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Subglacial erosional landforms and their relevance for the long-term safety of a radioactive waste repository

Guest editors: Jörg Lang, Anke Bebiolka, Sonja Breuer, and Maximilian Pfaff

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Image credit:
Elongated ice cave at the base of the Nordenskjöld glacier near its terminus in Billefjorden (Spitsbergen, Svalbard). The photo was taken in summer 2013. © Anke Bebiolka (BGR)

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Preface

Preface: Subglacial erosional landforms and their relevance for the long-term safety of a radioactive waste repository

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1 Long-term safety of radioactive waste repositories

Deep geological repositories are generally agreed to be the best way to dispose of radioactive waste and to isolate it from the biosphere. The long-term safety of such radioactive waste repositories has to be assessed for very long periods (up to 1 million years), implying that the impact of potential future cold stages and glaciations on the geological barrier of a repository needs to be considered. In glaciated regions, erosion beneath glaciers and ice sheets can mobilise and redistribute substantial amounts of rock and sediment. Subglacial erosional landforms such as overdeepened basins and tunnel valleys may attain depths of more than 500 m. Such deep erosion is within the depth range considered for repositories and could seriously affect the integrity of the geological barrier of a radioactive waste repository during future glaciations.

To address the topic of subglacial erosion and its relevance for the long-term safety of radioactive waste repositories, to bring the community together, and to assess the state of knowledge, a 2 d workshop was organised by the BGR (Bundesanstalt für Geowissenschaften und Rohstoffe – Federal Institute for Geosciences and Natural Resources) and BGE (Bundesgesellschaft für Endlagerung mbH – Federal Company for Radioactive Waste Disposal) in December 2021. The workshop attracted almost 200 participants from research institutes, universities, industry and the wider public. Although the workshop was held online, the combination of lectures, virtual posters and discussions provided an opportunity for interaction and exchange between the participants.

2 Contents of this special issue

This special issue presents a collection of studies on various aspects of subglacial erosion and its treatment with regard to site selection and the long-term safety of radioactive waste repositories.

Cohen et al. (2023) present the results of high-resolution ice-flow simulations of the Rhine glacier during the last glaciation. By estimating the location of subglacial drainage routes and linking these locations to zones of erosion, the model helps to explain the locations of overdeepened valleys in the northern foreland of the Alps.

Breuer et al. (2023) present a new overview map of the maximum depth of Pleistocene tunnel valleys in northern Germany. To assess the potential for future tunnel-valley formation, depth zones are mapped which may serve as a basis for defining a spatially variable additional depth to the minimum depth of a repository required by German legislation.

Müller et al. (2023) explain the role and execution of representative preliminary safety assessments in the site selection process for high-level radioactive waste in Germany. The example of tunnel-valley formation is used to demonstrate how possible future evolutions of potential disposal sites are developed from the understanding of past evolutions to minimise the consequences of error as far as possible.

Gegg and Preusser (2023) compare overdeepened valleys in the foreland of the Alps to tunnel valleys in northern central Europe. Their comparison highlights that foreland overdeepenings and tunnel valleys are connected to different types of ice masses but share many characteristics. The understanding of the formation and infilling of these features would benefit from more intense exchange and discussion between the respective scientific communities.

Lohrberg et al. (2022) mapped and updated Pleistocene tunnel valleys in the south-eastern North Sea based on seismic data. While marine seismic data are effective tools to understand tunnel-valley formation and filling, dense grids of seismic profiles are necessary to map the complex tunnel-valley networks and infills.

Acknowledgements. We would like to thank all the authors and reviewers for their contributions and efforts. Chief editor Christopher Lüthgens is thanked for his support. The Bundesanstalt für Geowissenschaften und Rohstoffe (BGR) and the Bundesgesellschaft für Endlagerung mbH (BGE) supported the production of this special issue.

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Research article

Subglacial hydrology from high-resolution ice-flow simulations of the Rhine Glacier during the Last Glacial Maximum: a proxy for glacial erosion

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Abstract: At the Last Glacial Maximum (LGM), the Rhine Glacier complex (Rhine and Linth glaciers) formed large piedmont lobes extending north into the Swiss and German Alpine forelands. Numerous overdeepened valleys there were formed by repeated glaciations. A characteristic of these overdeepened valleys is their location close to the LGM ice margin, away from the Alps. Numerical models of ice flow of the Rhine Glacier indicate a poor fit between the sliding distance, a proxy for glacial erosion, and the location of these overdeepenings. Calculations of the hydraulic potential based on the computed time-dependent ice surface elevations of the Rhine Glacier lobe obtained from a high-resolution thermo-mechanically coupled Stokes flow model are used to estimate the location of subglacial water drainage routes. Results indicate that the subglacial water discharge is high and focused along glacial valleys and overdeepenings when water pressure is equal to the ice overburden pressure. These conditions are necessary for subglacial water to remove basal sediments, expose fresh bedrock, and favor further erosion by quarrying and abrasion. Knowledge of the location of paleo-subglacial water drainage routes may be useful to understand patterns of subglacial erosion beneath paleo-ice masses that do not otherwise relate to the sliding of ice. Comparison of the erosion pattern from subglacial meltwater with those from quarrying and abrasion shows the importance of subglacial water flow in the formation of distal overdeepenings in the Swiss lowlands.

Kurzfassung: Während des letzteiszeitlichen Maximums (LGM) kam es zur Bildung von grossen Vorlandloben des Rheingletschersystems (Rhein- und Linthgletscher), die sich nordwärts in das Schweizer und deutsche Alpenvorland erstreckten. Durch wiederholte Vergletscherungen wurden dort zahlreiche übertiefte Täler ausgeschürft. Ein Merkmal dieser übertieften Tälern ist deren Lage in der Nähe des LGM-

Eisrands weit weg von den Alpen. Jedoch zeigen numerische Modellierungen des Eisfliessens des Rheingletschers eine schlechte Übereinstimmung der Gleitdistanz, Proxyindikator für glaziale Erosion, mit der Lage dieser Übertiefungen. Deshalb werden Berechnungen des hydraulischen Potenzials basierend auf zeitabhängigen Höhen der Eisoberfläche der Rheingletscherlobe, welche von einem hoch aufgelösten thermo-mechanisch gekoppelten Modell für Stoke'sches Fliessen resultieren, benutzt, um die Lage der Entwässerungsrouten unter dem Eis abzuschätzen. Die Resultate deuten darauf hin, dass der subglaziale Wasserabfluss gross ist und entlang glazialen Tälern und Übertiefungen geführt wird, wenn der Wasserdruck dem Eisüberlagerungsdruck entspricht. Dies sind notwendige Bedingungen, unter denen basale Sedimente wegtransportiert werden und frischer Fels freigelegt wird, um weitere glaziale Erosion zu begünstigen. Somit ist die Kenntnis der Lage von subglazialen Paläo-Entwässerungsrouten nützlich, um die Erosionsmuster unter Paläo-Gletschern zu verstehen, die nicht mit der Gleitbewegung des Eises in Verbindung gebracht werden können. Ein Vergleich der durch subglaziale Schmelzwässer erzeugten Erosionsmuster mit jenen, die durch direkte Gletschererosion entstanden sind, zeigt die Wichtigkeit der subglazialen Schmelzwasserflüsse für die Entstehung von Übertiefungen im alpenfernen Schweizer Vorland auf.

1 Introduction

At the Last Glacial Maximum (LGM), the Rhine Glacier complex (Cohen et al., 2018), combining the Rhine and Linth glaciers, descended well into the Swiss and German Alpine forelands, forming two large piedmont lobes. The Rhine Glacier lobe covered present-day Lake Constance and land to the north in southern Germany, and the Linth Glacier lobe advanced beyond the city of Zurich, Switzerland. Repeated glaciations since the Middle–Late Pleistocene (Preusser et al., 2011; Ellwanger et al., 2011) have carved numerous landforms in the Alps such as deep alpine valleys with prominent horns and ridges. In the forelands, the passage of glaciers left numerous imprints on the landscape such as terminal moraines, outwash deposits, and erratics (e.g., Schlüchter, 1988, 2004; Keller and Krayss, 2005a; Preusser et al., 2007; Beckenbach et al., 2014; Gaar et al., 2019); drumlin fields (Kamleitner, 2022); tunnel valleys (Reber and Schlunegger, 2016); and overdeepened valleys now filled with sediments such as the Thur, the Glatt, and the Aare valleys (e.g., Preusser et al., 2011; Dehnert et al., 2012; Dürst Stucki and Schlunegger, 2013) or with water such as Lake Constance and Lake Zurich. A surprising characteristic of these overdeepenings along these valleys is their proximity to the terminal position of past glacial maxima in a region where one would expect that the erosive power of ice, often associated with the rate of sliding at the base (e.g., Hallet, 1981; Humphrey and Raymond, 1994; MacGregor et al., 2000; Koppes et al., 2015; Herman et al., 2015), would be smaller in comparison to that of large alpine valleys where ice fluxes and sliding speeds are higher. The existence of these overdeepenings in a distal-foreland setting thus remains puzzling.

Previous models of ice flow of the Rhine Glacier (Cohen and Jouvet, 2017; Haeberli et al., 2020; Fischer et al., 2021; Seguinot and Delaney, 2021) indicate that the sliding distance, a proxy for glacial erosion based on the time-integrated sliding speed, or a power of it, during the advance and retreat of the Rhine Glacier lobe into the foreland, does not correlate well with the location of existing overdeepenings (Fig. 1) owing to the short time period over which ice covered these distal forelands. More recent work by Egholm (2022) using a landscape evolution model indicates that the formation of distal overdeepenings is possible if the rock material is significantly weaker there than the material in the Alpine massif. The model of Egholm (2022) used empirical parametrizations of abrasion and quarrying (Ugelvig et al., 2018) that were fitted against theoretical models (Hallet, 1981; Iverson, 2012). Egholm (2022)'s results, however, were mostly obtained using a steady-state climate model in which ice occupation time over the Swiss Alpine foreland was artificially long, in contrast with the geomorphic record for the Rhine Glacier (e.g., Keller and Krayss, 2005b) that suggests a maximum extent lasting a couple of thousand of years. This dilemma begs the question as to what other subglacial glaciological process could have contributed to the erosion of the northern Alpine forelands and the formation of these distal overdeepenings.

Several authors have noted the importance of subglacial water in glacial erosion (see review by Alley et al., 2019). Erosion by quarrying (Hooke, 1981; Iverson, 1991; Alley et al., 1999), whereby blocks of rock are plucked from the glacier bed by the moving ice, depends on subglacial water pressure fluctuations. Field experiments (Cohen et al., 2006) and models of quarrying (Iverson, 1991, 2012; Hallet, 1996; Beaud et al., 2014; Ugelvig et al., 2018) have shown a link between fluctuations in water pressure and the rate of opening of pre-existing bedrock cracks and fractures. Furthermore, large-scale investigations of the erosive power of glaciers (Koppes et al., 2015) indicate that temperate basal conditions, which imply the presence of subglacial water, are indicative of higher erosion rates, particularly at middle to high latitudes where seasonal and diurnal temperature variations are high and thus increase the possibility of significant water pressure fluctuations. Subglacial water is also key to the removal of sediments eroded and deposited at the base of glaciers (e.g., Hallet, 1996) that would otherwise blanket the bedrock, preventing any erosion process from occurring (Alley et al., 2019). Indeed, freeze-on conditions along the adverse slope of overdeepenings inhibit subglacial water flow and sediment removal, leading to decrease in erosion there (e.g., Hooke, 1991; Alley et al., 2003; Werder, 2016). Thus, subglacial water plays an important role in the erosive power of glaciers.

Lacking a clear link between sliding ice (or motion) and observed locations of overdeepenings in the Swiss Alpine forelands underneath the paleo-Rhine Glacier and paleo-Linth Glacier lobes, we seek to find whether subglacial water drainage distribution under the paleo-Rhine Glacier complex could help us understand the location of these overdeepenings. Our approach uses the numerical ice-flow model described in Cohen et al. (2018) but with a new climate signal to drive the ice model and a finer horizontal discretization to better resolve the ice flow. Calculated ice surface elevation, together with basal topography, allows us to compute the hydraulic potential and hydraulic head and the associated flow accumulation area during the course of advance and retreat of the Rhine Glacier around the LGM. Time integration of the flow accumulation area yields a metric for glacial erosion by subglacial water to test against the positions of observed overdeepenings. The underlying assumption in this analysis is that larger values of the time-integrated flow accumulation area indicate zones with significant subglacial water flow, which, over seasonal and annual timescales, have high potential for subglacial water fluxes that would enhance sediment removal and increase glacial erosion. We then compare patterns of relative erosion by quarrying, abrasion, and subglacial water flow to show the importance of subglacial water in the formation of distal overdeepenings in the Swiss lowlands.

The remainder of this paper is divided into three parts. Section 2 presents the methodology to compute the time-integrated flow accumulation area and the erosion by quarrying, abrasion, and subglacial water; Sect. 3 discusses the source of the data used for the computation; finally, Sect. 4 reports results and discusses them.

2 Methods

Subglacial water flow paths are calculated using the method described in Chu et al. (2016) (see also Arnold et al., 1998; Willis et al., 2012; Livingstone et al., 2013) and used in several studies to estimate subglacial water drainage locations in Greenland and Antarctica (e.g., Le Brocq et al., 2009; see also references in Chu et al., 2016) and in paleo-ice masses such as the Fennoscandian Ice Sheet (e.g., Shackleton et al., 2018). More sophisticated models of subglacial hydrology (e.g., GlaDS; Werder et al., 2013) coupled to ice-flow models (e.g., Gagliardini and Werder, 2018, with Elmer/Ice) are too computationally expensive to be applied to the long timescales and the large complex topography of the paleo-Rhine Glacier system.

Water flows down the hydraulic potential gradient, and the water flux, Q, can be calculated as (Shreve, 1972)

$$Q = -k\nabla\phi, \tag{1}$$

where k is the hydraulic conductivity of the subglacial water system and ϕ is the hydraulic potential, defined as

$$\phi = \rho_\mathrm{w} g Z_\mathrm{b} + p_\mathrm{w}. \tag{2}$$

Here $\rho_\mathrm{w} = 1000\,\mathrm{kg\,m^{-3}}$ is the density of water (constant), $g = 9.81\,\mathrm{m\,s^{-2}}$ is the acceleration due to gravity, Z_b is the elevation of the basal surface (glacier bed topography), and p_w is the water pressure. In glaciological applications, the water pressure at the bottom of the ice is often not known. In paleo-glaciological applications, no subglacial water pressure data exist, so it is often not computed (e.g., Cohen et al., 2018). Here we assume that the water pressure at the bed of the glacier is a fraction of the ice overburden pressure; i.e.,

$$p_\mathrm{w} = f\, p_\mathrm{i}, \tag{3}$$

where f is known as the flotation factor and

$$p_\mathrm{i} = \rho_\mathrm{i}\, g\, H, \tag{4}$$

with $\rho_\mathrm{i} = 917\,\mathrm{kg\,m^{-3}}$ being the ice density (constant) and H the ice thickness (variable). Substituting Eqs. (3) and (4) into Eq. (2) yields

$$\phi = \rho_\mathrm{w} g Z_\mathrm{b} + f \rho_\mathrm{i}\, g\, H, \tag{5}$$

where the hydraulic potential now only depends on the basal surface topography and the ice thickness. As an alternative we can use the hydraulic head, h, defined as

$$h = Z_\mathrm{b} + f\frac{\rho_\mathrm{i}}{\rho_\mathrm{w}}\, H. \tag{6}$$

The value of the flotation factor f depends on the basal water pressure. When water pressure equals ice overburden pressure, $f = 1$. Values greater than 1 are observed during times of intense surface melt that bring water through crevasses and moulins directly to the bed of the glacier (Meierbachtol et al., 2013; Wright et al., 2016). Since our ice-flow model (Cohen et al., 2018) does not simulate subglacial water pressures and given the uncertainties in both the basal surface during the last glacial cycle and estimates of the ice surface, we assume spatially fixed values of f over the entire glacier system and compute the hydraulic potential for three cases – $f = 0.6$, $f = 1$, and $f = 1.1$ – to study the effects of water pressure on subglacial water drainage paths. Low values of f imply

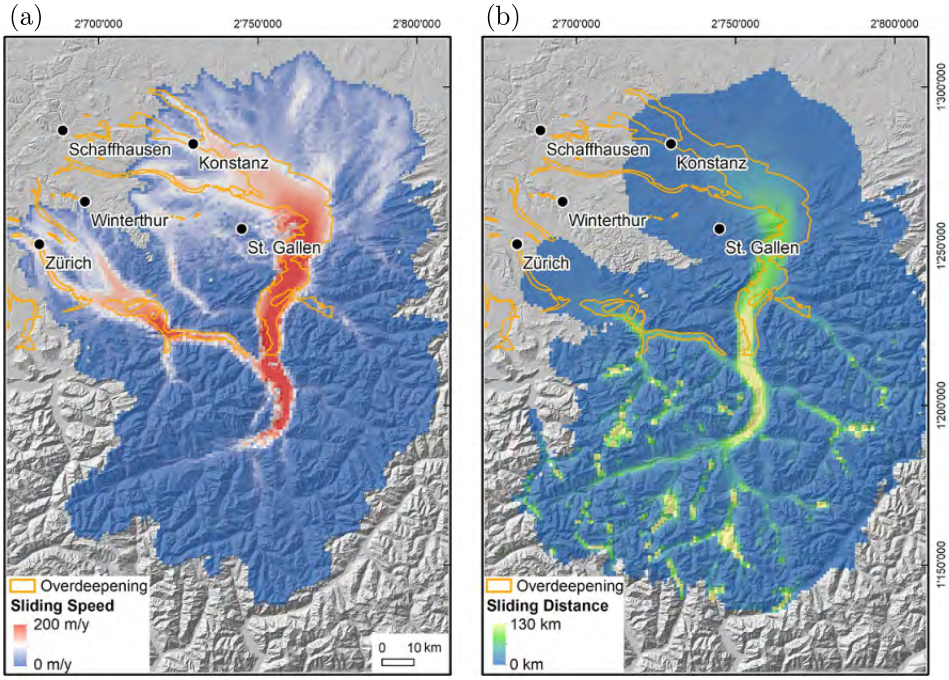

Figure 1. Map of **(a)** sliding speed and **(b)** sliding distance calculated by integrating the sliding speed over the advance and the retreat of the Rhine Glacier system during the last glacial cycle (see Cohen, 2017; Cohen and Jouvet, 2017; and Cohen et al., 2018, for details). The location of overdeepenings is indicated in orange. Reproduced from Fischer et al. (2021).

that basal topography dominates, while higher values emphasize the ice surface topography (see Eq. 6).

The subglacial water drainage paths can be obtained by computing the upstream accumulation area at each of the cells of the raster of the hydraulic head (Flowers and Clarke, 1999; Fischer et al., 2005; Chu et al., 2016; Pitcher et al., 2016) using the D_∞ algorithm of Tarboton (1997) to compute flow directions. Water drainage flow paths are continuous paths with high values of upstream accumulation areas.

On the long timescales over which erosion occurs (on the order of thousands of years), the ice thickness (H in Eq. 6) evolves with advances and retreats of the glacier. To include these transient effects, we compute the subglacial water flow paths at each available time step for the ice surface (every 10 years) and integrate the computed upstream accumulation area over time to obtain a time-integrated view of subglacial water drainage paths.

Estimates of erosion by subglacial water flow are calculated using the method of Kirkham et al. (2022) based on a model of subglacial channel erosion by Carter et al. (2017). Here we only compute the erosive component (see Kirkham et al., 2022, Eq. 3),

$$\dot{E}_s = K_1 \left(\frac{v_s}{\alpha_{cc}}\right) \left(\frac{\max(\tau_{cc} - \tau_k, 0)}{g(\rho_s - \rho_w) D_{15}}\right)^{3/2}, \quad (7)$$

where

$$\tau_{cc} = 0.125\, f_{cc}\, \rho_w\, u_c^2, \quad (8)$$

$$\tau_k = 0.025\, D_{15}\, g\, (\rho_s - \rho_w), \quad (9)$$

$$v_s = D_{15}^2 \frac{(\rho_s - \rho_w)\, g}{9\mu_w}. \quad (10)$$

Here τ_{cc} is the channel shear stress, τ_k is the critical shear stress, v_s is the mean sediment velocity, u_c is the water velocity in the channel, $f_{cc} = 0.07\,\mathrm{m}^{-2/3}\,\mathrm{s}^{-2}$ denotes the channel roughness parameters, $\rho_s = 2600\,\mathrm{kg\,m}^{-3}$ is the solid density of sediment grains, $D_{15} = 0.1$ mm is the 15th-percentile grain size, and $\mu_w = 1.78 \times 10^{-3}$ Pa s is the viscosity of water. The water flow velocity in the channel, u_c, is computed from the flow accumulation area weighted by the available meltwater at each cell, which is estimated by summing the surface melt rate and the basal melt rate. The meltwater is then scaled by multiplying it by the grid cell area divided by the cross-sectional area of the channel, assumed to be 30 m wide and 2 m deep, to obtain the channel water velocity. Although changing the channel size would affect the relationship between τ_{cc} and τ_k and thus where erosion occurs, it does not affect the overall results of erosion patterns as our calculations are meant to only look at erosion patterns and not calculate erosion magnitudes. The surface melt rate is calculated in the ablation area of the glacier and is proportional to the number of positive degree days (PDDs) (Hock, 2003). It is assumed to be zero in the accumulation area. The basal melt rate is computed from the geothermal heat flux map of Medici and Rybach (1995). Frictional heat generated by ice sliding over the substrate is an order of magnitude smaller than either surface or basal melt and is thus neglected.

To compare patterns of subglacial erosion due to ice motion, we also estimate erosion by quarrying, $\dot{E}_q$, and abrasion, $\dot{E}_a$, given, respectively, by (see Ugelvig et al., 2016, 2018)

$$\dot{E}_q = K_q \, p_e^3 v_s, \tag{11}$$

$$\dot{E}_a = K_a \, v_s^2, \tag{12}$$

where $p_e = p_w - p_i$ is the effective pressure, v_s is the glacier sliding speed, and K_q and K_a are constants.

