrat.	Mueller et al. (2020, 2024.), Gegg et al. (2023)
HT	NIE	Niederweningen	B	2670945	1261960	8.380295	47.50456	Eemian	Biostrat., Lumi., Lithostrat.	Anselmetti et al. (2010), Dehnert et al. (2012), Preusser et al. (2023), Thew (2024)
HT	HAB	Hablützel, Wilchingen-Hallau	O	2676546	1281538	8.458	47.680	MIS 6	Lumi., Lithostrat.	Preusser & Graf (2002), Lowick et al. (2015)
HT	QHST	QBO Hochfelden- Strassberg	B	2680479	1264843	8.5073493	47.52939236	> 0.204 ± 0.028	Lumi., Lithostrat.	Mueller et al. (2024), Buechi et al. (2024)
HT	BER	Beringen	O	2685915	1282780	8.583	47.690	≥ MIS 6	Lumi., Lithostrat.	Graf (2009b), Lowick et al. (2015)
HT	DAR	Dachsen	O	2688514	1279704	8.617	47.662	MIS 6	Lumi., Lithostrat.	Preusser & Graf (2002), Lowick et al. (2015)
OD	QUST	QBO Untersiggenthal	B	2659921	1262741	8.23408683	47.5126845	≥ MIS 6	Lumi., Lithostrat.	Gegg (2021), Mueller et al. (2024)
OD	QGBR	QBO Gebenstorf- Brüel	B	2660131	1259468	8.23643284	47.48322981	≥ MIS 6	Lumi., Lithostrat.	Gegg (2021), Mueller et al. (2024)
OD	NIE	Niederweningen	B	2670945	1261960	8.380295	47.50456	MIS6	Lumi., Lithostrat.	Anselmetti et al. (2010), Dehnert et al. (2012), Preusser et al. (2023), Thew (2024)
OD	QHST	QBO Hochfelden- Strassberg	B	2680479	1264843	8.5073493	47.52939236	0.6 ± 0.12	⁴ He/U-Th	Tomonaga et al. (2024)
OD	B1/12	Bülach-Hochfelden	B	2682368	1264901	8.53244	47.52968	MIS 6	Lumi., Lithostrat.	Buechi et al. (2017, 2018)

Morpho- strati- graphy	Acronym Sampling Location	Name Sampling Location	B: Borehole O: Outcrop	CH1903+ / LV95 (Bessel 1841)		Dezimalgrad		Age (Ma) or MIS	Dating method	Reference
				X_LV95	Y_LV95	X_WGS	Y_WGS			
OD	B08-7	Hardwald, Bassersdorf	B	2688536	1254499	8.61227	47.43533	> MIS 6	Lumi., Lithostrat.	Buechi et al. (2017, 2018)
OD	QTRU	QBO Trüllikon- Rudolfingen	B	2691937	1277353	8.66207372	47.64038546	≥ MIS 6	Lumi., Lithostrat.	Buechi et al. (2024)
OD	QANI	QBO Andelfingen- Niederfeld	B	2692453	1272633	8.66795524	47.5978688	≥ MIS 6	Lumi., Lithostrat.	Buechi et al. (2024)
OD	QKLA	QBO Kleinandelfingen- Laubhau	B	2692654	1274264	8.67096806	47.61250711	> MIS 6	Lumi., Lithostrat.	Buechi et al. (2024)
NT	BIR	Joriacher, Birr	O	2658678	1253597	8.21638901	47.43055601	0.026 - 0.022	Lumi., TCN, Lithostrat.	Graf (2009b), Gaar et al. (2019), Kamleitner et al. (2023), Reber et al. (2014)
NT	MÜB	Mülligen	O	2660889	1256677	8.24611101	47.45805601	MIS 4	Lumi., Lithostrat.	Gaar et al. (2019)
NT	HÜ	Hüntwangen, Rafzerfeld	O	2680600	1272000	8.51025377	47.5937394	0.027 – 0.022	¹⁴ C-dating	Preusser et al. (2007), Kamleitner et al. (2023)

A.2 Incision rates in Northern Switzerland along selected valleys

Incision rates were calculated along selected valleys in Northern Switzerland (Lower Aare Valley, Limmat Valley, Wehntal Valley, Töss Valley and Hochrhein Valley at two different locations [I and II], see Fig. A-1) with focus on bedrock incision. This calculation is based on simplified assumptions: firstly, the elevation of base of the Deckenschotter deposits (i.e. the bedrock elevation) is used, to address the net lowering of the local baselevel over time, independent of intermediate aggradation phases. Secondly, it is assumed that the age range reflects the deposition at a late stage, i.e. shortly before the subsequent incision. Combining an age derived from sediment deposits accumulated up to several tens of metres with the elevation of the bedrock marker below these sediments requires the third assumption: a "quasi-instantaneous" incision through the unconsolidated Quaternary deposits. In addition, instead of focusing on the lowest point of a certain Base Deckenschotter, their height uncertainty is incorporated (Hammer et al. 2019). To do so, only Deckenschotter occurrences for which this uncertainty was assigned either by Hammer et al. (2019) or during the update of the bedrock model (Section 2.4.1.2) were used. The data were collected with the outlines of swath profiles, respectively (Fig. A-1) and projected on the swath profile centre lines (long axis, see Fig. A-2, Fig. A-4, Fig. A-6, Fig. A-8, Fig. A-10, and Fig. A-12). Incision rates were calculated in a Monte Carlo sampling from the height difference between Base Deckenschotter and LEB (normal distribution) and the Deckenschotter ages (uniform distributed), assuming steady incision since HDS or TDS deposition, respectively. Where local age information was available, this was applied in addition to a full age range. The incision rates and corresponding age ranges are listed in Tab. A-2, based on chronology provided in Section 2.4.1.3 and summarised in Tab. A-1.

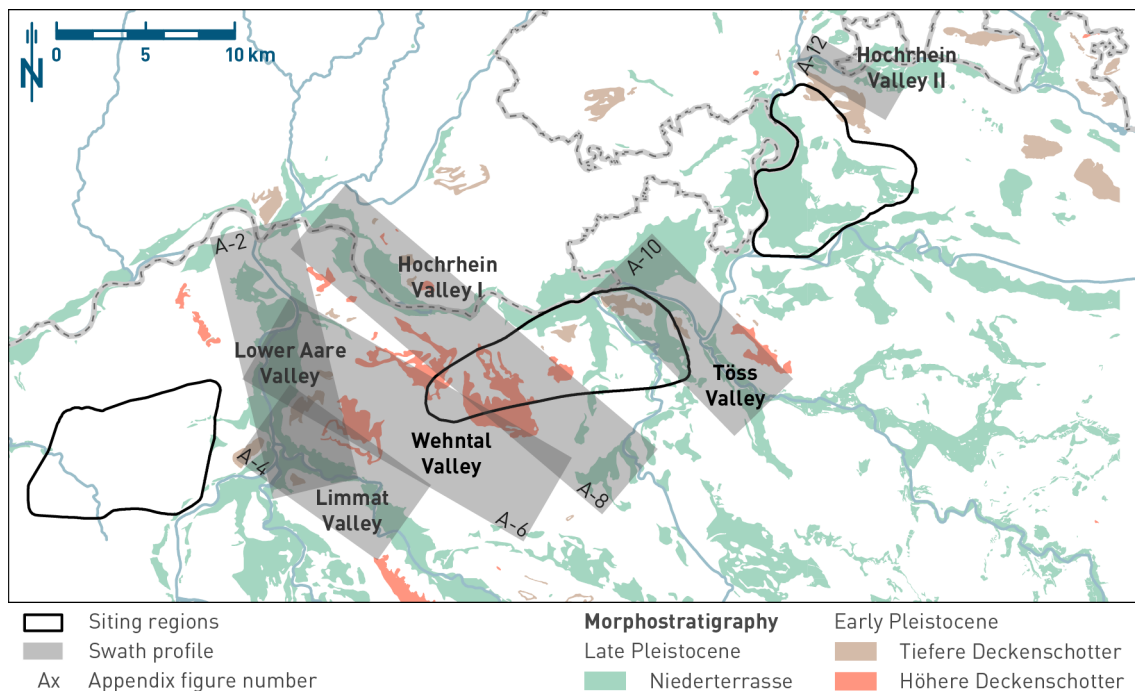


Fig. A-1: Overview figure showing the database for incision calculations

Map shows the locations of the swath profiles used for data collection of Base Deckenschotter and Niederterrasse surface. Please note that although the swath outlines appear as rectangles the data were only collected at the Swiss side of the border. Note that some swath profiles display partly the same morphostratigraphic marker (e.g. Fig. A-2 and Fig. A-4).

Tab. A-2: Incision rates along selected valleys in Northern Switzerland

Full range HDS: -2.1 Ma (Biostratigraphy) to -1.11 ± 0.37 Ma (Feusi base); Full range Dieleman et al.: -2.63 - 1.06 Ma (Irchel Hochwacht) to -0.99 ± 0.25 Ma (Siglistorf); Full range TDS: -1.18 - 0.09 Ma (Hungerbol) to -0.68 ± 0.16 Ma (Iberig); see Section 2.4.1.3. Please note that the values presented here deviate slightly from those presented in Section 2.4.1.3 and Tab. A-1 due to recently updated P-PINI recalculation. This difference is not expected to change the rates significantly. The elevations used for incision rate calculations are shown in the respective figures below. For incision rates, see the following figures (Fig. A-3, Fig. A-5, Fig. A-7, Fig. A-9, Fig. A-11, and Fig. A-13); B: Bowald; D-G: Dürn-Glând.