All erosion rates are time-integrated to yield a total erosion over the course of the advance and retreat of the glacier during the LGM. Because of uncertainties in model and parameter values, a direct comparison between quarrying, abrasion, and subglacial meltwater erosion is not possible. Instead, these quantities are used to look at relative patterns of erosion by normalizing them. This makes the exact values of the parameters K_1 (Eq. 7) and K_q and K_a (Eqs. 11 and 12, respectively) irrelevant.

3 Data

Only two surfaces are required to compute the hydraulic potential and subglacial water flow paths: the basal surface (Z_b in Eq. 6) and the ice surface from which the ice thickness, H, can be obtained (see Eq. 6). The ice surface is computed from ice-flow simulations of the paleo-Rhine Glacier complex around the LGM using Elmer/Ice, an open-source, finite-element multi-physics Fortran code (Gagliardini et al., 2013; Ruokolainen et al., 2020). Elmer/Ice computes the transient three-dimensional ice velocities, temperatures, and pressures as well as the elevation of the ice surface based on fully coupled thermo-mechanical equations by solving the non-linear Stokes flow equation together with the heat equation and the mass balance equation at the ice surface (see Cohen et al., 2018, for details). The mass balance is based on the positive-degree-day model (Hock, 2003) that computes ice melt as a function of air temperature. PDDs are computed using the method of Calov and Greve (2005) over month-long intervals using factors of 8 and 3 mm °C^{-1} d^{-1} for ice and snow, respectively. Air temperature and precipitation are obtained by linear interpolation using a glacial-index (GI) approach (Sutter et al., 2019) between LGM and pre-industrial (PI) climate states. The GI follows the δ^{18}O signal of the Antarctica EPICA ice core (Jouzel et al., 2007), rescaled to between 0 (no ice, PI) and 1 (maximum ice volume at the LGM). The LGM climate state provides 2 km resolution maps of the temperature and precipitation over the entire Alps computed from a regional climate model by dynamical downscaling of a global Earth system model (see details in Velasquez et al., 2020, 2022; Russo et al., 2022; Buzan et al., 2023).

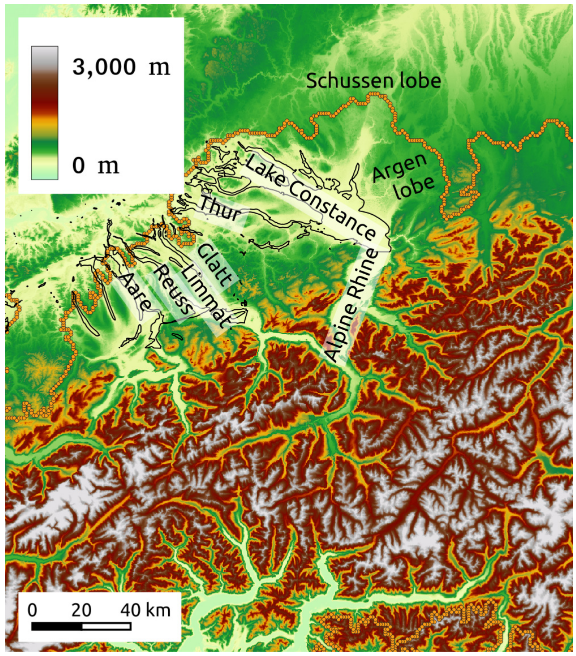

Figure 2. Basal topography with outline of overdeepened basins in black and extent of ice at the LGM with orange circles (Ehlers and Gibbard, 2008).

3.1 Basal surface

The basal surface is the present-day topography minus the computed ice thickness of major present-day glaciers using the model of Farinotti et al. (2019). It remains unchanged during the ice-flow simulation. Figure 2 shows the basal topography used in the model simulation.

3.2 Ice surface

The ice surface evolves with time during the advance of the Rhine Glacier towards its maximum LGM position and during its retreat. Figure 3 shows three ice surface elevations at 27 ka, at 22.9 ka (LGM), and at 14.2 ka. Ice surface elevations are available every 10 years from 27 to 14.2 ka.

3.3 Calculation of hydraulic potential and flow accumulation area

The hydraulic potential is first calculated on the unstructured triangular mesh of the Elmer/Ice finite-element model at every time step (every 10 years). Resolution in the mesh ranges from 800 to 1100 m, with increased resolution in areas of steeper slope gradients. The nodal values of the hydraulic potential are then linearly interpolated to a uniform raster at a resolution of 500 m. The calculation of the flow accumulation area is performed on the hydraulic potential using the open toolbox SAGA (System for Automated Geoscientific Analyses; Conrad et al., 2015) and summed over time to compute the time-integrated values.

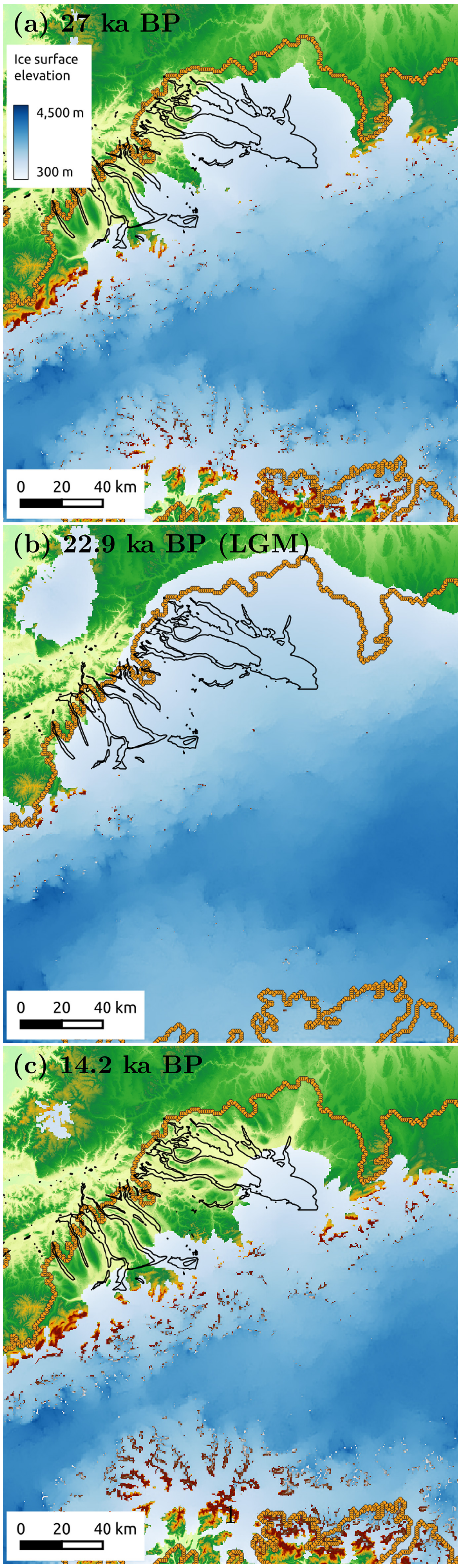

Figure 3. Ice surface elevation in blue at **(a)** 27 ka (start of simulation), **(b)** 22.9 ka (LGM), and **(c)** 14.2 ka (end of simulation). LGM ice extent is shown with orange circles (Ehlers and Gibbard, 2008).

4 Results and discussion

Figure 4 shows the subglacial water drainage routes based on the flow accumulation area of the hydraulic potential (or hydraulic head) at three different times for flotation values of $f = 0.6$ (Fig. 4, left panels), $f = 1.0$ (Fig. 4, center panels), and $f = 1.1$ (Fig. 4, right panels).

The flow accumulation area can be understood as a proxy for subglacial water flux that passes through a specific drainage channel because water flux to the glacier is an increasing function of the upstream flow accumulation area.

When $f = 0.6$ (Fig. 4, left panels), subglacial water drainages are mostly concentrated in the Alpine Rhein Valley in the Alps, in Lake Constance, and along major valleys in the forelands such as the drainage of Lake Zurich. Drainages are constrained by basal topography, following valley morphology. Lake Constance (a zone of constant elevation), where the Alpine Rhein Glacier flows, concentrates a significant flux of subglacial water according to the model. Water exits the Rhine Glacier lobe to the north via existing valleys in the German Alpine foreland, following the Überlingen branch of present-day Lake Constance. In the Swiss Alpine foreland, water drains out of Lake Constance via the Untersee and also by connecting to the Thur Valley, which coincides with an overdeepening there. In the Linth Glacier lobe, subglacial water also flows along Lake Zurich and the Limmat and Glatt valleys, where two overdeepenings are located (Preusser et al., 2010; Buechi et al., 2018). The Reuss and Aare overdeepenings are also pathways for subglacial water drainages further to the west. The drainage pattern does not change significantly with time (see Fig. 4a, c, e) except at $t = 14.2$ ka when the ice has retreated halfway out of Lake Constance and completely out of the distal overdeepenings in the Swiss forelands.

The drainage pattern is significantly different when $f = 1.0$ (Fig. 4, center panels), equivalent to zero effective pressure at the base. Water now drains out of the Lake Constance basin to the north in many concentrated flow channels, such as the Schussen Valley, against an adverse bedrock slope (see basal topography in Fig. 2) and where a few overdeepenings have been noted (Ellwanger et al., 2011). Subglacial flow is also more important along many valleys in the foreland, both in the Rhine Glacier (e.g., Thur) and in the Linth Glacier (e.g., Reuss) lobes. Because of higher water pressure, the hydraulic potential gradient conforms less to the basal (or bed) topography, and subglacial water moves more along the gradient determined by the slope of the glacier surface; subglacial water paths no longer follow topographic lows.

At a yet higher flotation value ($f = 1.1$, Fig. 4, right panels), when water pressure exceeds ice overburden pressure, subglacial drainages out of the Alpine Rhine are also diverted to the north across Lake Constance and then ascending the bedrock slope north of it. Water drainage, however, occurs over many more subglacial channels that fan out radially from the Lake Constance basin. Because of the high wa-

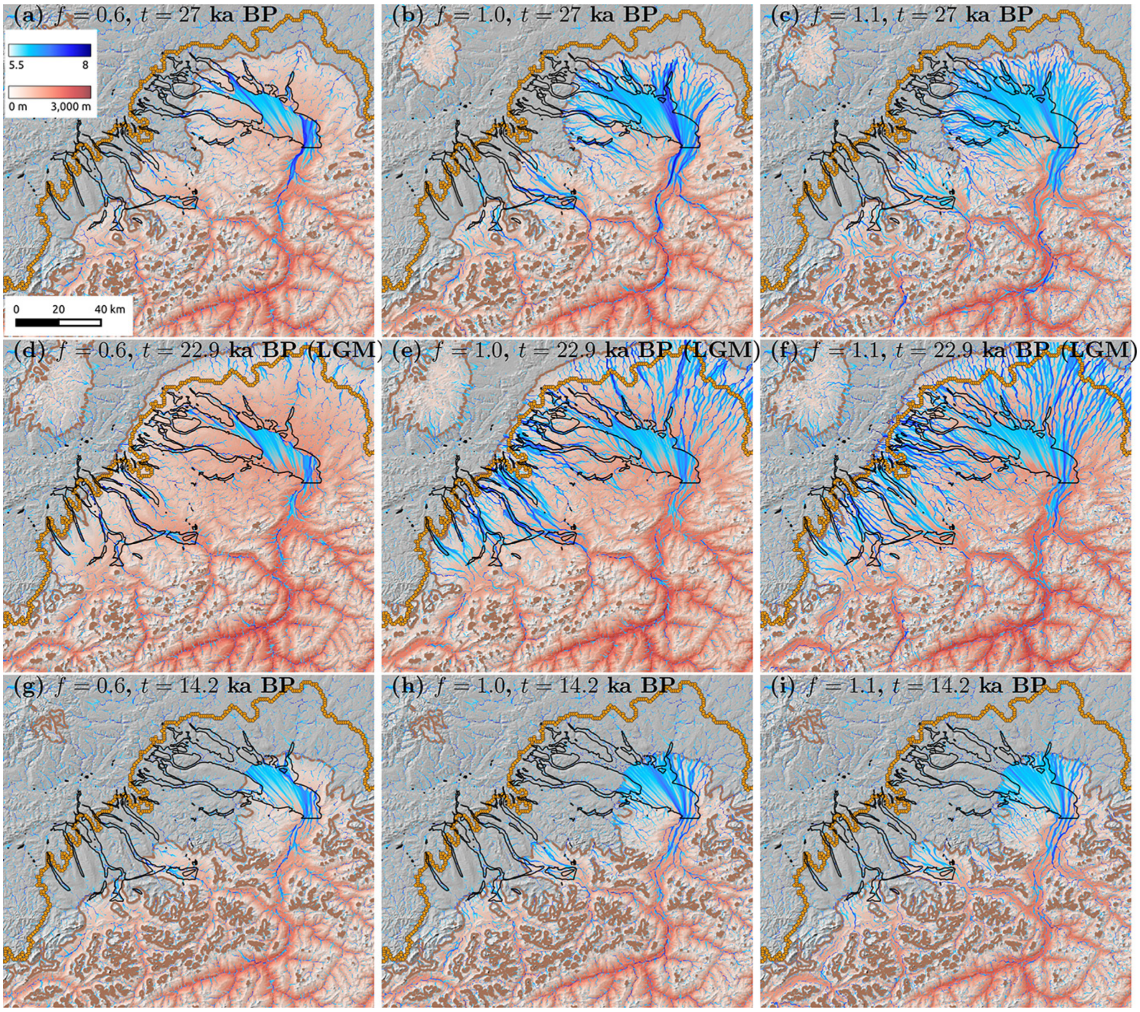

Figure 4. The $\log_{10}$ of computed flow accumulation area (in m^2) in blue at three different times – **(a, b, c)** 27 ka, **(d, e, f)** 22.9 ka (LGM), and **(g, h, i)** 14.2 ka – based on the hydraulic potential for a flotation factor of **(a, d, g)** $f = 0.6$, **(b, e, h)** $f = 1.0$, and **(c, f, i)** $f = 1.1$. Hillshade topography and ice thickness in red shades at the times indicated are shown in the background. Actual LGM ice extent is shown with orange circles. Modeled ice extent is shown with a brown line. Black outlines indicate locations of overdeepenings.

ter pressures, drainages are no longer constrained by basal topography and can move from one valley to the next over topographic high points such as from the Thur to the Lower Rhine overdeepenings out of the Untersee. High flotation values cause water drainage paths to cut across valleys and overdeepenings, such as, for example, in the Thur Valley, in the Rhine Glacier lobe, and in the Reuss and Linth lobes. In the Alps, water drainage paths are no longer constrained to the valley centers but meander on the sides of the valley and along the valley walls (contrast Fig. 4d and f), even cutting across mountain ranges during retreat (see Fig. 4i).

Basal conditions where the value of flotation is equal to 1 or greater are unlikely to occur over the entire glacier basal surface and for significant lengths of time. High basal water pressure arising from increased surface melting would promote the development of an efficient drainage system that would tend to reduce basal water pressure (e.g., Bartholomaus et al., 2008; Sundal et al., 2011; Andrews et al., 2014). These basal conditions, despite their limited duration, could have the potential to produce significant changes in the basal geomorphology, for example, because of substantial increases in the capacity of subglacial water streams to transport sediments and in their ability to further erode bedrock by processes of abrasion and quarrying associated with both larger fluctuations in basal water pressure and increased basal sliding speed (e.g., van de Wal et al., 2008; Schoof, 2010; Bartholomew et al., 2010; Hewitt, 2013).

The time integration of subglacial water flow paths over the 12.8 kyr of the simulation covering advance and retreat of the Rhine Glacier around the LGM is shown in Fig. 5 for the three values of flotation: $f = 0.6$, 1, and 1.1.

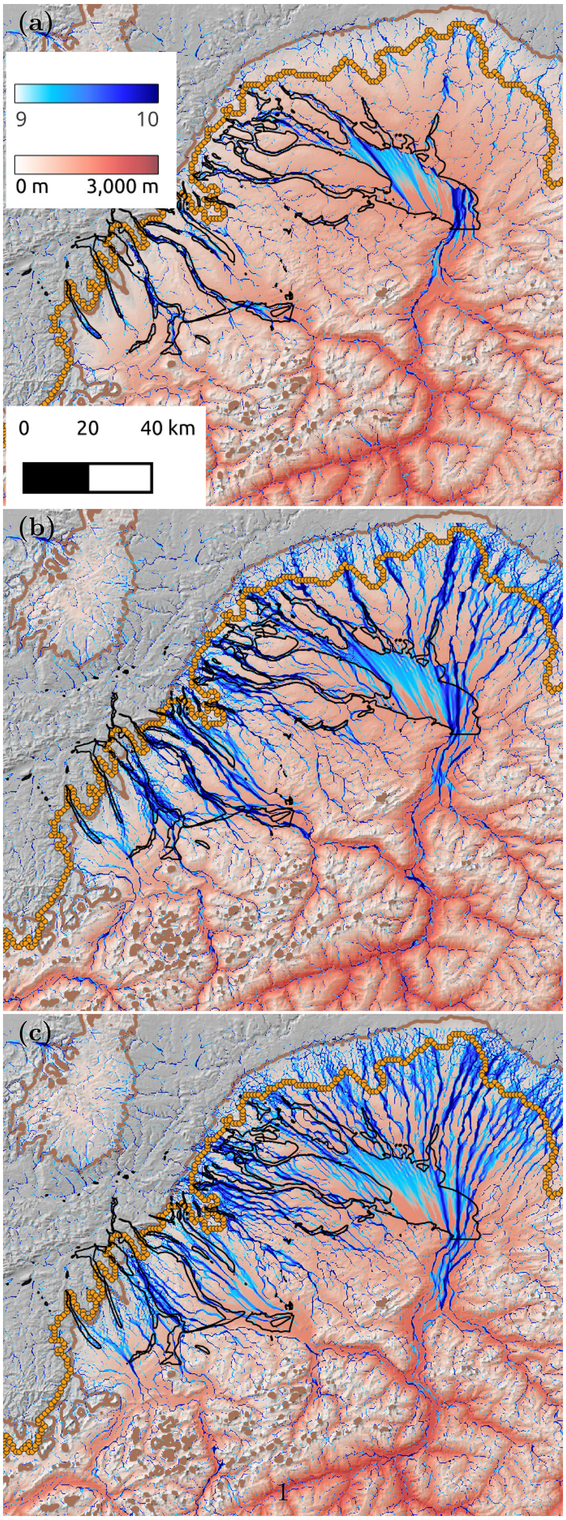

Figure 5. The $\log_{10}$ of the time-integrated flow accumulation area (in m^2) over 12.8 kyr (in blue) for **(a)** $f = 0.6$, **(b)** $f = 1$, and $f = 1.1$ with hillshade topography and ice thickness at the LGM (red colors) in the background. LGM ice extent is shown with orange circles. Model ice extent is shown with a brown line. Black outlines indicate location of overdeepenings.

Clear differences exist between a moderate flotation value ($f = 0.6$) and flotation values with zero effective pressure ($f = 1$) or where water pressure exceeds ice overburden pressure ($f = 1.1$). Although high water pressures in excess of hydrostatic pressure have been noted in both Alpine (e.g., Flowers and Clarke, 1999) and Greenland (e.g., Banwell et al., 2013; Lindbäck et al., 2015) settings and may be occurring in overdeepenings (Lawson et al., 1998), they may not be representative of long-timescale average values in large areas of the Rhine Glacier lobe. Subglacial water flow drainages with high water pressure bypass some topographic relief, producing near-straight channels that cut across relief, such as in the northern part of the Rhine Glacier lobe. These high-water-pressure subglacial drainage paths also cut across overdeepenings in the Thur, Glatt, and Limmat overdeepenings in the Rhine Glacier and Linth Glacier lobes. A value of f equal to unity yields the closest match with the actual mapping of overdeepenings, although many overdeepenings are also well matched for $f = 0.6$, particularly in the Linth Glacier lobe and the lower Thur Valley. Overdeepened regions north of Lake Constance require higher flotation values to be occupied by a subglacial water drainage path. When water pressure exceeds ice overburden pressure such as when $f = 1.1$, subglacial drainages are too distributed and no longer only conform to valleys, a situation that may occur during periods of significant melt (such as during retreat) but is likely not representative of the situation over longer timescales.

High values of the time-integrated flow accumulation area represent drainage pathways with large fluxes of water. Many of these high-discharge zones correspond to several locations of overdeepenings such as the Schussen, Thur, Limmat, Reuss, and Aare overdeepened valleys. In the Aare Valley, a tunnel valley system has been identified beneath the city of Bern (Dürst Stucki and Schlunegger, 2013; Reber and Schlunegger, 2016) with a branching pattern that resembles the pattern observed in our simulation. The presence of this tunnel valley system has been associated with an area of high water pressure. Our simulation also indicates that a flotation value of 1 is necessary for significant subglacial water drainage to occur in this area.

Figure 6 shows patterns of erosion for quarrying, abrasion, and subglacial water flow. All values of erosion have been scaled to between 0 and 1 by dividing the computed values by their respective maximum value. All erosion types have been time-integrated over the course of the advance and retreat of the ice during the LGM.

A clear difference in the erosion pattern emerges between erosion due to ice motion (quarrying and abrasion) and erosion due to subglacial meltwater flow. Erosion by ice motion is greatest at the base of alpine valleys where sliding is fastest and where the ice occupation time is longest. Limited durations of ice cover and lower sliding speeds near the margin limit erosion by ice motion there. Erosion by quarrying (Fig. 6a) shows more variability in the Rhine, Linth, and

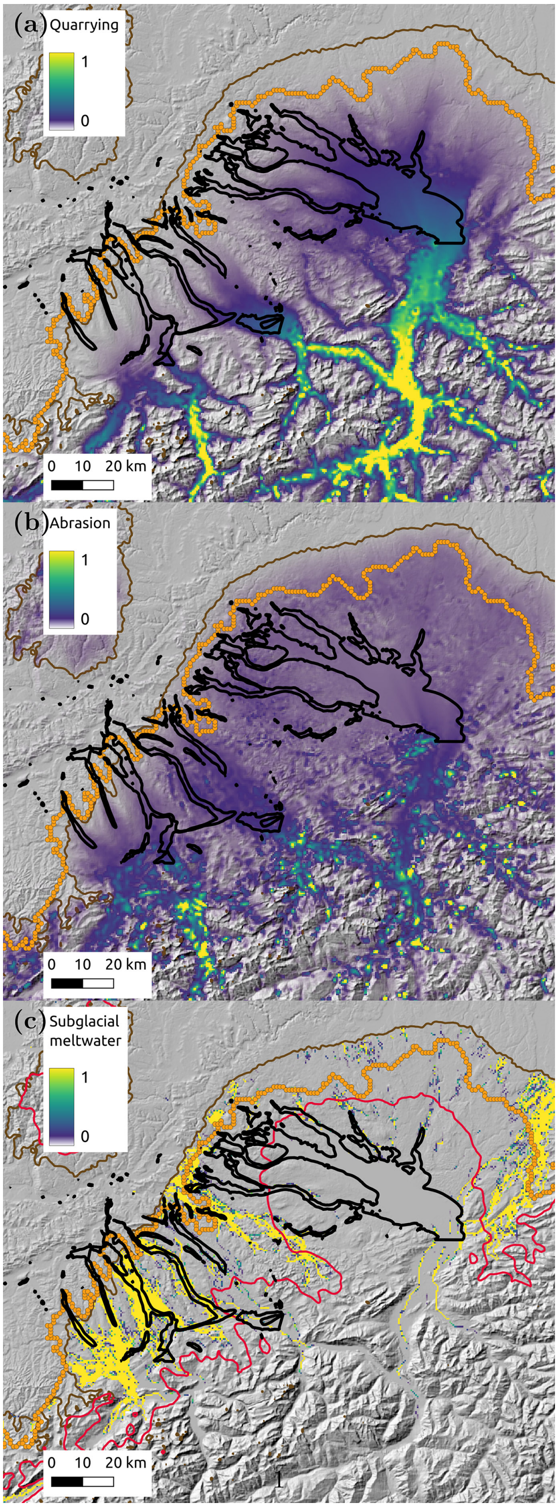

Figure 6. Normalized erosion by **(a)** quarrying, **(b)** abrasion, and **(c)** subglacial meltwater flow. Background shows hillshade topography. LGM margin is shown with orange circles; modeled maximum LGM extent is shown with a brown curve. Black curves indicate areas of overdeepenings. The red curve in **(c)** is the equilibrium line at the LGM separating accumulation (south) from ablation (north) in the Alps.

Reuss lobes owing to its cubic dependence on the ice thickness via the effective pressure term (see Eq. 11). Any variation in ice thickness along or across overdeepenings amplifies erosion by quarrying, resulting in a significant change in the total amount of quarrying in these areas. In contrast, erosion by abrasion (Fig. 6b), which only depends on the square of the sliding speed (see Eq. 11), is more uniform in these lobes. Erosion by subglacial meltwater flow (Fig. 6c) is significantly different: it is almost entirely focused near the ice margin, particularly to the northwest in the Reuss and Linth lobes, the Thur Valley, and the Argen lobe (see Fig. 2). This is due to the large available surface melt in these areas that have larger ablation areas. This is visible by comparing, for example, the position of the equilibrium line (Fig. 6c, red line) to the LGM margin (Fig. 6c, brown line). In the northwestern lobes and in the northeastern part of the Rhine lobe, the area for available surface meltwater is significantly larger then in the main Rhine lobe, explaining why most erosion by subglacial meltwater occurs preferentially there. Also, channel shear stress is more likely to exceed critical shear stress in areas where subglacial channel velocity is higher (see u_c in Eq. 8), which occurs where more surface melt is available and thus increases the potential for erosion by subglacial meltwater.

The different erosion patterns due to ice motion and subglacial meltwater suggest that the existence of distal-foreland overdeepenings in the Swiss lowlands is driven, at least in part, by erosion by subglacial meltwater. Furthermore, high water flux during high-water-pressure events along these valleys and the associated fluctuations in water pressure in the channels could have increased the rate of glacial erosion by quarrying (e.g., Cohen et al., 2006), a dependence that is not included in the simple quarrying erosion law in Eq. (11). This may have been further enhanced by pre-existing rock weaknesses due, for example, to a higher density of joints or faulting (e.g., Dühnforth et al., 2010; Hooyer et al., 2012) such as is found in the Lake Constance area (Fabbri et al., 2021). The patterns of subglacial water drainage paths computed from numerical simulations of ice flow and ice surface elevation indicate a possible link between the subglacial water flux and the existence of overdeepenings. The location of these overdeepenings suggests that these overdeepenings are controlled, in part, by higher rock erodibility in these areas, which, during glacial times, were preferentially excavated by the combined effects of water and ice.

5 Conclusions

We present a model of subglacial water drainages underneath the Rhine Glacier during the advance and retreat around the LGM. The model is based on high-resolution ice surfaces obtained from a numerical ice-flow model. The model computes ice velocities and ice surface elevations using a thermomechanical Stokes flow model coupled to a high-resolution regional climate model to give the most plausible representation of the Rhine Glacier during its advance and retreat around the LGM. The hydraulic potential (or head) calculated from transient elevations of the ice surface together with the basal topography yields a map of subglacial water flux during the LGM. Results of the calculation show that subglacial water drainage is strongly constrained by topography for low values of water pressure (small flotation factor), focusing water flow in valleys in the foreland. When water pressure equals ice overburden pressure (flotation value equal to 1), the subglacial drainages coincide with regions of intense glacial erosion as observed by numerous overdeepenings there. When water pressure exceeds ice overburden pressure and flotation is greater than 1, the pattern of subglacial drainages changes to straighter channels that bypass valleys, mountain ranges, and overdeepened basins. This suggests that the maximum potential for glacial erosion by subglacial water channels is obtained when water pressure underneath the Rhine Glacier and Linth Glacier lobes were, on average, close to the ice overburden pressure. These conditions favor high water flow volumes and maximize the potential for subglacial water fluctuations needed for enhanced erosion by ice and water. Finally, comparison of the erosion pattern from subglacial meltwater with those from quarrying and abrasion shows the importance of subglacial water channels in the formation of distal overdeepenings in the Swiss lowlands.

Code availability. The software package Elmer FEM version 9 which contains Elmer/Ice can be downloaded via Zenodo, https://doi.org/10.5281/zenodo.7892181 (Ruokolainen et al., 2023).

Data availability. Data are not publicly available as part of the climate data were obtained from another group (references cited in text) and they have not been made available publicly as of now. Other data sources are not publicly available as they were obtained from Nagra.

Author contributions. The research was conceived by DC, UHF, and AL. The ice-flow modeling was performed by DC with help from TZ and GJ. DC carried out hydrological modeling and interpretation. DC wrote the paper with inputs from all co-authors.

Competing interests. The contact author has declared that none of the authors has any competing interests.

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Research article

The past is the key to the future – considering Pleistocene subglacial erosion for the minimum depth of a radioactive waste repository

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Abstract: Erosion during potential future glaciations, especially the incision of deep tunnel valleys, is a major challenge for the long-term safety of a radioactive waste repository. Tunnel valleys are a common feature of formerly glaciated sedimentary basins and were incised by pressurised subglacial meltwater. Besides glaciological conditions, tunnel-valley formation depends strongly on the erodibility and hydraulic conductivity of the substratum. In northern Germany, tunnel valleys formed during the Pleistocene glaciations are widespread and may attain depths of almost 600 m. The Pleistocene record may provide an indication for the potential regional distribution and maximum depth of future glaciogenic erosion. We present a new overview map of the maximum depth of Pleistocene erosion in northern Germany. Depth zones were extracted from the existing data and maps provided by the state geological surveys. Based on the mapped depth zones, the potential for future tunnel-valley formation can be assessed. The map may serve as a base to define a spatially variable additional depth that should be added to the minimum depth of a repository required by legislation.

Kurzfassung: Erosion während möglicher zukünftiger Eiszeiten, insbesondere das Einschneiden tiefer subglazialer Rinnen, ist eine große Herausforderung für die Langzeitsicherheit eines Endlagers für hochradioaktive Abfälle. Subglaziale Rinnen sind in ehemals vergletscherten Sedimentbecken weit verbreitet und wurden durch subglaziales Schmelzwasser unter hohem Druck eingeschnitten. Außer durch glaziologische Faktoren wird die Bildung subglazialer Rinnen stark durch die Erodierbarkeit und die hydraulische Durchlässigkeit des Untergrundes bestimmt. In Norddeutschland sind pleistozäne subglaziale Rinnen weit verbreitet und erreichen Tiefen von fast 600 m. Die pleistozäne Überlieferung kann Hinweise auf die potenzielle regionale Verbreitung und maximale Tiefe zukünftiger glazigener Erosion liefern. Wir stellen eine neue Übersichtskarte der maximalen pleistozänen Erosionstiefe in Norddeutschland vor. Tiefenzonen wurden aus den existierenden Daten und Karten, die von den Staatlichen geologischen Diensten zur Verfügung gestellt wurden, extrahiert. Anhand der kartierten Tiefenzonen kann das Potenzial einer zukünftigen Bildung subglazialer Rinnen abgeschätzt werden.

Die Karte kann als Grundlage zur Festlegung eines räumlich variablen Aufschlags zur gesetzlich vorgeschriebenen Mindesttiefe eines Endlagers dienen.

1 Introduction

Tunnel valleys, although commonly no longer directly perceptible at the surface, are among the most impressive landforms related to former glaciations. Tunnel valleys are incised by pressurised meltwater beneath ice sheets and are an important component of the subglacial hydrological system (Kehew et al., 2012; van der Vegt et al., 2012). As overdeepened erosional landforms, tunnel valleys form independently of the regional base level and commonly reach depths of several hundreds of metres. The deepest Pleistocene tunnel valley in northern Germany, for example, has a depth of 554 m below sea level (b.s.l.; Schulz, 2002; Müller and Obst, 2008).

Subglacial erosion during potential future glaciations is a major challenge for the long-term safety of a repository for radioactive waste in deep geological formations. Future tunnel-valley incision is indirectly referred to as "intense erosion caused by an ice age" in the German Site Selection Act (StandAG, 2017). According to German legislation, the safety of a repository for high-level radioactive waste must be assessed for the next 1 million years (StandAG, 2017). Within this period, up to 10 glaciations could occur that may reach similar maximum extents as the Pleistocene glaciations (Fischer et al., 2021). Therefore, the impact of potential future glaciations on the safety of a repository must be considered. The Site Selection Act requires a minimum depth of the upper boundary of the effective containment zone (ECZ) of 300 m b.s.l. (StandAG, 2017). However, 300 m may be insufficient to safeguard the effective containment zone against the impact of deep subglacial erosion.

Assessing the potential for future tunnel-valley formation and its impact on a repository requires a thorough understanding of the processes and controlling factors of tunnel-valley formation. As the formation of tunnel valleys is controlled not only by geological but also by glaciological and climatic factors (Kehew et al., 2012; van der Vegt et al., 2012; Kirkham et al., 2022), predicting future incision is a major challenge. The dimensions and distribution of Pleistocene tunnel valleys may provide indications for future glaciogenic erosion. In this study, we describe a new approach to assessing the potential for future tunnel-valley formation based on the Pleistocene record. Our new approach is based on the assessment of the zones of maximum depth of the tunnel valleys and aims to recommend a minimum depth of the upper boundary of the effective containment zone (300 m or deeper), considering potential future tunnel-valley formation.

2 Processes of tunnel-valley formation

Tunnel valleys are characterised by undulating basal profiles, abrupt terminations, steep flanks and infills dominated by meltwater deposits (e.g. Kehew et al., 2012; van der Vegt et al., 2012). Various processes have been invoked for tunnel-valley formation, including erosion by meltwater, rivers or glaciers in subglacial or proglacial environments. Erosion by pressurised subglacial meltwater is now generally accepted as being responsible for tunnel-valley formation (Wingfield, 1990; Ó Cofaigh, 1996; Huuse and Lykke-Andersen, 2000b; Kehew et al., 2012; van der Vegt et al., 2012; Kirkham et al., 2022).

Subglacial meltwater conduits are crucial features in the drainage network of ice sheets, as the meltwater volumes are too large to be evacuated by groundwater flow (Piotrowski, 1997; Kehew et al., 2012). According to Clayton et al. (1999), tunnel channels can be defined as subglacial meltwater conduits with dimensions (width and depth) adapted to bankfull discharge. Tunnel channels are characterised by uniform dimensions along their course, lack tributaries and can commonly be linked to specific palaeo-ice-margin positions (Clayton et al., 1999). In contrast, tunnel valleys are higher-order, polyphase features, typically comprising multiple tunnel channels and their infills (Kehew et al., 2012). However, due to the wide variety of tunnel valleys and their infills, there is no single model that can explain all observations. Current models of tunnel-valley formation broadly fall into two groups: (i) erosion by steady-state meltwater discharge or (ii) erosion by sudden outbursts from meltwater reservoirs (cf. Kehew et al., 2012; van der Vegt et al., 2012; Kirkham et al., 2022).

– (i) Tunnel-valley formation by steady-state meltwater discharge is a gradual process that requires the substratum of the ice sheet to be permeable, erodible and poorly lithified (Boulton and Hindmarsh, 1987; Boulton et al., 2009; Kehew et al., 2012; van der Vegt et al., 2012; Kirkham et al., 2022). Incision is caused by downcutting of subglacial tunnel channels that switch laterally in space and time, forming an anastomosing or anabranching pattern near the base of the evolving tunnel valley (Catania and Paola, 2001; Ravier et al., 2015; Kirkham et al., 2022). Modelling results by Kirkham et al. (2022) imply that the meltwater responsible for tunnel-valley formation is mainly derived from melting at the surface of the ice sheet. Tunnel-valley formation is furthermore impacted by the interplay between subglacial channelised meltwater discharge, groundwater flow and remobilisation of the bed (Boulton and Hindmarsh, 1987; Piotrowski et al., 1999; Janszen et al., 2013; Ravier et al., 2015). If the water pressure in the subglacial conduits remains lower than in the substratum, the pressure gradient causes a flow from the substratum into the conduits (Boulton et al., 2009). Creeping of soft sediment into the evolving tunnel valley enlarges incipient subglacial channels (Boulton and Hindmarsh, 1987). If the pressure gradient becomes sufficiently high, fluidisation of the substratum may occur that strongly enhances erosion (Boulton et al., 2009; Janszen et al., 2013). Vice versa, pressurised meltwater within a subglacial conduit may trigger hydrofracturing in the substratum that also leads to sediment remobilisation (Ravier et al., 2015). The water pressure beneath a warm-based ice sheet is highly variable in both space and time (Piotrowski et al., 2004), and zones of increased water pressure are prone to brecciation, liquefaction and fluidisation, allowing for a remobilisation of the subglacial bed and the gradual incision of tunnel valleys (Janszen et al., 2013; Ravier et al., 2014, 2015).

– (ii) Outbursts of large volumes of stored meltwater with high flow velocities and discharge rates are another possible mechanism of tunnel-valley formation. Incision by meltwater outbursts is mainly related to retrograde erosion, causing the rapid enlargement of an initial meltwater pathway (Wingfield, 1990; Hooke and Jennings, 2006). Different models exist on the magnitudes, trigger mechanisms and recurrences of such subglacial outburst floods. Outburst floods may be responsible for the formation of individual tunnel channels (Hooke and Jennings, 2006; Sandersen et al., 2009) or tunnel valleys (Ehlers and Linke, 1989; Wingfield, 1990; Jørgensen and Sandersen, 2006). Meltwater storage in subglacial lakes is controlled by the hydraulic gradient, subglacial topography and the permeability of the substratum (Shreve, 1972; Piotrowski, 1994). The presence of permafrost, which lowers the permeability of the substratum and allows for the accumulation of meltwater reservoirs, is often considered a prerequisite for subglacial meltwater-reservoir formation (Piotrowski, 1994; Hooke and Jennings, 2006). Outbursts of stored meltwater occur if the pressure in the meltwater reservoir exceeds the strength of the impermeable substratum that forms the seal (Hooke and Jennings, 2006). Tunnel channels or valleys formed by meltwater outbursts afterwards serve as low-pressure meltwater conduits (Kehew et al., 2012). Episodic outbursts along the same pathway will eventually form larger tunnel valleys comprising multiple cut-and-fill sequences (Jørgensen and Sandersen, 2006).

Strong arguments exist both supporting and refuting the different models of tunnel-valley formation (cf. Ó Cofaigh, 1996; Kehew et al., 2012; van der Vegt et al., 2012; Kirkham et al., 2022). The widespread occurrence of tunnel valleys, however, suggests the existence of a common mechanism for their formation. In most instances, a formative model comprising quasi-steady-state meltwater discharge in combination with small outbursts is applicable to explain tunnel-valley formation (van der Vegt et al., 2012). In addition to erosion by meltwater, tunnel-valley evolution may be affected by erosion of the moving ice sheet and mass-wasting processes, which may enlarge tunnel valleys or modify their cross-sectional geometries and infills (Prins et al., 2020; Kirkham et al., 2021, 2022).

Tunnel-valley fills may indicate repeated episodes of erosion and deposition, which may relate to a single ice advance or multiple ice advances or extend across multiple glaciations (Piotrowski, 1994; Sandersen et al., 2009; Kirkham et al., 2022). The incision of tunnel valleys by meltwater is reflected by their infills, which are commonly dominated by meltwater deposits. Typically, tunnel-valley fills may be subdivided into synglacial and postglacial deposits (e.g. Piotrowski, 1994; Huuse and Lykke-Andersen, 2000b; Lang et al., 2012; van der Vegt et al., 2012; Janszen et al., 2013). Synglacial deposits comprise subglacial and ice-contact deposits formed during or immediately after incision and proglacial successions deposited during ice-sheet retreat. Meltwater deposits typically display fining-upward trends due to deposition at an increasing distance from the ice margin, with fine-grained glaciomarine or glaciolacustrine deposits forming the uppermost part of the synglacial successions. After deglaciation, remnant tunnel valleys commonly form local depocentres for postglacial marine, lacustrine or fluvial deposition (Ehlers and Linke, 1989; Piotrowski, 1994; Lang et al., 2012; Janszen et al., 2013; Lang et al., 2015; Steinmetz et al., 2015).

3 Controlling factors of tunnel-valley formation

Tunnel-valley formation is controlled by glaciological and geological factors. Glaciological factors include the thickness and temperature of the ice sheet, which control the availability and pressure of the meltwater. Furthermore, tunnel valleys develop broadly parallel to the ice-flow direction. The geological control on tunnel-valley formation is mainly exerted by the erodibility and hydraulic conductivity of the substratum (Ó Cofaigh, 1996; Huuse and Lykke-Andersen, 2000b; Kehew et al., 2012; van der Vegt et al., 2012).

3.1 Substratum control

Tunnel valleys are generally restricted to areas with an easily erodible substratum as the infill of sedimentary basins. In areas characterised by resistant (e.g. crystalline) bedrock, eskers form instead of tunnel valleys (Boulton et al., 2009; Kehew et al., 2012).

In northern central Europe, the highest density of Pleistocene tunnel valleys and the deepest incisions occur in areas with thick, largely unlithified Cenozoic deposits (Hinsch, 1979; Kuster and Meyer, 1979; Ehlers and Linke, 1989; Stackebrandt et al., 2001; Stackebrandt, 2009; Lohrberg et al., 2020; Ottesen et al., 2020). Tunnel valleys, which were incised into more resistant (Mesozoic) rocks, are generally relatively shallow (Stackebrandt, 2009; Sandersen and Jørgensen, 2012). According to Stackebrandt (2009), Pleistocene tunnel valleys in northern Germany are concentrated in the area of the subsiding basin axis of the North German Basin, which was oriented approximately normal to the main ice-advance directions.

Tunnel-valley formation also depends on the hydraulic conductivity of the substratum. Initial subglacial incision commonly occurs if the substratum is unable to drain meltwater by groundwater flow, e.g. in areas of high meltwater production or low-permeability substrata (Piotrowski, 1997; Huuse and Lykke-Andersen, 2000b; Boulton et al., 2009; Kehew et al., 2012; Sandersen and Jørgensen, 2012). A layer-cake stratigraphy of high- and low-permeability strata favours the build-up of high pressures in subglacial aquifers, allowing for remobilisation of the sediment at the margins of tunnel valleys (Janszen et al., 2013; Ravier et al., 2014; Ravier et al., 2015).

3.2 Structural control

Faulting modifies the hydraulic conductivity and resistance to erosion of the affected rocks, thus creating structural weaknesses, which may act as preferential pathways of tunnel-valley incision (Wenau and Alves, 2020; Sandersen and Jørgensen, 2022). Sandersen and Jørgensen (2022) demonstrated a close correlation between the orientations of faults and tunnel valleys in Denmark, suggesting that the locations of many tunnel valleys are fault controlled. However, there are few studies showing a clear connection between faults and the incision of tunnel valleys (e.g. Al Hseinat et al., 2016; Wenau and Alves, 2020; Brandes et al., 2022). Furthermore, Stackebrandt (2009) suggested that also the terminations of tunnel valleys might be controlled by (neotectonically active) faults based on the observation of the simultaneous beginning and ending of parallel tunnel valleys.

Different interpretations exist on the impact of salt structures on tunnel-valley formation, based on observations that tunnel valleys may cross-cut, be parallel to or evade salt structures. Some tunnel valleys cut across salt structures, crestal faults and rim synclines without any perceptible interrelation (Grube, 1983; Ehlers and Linke, 1989; Sonntag and Lippstreu, 2010). In other cases, tunnel-valley incision is interpreted as having exploited pre-existing weaknesses related to salt structures, such as crestal faults or depressions related to the subsurface dissolution of evaporites (Kuster and Meyer, 1979; Ehlers and Linke, 1989; Huuse and Lykke-Andersen, 2000b; Lang et al., 2014; Wenau and Alves, 2020). Based on the interpretation of 3D seismic data from the southern North Sea, Wenau and Alves (2020) demonstrated that some tunnel valleys follow the trend of underlying salt structures and their crestal graben faults. In contrast, Kristensen et al. (2007) mapped tunnel valleys that circle around salt diapirs. Such changes in tunnel-valley trends may be explained by the uplift of more resistant sedimentary rocks at the flanks of salt structures (Kuster and Meyer, 1979) or by hydrogeological changes near salt structures (Piotrowski, 1997).

4 Pleistocene tunnel valleys in northern Germany

Northern Germany is one of the classic areas for tunnel-valley research, and tunnel valleys are a common feature in the areas that were covered by the Pleistocene ice sheets. Three major glaciations referred to as the Middle Pleistocene Elsterian and Saalian and the Late Pleistocene Weichselian glaciations are well documented (e.g. Litt et al., 2007; Ehlers et al., 2011). Pleistocene tunnel valleys in northern Germany are several hundred metres deep; are several hundred metres to a few kilometres, in extreme cases 8–12 km, wide; and can be longer than 100 km (e.g. Kuster and Meyer, 1979; Ehlers and Linke, 1989; Smed, 1998; Stackebrandt, 2009; Gegg and Preusser, 2023, and references therein).