Valley	Key marker	Name of date	Age range [Ma]	Mean incision rate
Lower Aare Valley	HDS	Tromsberg	-1.32 ± 0.13	0.21 mm/yr
Lower Aare Valley	HDS	Full range	-2.1 to -0.74	0.19 mm/yr
Lower Aare Valley	TDS	Iberig	-0.68 ± 0.16	0.3 mm/yr
Lower Aare Valley	TDS	Full range	-1.27 to -0.52	0.2 mm/yr
Limmat Valley	HDS	Tromsberg	-1.32 ± 0.13	0.2 mm/yr
Limmat Valley	HDS	Full range	-2.1 to -0.74	0.15 mm/yr
Limmat Valley	TDS	Iberig	-0.68 ± 0.16	0.32 mm/yr
Limmat Valley	TDS	Full range	-1.27 to -0.52	0.22 mm/yr
Wehntal Valley	HDS (B/D-G)	Feusi	-1.11 ± 0.37	0.23 mm/yr
Wehntal Valley	HDS (B/D-G)	Full range	-2.1 to -0.74	0.17 mm/yr
Wehntal Valley	HDS (Egg)	Feusi	-1.11 ± 0.37	0.25 mm/yr
Wehntal Valley	HDS (Egg)	Full range	-2.1 to -0.74	0.19 mm/yr
Töss Valley	HDS	Biostratigraphy	-2.1 to -1.8	0.17 mm/yr
Töss Valley	HDS	Full range Dieleman et al.	-3.69 to -0.74	0.14 mm/yr
Töss Valley	HDS	Full range	-2.1 to -0.74	0.24 mm/yr
Töss Valley	TDS	Full range	-1.27 to -0.52	0.28 mm/yr
Hochrhein Valley I	HDS	Feusi	-1.11 ± 0.37	0.25 mm/yr
Hochrhein Valley I	HDS	Full range	-2.1 to -0.74	0.19 mm/yr
Hochrhein Valley II	TDS	Cholfirst	-0.83 ± 0.15	0.28 mm/yr
Hochrhein Valley II	TDS	Full range	-1.27 to -0.52	0.25 mm/yr
Northern Switzerland	LEB	Minimum age HT	-0.204 ± 0.028	Not applicable

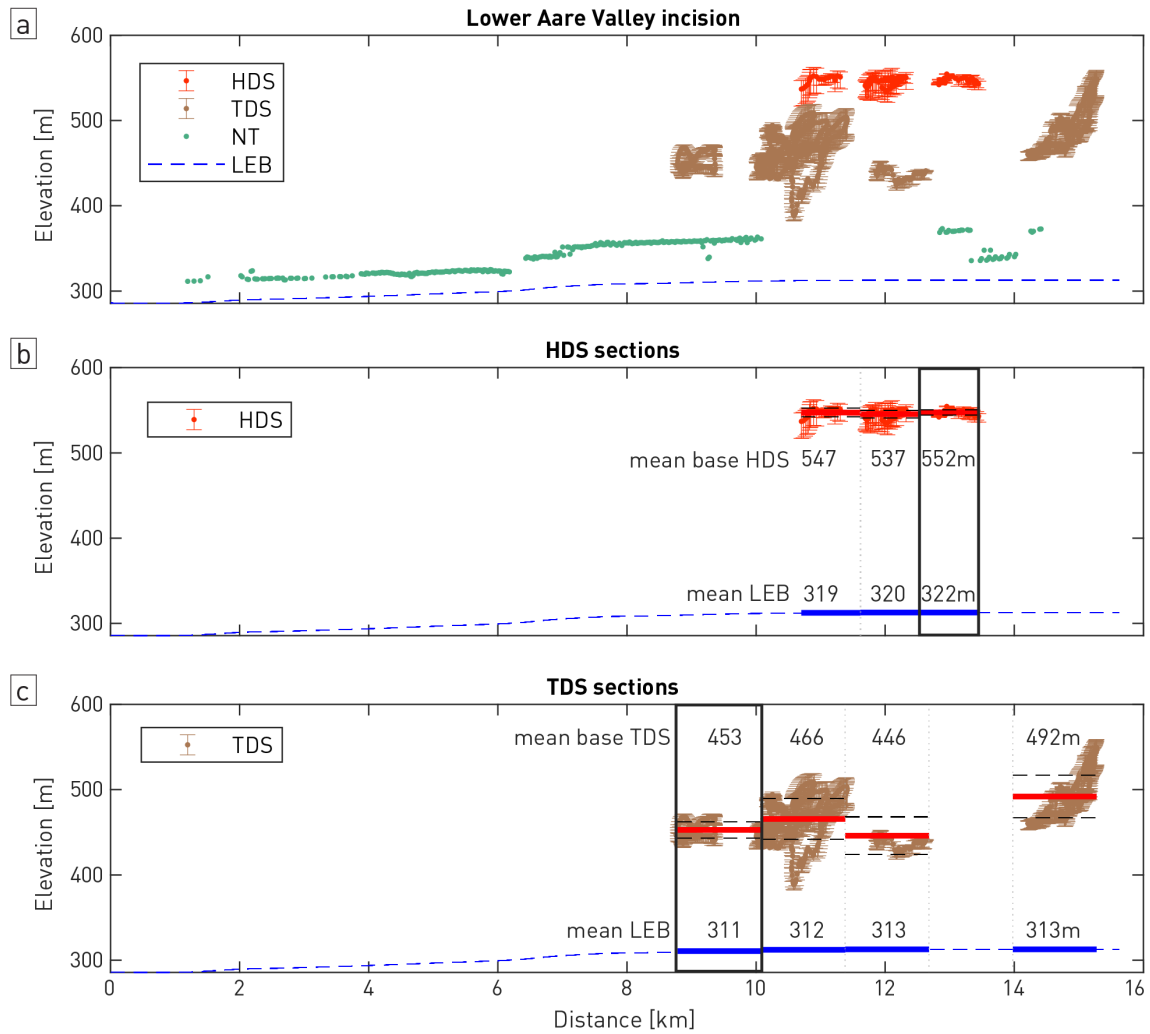


Fig. A-2: Distribution of marker horizons and section-wise statistics for incision calculation along the Lower Aare Valley

(a) A swath profile aligned along the Lower Aare Valley captured the data (see rectangle in Fig. A-1 for location and colour-code of the morphostratigraphic unit), which were projected along a profile in the centre of the swath along its long axis. Displayed are Base HDS and TDS with their uncertainty marks (after Hammer et al. 2019), the Niederterrasse surfaces, and the local erosion base (LEB). The latter corresponds to the lowest fluvial incision level. (b) Similar section with focus on Base HDS. For three sections of rather homogeneous marker elevation, the mean (red line with numbers shown) and standard deviation (black dashed line) were calculated. Mean elevation of the LEB is shown for the same sections (c) Similar section with focus on Base TDS. Four sections were chosen for statistic calculations. For both, HDS and TDS, the data highlighted by the black rectangle were used for rate calculations (sections with lower uncertainty were preferred) between elevations of Base HDS or TDS and LEB, respectively shown in Tab. A-2 and Fig. A-3.