Deep tunnel valleys characterise the base of the Quaternary across northern Germany and are generally interpreted as having formed during the Elsterian glaciation (Grube, 1979; Hinsch, 1979; Kuster and Meyer, 1979; Ehlers and Linke, 1989; Stackebrandt, 2009). The major trend of the tunnel valleys is north–south in northwestern Germany and northeast–southwest in northeastern Germany, which is generally interpreted as pointing to different ice-advance directions and probably a temporal change in the ice-advance directions (Ehlers and Linke, 1989; Stackebrandt, 2009). However, the tunnel valleys form a complex net-like anastomosing pattern and the controls on tunnel-valley orientations may be more complex. Major tunnel valleys display a regular spacing of 25 to 30 km (Stackebrandt, 2009). The deepest mapped Elsterian tunnel valley has a maximum depth of 554 m b.s.l. (Schulz, 2002; Müller and Obst, 2008). The deep ($>$ 200 m) Elsterian tunnel valleys were formed at a distance of at least 100 km inside the former ice margin. Shallower ($<$ 100 m) Elsterian tunnel valleys are known from areas close to the former ice-marginal position (e.g. Eissmann, 2002; Lang et al., 2012).

In contrast to the many and deep Elsterian tunnel valleys, Saalian tunnel valleys are generally considered rare (Passchier et al., 2010). The few tunnel valleys that have been attributed to the Saalian glaciation are commonly isolated and shallow ($<$ 100 m) features (e.g. Grube, 1979; Piotrowski, 1994; Piotrowski et al., 1999). Passchier et al. (2010) suggested that the lack of Saalian tunnel valleys may be related to different glaciological and hydrogeological conditions compared with the Elsterian glaciation. However, the attribution of the majority of tunnel valleys to the Elsterian glaciation is almost entirely based on lithostratigraphy.

The Weichselian glaciation had a lesser maximum extent than the Middle Pleistocene glaciations, covering only parts of northern Germany (Ehlers et al., 2011). Weichselian tunnel valleys are shallower (< 100 m), shorter and narrower than their Elsterian counterparts (Smed, 1998; Stackebrandt, 2009). Weichselian tunnel valleys are visible in the present-day landscape as narrow depressions, commonly leading to the formation of lakes. In some instances, Weichselian tunnel valleys can be linked to palaeo-ice-marginal positions and outwash fans (Smed, 1998; Jørgensen and Sandersen, 2006).

5 Data

As a database for this study, maps and models showing the base of the Quaternary provided by the state geological surveys of Schleswig-Holstein, Hamburg, Bremen, Lower Saxony, Saxony-Anhalt, Mecklenburg-Western Pomerania, Brandenburg and Berlin were used (Table 1). Most of the maps showing the base of the Quaternary were provided as line shapefiles by the state geological surveys.

Currently, all northern German state geological surveys are in the process of revising their base Quaternary maps or creating 3D models of the medium-depth subsurface. The results of new boreholes and geophysical investigations and their interpretations will be used to update the maps of the base Quaternary and adapt them to the current state of research. However, the progress and planned completion dates of the individual federal states vary, so it was decided to work with the data that were available at the starting point of this project.

The base Quaternary maps were used as the main input data sets. The base of the Quaternary actually represents a diachronous basal surface of glaciogenic erosion, which can mostly be attributed to the Middle Pleistocene ice advances. As the focus of this project is on the areas with the deepest erosion, the investigations concentrate on the mapped tunnel valleys. For this purpose, the tunnel valleys were extracted from the respective base Quaternary maps of the federal states.

For data preparation, ArcMap (version 10.8.1) and ArcGIS Pro (version 2.9.1) were used. Merging the depth-contour maps of the individual federal states into a common map proved a major challenge. The use of different data sets, resolutions and mapping approaches led to major discontinuities at the state borders. Without data harmonisation and/or reinterpretation of the border areas, the maps could not be merged. However, such a reinterpretation was not possible within the time frame of our project. Therefore, the data for each federal state were considered individually in the first step.

6 Methods

The focus of this study was on the evaluation of areas with deep tunnel valleys. Therefore, the areas without clearly defined tunnel valleys and with an erosion depth of less than 100 m were removed from the data sets. After this manual processing, only the deep tunnel valleys remain visible on the map (Fig. 1).

To further reduce the complexity of the base Quaternary maps, a geographic information system (GIS) workflow was developed to extract the thalweg lines of the tunnel valleys. The GIS workflow made it possible to extract the deepest points along the tunnel valleys. Details on the GIS workflow are provided in the Supplement (Figs. S1 and S2). Based on the map of the tunnel-valley thalwegs and their depths, zones of similar maximum tunnel-valley depths were defined.

7 Results

Removing the areas from the base Quaternary maps that are shallower than 100 m b.s.l. provides a comprehensive overview of the Pleistocene tunnel-valley network (Fig. 1). The deepest parts and southern terminations of the tunnel valleys are both aligned in an approximately northwest–southeast-trending zone.

The extracted thalweg map allows for an easier recognition of individual tunnel valleys and their respective depths (Fig. 2). As each wide tunnel valley is now represented by a line of defined depth (Fig. 2), it becomes easier to identify zones of similar maximum depth (Fig. 3). Furthermore, the thalweg lines will be intersected with other geological structures and units to analyse their relationships and potential impact on the formation of the tunnel valleys in a subsequent step of our project. The visual analysis of the data showed that the tunnel valleys could be divided into five depth zones: no tunnel valley deeper than 100 m, up to 200 m, up to 300 m, up to 400 m and up to 600 m (Fig. 3), with the zone "no tunnel valley deeper than 100 m" extending to the maximum ice margin. The transitions between the depth zones are always placed at the terminations of the respective tunnel-valley sections and therefore occur abruptly. The depth zones of the tunnel valleys display a clear northwest–southeast-trending pattern. The deepest zone occurs in an area approximately between Hamburg and Berlin (Fig. 3). Towards the northeast and southwest, the maximum depths decrease successively.

8 Discussion

We developed an effective and timesaving method to extract an overview map of the Pleistocene tunnel-valley network from the available data sets. The focus of the new map is on the distribution of the maximum depths rather than on individual tunnel valleys and their geometries.

Table 1. The input data (maps and models of the base of the Quaternary) for each federal state, with data type, scale, year of publication and reference, as far as this information is known. "/" denotes no scale.

	Data type	Scale	Publication date	Reference
Schleswig-Holstein (SH)	line shapefile	1 : 200 000	2016	provided by the geological survey
Hamburg (HH)	surface (.ts)	/	2018	provided by the geological survey
Bremen (HB)	surface (.ts)	/	2016	provided by the geological survey
Lower Saxony (NDS)	line shapefile	1 : 500 000	2011	provided by the geological survey
Mecklenburg-Western Pomerania (MVP)	line shapefile	unknown	2002	Brückner-Röhling et al. (2002)
Saxony-Anhalt (SA)	line shapefile	1 : 50 000	1993–2014	provided by the geological survey
Berlin and Brandenburg (BB)	point shapefile	1 : 1 000 000	2010	Noack et al. (2010)

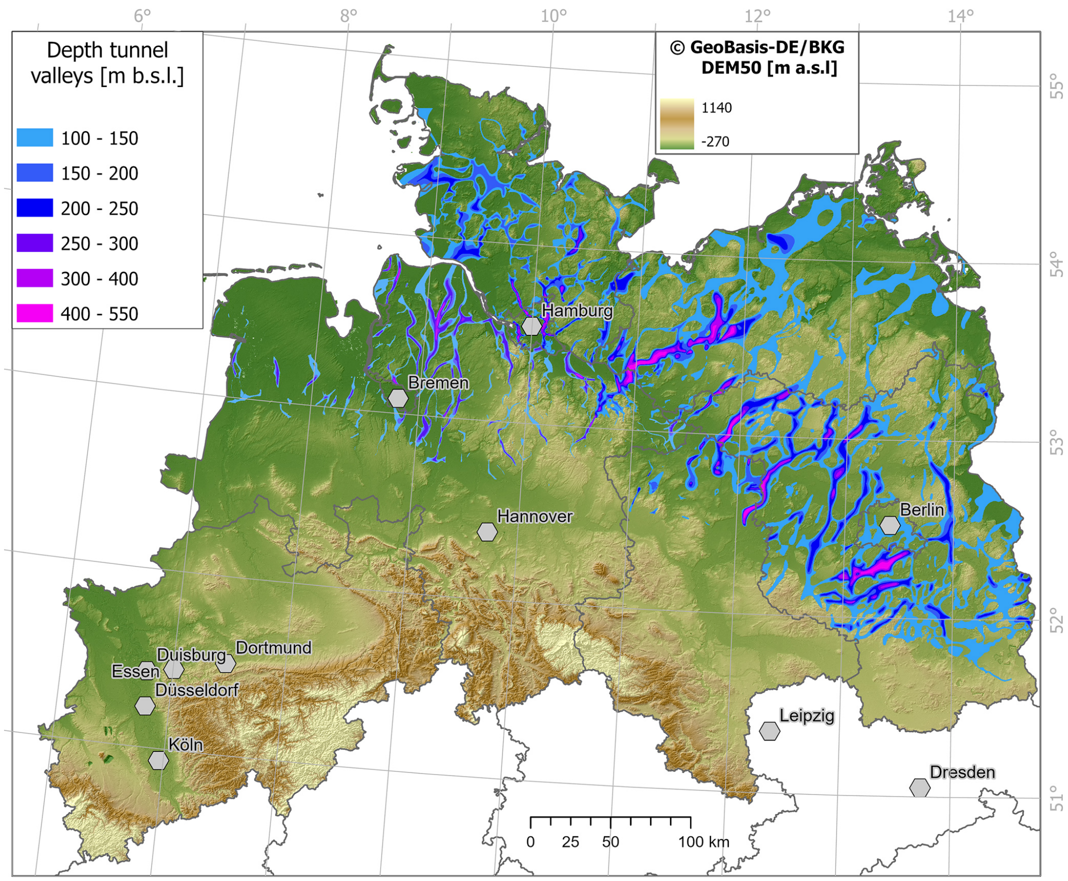

Figure 1. Map showing the digital elevation model (DEM) of northern Germany with a cell size of 50 m (© GeoBasis-DE/BKG) and a raster of the tunnel valleys with a cell size of 100 m extracted from the base Quaternary data sets.

By calculating the thalweg lines, the map of the distribution and maximum depths of the tunnel valleys becomes much clearer and focussed on the major incisions (Fig. 2). When interpreting the course of the thalweg lines, however, it is important to bear in mind that the "Flow Accumulation" tool is a hydrological tool that always searches for connections downslope. Although tunnel valleys represent former drainage pathways, the tunnel-valley floors display undulating morphologies. This undulating morphology is characteristic of tunnel valleys and caused by the flow of pressurised water below an ice sheet (Kehew et al., 2012; van der Vegt et al., 2012). The derived thalweg maps represent a simplification of the complex tunnel-valley network. Therefore, the map of the thalweg lines should always be considered in combination with contoured depth maps, as otherwise the continuity of the tunnel valleys is underestimated.

The ultimate aim of our research is to assess the long-term safety of a repository for highly radioactive waste and, in particular, effects on the integrity of the effective containment zone (ECZ). The deepest tunnel valleys attain depths

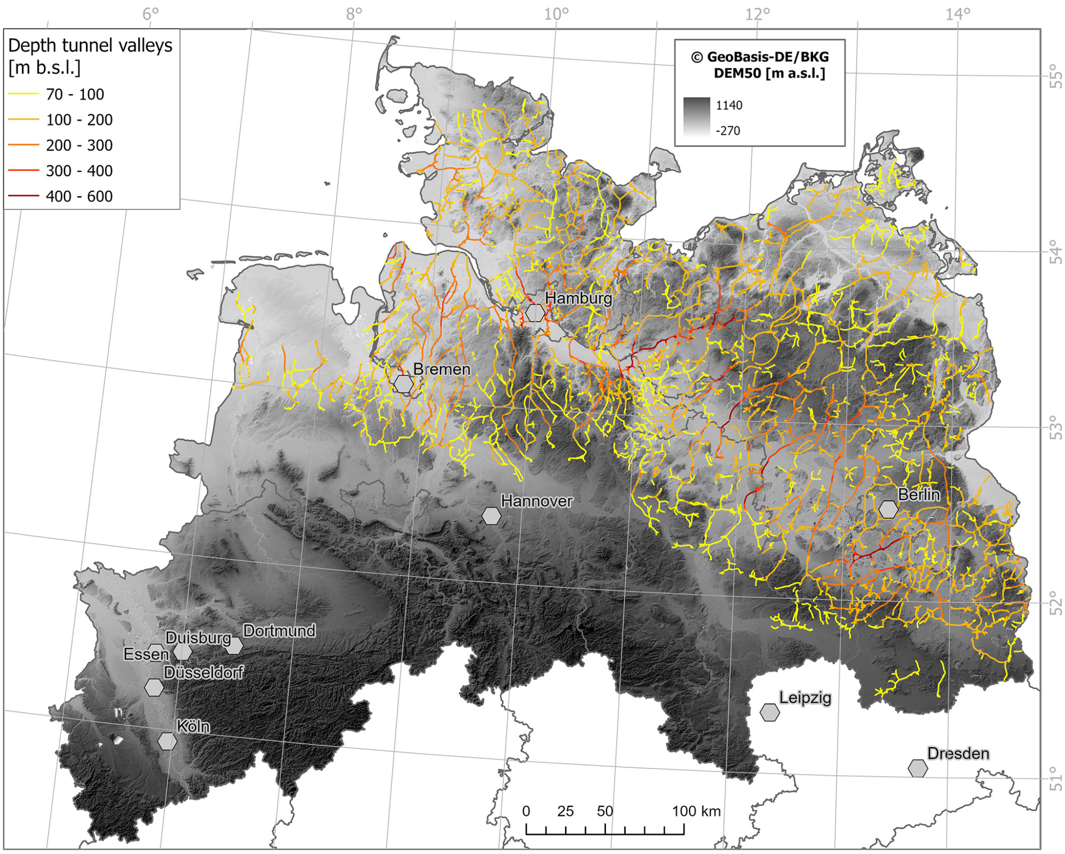

Figure 2. Map showing the digital elevation model (DEM) of northern Germany with a cell size of 50 m (© GeoBasis-DE/BKG) and the result of the thalweg extraction. The depths of the thalweg lines are colour-coded.

of more than 500 m (Fig. 3). Potential future erosion with similar depths may thus pose a risk to the ECZ. However, legislation only stipulates a required minimum depth of the top of the ECZ of 300 m b.s.l. (StandAG, 2017).

The depths of Pleistocene tunnel valleys clearly exceed the legally required minimum depth of 300 m. To minimise the risk posed by future tunnel-valley formation, an additional depth might be added to the minimum depth requirement. The mapped depth zones can be used to guide the definition of such an additional depth (Figs. 4 and 5).

The block diagram (Fig. 4) represents a schematic cross section across the Pleistocene tunnel valleys of northern Germany. Within the block diagram the geological bedrock is not further divided and therefore consists only of preglacial rocks and syn- and postglacial deposits. A dashed black line indicates the legally required minimum depth (A) of the repository (300 m). The dashed red line represents the envelope of the tunnel-valley bases. Deep tunnel valleys (> 100 m) only occur near the maximum extent of the Pleistocene ice sheets. The greater the distance from the maximum ice-sheet extent, the greater the number and depth of the tunnel valleys. The calculated depth zones of the tunnel valleys (Fig. 3) are shown in the block diagram. Since the deepest tunnel valleys in northern Germany have only been formed in a spatially limited area, it is postulated that this will also be the case in the future. In the central area with the greatest tunnel-valley depths (up to 600 m), the final depth of the repository must be significantly below the depth of the tunnel valleys.

It has long been observed that tunnel valleys tend to form in certain areas that provide favourable conditions (e.g. Boulton et al., 2009; Stackebrandt, 2009; Kehew et al., 2012). Considering the depth and distribution of the tunnel valleys together with the pre-Quaternary subcrop map (i.e. a map without the Quaternary deposits; Fig. 5), it becomes clear that the majority of, and deepest, tunnel valleys occur in the central area of the basin, where poorly lithified and easily erodible Neogene sediments of the Pliocene and Miocene underlie the Quaternary deposits. In northeastern Mecklenburg-Western Pomerania, where more resistant Jurassic and Cretaceous rocks underlie the Quaternary deposits, significantly fewer and shallower tunnel valleys (Fig. 5) are present. Figure 6 shows the depth to near the base of the Cenozoic succession (top Danian Chalk Group; Doornenbal and Stevenson, 2010) together with the tunnel-valley thalwegs. For clarity, only thalweg depths greater than 200 m are shown. The

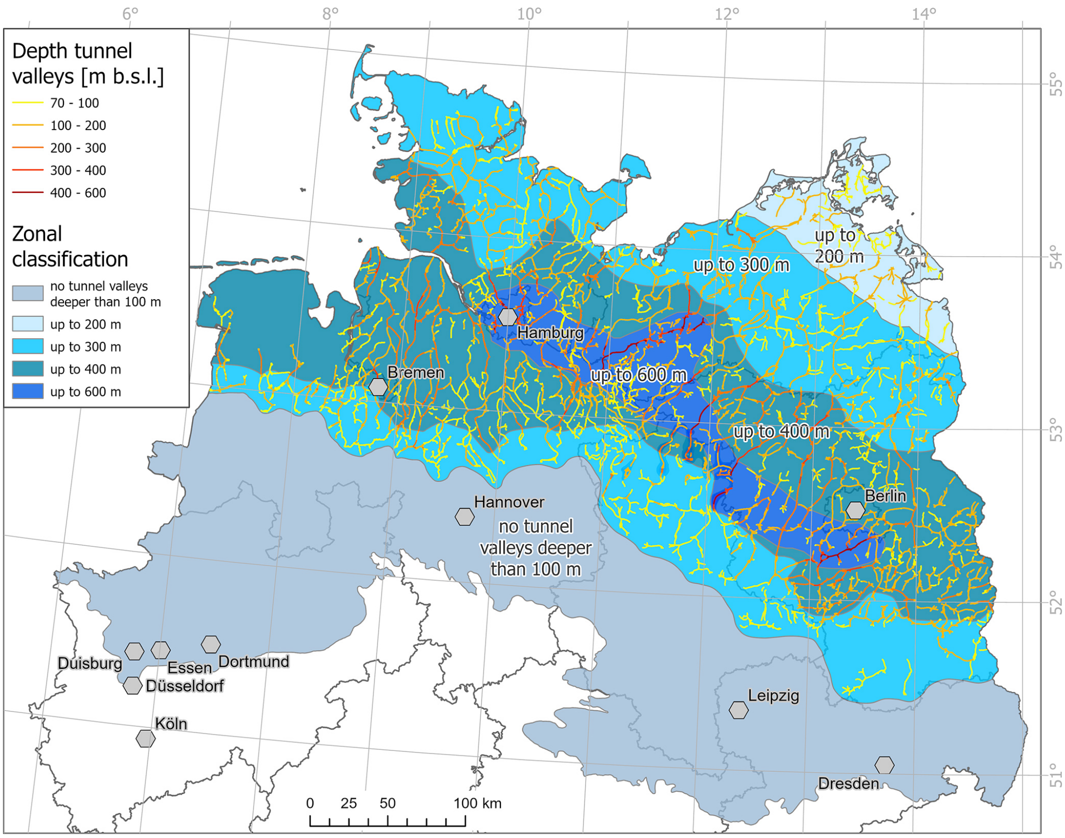

Figure 3. Map showing the analysed depth zones of the tunnel valleys defined from the spatial distribution of the thalweg depths. Five different depth zones were defined.

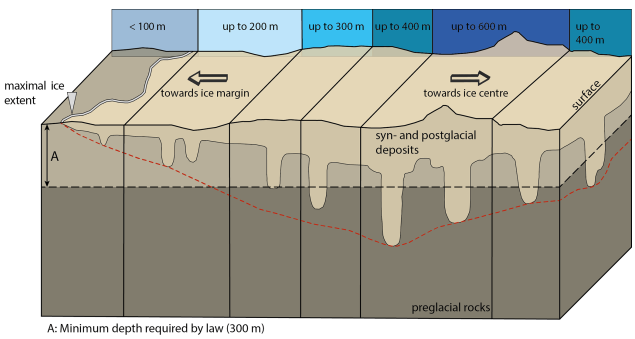

Figure 4. Block diagram showing a schematic section across the five zones of tunnel-valley depths.

map can be regarded as an approximate thickness map of the Cenozoic succession.

The greatest depths of the tunnel-valley thalwegs are reached in the central basin area where the thickness of the Cenozoic sediments is highest. The location of the basin centre correlates with the majority of the deep tunnel valleys between 200 and 558 m b.s.l. (Fig. 6). The Palaeogene and Neogene units consist mainly of poorly lithified sands, silts and clays (Doornenbal and Stevenson, 2010). The greater erodibility of the poorly lithified sediments most likely favoured deep tunnel-valley incision (cf. Stackebrandt, 2009; Kehew et al., 2012).

Deep glaciogenic erosion is also caused by the formation of basins associated with glaciotectonic thrust complexes. Basins are formed beneath the ice sheet where thrust sheets are removed and transported to the ice margin to form glaciotectonic ridges (van der Wateren, 1995; Huuse and Lykke-Andersen, 2000a; Aber and Ber, 2007). Individual thrust sheets can be up to 250 m thick and can be transported for several kilometres (Kupetz, 1997; Huuse and Lykke-Andersen, 2000a; Winsemann et al., 2020). The depth of the basins and the thickness of the thrust sheets are controlled by the depth of the detachment, which typically consists of mechanically weak beds of clay, chalk or organic-rich sediments (Huuse and Lykke-Andersen, 2000a; Andersen et al., 2005; Aber and Ber, 2007). Detachments may occur in depths of up to 350 m (Huuse and Lykke-Andersen, 2000a; Aber and Ber, 2007). In northern Germany, glaciotectonic complexes with associated basins have formed during all major ice advances (Meyer, 1987; Van Der Wateren, 1995; Kupetz, 1997; Winsemann et al., 2020; Gehrmann et al., 2022). Glaciotectonic basins attain maximum depths between 120 and 290 m (Kupetz, 1997; Winsemann et al., 2020). However, as these

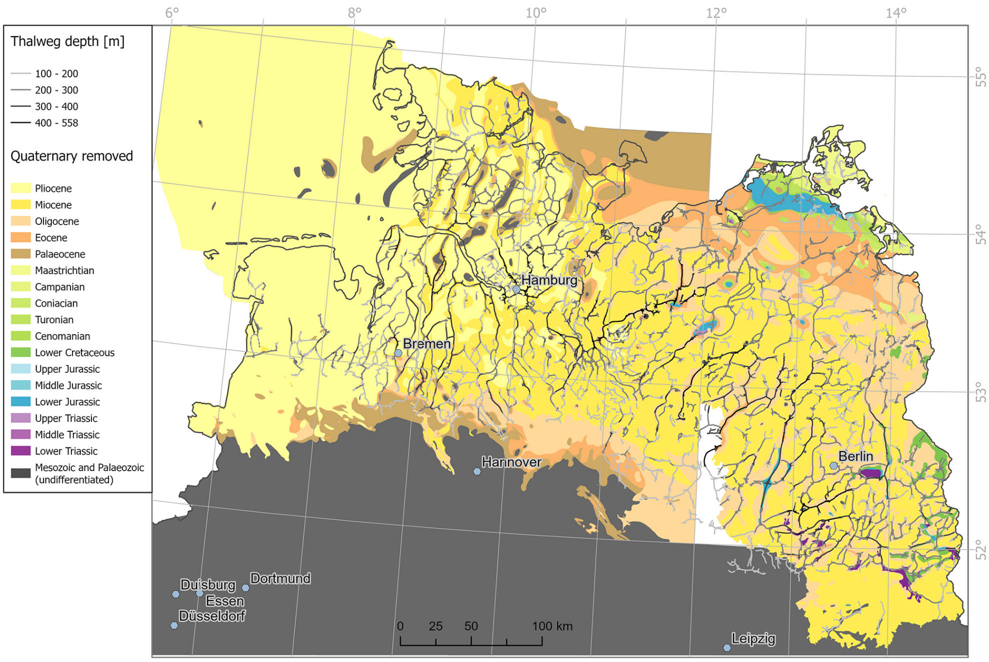

Figure 5. Pre-Quaternary subcrop map (i.e. a map without the Quaternary deposits), which is a combination of the pre-Quaternary subcrop map of the Gorleben project (Brückner-Röhling et al., 2002) and the pre-Quaternary subcrop map of Mecklenburg-Western Pomerania at 1 : 500 000 (Schütze, 2006) and the pre-Quaternary subcrop map of Brandenburg from the *Atlas zur Geologie von Brandenburg* (Stackebrandt et al., 2010) at 1 : 1 000 000.

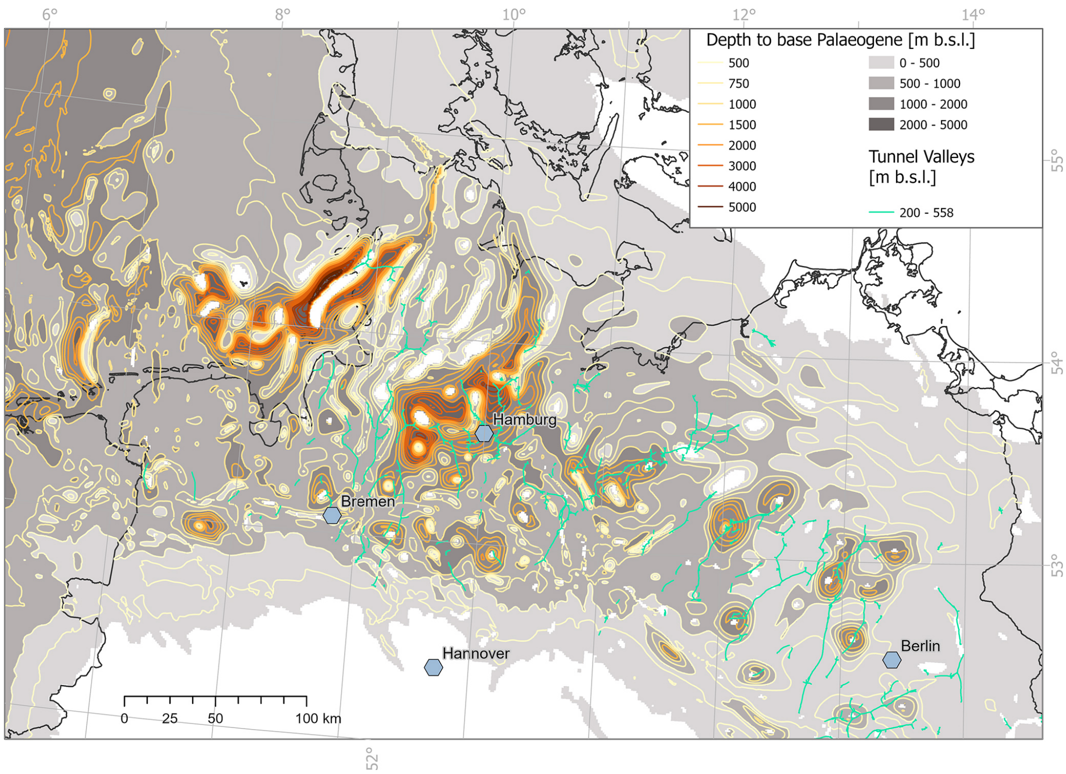

Figure 6. Map showing the near base of the Cenozoic (top Danian Chalk Group) derived from the Southern Permian Basin Atlas (Doornenbal and Stevenson, 2010). The green lines show the thalweg lines of the tunnel valleys with depths ranging from 200 to 558 m.

basins are shallower than many tunnel valleys, they are not clearly visible on most maps of the base Quaternary.

Predicting future tunnel-valley formation is extremely challenging due to the complex glaciological and geological factors controlling tunnel-valley formation. The analysis of the Pleistocene record provides valuable insights into the patterns of tunnel-valley formation that will apply to the future processes. It seems plausible that areas where deep incision occurred during the Pleistocene glaciations may also be affected by future glaciogenic erosion. The favourable conditions for tunnel-valley formation are related to the overall regional geological setting, e.g. the setting in a sedimentary basin filled by poorly lithified and erodible deposits. The overall geological setting is unlikely to undergo major changes in the next 100 000 to 1 000 000 years. However, glaciations will strongly modify the morphology and near-surface geology. Therefore, in the case of multiple future glaciations, an assessment of potential tunnel-valley formation becomes even more challenging. Furthermore, multiple occupations of tunnel valleys may occur (e.g. Piotrowski, 1994; Jørgensen and Sandersen, 2006) and a reactivation of tunnel valleys during future glaciations cannot be ruled out.

For the recommendation of the depth zones, several assumptions need to be made with regard to (i) the available data of the study area, (ii) maximum tunnel-valley depths and (iii) the extent of future glaciations. The study area is sufficiently well explored, and maps of the existing tunnel valleys are available from the state geological surveys and other publications. The source data are very heterogeneous, with regional and local variations, as the published maps of the base Quaternary have been under constant revision since the 1970s. The overall density of data in the study area is very high, and we can assume that no major structures have been overlooked that would change the observed broad trends. However, new details of the Pleistocene tunnel-valley network are still being revealed by ongoing mapping (e.g. Hese et al., 2021; Bruns et al., 2022; Lohrberg et al., 2022a, b). With maximum depths of 554 m b.s.l., the tunnel valleys in northern Germany are among the deepest known tunnel valleys (Schulz, 2002; Müller and Obst, 2008; Stackebrandt, 2009). Similar extreme depths are known from the Netherlands (ten Veen, 2015) and the continental shelf off the east coast of North America (Macrae and Christians, 2013). However, the vast majority of tunnel valleys are shallower than 400 m (Kehew et al., 2012; van der Vegt et al., 2012; Ottesen et al., 2020). Therefore, very deep tunnel valleys can be considered extreme examples formed under favourable conditions. Future research should investigate these favourable conditions in more detail. A similar assumption applies to the potential extent of future glaciations, where the well-studied Pleistocene ice-sheet extents (e.g. Ehlers et al., 2011; Batchelor et al., 2019) serve as analogues for probable future glacial limits.

9 Conclusions

Subglacial tunnel valleys are among the deepest erosional landforms and are ubiquitous features of formerly glaciated sedimentary basins. Tunnel valleys are incised by pressurised meltwater and are thus formed independently of any regional base level. In addition to glaciological factors, tunnel-valley formation is strongly controlled by the geology of the substratum. The main geological controls are the resistance to erosion and the hydraulic conductivity.

Tunnel-valley formation during potential future glaciations needs to be included in the long-term safety assessment of radioactive waste repositories. As prediction of the location of future tunnel valleys is a major challenge, the new depth-zonation map provides a straightforward approach to assessing the potential for tunnel-valley formation based on the Pleistocene record. When recommending the minimum depth for a repository, the geology of the substratum should also be considered and its potential impact on tunnel-valley formation assessed.

Data availability. The data that support the findings of this study are available from the corresponding author upon reasonable request.

Supplement. The supplement related to this article is available online at: https://doi.org/10.5194/egqsj-72-113-2023-supplement.

Author contributions. Concept and design of the study: SB and JL; data compilation and analysis: SB; interpretation and discussion of the results: SB, AB, VN and JL; first draft of manuscript and figures: SB and JL. All authors approved the submitted version.

Competing interests. The contact author has declared that none of the authors has any competing interests.

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Express report

Expected and deviating evolutions in representative preliminary safety assessments – a focus on glacial tunnel valleys

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1 Introduction

Germany gave a fresh start to its site selection procedure for a repository for high-level radioactive waste with the Repository Site Selection Act (StandAG, 2017) in 2017. The procedure consists of three phases; Phase I is split into two steps. BGE published the Sub-areas Interim Report in 2020 to conclude Step 1 (BGE, 2020). Ninety areas passed the geoscientific criteria and were declared sub-areas with a generally favourable geological situation.

To select a site for a high-level radioactive waste repository, the performance of the possible sites as disposal systems that safely contain radionuclides over 1 million years has to be investigated. Hence, in Step 2 of Phase I, representative preliminary safety assessments are an important part of the regulatory toolbox for finding the most suitable areas. They are performed according to the Disposal Safety Analysis Ordinance (EndlSiAnfV, 2020; EndlSiUntV, 2020) of 2020. Their results will help in narrowing down the sub-areas, which currently cover approximately 54 % of the German land surface, to a small number of siting regions that will be subject to surface exploration in Phase II. Preliminary safety assessments are performed in each phase of the German site selection procedure. The representative preliminary safety assessments of Phase I have a limited scope compared to the assessments in Phases II and III (EndlSiAnfV, 2020). For a more detailed overview of the site selection process and the role of the preliminary safety assessments, see Hoyer et al. (2021).

2 Evolutions of the disposal system

The safety of a disposal system is not only dependent on the present-day geological situation of its site, a good repository design, and an excellent technological implementation during construction and operation of the repository but also especially on the future behaviour of the whole system. Since safety has to be achieved for 1 million years, a well-founded, circumspect estimate of the processes acting in and on the repository becomes vital. In the representative preliminary safety assessments, the focus is on future processes in the geosphere, with the challenge to estimate and attach probabilities and risks to any of these geogenic processes[1].

The range of possible evolutions of the system is the core of a safety assessment. They are ordered by likelihood (§ 3 EndlSiAnfV):

1. the expected evolution is the most likely development from the most likely initial state of the disposal system, and

[1] A term newly used in the context of nuclear waste disposal in the Site Selection Act. It is interpreted as those *geogenic* processes that would occur independently of the repository and is used to distinguish them from *technogenic* processes, which are dependent on the existence of the repository structures.

2. deviating evolutions are not expected, but they are considered as far as they cannot be ruled out from occurring in the future.

For now, the geogenic processes are the starting point for developing the evolutions (§ 7 EndlSiUntV). In later phases, all components and processes in the disposal system can be a starting point for deviating evolutions.

Any attempt at forecasting the future is likely to be wrong to some degree. The aim of a safety assessment structured around the scenario method (a particular method of thinking about the future, further explained below) is to optimise the repository in such a way that deviating unexpected developments are very unlikely to compromise the safety of the repository – as far as it is humanly possible to tell. To achieve this, it is necessary to stretch the imagination beyond extrapolating past or current trends and data observations, to examine all assumptions, and to question the sufficiency of the data used.

Sandia National Laboratories first applied the scenario methodology in the context of nuclear disposal in the USA, after the method was previously used for policy strategy development for both governments and businesses (Cranwell et al., 1990). As part of an OECD-NEA (Organisation for Economic Co-operation and Development Nuclear Energy Agency) initiative, several national nuclear waste disposal organisations started using the scenario methodology. Since then, it has become a standard instrument for the development of a safety case (IAEA, 2012).

A scenario study requires

1. a clearly defined scope (area of interest, time period, goals),
2. a compilation of knowledge about the current state (relevant actors, facts, well-known processes and trends in science and society), and
3. an understanding of which elements of the current state are least understood or least predictable in their properties or behaviour while posing the greatest risk to the goals (MacKay and McKiernan, 2018).

In a safety assessment for a potential nuclear waste disposal site, the scope is defined by national regulations. In the German site selection process, the areas of interest are the subareas identified in Step 1 of Phase I. The period of interest is 1 million years after closure of the repository; the goal is a disposal system that keeps within the limits of radionuclide mass and number of atoms transported beyond the essential barriers as given by the Disposal Safety Requirements Ordinance (EndlSiAnfV, 2020; StandAG, 2017). Dose calculations are excluded from the representative preliminary safety assessments but will be part of later assessments.

The compilation of knowledge about the current state and trends is limited to a description of the components and processes in the geosphere and the repository by the EndlSiUntV. Future human actions will not be taken into account, and at this stage, it is permitted to assume that the repository was successfully constructed and closed according to the current repository concept. Possible risks related to these aspects will be included in the preliminary safety assessments from Phase II onward.

Safety assessments for potential nuclear waste disposal sites usually build up a database called a "FEP catalogue" to document the description of the features, events, and processes studied in the safety assessment. SKB developed this highly structured system description for their safety assessments (Swedish Nuclear Fuel and Waste Management Co., 1989). Today, the OECD-NEA provides the International FEP List as a starting point or auditing instrument for new safety assessment projects (Capouet et al., 2019).

A catalogue entry of a feature, event, or process will typically include a short definition and a longer description, a statement on whether the element is relevant to the repository and why, the times at which the element is relevant, and its influence on other elements with a record of the reasoning behind the identified influence (Fig. 1).

The complete FEP catalogue holds a network of direct and indirect influences that have to be assessed for their safety relevance. This is done by analysing the performance of safety functions of different barriers in the disposal system over time.

The third part of a scenario study is an examination of knowledge, lack of knowledge, and risk. Uncertainties for all features and processes under consideration need to be mapped out and quantified wherever possible.

3 Example – subglacial tunnel valley erosion

Subglacial tunnel valley erosion is a geogenic process (occurring independently of the existence of a repository system) that has produced large-scale linear erosional features during past glaciations of what is now Germany (e.g. Weitkamp and Bebiolka, 2017). The process that leads to the development of subglacial tunnel valleys is still under debate (Weitkamp and Bebiolka, 2017). Indications that tunnel valleys are filled in quickly (most commonly with coarse material or locally reworked sediments) but can be reactivated in successive glaciations have been observed (Kuster and Meyer, 1979).

The minimum depth for a containment-providing rock zone (CRZ)[2] is 300 m below present ground surface. However, the integrity of the CRZ has to hold for 1 million years. Therefore, the depth of the repository must be below the depth reached by direct and indirect effects of exogenic processes, and there must be no reasonable doubt about the long-term integrity of the barriers in the disposal system, particularly the host rock (§ 23 StandAG). If subglacial tunnel valley erosion can be expected to occur with any glaciation, the

[2] The part of the host rock that ensures the safe containment of the radioactive waste in interaction with technical and geotechnical barriers (§ 2 Nr. 9 StandAG).

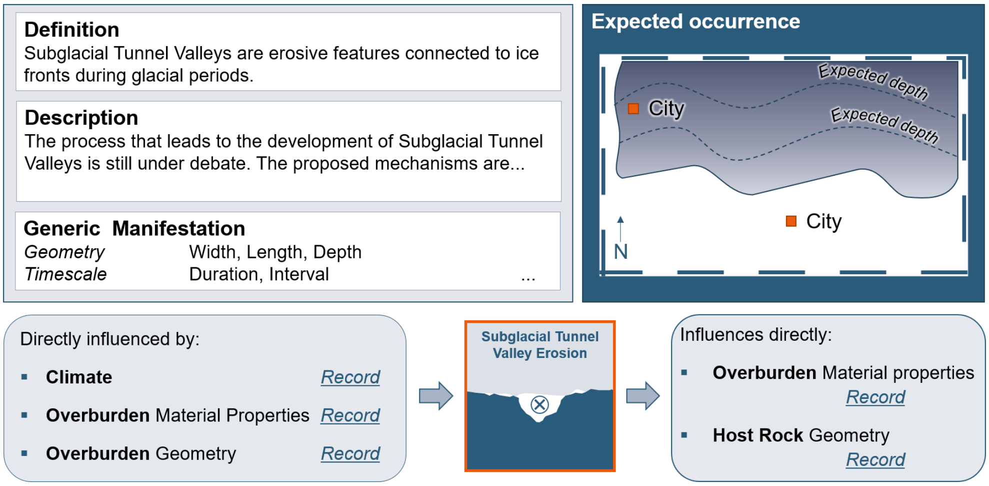

Figure 1. Sketch of FEP catalogue entry for the process of subglacial tunnel valley erosion. These are only examples of possible influences within the system. The final FEP catalogue entry in the representative preliminary safety assessments performed by BGE may look different and include different conclusions.

likely intensity of the process (erosional depth) is the key to estimating whether the siting of a repository at a certain location and depth can ensure the long-term integrity of the CRZ.

In the safety assessment, the process will first be entered into the FEP catalogue (Fig. 1). Its definition distinguishes it from other forms of erosion. The description contains a summary of the current understanding of the process in the scientific community, as well as the areas affected by the process in the past.

Next, the properties of features in the catalogue that can directly influence the manifestation of the process are identified, as are the properties of features that are in turn directly affected by the process. Tunnel valley erosion can locally reduce the thickness of the overburden of the CRZ, as well as change the permeability when the valley is filled up again. Assuming, for instance, 10 future glaciations over 1 million years, then multiple potential erosion "events" can affect a potential disposal system.

4 Consequences of expected and deviating evolutions

Uncertainty about the processes that form subglacial tunnel valleys could invite the definition of a vertical safety margin, for example 100 m. This strategy, though, has issues: to start with, increasing the depth of the repository may offer more long-term safety, but it can increase short-term risks during the operational phase, depending on the host rock, since the drilling of the mine can become more difficult.

Another issue is the problem of missing knowledge (epistemic): how can we know that we have observed the deepest possible tunnel valley? For example, in the Netherlands, a maximum erosion depth of 500 m was known. Reprocessing of existing seismic data, however, led to the discovery of "new" tunnel valleys with depths of nearly 600 m below ground surface (Ten Veen, 2015). This necessitated a re-evaluation of the risk tunnel valley erosion posits for radioactive waste disposal, increasing the maximum observed depth by approximately 100 m.

The deepest known tunnel valley in Germany is the so-called Hagenower Rinne with a depth of 584 m below present ground surface (Kuster and Meyer, 1979), but there is no way of knowing at this time whether that is the deepest valley. Tunnel valley research in the North Sea allows for the conclusion that tunnel valleys will be found when specifically (re)processing data, e.g. seismic data (e.g. Lutz et al., 2009). An absence of tunnel valleys on a thematic map does not imply that there are none present. It may just mean no appropriate study has been conducted there to gain significant data, or existing data were not analysed for this special purpose.

This leads to our understanding of the risks of tunnel valley erosion and how we assess which evolution is expected and which is deviating. It becomes hard to judge whether a safety buffer of 100 m below the deepest known tunnel valley is actually a sensible conservative measure. To account for this, Weitkamp and Bebiolka (2017) suggest zones of different representative predicted depths of erosion. Even so, this could either exclude possibly safe sites from further consideration by strongly overestimating tunnel valley erosion depths or include too many by underestimating them.

5 Conclusions

Considering the future with scenario methods should inevitably call into question the extent and certainty of our knowledge and understanding regarding both the past and the present. Particularly in high-reliability institutions such as a disposal system for high-level radioactive waste, the consequences of error need to be minimised as far as possible. A disposal system needs to be sited and conceptualised with sufficient safety reserves to accommodate a range of future evolutions beyond the expected evolution.

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Research article

Comparison of overdeepened structures in formerly glaciated areas of the northern Alpine foreland and northern central Europe

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Abstract: Overdeepened structures occur in formerly and presently glaciated regions around the earth and are usually referred to as overdeepenings or tunnel valleys. The existence of such troughs has been known for more than a century, and they have been attributed to similar formation processes where subglacial meltwater plays a decisive role. This comparison highlights that (foreland) overdeepenings and tunnel valleys further occur in similar dimensions and share many characteristics such as gently sinuous shapes in plan view, undulating long profiles with terminal adverse slopes, and varying cross-sectional morphologies. The best explored examples of overdeepened structures are situated in and around the European Alps and in the central European lowlands. Especially in the vicinity of the Alps, some individual troughs are well explored, allowing for a reconstruction of their infill history, whereas only a few detailed studies, notably such involving long drill core records, have been presented from northern central Europe. We suggest that more such studies could significantly further our understanding of subglacial erosion processes and the regional glaciation histories and aim to promote more intense exchange and discussion between the respective scientific communities.