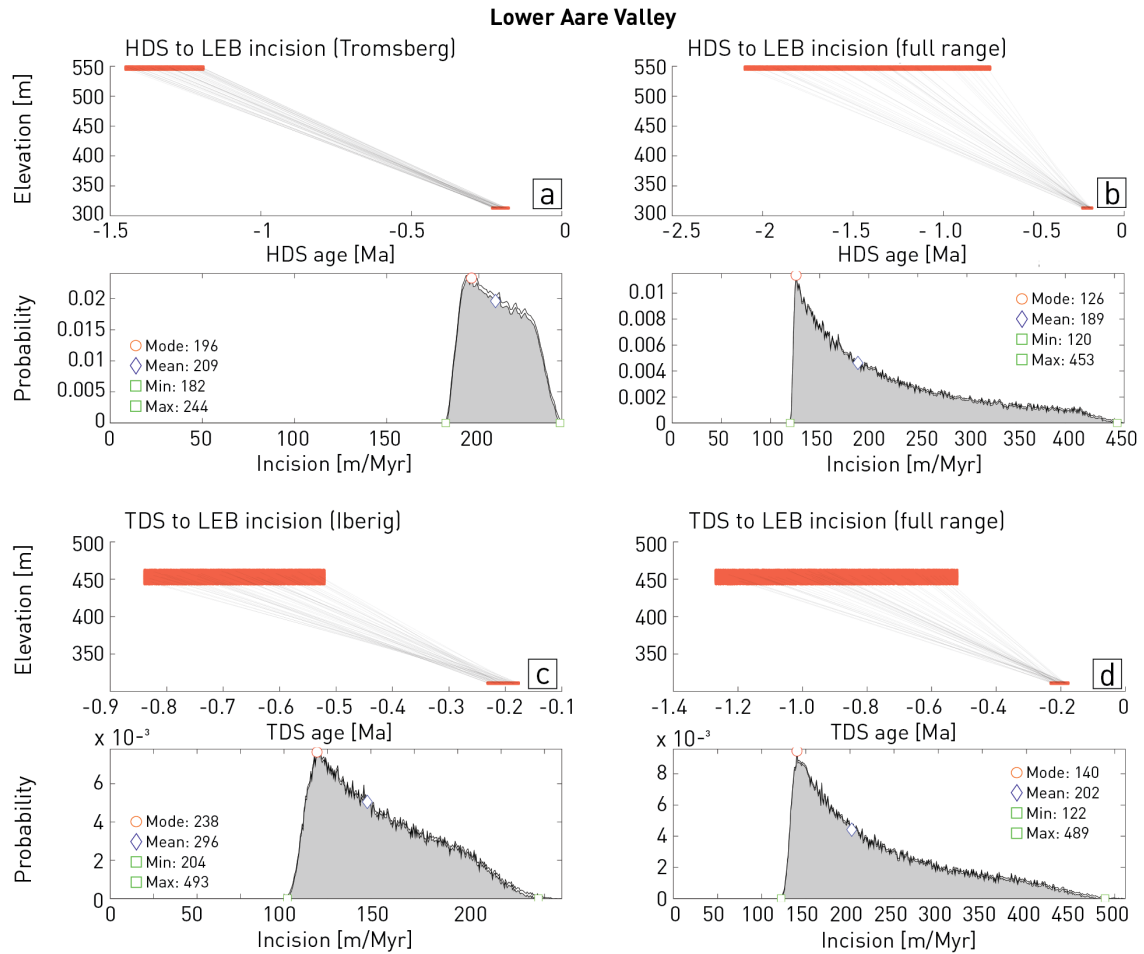


Fig. A-3: Incision rates calculated along the Lower Aare Valley

Plots show Monte Carlo-derived incision rates based on the Base Deckenschotter and LEB elevations along the Lower Aare Valley and HDS and TDS ages (see Tab. A-2 for ages, Fig. A-1 for location of data swath profile and Fig. A-2 for data). Always two plots on top of each other belong together: the upper one illustrating the sampling from the age-elevation pairs (boxes) and the lower one showing the probability density function (2-sigma range) of the associated incision rate. Incision rates were calculated from the height difference between Base Deckenschotter and LEB and the Deckenschotter ages, assuming steady incision since HDS or TDS deposition, respectively. Samples were taken from a uniform age distribution and normal distribution of elevation. (a) Sampling using height estimates of Base HDS measurements were paired with the locally-derived Tromsberg age and the corresponding incision rate (with respect to Fig. A-2b). (b) Sampling using the same heights, but the full range of HDS age estimate. (c) Sampling using height estimates of Base TDS measurements were paired with the locally derived Iberig age and the corresponding incision rate (b) Sampling using the same heights, but the full range of TDS age estimate.

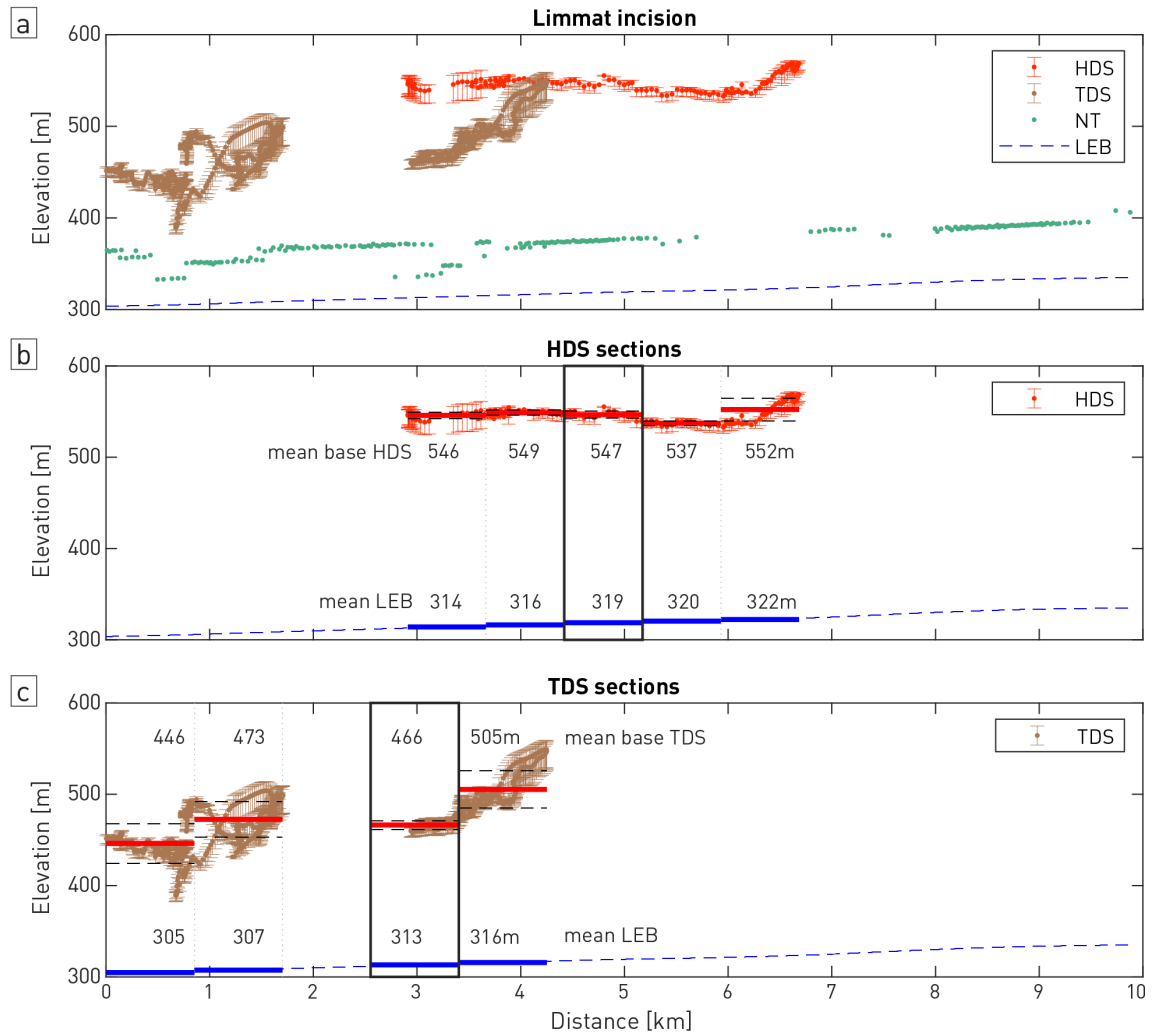


Fig. A-4: Distribution of marker horizons and section-wise statistics for incision calculation along the Limmat Valley

(a) A swath profile aligned along the Limmat Valley captured the data (see rectangle in Fig. A-1 for location and colour-code of the morphostratigraphic unit), which were projected along a profile in the centre of the swath along its long axis. Displayed are Base HDS and TDS with their uncertainty marks (after Hammer et al. 2019), the Niederterrasse surfaces, and the local erosion base (LEB). The latter corresponds to the lowest fluvial incision level. (b) Similar section with focus on Base HDS. For five sections of rather homogeneous marker elevation, the mean (red line with numbers shown) and standard deviation (black dashed line) were calculated. Mean elevation of the LEB is shown for the same sections (c) Similar section with focus on Base TDS. Four sections were chosen for statistic calculations. For both, HDS and TDS, the data highlighted by the black rectangle were used for rate calculations (sections with lower uncertainty were preferred) between elevations of Base HDS or TDS and LEB, respectively shown in Tab. A-2 and Fig. A-5.