Kurzfassung: In den ehemals und gegenwärtig vergletscherten Regionen der Erde finden sich übertiefte Beckenstrukturen, die üblicherweise als Übertiefungen oder Tunneltäler angesprochen werden. Die Existenz dieser Tröge ist seit mehr als einem Jahrhundert bekannt, und sie wurden ähnlichen Erosionsprozessen zugeschrieben, bei denen subglaziales Schmelzwasser eine entscheidende Rolle spielt. Mit diesem Vergleich möchten wir zeigen, dass (Vorland-)Übertiefungen und Tunneltäler in vergleichbaren Dimensionen auftreten, und eine Reihe weiterer gemeinsamer Merkmale aufweisen, zum Beispiel sanft gewundene Längsachsen, undulierende Profile mit Gegensteigungen am distalen Ende, und variable Formen im Querschnitt. Die bestuntersuchten Beispiele für übertiefte Becken befinden sich im europäischen Alpenraum, und in zentraleuropäischen Tiefländern. Vor allem im Alpenraum wurden einige übertiefte Strukturen detailliert untersucht, was eine Rekonstruktion ihrer Verfüllung erlaubt. Im Gegensatz dazu existieren nur wenige (bohrungsbasierte) Detailstudien im nördlichen Mitteleuropa. Wir betonen, dass solche Untersuchungen zu unserem Verständnis subglazialer Erosionsprozesse, aber auch regionaler Vergletscherungsgeschichten beitragen können, und möchten Austausch und Diskussion unter den entsprechenden wissenschaftlichen Lagern anregen.

1 Introduction

The repeated glaciations during the Quaternary period shaped large swathes of the earth's surface, first, by eroding and removing large rock masses from mountain areas and other regions covered by ice; glacially polished bedrock surfaces and deep valley flanks are some of the most striking features in this context. Second, the accumulation of rock debris transported by ice and meltwater led to the formation of characteristic landscape features such as moraine ridges and gravel terraces. However, a third feature that represents a combination of both erosive and aggradational processes is less well known to a broader public: geological structures deeply incised by subglacial erosion into pre-existing landscapes but hidden below the present land surface after being filled up by sediment and/or occupied by large bodies of water. First recognised within and around the European Alps as well as below the central European lowlands, these troughs are referred to as overdeepenings and tunnel valleys, respectively.

While these structures have been known since the late 19th century, for a long time they attracted relatively little scientific attention as their investigation is costly due to the fact that it requires the use of either deep drilling or geophysics – ideally a combination of both. Since the beginning of the 21st century, scientific interest in glacially eroded, and specifically overdeepened, structures increased considerably, to some extent motivated by applied aspects. On the one hand, these structures have a poorly explored potential as groundwater sources and for geothermal energy production. On the other hand, planning nuclear waste repositories requires the identification of areas that will ideally not be affected by deep glacial erosion in the near geological future (i.e. the next million years), as this may challenge the intactness of the disposal site.

Advances in geophysical techniques together with the increased interest in glacially eroded structures have triggered several projects into the subject during the past 2 decades notably in the aforementioned areas in central Europe. These can thus be regarded as the best explored regions with regard to overdeepened structures and, due to their proximity, should be subject to similar evolutionary trends over time. However, there has been very limited exchange among the communities working in and around the Alps and the northern central European lowlands, respectively. This is in contrast to the common assumption that all these structures were formed by similar erosion processes, and, hence, it should be expected they share many similarities.

The aim of this article is to summarise and compare the current knowledge about overdeepenings and tunnel valleys in general but with special regard to the northern Alpine foreland and the wider North Sea area. This comparison is done specifically in light of the question whether there are relevant differences in the morphology and/or filling history of these troughs or whether they are just different examples of basically the same type of feature. We thus touch only lightly upon the concepts of the erosion mechanisms that have been intensively discussed by others (e.g. Alley et al., 2019; Cook and Swift, 2012, and references therein).

2 History of recognition and state-of-the-art research

The term "overdeepening" is attributed to Albrecht Penck, one of the pioneers and most prominent representatives of Alpine Quaternary geology. He used it to describe intensive and deep-reaching glacial erosion focused in the main Alpine trunk valleys whose valley floors lie significantly below their off-cut tributaries (Penck and Brückner, 1909). The concept of deep-reaching subglacial erosion had been subject to intensive discussion (e.g. Heim, 1885) but was lastly confirmed in 1908 by a tragic accident during tunnel works in central Switzerland (Lötschberg), when blasting their way into an infilled overdeepening ~ 170 m below ground cost the lives of 24 workers, as the liquefied valley infill flooded the tunnel (Waltham, 2008). Over the following decades, boreholes and geophysical data revealed the existence of such deep infilled bedrock troughs below the majority of the present-day valleys not only in the Alps and the Alpine foreland (cf. Haeberli et al., 2016; Preusser et al., 2010) but also in other mountain regions on earth (e.g. Carrivick et al., 2016; Gao, 2011; James et al., 2019; Magnússon et al., 2012; Smith, 2004). First systematic compilations of these deeply incised features were provided by Van Husen (1979) and Wildi (1984), who noted that large sections of these troughs lie deeper than the bedrock at the respective valley outlet. This closed basin shape with a terminal adverse slope is today generally regarded as defining an overdeepening (Alley et al., 2019; Cook and Swift, 2012; Patton et al., 2016). Notably – but not exclusively – in Switzerland, overdeepenings have since been intensively studied based on scientific drillings (Hsü and Kelts, 1984; Preusser et al., 2005; Schlüchter, 1989; Schwenk et al., 2022), geophysical campaigns (Bandou et al., 2022; Burschil et al., 2018, 2019; Finckh et al., 1984; Nitsche et al., 2001; Reitner et al., 2010), or a combination of both (Buechi et al., 2018; Dehnert et al., 2012; Gegg et al., 2021; Pomper et al., 2017).

The recognition of tunnel valleys goes similarly far back in time, to the late 19th century, when Danish and north German scholars identified deep subsurface troughs infilled with Quaternary sediments or hosting lakes (cf. van der Vegt et al., 2012). Jentzsch (1884) already hypothesised about an origin by subglacial fluvial erosion, a concept that was refined by Ussing (1904), who attributed these troughs, characterised by internal and terminal adverse slopes, to the action of pressurised subglacial meltwater. The model of Ussing (1904) became increasingly widely accepted, and the "tunnel valleys" (Madsen, 1921) became a subject of increasing scientific interest. Systematic regional-scale mapping and investigation started in the 1960s (e.g. Kuster and Meyer, 1979) and was expanded to other regions such as Great Britain (e.g. Woodland, 1970) and North America (e.g. Wright, 1973). Soon, the widespread availability of high-resolution marine seismic data acquired for oil and gas exploration marked a breakthrough in the recognition of tunnel valleys: they were encountered in previously glaciated shelf areas all around the world and could be temporally efficiently mapped on reflection-seismic sections (Boyd et al., 1988; Destombes et al., 1975; Kunst and Deze, 1985). Tunnel valley characterisation made a second major leap forward with the establishment of 3D seismic acquisition (Praeg, 1997). Based on these datasets, the longitudinal and transverse morphology of individual structures can be imaged and analysed in high detail and free from gaps (Kirkham et al., 2021; Kristensen et al., 2008; Ottesen et al., 2020; Stewart et al., 2013). Further insights derive from electromagnetic (e.g. Bosch et al., 2009; Tezkan et al., 2009) as well as gravimetric studies (e.g. Barker and Harker, 1984; Götze et al., 2009), while only little detailed borehole information is available (e.g. Piotrowski, 1994).

3 Geological and glaciological settings

In extensively glaciated mountain ranges, such as the European Alps, large overdeepenings occur in two endmember settings: in the major intermontane trunk valleys and in the mountain forelands (Dürst Stucki and Schlunegger, 2013; Magrani et al., 2020; Preusser et al., 2010). Within the mountain range, the bedrock is generally quite resilient towards erosion (Kühni and Pfiffner, 2001), and glacial erosion occurs preferentially along zones of weakness such as faults (Dürst Stucki and Schlunegger, 2013). This is a self-promoting process: with ongoing downcutting, drainage of water and ice along the deepening valley becomes increasingly efficient, and erosion is further focused on the valley floor, the result being a spatially stable, deeply incised valley, where ice flows comparatively rapidly and under strong lateral confinement (Egholm et al., 2012; Herman et al., 2011; Ugelvig et al., 2016). Glacial erosion sensu stricto, i.e. quarrying and abrasion, probably plays an important role in this setting, aided by subglacial meltwater stripping debris off the glacier base (Alley et al., 2019). It culminates in the formation of overdeepenings, for example at confluences or valley constrictions where ice flow is accelerated (Cook and Swift, 2012; Dürst Stucki and Schlunegger, 2013; Herman et al., 2015; Preusser et al., 2010).

Foreland overdeepenings, in contrast, develop beneath the piedmont tongues of valley glaciers. There, although its large-scale pattern is defined by the locations of mountain valley outlets, the ice flow is topographically much less constrained and potentially diffluent and the ice thickness is significantly smaller (Bini et al., 2009), and as a result, glacial erosion sensu stricto should be rather limited. However, with increasing catchment area, basal water availability also increases towards the glacier termini. This subglacial water, pressurised by the englacial water column, is regarded as a driver of overdeepening erosion in the foreland setting (Alley et al., 1997; Dürst Stucki et al., 2010; Dürst Stucki and Schlunegger, 2013; Gegg et al., 2021), where the bedrock commonly consists of poorly consolidated Molasse-type sediments (Kühni and Pfiffner, 2001). As these deposits are generally readily eroded, the occurrence of faults does not appear to play a significant role facilitating their erosion (Dürst Stucki and Schlunegger, 2013; Gegg et al., 2021). With the effect of structural preconditioning being low and ice flow being largely unconfined, the focusing of subglacial erosion may shift over time, evidenced for example by branching and off-cutting overdeepened troughs (Buechi et al., 2018; Ellwanger et al., 2011).

The primary erosive agent of tunnel valleys is, by definition, inferred to be basal meltwater (Cofaigh, 1996; Kehew et al., 2012; van der Vegt et al., 2012) at the margins of continental-scale ice sheets, where topographical confinement of ice flow is significantly lower than within mountain ranges (Schoof and Hewitt, 2013). Still, discharge at ice sheet margins is concentrated along ice streams, corridors of enhanced ice flow (Margold et al., 2015; Rignot et al., 2011) that develop where basal meltwater abundance is high and facilitates tunnel valley incision (Jennings, 2006; Lelandais et al., 2018). Further, where tunnel valleys occur near highland areas, some lie distinctly in extension of fjord valleys that direct the larger-scale flow pattern of ice and water (e.g. Bradwell et al., 2008; Kearsey et al., 2019). Large tunnel valleys occur predominantly within rather soft sedimentary substrata, while troughs in more resistant lithologies tend to be smaller and narrower (Janszen et al., 2012; Jørgensen and Sandersen, 2006). In resilient crystalline bedrock, tunnel valleys are typically lacking but are replaced by large-scale esker systems that are interpreted to reflect meltwater streams cut into the basal ice instead of the substratum (Boulton et al., 2009; Clark and Walder, 1994).

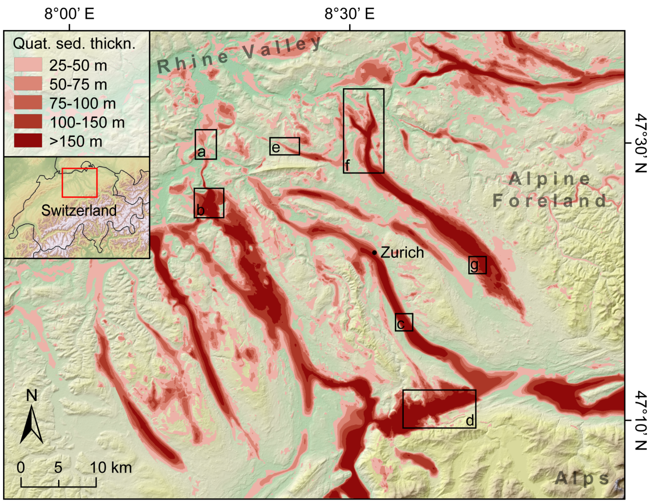

Figure 1. Foreland overdeepenings in central northern Switzerland highlighted as areas of increased Quaternary sediment thickness. Data from Pietsch and Jordan (2014), Gegg et al. (2021), and references therein. Black boxes highlight study areas mentioned in the text: (*a*) Lower Aare Valley (Gegg et al., 2021), (*b*) Birrfeld (Graf, 2009; Nitsche et al., 2001), (*c*) Lake Zurich (Lister, 1984a, b), (*d*) Richterswil (Wyssling, 2002), (*e*) Wehntal (Anselmetti et al., 2010; Dehnert et al., 2012), (*f*) Lower Glatt Valley (Buechi et al., 2018), and (*g*) Uster (Wyssling and Wyssling, 1978).

4 Morphological comparisons

4.1 Large-scale morphologies

Foreland overdeepenings are up to 15 km wide and 850 m deep bedrock incisions that reach lengths of >100 km (James et al., 2019; Magrani et al., 2020; Smith, 2004). They have been described as slightly sinuous troughs that may be arranged in seemingly anastomosing patterns (Fig. 1; Dürst Stucki and Schlunegger, 2013; Preusser et al., 2010). While some troughs have been re-excavated by multiple phases of ice advance (e.g. Birrfeld overdeepening; Nitsche et al., 2001; Lower Glatt Valley overdeepening; Buechi et al., 2018), the focus of overdeepening has in other areas laterally shifted over the course of several glaciations and produced subparallel (e.g. Reuss Valley; Jordan, 2010) or radially diverting branch basins (e.g. Lake Constance area; Ellwanger et al., 2011). The spacing between individual troughs is 5–20 km (Cummings et al., 2012; see also Jordan, 2010), and their sinuosities are around 1.05–1.10 (Gegg et al., 2021). In the longitudinal section, foreland overdeepenings often consist of several distinct sub-basins separated by abrupt bedrock swells and terminating with mostly gentle (~ 1–$2°$) adverse slopes (Jordan, 2010; Magrani et al., 2020; Preusser et al., 2010). While typically shifted upstream in intra-mountain overdeepenings, the deepest points in most foreland overdeepenings lie roughly at their central position (Magrani et al., 2020).

Tunnel valleys have maximum widths of ~ 12 km and maximum depths of ~ 500 m and can be over 150 km long (Cameron et al., 1987; Lutz et al., 2009; Praeg, 1997; Stewart et al., 2013). In plan view, they have straight to slightly sinuous courses and can occur in swarms or in pseudo-anastomosing (i.e. different branches of seemingly individual valleys belong to different generations of valley formation; Fig. 2; Jørgensen and Sandersen, 2006; Kristensen et al., 2008), radiating or tributary networks (Cofaigh, 1996; Kehew et al., 2012; van der Vegt et al., 2012). The lateral spacing between individual structures typically ranges between 5 (Livingstone and Clark, 2016) and 20–30 km (Kehew et al., 2012; van der Vegt et al., 2012), while Lutz et al. (2009) specifically note that tunnel valleys are not uniformly distributed and may in some areas be lacking entirely. According to data by Streuff et al. (2022), typical tunnel valley sinuosities are around 1.05. Tunnel valleys characteristically start and terminate abruptly (with terminal adverse slope angles in some cases exceeding 10°; Kristensen et al., 2008) and have undulating long courses comprising sub-basins as well as steep thresholds (Cofaigh, 1996; Kristensen et al., 2008; Lutz et al., 2009). The deepest points along the thalweg frequently lie far downstream from the valley centre (Jørgensen and Sandersen, 2006).

Based on datasets by Magrani et al. (2020) and Kristensen et al. (2008), Swiss foreland overdeepenings tend on average to be shorter, shallower, and narrower than Danish tunnel valleys, although only the difference in depth is statistically significant (unequal-variance *t* tests; Fig. 3). It should be noted, however, that the dataset by Kristensen et al. (2008) refers only to a restricted area and that other studies in the North Sea found tunnel valleys that generally tend to be shallower (Andersen et al., 2012) and narrower (Jørgensen and Sandersen, 2006; Stewart et al., 2013). The overall size of tunnel valleys as well as their form ratio appears to be influenced mainly by the erosional resistance of the substratum, whereas the valley width has been linked rather to glaciological parameters (e.g. ice thickness; van der Vegt et al., 2012). The maximum size of foreland overdeepenings on the other hand is clearly limited by the spatial extent of the respective piedmont glaciers, which were considerably smaller for example in front of the Eastern or Southern Alps (or in other mountain ranges) than in northern Switzerland (e.g. Preusser et al., 2010). Still, the comparison shows that, although the corresponding ice masses differ vastly in size, the dimensions of tunnel valleys and foreland overdeepenings are not necessarily greatly different.

4.2 Detailed cross-sectional morphologies

Cross-sections of foreland overdeepenings are frequently asymmetric and tend towards a U shape, but especially in more resistant bedrock lithologies, steep-angled (up to 60°) V-shaped valleys have been observed (Fig. 4; Dürst Stucki and Schlunegger, 2013; Jordan, 2010; Preusser et al., 2010),

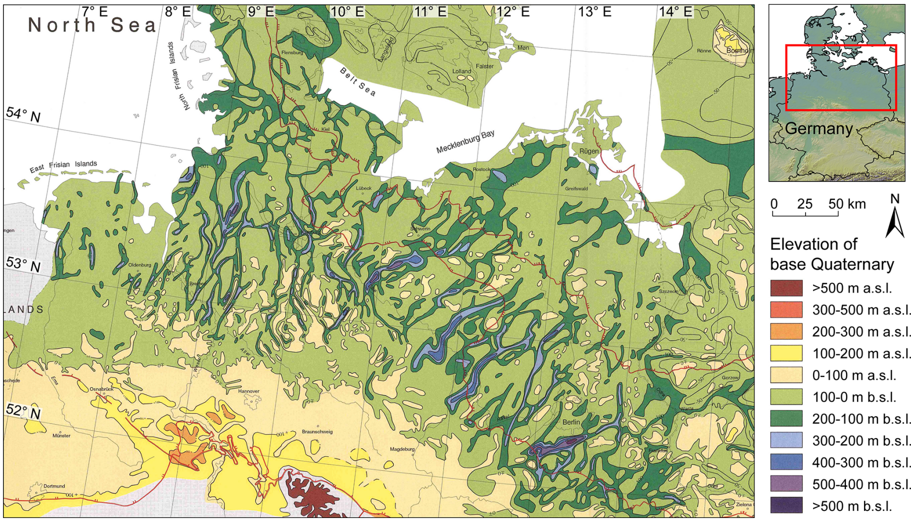

Figure 2. Buried tunnel valleys in northern Germany. Red lines mark, from NE to SW, the ice extents of the Weichselian, Saalian, and Elsterian main glacial stages. Base map (Stackebrandt et al., 2001) courtesy of the Geological Survey of Brandenburg (Landesamt für Bergbau, Geologie und Rohstoffe; LBGR), Germany (https://lbgr.brandenburg.de/lbgr/de/, last access: 5 November 2022), © LBGR 2001.

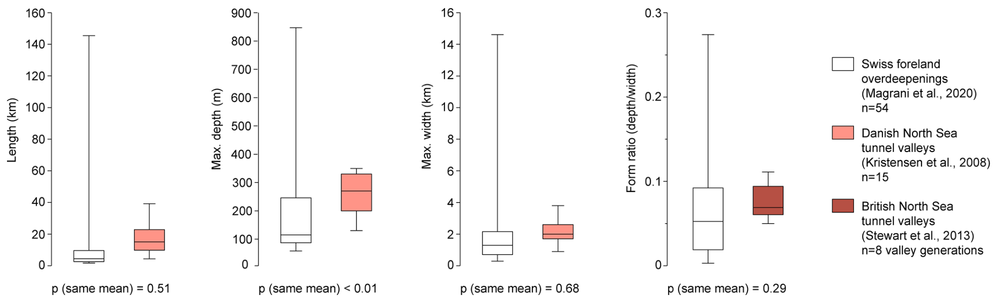

Figure 3. Comparison of lengths, depths, widths, and form ratios of Swiss foreland overdeepenings and North Sea tunnel valleys.

and both morphologies may occur within a single overdeepening (Gegg et al., 2021). Recent highly resolved geophysical studies have further revealed irregular and stepped flanks. The same morphologies are also encountered in tunnel valleys (Fig. 4). In a dataset comprising >900 Danish tunnel valleys, Andersen et al. (2012) identified $\sim 65\,\%$ of mainly U-shaped or flat-bottomed and $\sim 35\,\%$ of V-shaped structures. Flank slopes vary between $<5°$ and $>55°$ (Cofaigh, 1996; Huuse and Lykke-Andersen, 2000), and van der Vegt et al. (2012) mention examples of overhanging valley flanks. Individual valleys exhibit downstream transitions in a cross-section, for example, from a V to a U shape (Giglio et al., 2022). In both cases, overdeepenings and tunnel valleys, U-shaped morphologies with flat valley bottoms may be linked to lithological boundaries in the substratum (Gegg et al., 2021; Janszen et al., 2012) but occur also where the bedrock is seemingly rather homogeneous (e.g. Fig. 4a, d)

5 Post-erosional history

Concrete information regarding the sedimentary fill of overdeepened structures is rather limited and is often derived from commercial or geotechnical drillings. Under such circumstances, typically no explicit sedimentological descriptions or even further reaching analyses (e.g. pollen analyses, dating) have been carried out. As a result, for the time being, detailed knowledge about the infilling history of overdeepenings is limited to a few case studies. As stated previously, a majority of detailed studies stem from central Europe. This applies especially to the Alpine realm, while fewer well-documented drillings exist for the North Sea region. There, the reconstruction of the sedimentary filling is hence often based on the interpretation of 2D or 3D seismic data that are frequently available from hydrocarbon exploration. We would like to highlight that although we focus on examples from central Europe, a variety of similar case studies on overdeepened structures in other areas exist, e.g. Iceland (Andrews et al., 2000; Gregory, 2012; Quillmann et al., 2010), North America (Atkinson et al., 2013; Smith,

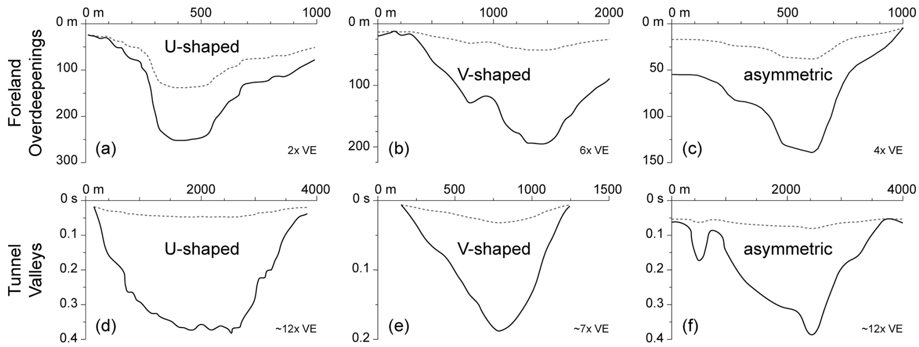

Figure 4. Comparison of overdeepening **(a–c)** and tunnel valley **(d–f)** cross-sections. *x* axis: horizontal distance, *y* axis: depth **(a–c)**/two-way travel time **(d–f)**. VE: vertical exaggeration (dotted lines plot the respective cross-sections without vertical exaggeration). **(a)** Basadingen trough, northern Switzerland (after Anselmetti et al., 2022; Brandt, 2020); **(b)** Tannwald basin, southern Germany (after Burschil et al., 2018); **(c)** Gebenstorf-Stilli Trough, northern Switzerland (after Gegg et al., 2021); **(d)** offshore tunnel valley, southern North Sea (after Benvenuti and Moscariello, 2016); **(e)** offshore tunnel valley, central North Sea (after Kirkham et al., 2021); **(f)** offshore tunnel valley, southeastern North Sea (after Lohrberg et al., 2020). Note frequently occurring stepped valley flanks.

2004; Russell et al., 2003), Patagonia (Bertrand et al., 2017; Boyd et al., 2008; Moernaut et al., 2009), and even regions glaciated during the Palaeozoic (Clerc et al., 2013; Vesely et al., 2021).

5.1 Examples from the northern Alpine foreland

A number of well-studied profiles exist in the greater Bern area/Aare Valley, Switzerland (cf. Schlüchter, 1979). The sequence of Thalgut shows a complex deposition and erosion history and is considered one of the most complete Quaternary sequences of the northern Alpine foreland (Schlüchter, 1989). The bottom part of the sequence is composed of lake deposits with pollen spectra typical of the Holsteinian Interglacial (Welten, 1982, 1988), usually attributed to Marine Isotope Stage (MIS) 11 (424–374 ka; cf. Cohen and Gibbard, 2019). Hence, the formation (or re-excavation) of the overdeepening likely occurred during MIS 12 (478–424 ka). Close to the city of Bern, the Rehhag scientific drilling targeted a tributary overdeepening feeding into the central Aare Valley and recovered a diverse, 240 m thick Quaternary succession (Schwenk et al., 2022). It is subdivided into two basin fill sequences of glacial and (glacio-)lacustrine deposits, the upper being attributed to MIS 8–7 (300–191 ka). Just north of Bern, the core of Meikirch starts with till, followed by a 70 m succession of lacustrine sediments comprising a complex vegetation history with three pronounced warm phases (Welten, 1982; Preusser et al., 2005). Based on luminescence dating, this part of the sequence was assigned to MIS 7 (243–191 ka; Preusser et al., 2005). Above the lake deposits, two successions from glaciofluvial to glacial deposits are recorded. In the overdeepened lowermost Aare Valley, at the confluence with Reuss and Limmat, the sediment filling is dominated by lacustrine sand overlying a thin layer of coarse-grained debris at the trough base (Gegg et al., 2021, and references therein). Towards the distal part of the trough, the glaciolacustrine sand interfingers with deltaic gravel. This rather unusual infill pattern is most likely related to the local confluence situation as well as the overdeepening's narrow cross-section combined with large discharge of meltwater (see also Gegg et al., 2020). The Birrfeld basin in the lower Reuss Valley contains a multiphase infill that has been attributed to up to five different ice advances (Graf, 2009; Nitsche et al., 2001).

At a water depth of 180 m, a drilling in Lake Zurich recovered a thick succession of Late Pleistocene sediments starting with coarse-grained debris interpreted as till and subglacial outwash (Lister, 1984a, b). This debris is overlain by > 100 m of diamictic glaciolacustrine muds that repeatedly show traces of glaciotectonic deformation. They transition into laminated, presumably varved, basin fines and are followed by ~ 30 m of postglacial lake deposits. The complex filling of the Richterswil trough, west of Lake Zurich, indicates deposition possibly related to three individual glaciations, but this evidence lacks detailed sedimentological and further analyses (Wyssling, 2002; Preusser et al., 2010; Fig. 5).

Several cores taken from the Wehntal trough (Niederweningen) show that the infill starts with glacial deposits followed by lake sediments, first containing dropstones (Anselmetti et al., 2010; Dehnert et al., 2012). Increased shear strength in the lacustrine deposits implies a second glacial advance, grounding in the lake, followed by a final lake and subsequent bog stage. According to luminescence dating, the carving of the basin occurred during early MIS 6 (around 185 ka), whereas the second ice advance is assigned to ca. 140 ka.

In the Lower Glatt Valley, Buechi et al. (2018) distinguished up to nine separate depositional sequences within the infills of the Kloten Trough and its branch basins, three of which represent episodes of overdeepened basin fill. These are characterised by successions of tills and gravel fining up-

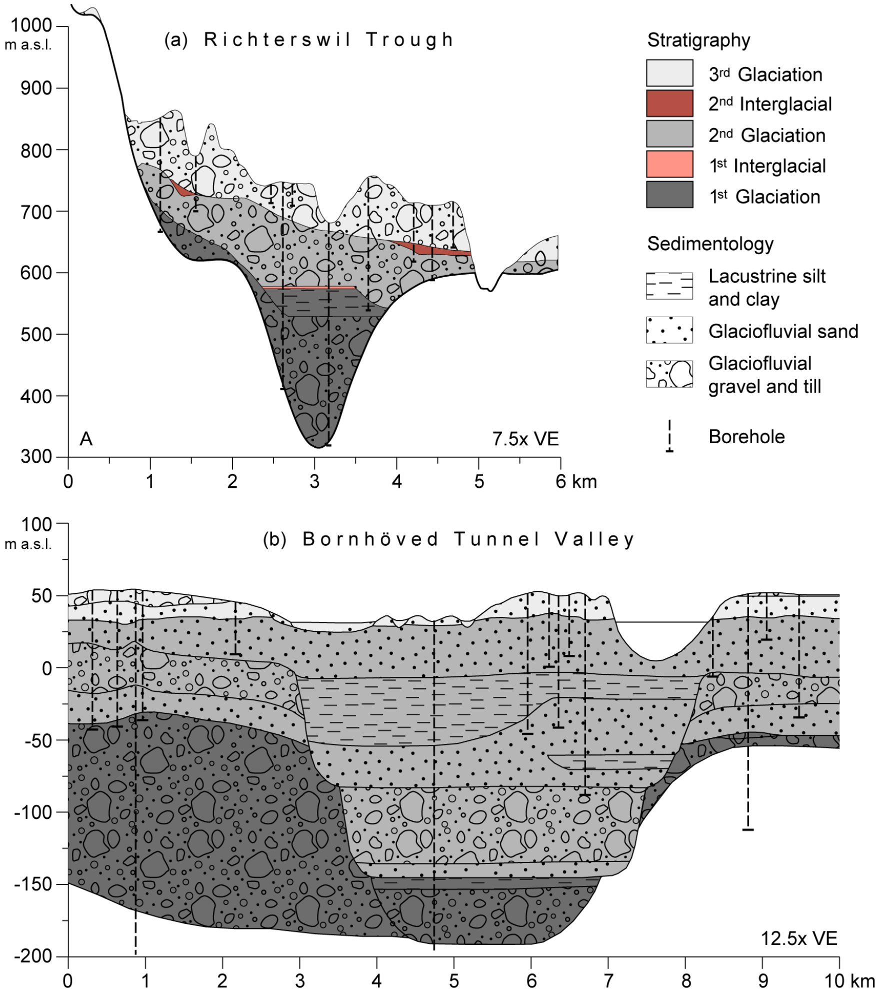

Figure 5. Examples of borehole-constrained sedimentary infills of a central Swiss foreland overdeepening (**a**: Richterswil trough; after Wyssling, 2002, in Preusser et al., 2011) and a northern German tunnel valley (**b**: Bornhöved tunnel valley; after Piotrowski, 1994). VE: vertical exaggeration.

wards into glaciodeltaic to glaciolacustrine deposits, the oldest dating back to at least MIS 8 (300–243 ka). In the middle part of the Glatt Valley, the > 200 m deep trough of Uster is filled by till and partially covered by ice decay and meltwater deposits, followed by laminated lake sediments with a thickness of 100–150 m that are interpreted to represent varved late-glacial deposits (Wyssling and Wyssling, 1978; Preusser et al., 2011). The infill of the trough is covered by the deposits of one or possibly two later glacial advances.

In the Lake Constance area, Ellwanger et al. (2011) distinguish three major generations of overdeepenings, with the drilling in the Tannwald basin being the best documented for the time being (LGRB, 2015; Burschil et al., 2018; Anselmetti et al., 2022). This basin belongs to the oldest generation recognised so far. The sequence starts with sheared allochthonous bedrock slabs on top of the bedrock contact (Upper Marine Molasse), followed by gravel with molasse components and diamictic fines. The subsequent fine-grained lacustrine deposits are correlated with a nearby sequence that contains pollen assemblages assigned to the Holsteinian (Ellwanger et al., 2011; Hahne et al., 2012). The upper part of the Tannwald core comprises evidence for further glacial advances. In Lake Constance itself, as revealed by seismic surveys, up to 150 m of sediment has accumulated, the uppermost 24 m of which was recovered by a recent drilling campaign (Schaller et al., 2022). It consists of 12 m of coarse lacustrine sands of the late-glacial period, overlain by Holocene basin fines.

The basin of Wolfratshausen in Bavaria reveals multiple basal tills and lacustrine fines, evidence for a complex erosion and infill history attributed to three individual glaciations (Jerz, 1979). Fiebig et al. (2014) report a core taken in the Salzach basin containing ∼ 90 m of lacustrine deposits, sand and fines, on top of an Alpine till. Luminescence dating of the lake deposits indicates that the initial formation and filling of the overdeepened trough date to more than 220 ka.

5.2 Examples from northern central Europe

While tunnel valleys have also been described from the British Isles (e.g. Eyles and McCabe, 1989; Coughlan et al., 2020) as well as Poland (e.g. Salamon and Mendecki, 2021), we are here concentrating on the southern North Sea region including the bordering mainland (Denmark, northern Germany, the Netherlands).

For the Bornhöved tunnel valley, Piotrowski (1994) reconstructed a polygenetic evolution based on borehole evidence (Fig. 5). This trough, situated in a peripheral sink between two Permian salt diapirs, was presumably eroded during the Elsterian Glaciation (MIS 12, 478–424 ka) and was filled by fine-grained glaciolacustrine sediments during ice retreat and marine sediments during the Holsteinian Interglacial. These sediments were reworked and redeposited in the shape of a 200 m thick glaciotectonic melange by the first Saalian advance (MIS 10?, 374–337 ka). While the second Saalian advance had little impact, the tunnel valley was reactivated as a subglacial channel during the last, Weichselian, glaciation. In Vendsyssel, northern Denmark, tunnel valleys attributed to the main Weichselian advance (ca. 23–21 ka) are infilled by glaciolacustrine sand and fines (Sandersen et al., 2009). This implies that these tunnel valleys not only have formed rapidly but were also infilled in a few hundred years, or less, before the late-glacial marine inundation (ca. 18 ka).

In the North Sea basin, over 2200 tunnel valleys attributed to up to seven generations were identified using 3D seismic and magnetic data (Stewart and Lonergan, 2011; Ottesen et al., 2020). It appears that each of the seven generations of tunnel valleys, attributed mostly to the Middle Pleistocene, has been excavated and infilled during a separate cycle of ice-sheet advance and retreat (as already suggested by others, e.g. Kristensen et al., 2008), partially reaching into the following interglacial (e.g. Hepp et al., 2012; Janszen et al., 2013; Lang et al., 2015; Steinmetz et al., 2015). Moreau and Huuse (2014) interpret the infill process of tunnel valleys off the shore of the Netherlands to be separate from the incision. Rather than filled by subglacial deposits, Moreau and Huuse (2014) expect the infill sediments were supplied from the southeast by the Rhine–Meuse systems which consistently flowed towards the North Sea basin during glacial periods. This view has been challenged by Benvenuti et al. (2018) who inferred from provenance analysis that the infill of a large Elsterian tunnel valley consists mainly of reworked glacially derived sediment. This is in line with observations by Kirkham et al. (2021) demonstrating that over 40 % of the examined tunnel valleys in the North Sea between Scotland and Norway contain buried glacial landforms such as eskers, crevasse-squeeze ridges, glaciotectonic structures, and kettle holes. Hence, grounded ice must have played an active role not only in the incision but also in the (onsetting) infilling.

5.3 Synthesis

Fillings of overdeepenings in the northern Alpine foreland are clearly dominated by glacial and proglacial deposits. The typical sedimentary succession starts with glacigenic diamicts that are frequently referred to as tills. However, only a few studies (e.g. Buechi et al., 2017) investigated these deposits in sufficient detail to clearly identify them as subglacial ice-contact deposits (cf. Evans, 2007, and references therein), and thus their palaeo-glaciological significance is often not clear. These diamicts are overlain by extensive glaciolacustrine deposits, sands or fines, that at their base frequently contain dropstones. In some basins, several such infill cycles are recorded. Interglacial sediments are in most cases lacking or occur only near the ground surface, known exceptions being the successions of Thalgut, Meikirch, Uster, and Richterswil. This suggests that overdeepened troughs are in most cases infilled and silted up under still periglacial conditions, i.e. in a brief time interval (cf. Van Husen, 1979; Pomper et al., 2017). In contrast, overdeepenings incised during the last glaciation host a number of large and deep contemporaneous, i.e. interglacial, lakes that have not yet been infilled completely (Fig. 6). Whether this is due to exceptionally large basin sizes, comparatively small sediment influxes, or other geological or climatic factors is difficult to assess.

The sediment filling of tunnel valleys and the related depositional processes in northern central Europe are less constrained and remain more poorly understood, although it appears that interglacial sediments are more frequently observed than in the Alpine realm. This could imply a longer "lifetime" of tunnel valleys before being entirely infilled. However, potential tunnel valleys of the last glaciation are completely infilled (e.g. Sandersen et al., 2009) or at least infilled to an extent that does not allow the formation of deep lakes anymore. While some large lakes also exist in the formerly glaciated areas of northern Germany and neighbouring countries, they are significantly shallower than their Alpine counterparts (Fig. 6); do not exhibit the typical elongated plan view trough morphology; and might not be products of overdeepening erosion at all but rather, for example, ice decay structures.

6 Conclusions

Foreland overdeepenings and tunnel valleys are connected to two very different types of ice masses but occur in generally similar palaeo-glaciological and topographic settings. They are formed at the termini of large ice bodies, where ice flow is little constrained, but the abundance of meltwater as well as the ice surface slope, and thus the subglacial pressure gradient towards the margin, are comparatively high. It is this pressurised meltwater that the incision of the spectacular subglacial landforms is mainly attributed to. Both foreland overdeepenings and tunnel valleys share the same morpho-

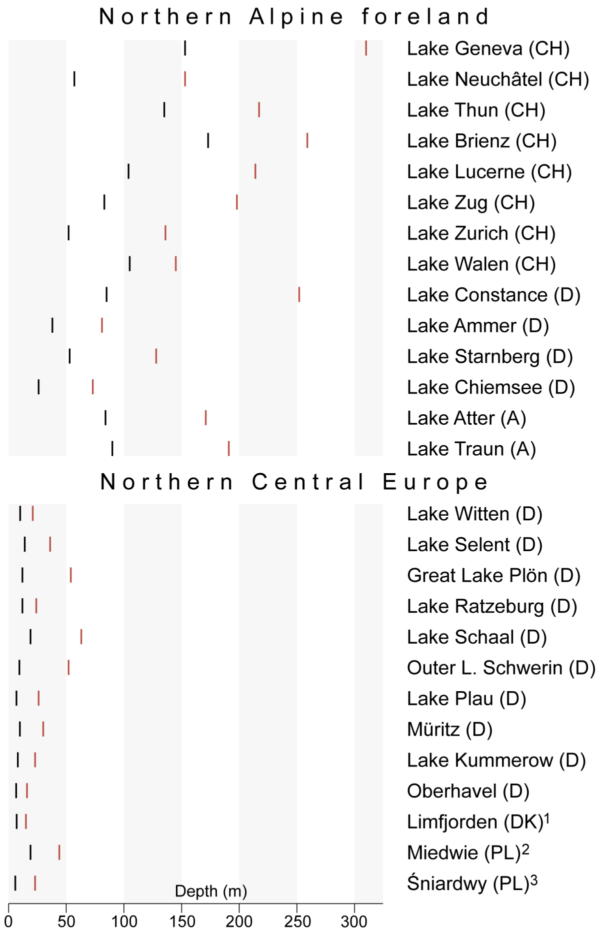

Figure 6. Comparison of mean (black) and maximum (red) depths of large contemporaneous lakes in formerly glaciated areas in the northern Alpine foreland and northern central Europe (Switzerland, CH: BAFU, 2016; Germany, D: Nixdorf et al., 2004; Austria, A: Beiwl and Mühlmann, 2008; Denmark, DK: [1]Riemann and Hoffmann, 1991; Poland, PL: [2]Czerniejewski et al., 2004; [3]Robbins and Jasinski, 1995).

logical characteristics, including anastomosing courses, undulating longitudinal profiles with swells and adverse slopes, and a variety of different cross-sectional shapes. And, perhaps surprisingly, the absolute dimensions of both groups of troughs also appear not to be significantly different from each other.

However, a review of individual troughs and their fillings in central Europe, where these structures have the longest history of recognition and exploration, reveals some notable disparities. In the northern Alpine foreland, multiple well-studied cores exist that allow the characterisation of a typical overdeepening-fill succession in the shape of glacigenic diamicts overlain by glaciolacustrine deposits. In contrast, the infills of tunnel valleys and the related processes have received only little attention and remain rather poorly understood. Further, while the Middle Pleistocene sedimentary record would imply that foreland overdeepenings are more quickly infilled and silted up than tunnel valleys (i.e. typically before interglacial conditions ensue), the characteristics of present-day lakes suggest that the exact opposite has been the case following the last glaciation.

Detailed borehole records are an important tool to better characterise the formation and infilling process of overdeepenings and tunnel valleys (e.g. by process-oriented investigations of the basal "tills"). This is the central objective of the current International Scientific Continental Drilling Program project Drilling Overdeepened Alpine Valleys (DOVE; Anselmetti et al., 2022). Similar drilling projects targeting tunnel valleys are in dire need in order to better understand the overdeepening erosion and glaciation history of northern central Europe. We would like to stress that ongoing and future work of the respective scientific communities may be significantly facilitated and improved by more intensive exchange and discussion, as stimulated for example by the "Subglaziale Rinnen" workshop. The relevance of this exchange is highlighted by the socio-economic aspects of overdeepenings and tunnel valleys, especially in the context of radioactive waste disposal.

Data availability. We utilised datasets by Magrani et al. (2020; https://www.sciencedirect.com/science/article/pii/S0277379120304455; last access: 13 January 2023) and Streuff et al. (2021; https://doi.org/10.1594/PANGAEA.937782).

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Research article

Tunnel valleys in the southeastern North Sea: more data, more complexity

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Abstract: Large Pleistocene ice sheets have produced glacial structures both at and below the surface in northern Europe. Some of the largest and most erosive structures are so-called tunnel valleys (TVs): large and deep channels (typically up to 5 km wide and up to 400 m deep, with lengths up to 100 km), which formed below ice sheets. Although the subject of many studies, the details of their formation and fill are still not well understood. Here, we present an update on the distribution of TVs in the southeastern North Sea between Amrum and Heligoland based on a very dense grid of high-resolution 2D multi-channel reflection seismic data (400 m line spacing). The known tunnel valleys (TV1–TV3) in that area can now be traced in greater detail and further westwards, which results in an increased resolution and coverage of their distribution. Additionally, we were able to identify an even deeper and older tunnel valley, TV0, whose orientation parallels the thrust direction of the Heligoland Glacitectonic Complex (HGC). This observation implies a formation of TV0 before the HGC during an early-Elsterian or pre-Elsterian ice advance. For the first time, we acquired high-resolution longitudinal seismic profiles following the thalweg of known TVs. These longitudinal profiles offer clear indications of an incision during high-pressure bank-full conditions. The fill indicates sedimentation in an early high-energy environment for the lower part and a subsequent low-energy environment for the upper part. Our results demonstrate that a very dense profile spacing is required to decipher the complex incisions of TVs during multiple ice advances in a specific region. We also demonstrate that the time- and cost-effective acquisition of high-resolution 2D reflection seismic data holds the potential to further our understanding of the incision and filling mechanisms as well as of the distribution, complexity and incision depths of TVs in different geological settings.

Kurzfassung: Große pleistozäne Eisschilde haben in Nordeuropa glaziale Strukturen sowohl an der Oberfläche als auch unter der Oberfläche hinterlassen. Einige der größten und erosivsten Strukturen sind sogenannte Tunneltäler (TV); große und tiefe Rinnen (typischerweise bis zu 5 km breit, bis zu 400 m tief, mit einer Länge von bis zu 100 km), die sich unter Eisschilden gebildet haben. Die Bedingungen ihrer Entstehung und Verfüllung sind jedoch noch immer nicht genau verstanden. Hier zeigen wir eine Erweiterung der Verteilung von Tunneltälern in der südöstlichen Nordsee zwischen Amrum und Helgoland auf der Grundlage eines sehr dichten Netzes von hochauflösenden 2D-Mehrkanal-Reflexionsseismikdaten (400 m Profilabstand). Die bekannten Tunneltäler (TV1–TV3) in diesem Gebiet können nun detaillierter und weiter westlich verfolgt werden, was zu einer erhöhten Auflösung und Abdeckung ihrer Verteilung führt. Darüber hinaus konnten wir ein tieferes und älteres Tunneltal TV0 identifizieren, dessen Ausrichtung parallel zur Schubrichtung des Helgoland Glazialtektonischen Komplex (HGC) verläuft. Diese Beobachtung deutet auf eine Entstehung von TV0 vor dem HGC während eines früh- oder vor-elsterzeitlichen Eisvorstoßes hin. Zum ersten Mal haben wir hochauflösende seismische Profile genau entlang des Verlaufs der bekannten Tunneltäler aufgenommen. Diese Längsprofile liefern eindeutige Hinweise auf einen Einschnitt unter hohen Drücken. Die Füllung deutet auf eine Sedimentation während einer frühen hochenergetischen Phase für den unteren Teil und einer nachfolgenden niederenergetischen Phase für den oberen Teil hin. Unsere Ergebnisse zeigen, dass ein sehr dichter Profilabstand erforderlich ist, um die komplexen Einschnitte von Tunneltälern während mehrerer Eisvorstöße in einer ausgewählten Region zu entschlüsseln. Wir zeigen auch, dass die zeit- und kosteneffiziente Erfassung von hochauflösenden 2D-reflexionsseismischen Daten das Potenzial hat, unser Verständnis für die Erosions- und Füllmechanismen sowie für die Verteilung, Komplexität und Einschnitttiefen von Tunneltälern in verschiedenen geologischen Umgebungen zu verbessern.