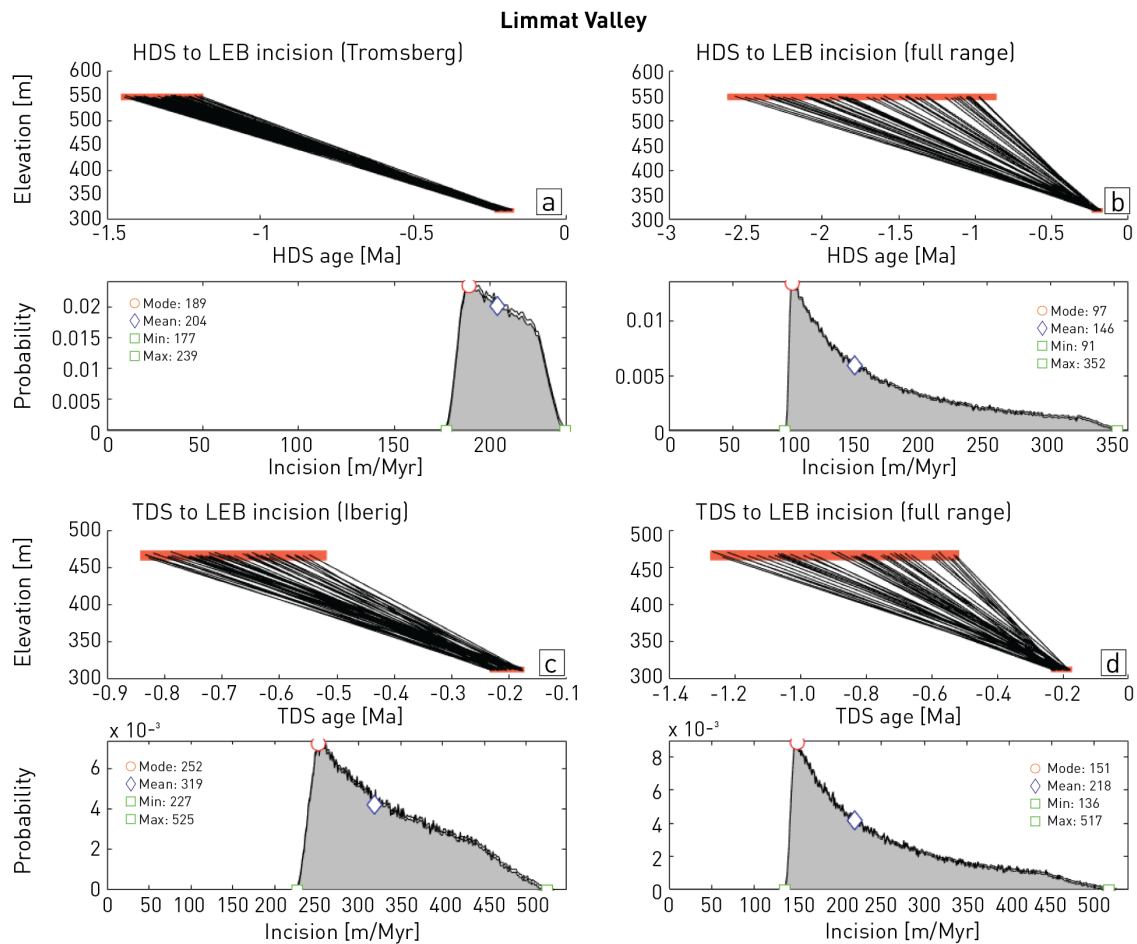


Fig. A-5: Incision rates calculated along the Limmat Valley

Plots show Monte Carlo-derived incision rates based on the Base Deckenschotter and LEB elevations along the Limmat Valley and HDS and TDS ages (see Tab. A-2 for ages, Fig. A-1 for location of data swath profile and Fig. A-4 for data). For general description see Fig. A-3. (a) Sampling using height estimates of Base HDS measurements were paired with the locally derived Tromsberg age and the corresponding incision rate (with respect to Fig. A-4b). (b) Sampling using the same heights, but the full range of HDS age estimate. (c) Sampling using height estimates of Base TDS measurements were paired with the locally-derived Iberig age and the corresponding incision rate (with respect to Fig. A-4c) (b) Sampling using the same heights, but the full range of TDS age estimate. The most likely rate shown in the associated probability density function is 150 m/Myr.

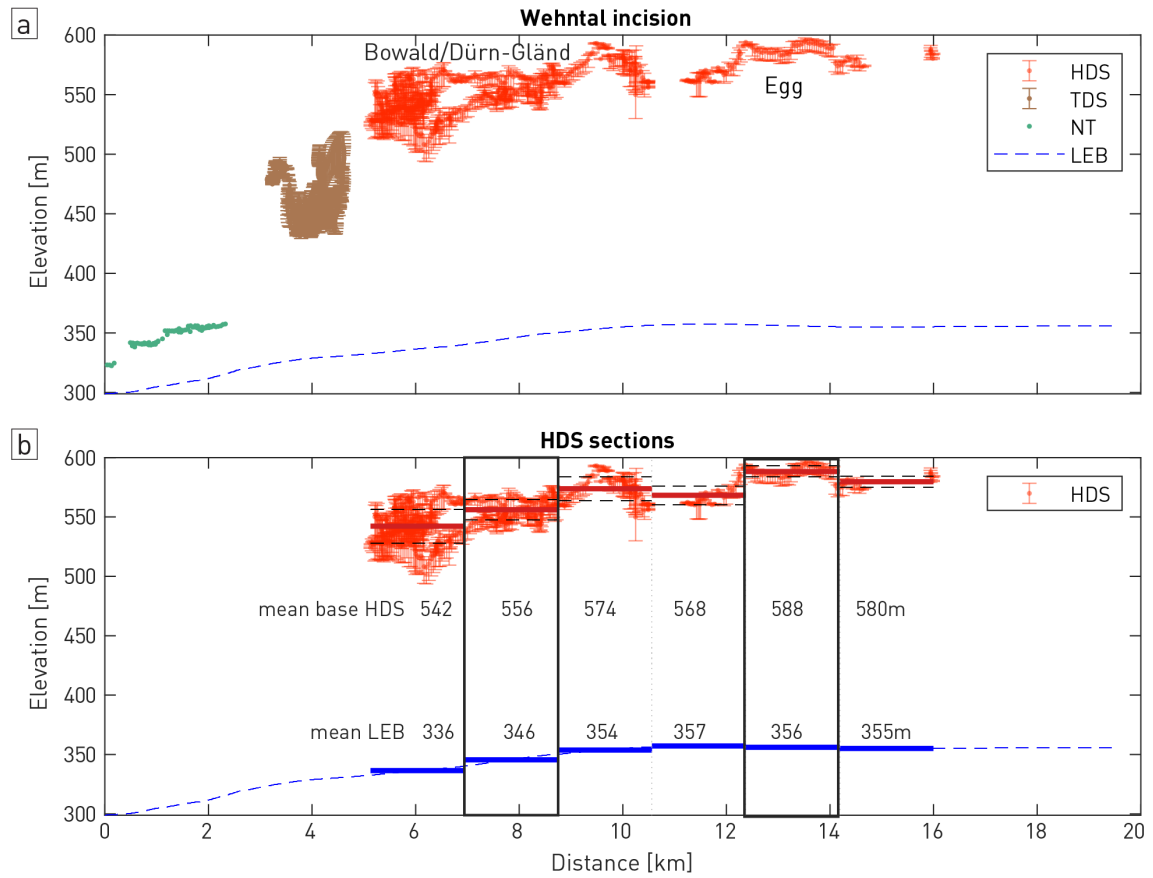


Fig. A-6: Distribution of marker horizons and section-wise statistics for incision calculation along the Wehntal Valley

(a) A swath profile aligned along the Wehntal Valley captured the data (see rectangle in Fig. A-1 for location and colour-code of the morphostratigraphic unit), which were projected along a profile in the centre of the swath along its long axis. Displayed are Base HDS and TDS with their uncertainty marks (after Hammer et al. 2019), the Niederterrasse surfaces, and the local erosion base (LEB). The latter corresponds to the lowest fluvial incision level.

(b) Similar section with focus on Base HDS. For six sections of rather homogeneous marker elevation, the mean (red line with numbers shown) and standard deviation (black dashed line) were calculated. Mean elevation of the LEB is shown for the same sections. Only for HDS, the data highlighted by the black rectangles (capturing two different HDS landmarks, i.e. Bowald/Dürn-Gländ and Egg) were used for rate calculations between elevations of Base HDS and LEB, respectively shown in Tab. A-2 and Fig. A-7.

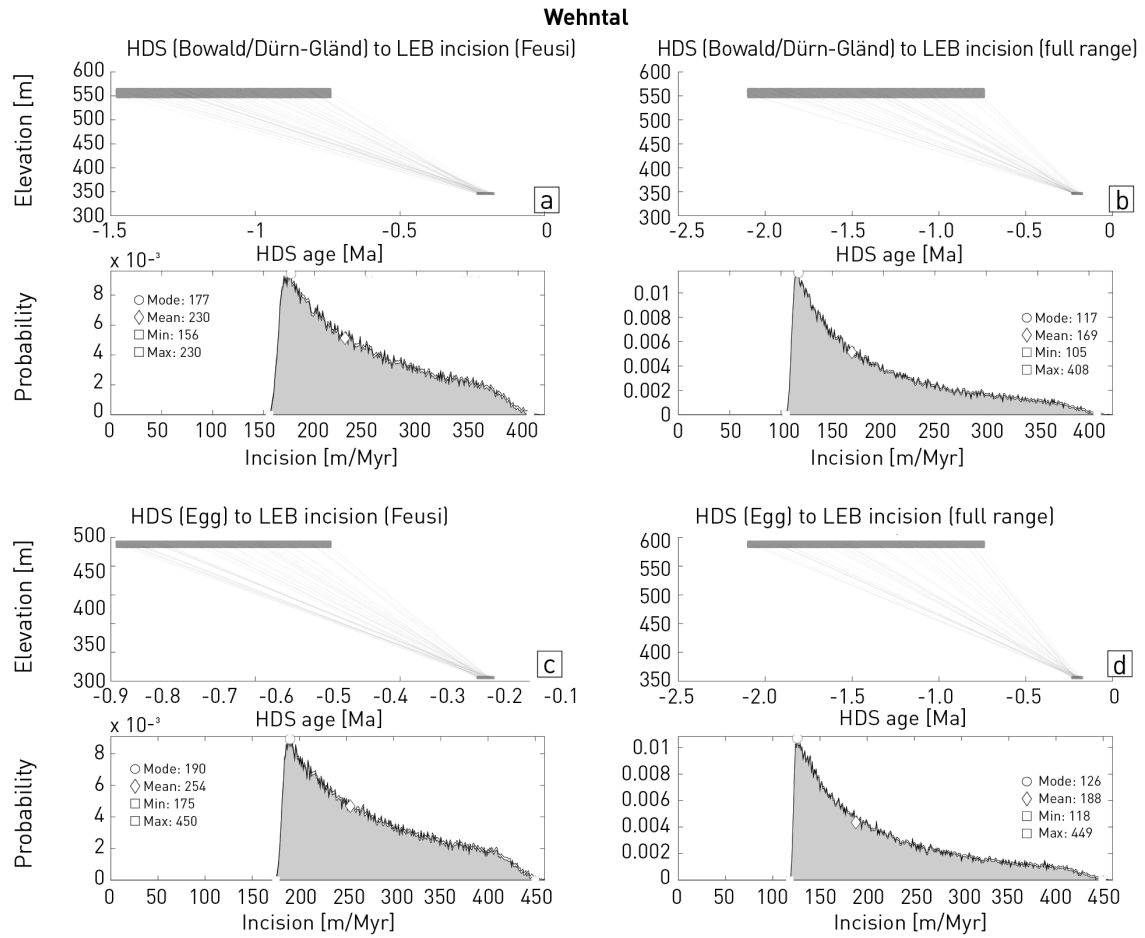


Fig. A-7: Incision rates calculated along the Wehntal Valley

Plots show Monte Carlo-derived incision rates based on the Base Deckenschotter and LEB elevations along the Wehntal Valley and HDS and TDS ages (see Tab. A-1 for ages, Fig. A-1 for location of data swath profile and Fig. A-6 for data). For general description see Fig. A-3. (a) Sampling using height estimates of Base HDS (Bowald/Dürn-Gländ landmark) measurements were paired with the locally derived Feusi age and the corresponding incision rate (with respect to Fig. A-6b). (b) Sampling using the same heights, but the full range of HDS age estimate. (c) Sampling using height estimates of Base HDS (Egg landmark) measurements were paired with the locally derived Feusi age and the corresponding incision rate (with respect to Fig. A-6c). (d) Sampling using the same heights, but the full range of HDS age estimate.

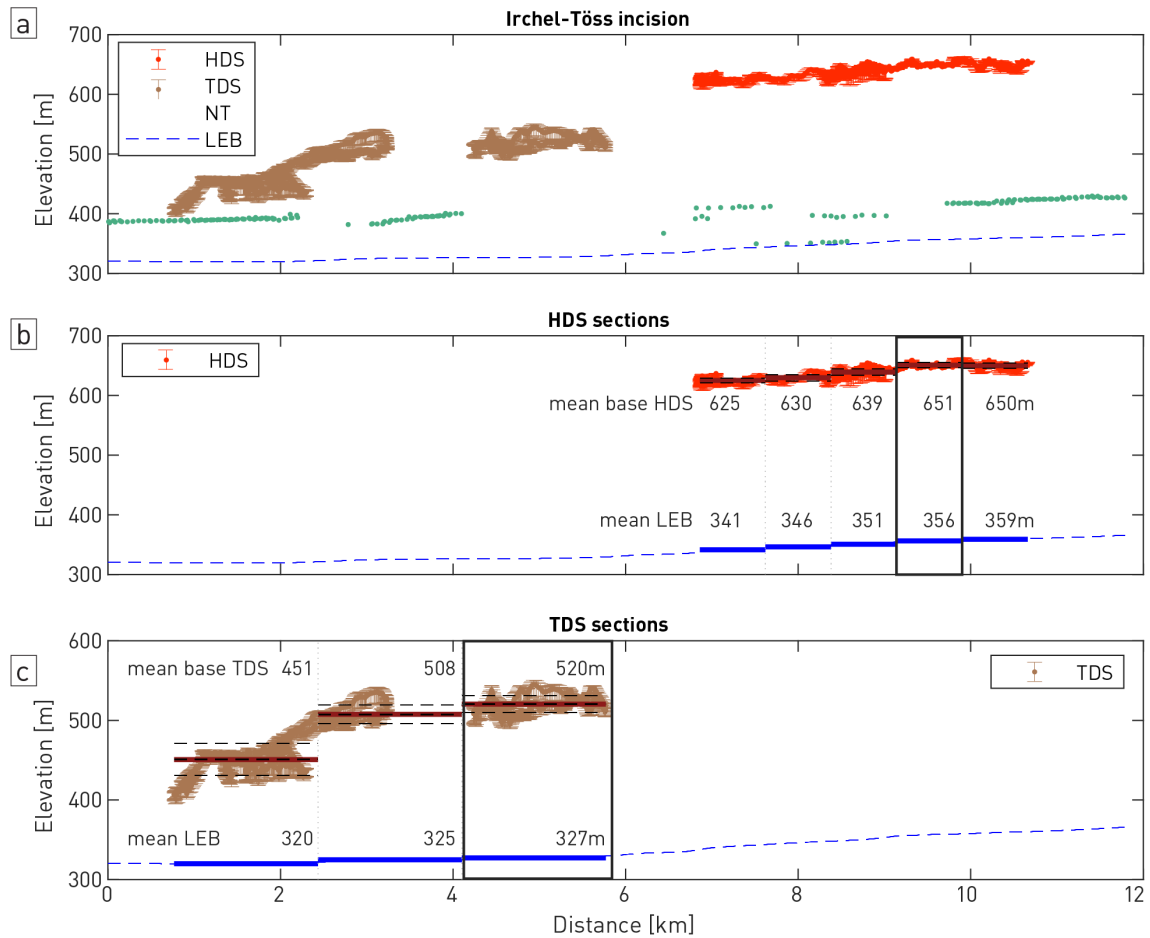


Fig. A-8: Distribution of marker horizons and section-wise statistics for incision calculation along the Töss Valley

(a) A swath profile aligned along the Töss Valley captured the data (see rectangle in Fig. A-1 for location and colour-code of the morphostratigraphic unit), which were projected along a profile in the centre of the swath along its long axis. Displayed are Base HDS and TDS with their uncertainty marks (after Hammer et al. 2019), the Niederterrasse surfaces, and the local erosion base (LEB). The latter corresponds to the lowest fluvial incision level. (b) Similar section with focus on Base HDS. For five sections of rather homogeneous marker elevation, the mean (red line with numbers shown) and standard deviation (black dashed line) were calculated. Mean elevation of the LEB is shown for the same sections (c) Similar section with focus on Base TDS. Three sections were chosen for statistic calculations. For both, HDS and TDS, the data highlighted by the black rectangle were used for rate calculations (sections with lower uncertainty were preferred) between elevations of Base HDS or TDS and LEB, respectively shown in Tab. A-1 and Fig. A-9.

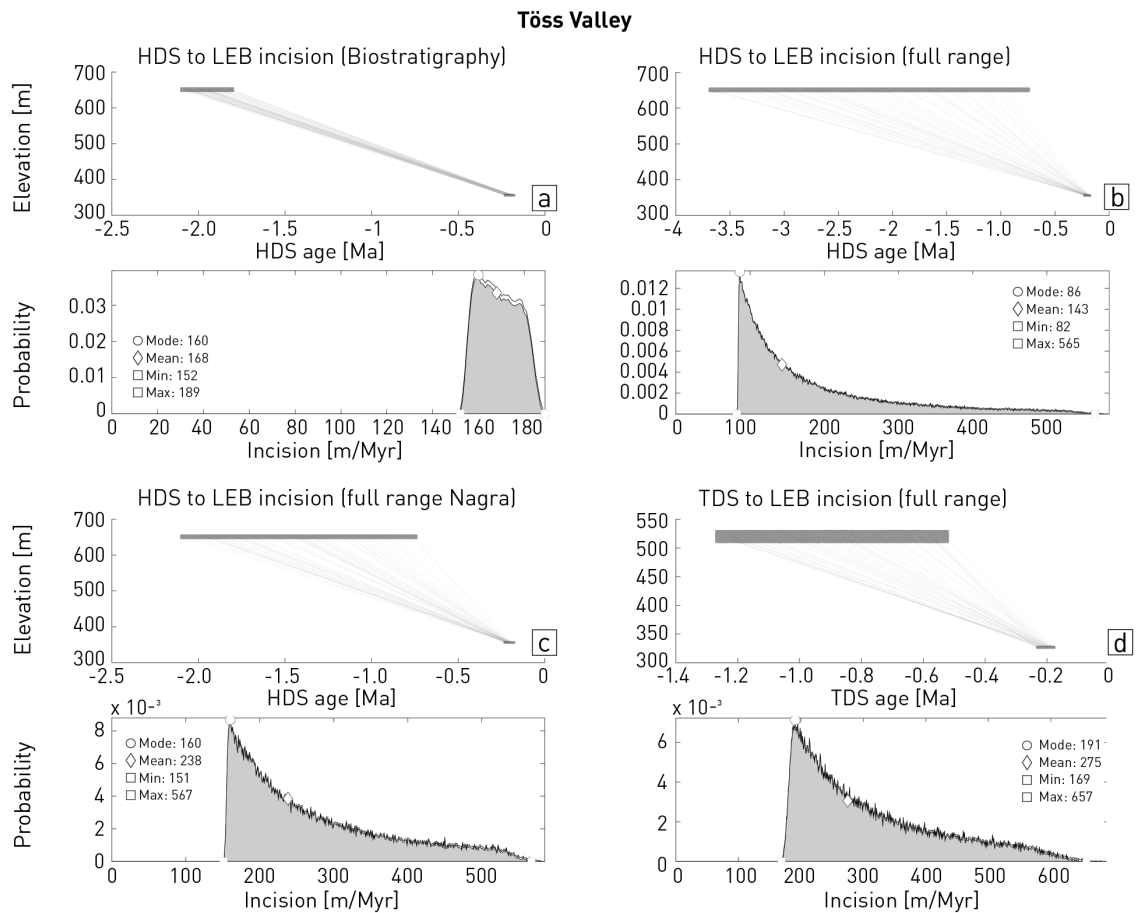


Fig. A-9: Incision rates calculated along the Töss Valley

Plots show Monte Carlo-derived incision rates based on the Base Deckenschotter and LEB elevations along the Töss Valley and HDS and TDS ages (see Tab. A-1 for ages, Fig. A-1 for location of data swath profile and Fig. A-8 for data). For general description see Fig. A-3. (a) Sampling using height estimates of Base HDS measurements were paired with the locally derived biostratigraphic age (Irchel) and the corresponding incision rate (with respect to Fig. A-8b). (b) Sampling using the same heights, but the full range of HDS age estimate. (c) Sampling using the same heights paired with the locally derived HDS ages of Dieleman et al. and the corresponding incision rate (with respect to Fig. A-8c). (d) Sampling using height estimates of Base TDS measurements and the full range of TDS age estimate.

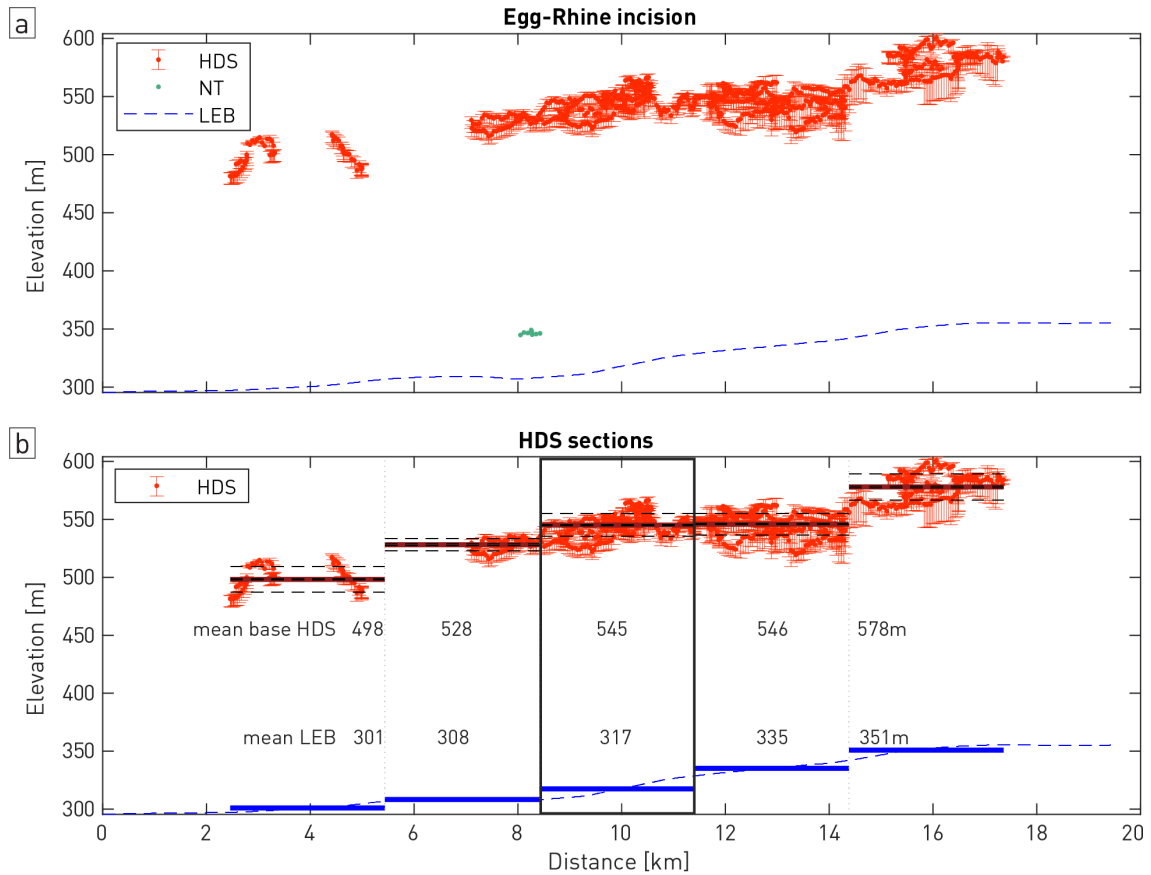


Fig. A-10: Distribution of marker horizons and section-wise statistics for incision calculation along the Hochrhein Valley (Egg)

(a) A swath profile aligned along the Hochrhein Valley captured the data (see rectangle in Fig. A-1 for location and colour-code of the morphostratigraphic unit), which were projected along a profile in the centre of the swath along its long axis. Displayed are Base HDS with their uncertainty marks (after Hammer et al. 2019), a small remnant of the Niederterrasse surfaces, and the local erosion base (LEB). The latter corresponds to the lowest fluvial incision level. (b) Similar section with focus on Base HDS. For five sections of rather homogeneous marker elevation, the mean (red line with numbers shown) and standard deviation (black dashed line) were calculated. Mean elevation of the LEB is shown for the same sections. The data highlighted by the black rectangle were used for rate calculations (sections with lower uncertainty were preferred) between elevations of Base HDS and LEB, shown in Tab. A-2 and Fig. A-11.

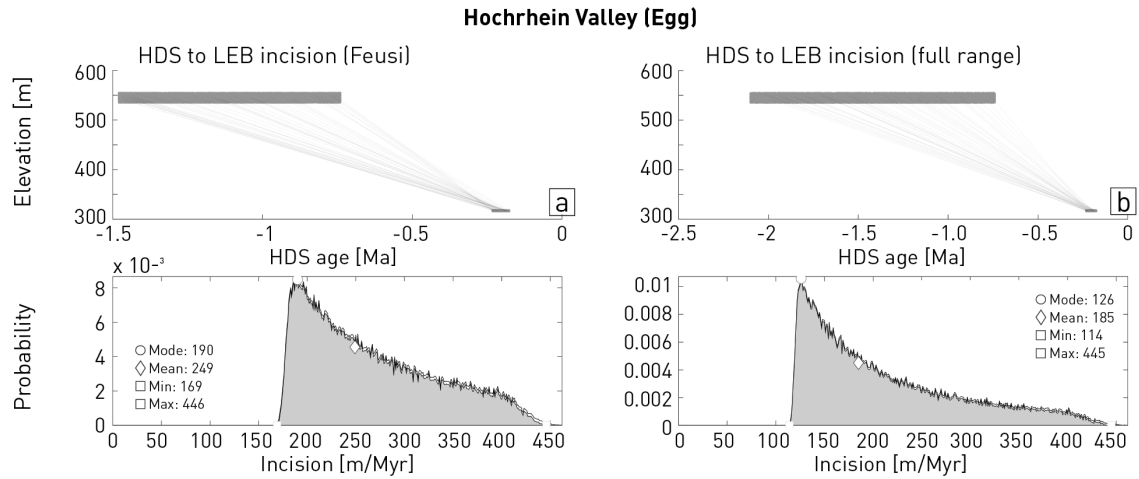


Fig. A-11: Incision rates calculated along the Hochrhein Valley (Egg)

Plots show Monte Carlo-derived incision rates based on the Base Deckenschotter and LEB elevations along the Hochrhein Valley and HDS and TDS ages (see Tab. A-2 for ages, Fig. A-1 for location of data swath profile and Fig. A-10 for data). For general description see Fig. A-3. (a) Sampling using height estimates of Base HDS measurements were paired with the locally derived Feusi age and the corresponding incision rate (with respect to Fig. A-10b). (b) Sampling using the same heights, but only the full range of HDS age estimate.

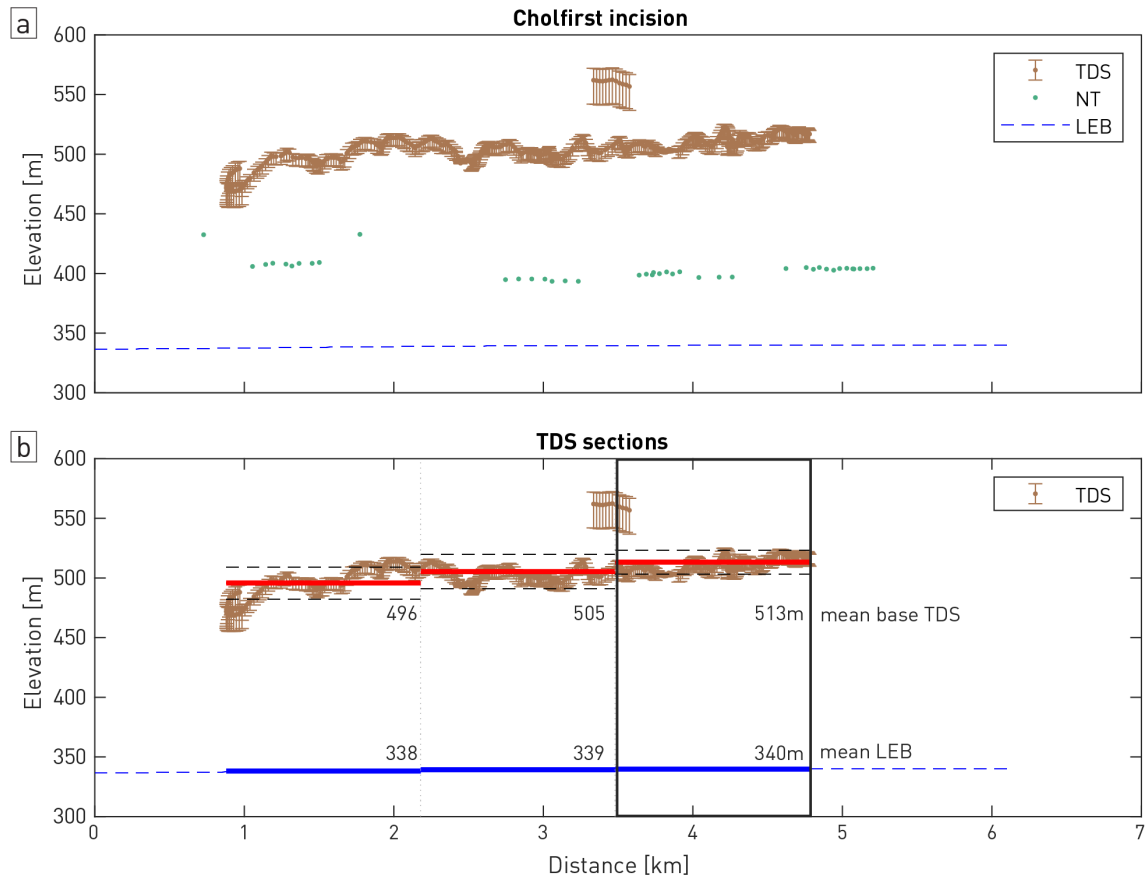


Fig. A-12: Distribution of marker horizons and section-wise statistics for incision calculation along the Hochrhein Valley (Cholfirst)

(a) A swath profile aligned along the Cholfirst landmark in the Hochrhein Valley captured the data (see rectangle in Fig. A-1 for location and colour-code of the morphostratigraphic unit), which were projected along a profile in the centre of the swath along its long axis. Displayed are Base TDS with their uncertainty marks (after Hammer et al. 2019), the Niederterrasse surfaces, and the local erosion base (LEB). The latter corresponds to the lowest fluvial incision level. (b) Similar section with focus on Base TDS. For three sections of rather homogeneous marker elevation, the mean (red line with numbers shown) and standard deviation (black dashed line) were calculated. Mean elevation of the LEB is shown for the same sections. The data highlighted by the black rectangle were used for rate calculations (sections with lower uncertainty were preferred) between elevations of Base TDS and LEB, shown in Tab. A-2 and Fig. A-13.

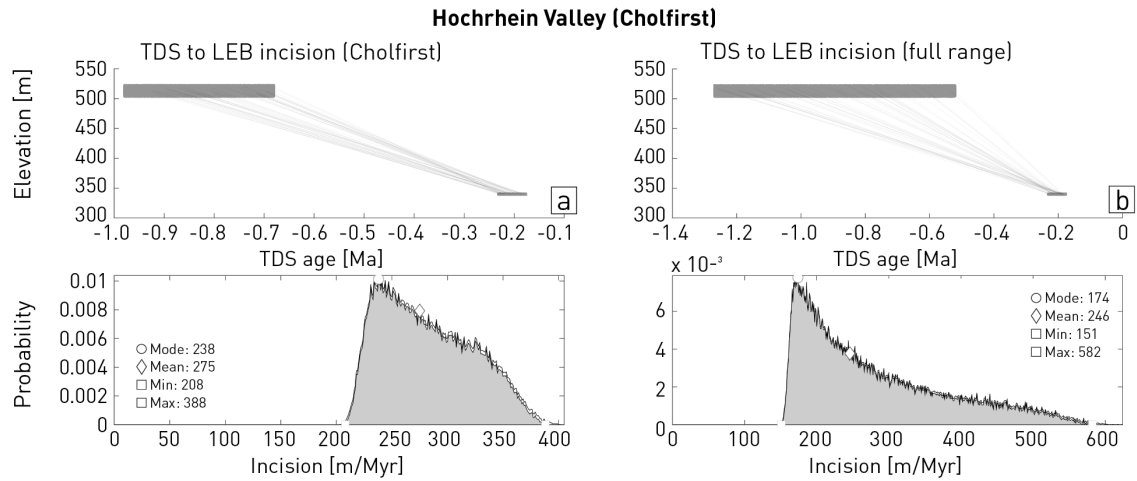


Fig. A-13: Incision rates calculated along the Hochrhein Valley (Cholfirst)

Plots show Monte Carlo-derived incision rates based on the Base Deckenschotter and LEB elevations along the Cholfirst landmark in the Hochrhein Valley and TDS ages (see Tab. A-2 for ages, Fig. A-1 for location of data swath profile and Fig. A-12 for data). For general description see Fig. A-3. (a) Sampling using height estimates of Base TDS measurements were paired with the locally derived Cholfirst age and the corresponding incision rate (with respect to Fig. A-12b). (b) Sampling using the same heights, but the full range of TDS age estimate.

App. B Additional information related to uncertainty management

B.1 Types of uncertainties

In the context of future erosion modelling, as in many other areas of science and engineering, modelling has to be carried out against a background of pronounced uncertainties. Determining the erosion at the siting regions for time spans of a million years involves navigating significant uncertainties. These uncertainties are multifaceted and vary in type and magnitude.

Parametric Uncertainties: These relate to variability in the input parameters used in models, such as environmental conditions or geological characteristics. Since these parameters can fluctuate, they introduce uncertainty into the model outputs. Parametric uncertainties may stem from limited data (epistemic uncertainty) or natural variability (aleatory uncertainty) and directly influence the model's accuracy. In addition, the type of these types of uncertainty may also change over time.

Aleatory Uncertainties: Also known as "irreducible" or "random" uncertainties, these are due to the intrinsic randomness of natural processes, like the frequency and intensity of seismic activity. Aleatory uncertainties can only be described probabilistically (e.g. through probability distributions), as they reflect the stochastic nature of the processes themselves. No amount of data can reduce aleatory uncertainties; rather, increased data collection helps in refining the probability distribution that describes them.

Epistemic Uncertainties: In contrast to aleatory uncertainties, epistemic uncertainties are due to a lack of knowledge or incomplete data, meaning they are "reducible." These uncertainties represent gaps in understanding that, in principle, could be minimised with additional data or improved modelling techniques. For example, uncertainties in fault movement can be reduced through more detailed geological surveys or advanced instrumentation. However, epistemic uncertainties cannot always be fully eliminated, as practical and resource limitations may restrict data collection efforts.

Model Uncertainties arise from the inherent limitations of the involved models. No model can perfectly represent all complex natural processes over long timescales. For instance, models predicting landscape evolution or tectonic uplift may miss certain interactions or fail to capture rare events. This type of uncertainty highlights the challenge of trusting any single model to fully describe dynamic, long-term changes in nature. Although we do not directly include model uncertainties quantitatively, we highlight the reasoning for the model selection (e.g. Sections 3.3 and 3.4) and discuss potential limitations when addressing the plausibility of the modelling results.

Each type of uncertainty requires specific strategies for assessment and management, underscoring the complexity involved in long-term assessments and the need for a nuanced approach to risk analysis.

For **real-value model parameters** (e.g. overdeepening depth), both aleatory and epistemic uncertainties are present. Further research (like more boreholes, better age control) might reduce epistemic uncertainty, but aleatory uncertainty remains due to the unpredictable nature of future events. Distinguishing between these uncertainty types within the model is challenging. Another example is assessing future glaciation or overdeepening events. More data can improve the probability estimates of these events. However, whether these events will occur within a specific timeframe remains an aleatory uncertainty, as it is inherently unpredictable. In the current model, for these discrete parameters, epistemic uncertainties are not separately accounted for; instead, a fixed probability is used, representing the expected value of the uncertainty. For all discrete parameters in the model, the modeling of the uncertainty about the expected value (i.e. the

epistemic uncertainty) does not have an impact on the uncertainty about the model output. Thus, the uncertainty about the expected values of the discrete parameters is not modeled explicitly and the uncertainty in these events is treated solely as aleatory.

However, as the separation of epistemic and aleatory uncertainties is important in the context of a sensitivity analysis (as only epistemic uncertainties can be reduced), the potential impact of (explicitly representing) the epistemic uncertainties of the discrete model parameters was investigated in an alternative extended probabilistic model. It was found that the sensitivity values associated with these uncertainties are small/negligible.

B.2 Information related to expert elicitation

Whenever there is not enough observational data to determine these uncertainties quantitatively, it is unavoidable to employ expert knowledge for this purpose (Oppenheimer et al. 2016). Within the solid-earth geosciences, this is particularly the case for seismic and volcanic hazard and risk assessments, in particular when nuclear facilities are involved. However, more recently, expert elicitation has been used in assessment of climate parameters and models (e.g. Yucca Mountain; Dewispelare et al. 1995, Sebok et al. 2022, Bamber et al. 2019) or even in upcoming topics such as carbon capture and sequestration (Moore et al. 2024).

Inclusion of expert knowledge must be carried out in a structured manner, as otherwise the results can easily become unreliable and not reproducible (Budnitz et al. 1997, Oppenheimer et al. 2016). This realisation has led to the development of a number of methodological approaches to structured expert elicitation, of which the protocol of the Senior Seismic Hazard Analysis Committee (or *SSHAC*; Budnitz et al. 1997) and the Sheffield Elicitation Framework (*SHELF*; Gosling 2018) are two well-tested examples. The aim of both methods is to describe uncertain quantities of interest (QoI) as probability distributions that represent the expert knowledge of a group, based on *expert judgment* of their members. It is important to note that there is a certain vocabulary inherent that may be uncommon. To avoid misunderstandings, brief explanations of the most common terms are added below (O'Hagan 2019).

Expert judgement (also expert opinion) describes the individual representation of one (or a group of) expert's scientific judgement on a specified topic, in this case for instance the uncertainty of a model input parameter. This rational judgement is based on available data and the expert's scientific know-how and experience and is provided after a training on the influence of cognitive biases (Kahneman et al. 1982) in decision making, especially such as anchoring, availability bias or overconfidence but also the more general aspect of group dynamics.

The probability concept behind expert judgement is that of *subjective probability*. Subjective probabilities (Cooke 1991) are the expert's *degrees-of-belief* formally expressed as probabilities. They are called "subjective" because they follow individual (or group's) judgements. These probabilities are nevertheless informative and scientific (O'Hagan 2019) and one of the existing legitimate probability concepts that meet Kolmogorov's axioms (Galavotti 2005).

Elicitation describes the process of gaining the expert's judgement. The uncertainty elicitation for the parameters relevant for the erosion assessment by Nagra followed the SHELF protocol (Gosling 2018), which is a two-step structured group elicitation framework for carrying out the elicitation of subjective probability distributions for quantities of interest (QoI). The actual elicitation is done in workshops which are led by an experienced facilitator (moderator) who has been trained in all aspects of the SHELF protocol.

The first step requires the individual elicitation of a set of quantiles (e.g. median, tertiles/quantiles, upper and lower bounds) for the uncertainty distribution of interest, one for each expert. In the second step, the expert group derives a single group distribution, which reflects an integrated view of the uncertain quantity. This means that all arguments brought up during the group discussions

are considered, as they would be seen through the eyes of a *rational impartial observer (RIO)* (Gosling 2018, O'Hagan 2019) and requires that all experts exchange their individual judgments with the role of the RIO. The RIO perspective takes into account the combined knowledge of the group but gives some (perhaps more convincing) arguments more weight than others. The facilitator "acts" on behalf of the RIO (and thus in agreement and collaboration with the experts), when aggregating the distributions.

The *RIO distributions* were fitted with parametric distributions that best reflected the elicited quantile values. Importantly, these RIO distributions comprise the uncertainty that is propagated into the model (Fig. 3-3). In Fig. 3-3, all input parameters (left side of each model) are quantified using expert elicitation workshops. Not all quantities, however, were assessed as probability distributions, some were provided as scenario weights or event probabilities. Because the elicitations were conducted within the erosion team at Nagra with usually 3 – 5 experts per workshop, it was particularly important to include both the expertise in local geology and geomorphology as well as expertise in process understanding. The facilitator was supported by an external expert with a strong background in statistics and probabilities, who was trained in all SHELF aspects. The high number of parameter uncertainties required prioritisation regarding which parameters to treat with most effort (e.g. a higher number of experts, more time, etc.). This prioritisation was supported by a sensitivity study. In exceptional cases, a pragmatic deviation from the protocol was allowed (e.g. derivation of K values).

It is important to recapitulate that a rigorous evaluation of prior and present conditions of the landscape and the processes that shaped it was conducted by an extensive scientific programme over several years. The available outcomes of this programme were provided to all the experts prior to the workshops to ensure that all experts were 'reading from the same page' and that they had access to the same factual basis.

App. C Additional information about specific output values

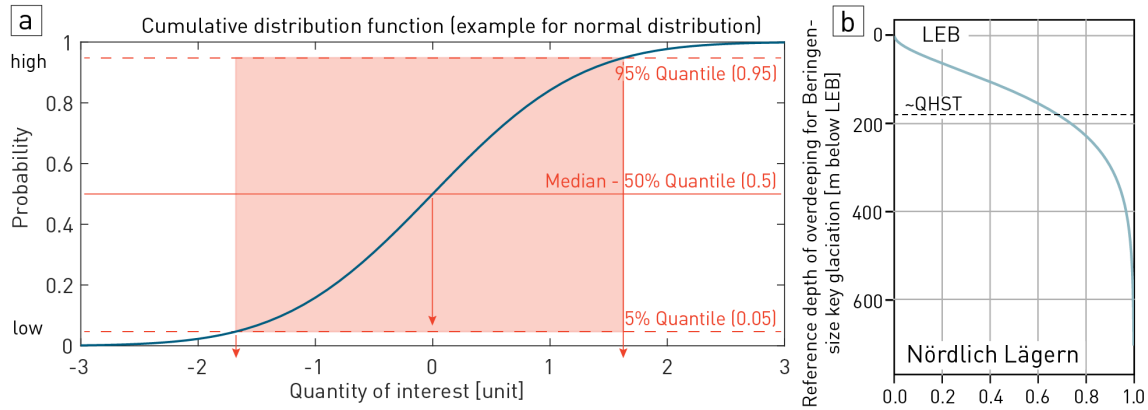


Fig. C-1: General description of a cumulative distribution function (CDF)

A number of our total assumed probabilities and final results, i.e. remaining overburden thickness, are presented as CDF. Here, it is shown as an example how to derive the most important information from such visualisation. (a) The example shows the CDF (blue curve) of a normal distribution of the uncertainty distribution of a “quantity of interest” (QoI), whose values are depicted along the x-axis. The y-axis shows the cumulative probability with a value of 1 representing 100% of the distribution. A typical, often used metric is the median, which is represented by the 50th percentile (= a value of 0.5 or 50% probability of this value), here marked by the horizontal red line. The median bisects the probability and corresponds to the QoI value at the intercept with the blue curve. This means that values associated with the lower half of the CDF (here negative values left of zero) are equally likely as values from the upper half of the CDF (positive values). Similarly, any other quantile value can be derived. Shown by the dashed red lines are the 5th to 95th quantiles, which throughout this report are referred to as “expected” range. The associated range of QoI values is highlighted by the red shaded box (note that similar, but grey shaded boxes highlight the 5 – 95% range of the final remaining overburden estimates in Fig. 4-1b, c and Fig. 4-2b, c). (b) To demonstrate a common use of a CDF to represent our parameter uncertainties, we show as an example the CDF of reference overdeepening depth for a Beringen-size glaciation (compare Fig. 3-40e). This time, the probability is shown on the x-axis and the depth below LEB (which represents the QoI) on the y-axis. Also shown is the depth of the Quaternary borehole QHST (Hochfelden Strassberg Trough), which approximately intersects the 70th percentile. This means that in 30% of the cases, we allow an overdeepening to be deeper than the observed depth in QHST (see Section 3.5.3.1 for argumentation).

Tab. C-1: Incision estimates based on SPIM modelling after 100'000 and one million years

Model results describe incision into the uplifting rock for the three downstream positions of each repository site (Sissle, Aare and Thur outlets). The first value corresponds to 100'000 years and the second value to incision after one million years. For consistency, the values presented here correspond to the HLW depth of the repository also for the 100'000-year time span. However, deviations are in the order of only one metre, if any.

Percentile	Incision at JO (Sissle outlet)	Incision at NL (Aare outlet)	Incision at ZNO (Thur outlet)
5%	10 m / 78 m	8 m / 81 m	6 m / 45 m
25%	17 m / 124 m	14 m / 145 m	9 m / 121 m
50%	26 m / 164 m	22 m / 196 m	12 m / 175 m
75%	38 m / 214 m	33 m / 263 m	15 m / 243 m
95%	52 m / 298 m	54 m / 386 m	27 m / 376 m
Mode	16 m / 151 m	14 m / 171 m	11 m / 157 m

Tab. C-2: Remaining overburden with respect to the local erosion base

In addition to the river profile evolution, the simulations account for uncertainty in rock uplift that may affect disposal sites in the region. Based on this uncertainty, expected overburden thickness above the repository sites relative to the local erosion base was estimated. The first value corresponds to 100'000 years and the second value to remaining overburden with respect to the local erosion base after one million years. For consistency, the values presented here correspond to the overburden above the HLW provisional disposal areas also for the 100'000-year time span.

Percentile	JO	NL	ZNO
5%	201 m / -44 m	699 m / 368 m	580 m / 234 m
25%	217 m / 41 m	720 m / 491 m	593 m / 366 m
50%	228 m / 90 m	731 m / 556 m	596 m / 433 m
75%	237 m / 130 m	738 m / 609 m	598 m / 487 m
95%	245 m / 176 m	744 m / 671 m	602 m / 562 m