1 Introduction

Extensive seismic studies have shown a high abundance of subglacially produced channels – so-called tunnel valleys (TVs) – in many regions of the North Sea (Fig. 1; Huuse and Lykke-Andersen, 2000; Lonergan et al., 2006; Kristensen et al., 2007; Lutz et al., 2009; Andersen et al., 2012; Hepp et al., 2012; Stewart et al., 2013; Lohrberg et al., 2020; Kirkham et al., 2021). TVs typically have widths of 1 to 5 km, incision depths of up to 400 m and lengths of up to 100 km (van der Vegt et al., 2012). Widths of up to 10 km and depths of up to 500 m have been observed locally in the North Sea (Ottesen et al., 2020). In the German sector of the North Sea in particular, Lutz et al. (2009) reported lengths of up to 60 km, widths of up to 8 km and depths of up to 400 m. Most of these TVs are now filled with sediments and often buried beneath a drape of glacial and interglacial deposits. The evaluation of their distribution with respect to different subsoil conditions may provide details on their incision process assuming that glacial conditions may have been comparable for larger regions of the North Sea. Constraints on the maximum incision depth and widths of TVs are needed to evaluate the long-term stability and safety of subsurface storage sites in formerly glaciated terrain due to the potential of direct hydraulic connections to otherwise sealed systems as a consequence of erosion and subsequent filling. Considering that glaciations of similar magnitude to those of the Pleistocene are likely to occur in the future (Loutre and Berger, 2000), this factor has increased in significance during the site selection process for radioactive waste. This is due to the fact that radioactive waste has half-lives of millions of years, such that German legislation requires safe storage for at least 1 million years.

Many stages of advancing and retreating ice sheets are known from the Pleistocene and correlated with the so-called marine isotope stages (MISs) derived from the analysis of ice

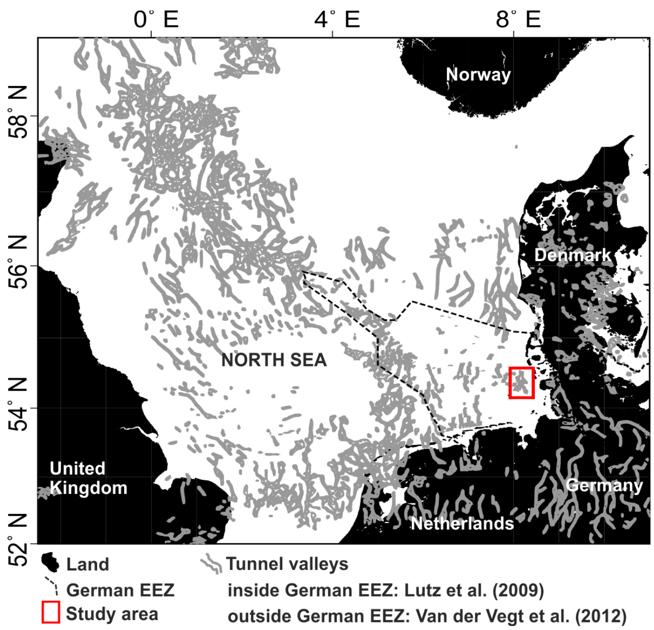

Figure 1. Location of the study area and distribution of tunnel valleys (TVs) in the North Sea and adjacent countries (adapted from Lohrberg et al., 2020).

and sediment cores all over the globe. Using the MISs, the remnants of glacially produced structures were used to correlate the cover of ice sheets in different regions during different stages of glaciations. In particular, the Scandinavian Ice Sheet (SIS) advanced into the North Sea during MIS 2 (Weichsel), MIS 6–8 (Saale), MIS 10 (Elster) and possibly MIS 16 (Cromer) (Ehlers, 1990; Ehlers et al., 2011; Batchelor et al., 2019). Only the latest glaciation during MIS 2 did not result in the full coverage of the North Sea as evidence for Weichselian ice sheets is missing in the southeastern North Sea (Batchelor et al., 2019).

Based on 3D seismic data, several authors showed that TVs often cross-cut and that they exist in different stratigraphic levels, such that a sequence of their formation can be derived (Kristensen et al., 2007, 2008; Stewart and Lonergan, 2011; Kirkham et al., 2021). In fact, Stewart and Lonergan (2011) were able to show that the formation of all TVs in their study area in the central North Sea can be correlated with at least seven phases of ice advance-and-retreat cycles in different directions. These results imply that TVs tend to form in multiple phases during a number of ice advances. As a consequence, a clear attribution of single TVs to the maximum extents of the three major ice advances in the North Sea (i.e. Elster, Saale and Weichsel) is not feasible. Instead, the occurrence of different phases of TVs likely represents a multitude of different ice advances and ice lobes in a specific area (Kehew et al., 1999).

The multitude of extensive 3D and high-resolution 2D seismic data provided good detail on the distribution of TVs in the North Sea (Huuse and Lykke-Andersen, 2000; Kristensen et al., 2007; Lutz et al., 2009; Stewart et al., 2013; Ottesen et al., 2020). Yet, this distribution has wide blank spots where costly 3D seismic data are not available and where 2D seismic data were acquired with a focus on potentially hydrocarbon-bearing structures hundreds of metres deeper than Pleistocene sediments. Extensive high-resolution 2D seismic studies using dense profile spacing of 800 m or less were able to show that a high-density 2D approach is well suited to assessing buried TVs (Hepp et al., 2012; Lohrberg et al., 2020). Here, we provide an update of the tunnel valley distribution between Amrum and Heligoland in the southeastern North Sea (Lohrberg et al., 2020; Fig. 2a). For the first time, we have acquired longitudinal seismic profiles following the thalweg of known TVs to image their fill over several kilometres in high resolution and to evaluate existing hypotheses for the incision and filling mechanism.

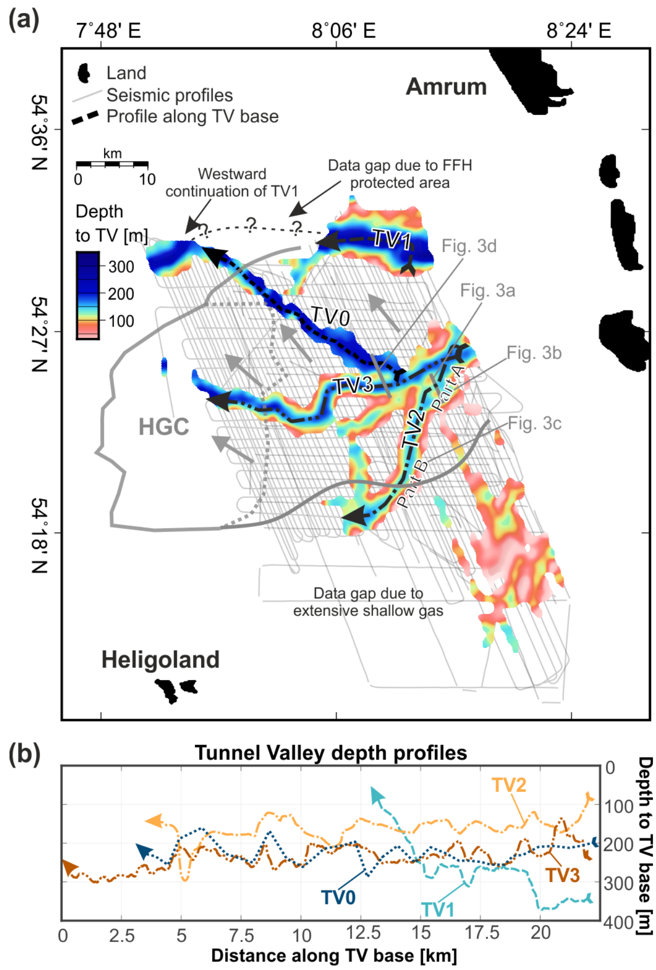

Figure 2. Overview of the study area. **(a)** Detailed distribution of large tunnel valleys (TVs) in the study area in the form of a depth grid based on all available high-resolution 2D reflection seismic profiles (with the sea level as depth reference). The limits of the Heligoland Glacitectonic Complex (HGC) are plotted for reference, where the arrows indicate the thrust direction towards the northwest (Winsemann et al., 2020; Lohrberg et al., 2022). **(b)** Depth profiles for the base of TV1–TV3 inferred from longitudinal seismic profiles following the thalweg of the TVs. Note that the depth profile for TV0 has been generated from the updated tunnel valley grid as it was unknown during data acquisition.

2 Materials and methods

We acquired 2D reflection seismic data to image subsurface structures and landforms from 0 to 800 ms two-way travel time (approximately 600 m) beneath the seafloor between Amrum and Heligoland in the southeastern North Sea. During cruise AL496 with R/V *Alkor* in July 2017, we acquired a total of 1058 km of 2D high-resolution multi-channel reflection seismic data in a closely spaced 2D grid covering approximately 800 km^2 with a mean profile spacing of 800 m. During cruise MSM98/2 with R/V *Maria S. Merian* in February 2021, we acquired an additional 1600 km of 2D high-resolution multi-channel reflection seismic data filling the gaps of the previous survey (AL496) to achieve a combined line spacing of approx. 400 m and to extend the data set westwards. Based on earlier results (Lohrberg et al., 2020), we were able to plan selected survey lines precisely along the thalweg of known TVs to produce longitudinal seismic profiles.

The acquisition setup and data processing were highly similar for both surveys, and details are described in Lohrberg et al. (2020, 2022). We used a sound velocity of 1600 m s^{-1} for the depth conversion for lack of a detailed velocity model, and depths are provided in reference to the water level. The frequencies used to image the subsurface range between 70 and 1000 Hz with the main frequency around 300 Hz, which results in a vertical resolution in the metre range for the upper tens of metres below the seafloor, whereas the resolution decreases with increasing penetration due to the absorption of high frequencies.

The fully processed seismic sections were loaded into IHS Markit Kingdom software (2018) for interpretation. The Kingdom software was used to trace the horizons and the morphology of the subsurface landforms. The horizons were then exported for interpolation and plotting using the Generic Mapping Tools (GMT) open-source software.

3 Results

We updated our previous results for the distribution of TVs for the study area (Lohrberg et al., 2020) using the newly acquired data, which resulted in an updated map (Fig. 2a). Due to the increased profile density, we were able to close gaps that previously led to an ambiguous tracing of the TVs. Although similar to the results of Lohrberg et al. (2020), the updated map allows for the tracing of previously identified TV1, TV2 and TV3 in greater detail and approx. 10 km farther towards the west. Based on the updated map, we are confident that we imaged the westward continuation of TV1 in the northwest of the study area (Fig. 2a). Furthermore, TV2 can now be traced along its thalweg over 17 km and TV3 can now be traced along 22 km (Fig. 2b). Owing to the increased profile density, we were able to identify a new tunnel valley, TV0, which lies in a deeper stratigraphic level than TV1–TV3. The average depth of its thalweg is around 250 m, and the thalweg shows minor undulation. The fill of TV0 shows less stratification, yet it can be separated into a lower and an upper part in some profiles, despite strongly undulating reflectors in its upper part. Its flanks show more gentle slopes than the flanks of TV1–TV3, and both its beginning and its end have been crossed and eroded by other TVs. In contrast to TV1–TV3, TV0 does not show clear “shoulders” at its top, and its orientation differs. TV0 is oriented in a SE–NW direction and therefore parallels the identified thrust direction of the Heligoland Glacitectonic Complex (HGC) towards the northwest postulated in that area (Fig. 2; Winsemann et al., 2020; Lohrberg et al., 2022).

For the first time, we were able to generate high-resolution seismic profiles oriented along the thalweg of TVs by planning profiles based on earlier results (Lohrberg et al., 2020). Figure 3a shows a typical transverse profile of TV2 in which the TV infill can be delineated into two parts based on its acoustic properties (TV2f1 and TV2f2). Figure 3b and c display longitudinal profiles following TV2’s thalweg, in which we traced the base of the two fills (TV2e1 and TV2e2). From Fig. 3b and c it is evident that TV2’s thalweg (TV2e1) significantly undulates along the profile. The depth profiles for the thalwegs of TV1–TV3 are shown in Fig. 2b for comparison, and Table 1 provides basic values for the incision depth with respect to sea level. Based on our data, it is clear that TV2 incised through Miocene and late Paleogene strata while overcoming significant barriers of over 75 m height difference during incision (Fig. 3b and c). Part A of the longitudinal profile following TV2 (Fig. 3b) shows increased undulation and partly very deep incision of the thalweg. The dip of the thalweg parallels the dip of the underlying strata in limited areas with frequent deeper incisions on several occasions (Fig. 3b and c). The base of the upper fill (TV2f2) follows this trend and undulates more strongly in areas of deeper incision (Fig. 3b). Part B further towards the west of the study area shows slightly less undulation of the thalweg and an almost flat base of the upper fill (TV2f2; Fig. 3c). Neither TV2f1 nor TV2f2 shows a distinct stratigraphic pattern, except for a faint layering, which follows TV2e2. Excluding the undulation of the thalweg (TV2e1), the mean incision depth of TV2 is close to 160 m over the whole course of the TV, despite a significant dip of the underlying strata towards the west (Fig. 3b and c).

Table 1. Basic values for the incision depth of the largest TVs in the study area with respect to sea level.

	Minimum depth (m)	Mean depth (m)	Maximum depth (m)
TV0	161	226	288
TV1	73	271	372
TV2	75	162	297
TV3	110	237	301

4 Discussion and conclusions

Different incision mechanisms have been described for TVs, and there has long been a debate about whether the incision could be explained with fluvial erosion (Donovan, 1972; Salomonsen, 1995; Sørensen and Michelsen, 1995) or whether pressurized subglacial erosion is the more likely mechanism (Jentzsch, 1884; Kuster and Meyer, 1979; Piotrowski, 1994). Our high-resolution longitudinal profiles along a single-incision TV show that the incision has overcome significant morphological highs of over 75 m over a 500 m distance, which is a significant gradient when compared to earlier studies by Stewart et al. (2013), who showed a maximum gradient of approx. 100 m over a 1 km distance. These large gradients are impossible to reconcile with a gravity-driven fluvial erosion, as any water flow would stagnate at either of the morphological barriers (Ó Cofaigh, 1996; van der Vegt et al., 2012). Therefore, we conclude that fluvial erosion is not the primary process that led to the formation of the TVs in our study area. Instead, we conclude and confirm interpretations that bank-full pressurized drainage beneath an ice sheet was responsible for their formation (Piotrowski, 1994; Huuse and Lykke-Andersen, 2000) as bank-full conditions would be capable of overflowing such barriers. High hydraulic heads beneath kilometre-thick ice sheets in combination with abrasion at their base provided a high

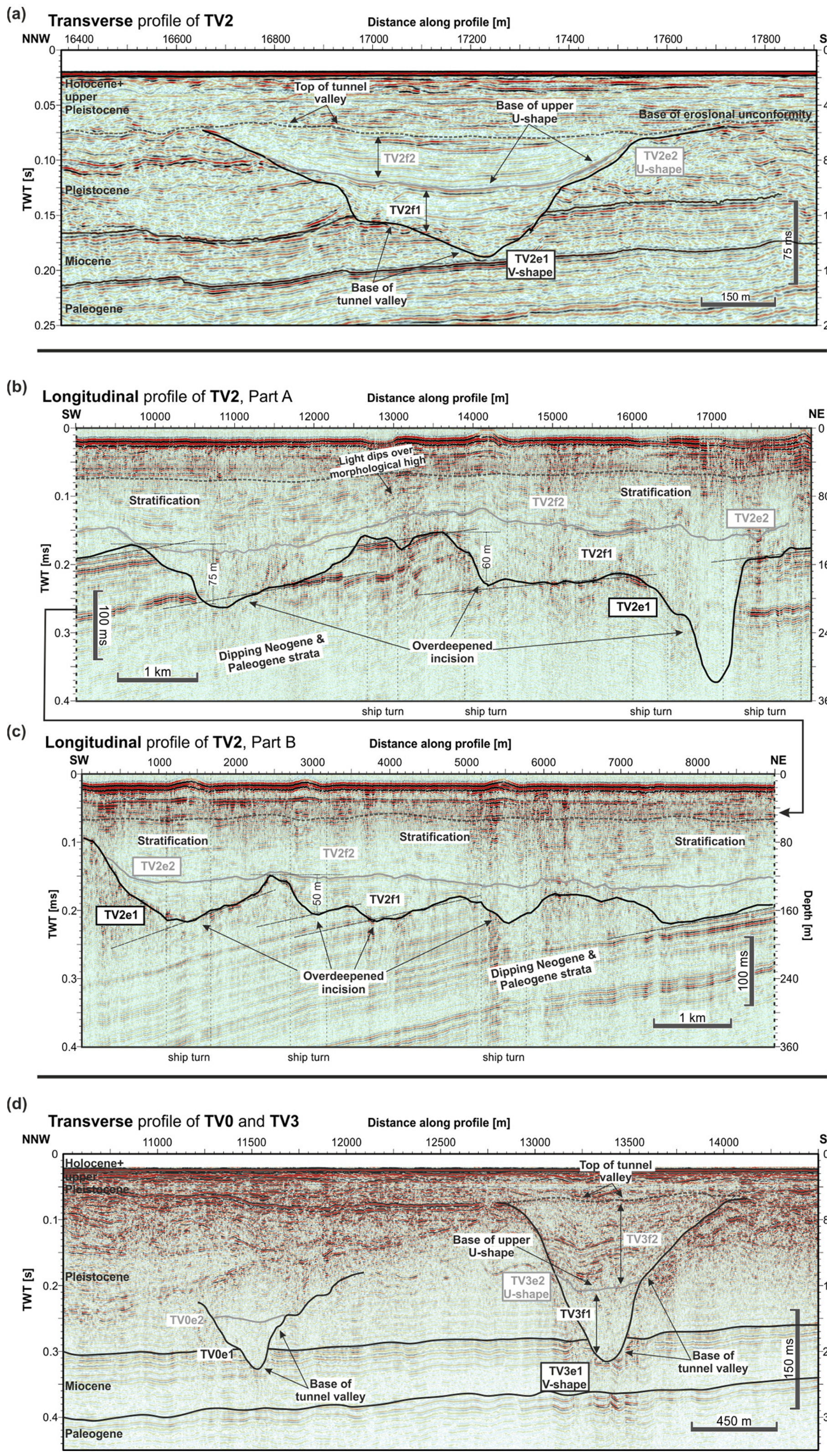

Figure 3. Seismic sections showing transverse and longitudinal profiles of a tunnel valley (TV2). The location of the profiles is indicated in Fig. 2a. **(a)** Typical transverse profile imaging TV2 perpendicular to its thalweg. **(b)** Part A and **(c)** Part B of a longitudinal profile following the thalweg of TV2 known from earlier results. **(d)** Transverse profile imaging TV0 and TV3 perpendicular to their thalweg. Turns of the ship needed to follow the TV introduce a slight decrease into the imaging quality and are marked in the figure. TV#e# refers to the respective erosional base of the respective upper or lower fill of the TV denoted as TV#f#, where # denotes a number.

efficiency for the erosion of the underlying strata. Based on the morphology of the thalweg and on the lower fill of the TVs alone, we are unable to answer whether the incision followed a "catastrophic" or rather a longer-term "steady-state" erosion process. Yet, the strongly undulating thalweg and barriers can be seen as strong indications of different stages of catastrophic erosion during different cycles of glacier retreat and advance, which would directly control the meltwater pressures at the base. Because of these observations, we consider a catastrophic meltwater outburst scenario more likely than steady-state erosion of the thalweg.

Following their incision, most TVs of the North Sea have been filled. Contrary to our expectation, we do not observe patterns that are a diagnostic for specific sedimentary processes in the fill of the TVs along their thalweg, except for a segmentation into an upper part with increased stratification and a lower part with decreased or absent stratification (TV2f1, TV2f2; Fig. 3b and c). In particular, we do not observe clinoforms or other structures postulated before as a diagnostic for the process of "backfilling" (Praeg, 1996), which refers to the near-simultaneous up-ice-directed meltwater erosion and deposition of the soil during ice sheet retreat. Rather the increased stratification in the fill's upper part (TV2f2) indicates a contemporaneous filling of the valley in a low-energy environment, possibly paralleling water level rise during melting ice sheets (Piotrowski, 1994; Huuse and Lykke-Andersen, 2000; Stewart et al., 2012; van der Vegt et al., 2012). Considering the decreased stratification in the lower part of the fill (TV2f1), we consider it likely that the early/lower fill of the TVs was deposited in a high-energy environment, which explains larger and less sorted grain sizes, such as coarse sands, gravels and boulders, being often observed in drilling campaigns in northern Germany (Hepp et al., 2012). The absence of glaciotectonic deformation along the thalweg precludes the sedimentation of TV2f1 and TV2f2 during the ice advance (van der Vegt et al., 2012). Consequently, we see evidence only for the hypothesis of filling during ice retreat in a glaciomarine or glaciolacustrine environment.

Owing to the increased density of seismic profiles, we were able to identify the hitherto unknown deep tunnel valley TV0. Likely due to a subsequent overriding ice sheet, TV0's shoulders were eroded, and only its middle and lower parts are preserved. Yet, its NW–SE orientation can be clearly derived from the data. This orientation parallels the thrust direction postulated for the HGC (Winsemann et al., 2020; Lohrberg et al., 2022). Combining its deeper stratigraphic location with thrusts occurring stratigraphically higher/later than the top of TV0, it seems likely that its incision dates back to the same ice lobe responsible for the formation of the HGC or is older (Fig. 2a). The clear separation from TV1–TV3, both in depth and in orientation, also indicates a substantial time between the formation of TV0 and TV1–TV3 as observed for other TVs in the North Sea (Stewart and Lonergan, 2011). These observations further strengthen the conclusion that an early-Elsterian or pre-Elsterian ice lobe reached the study area from the southeast (Winsemann et al., 2020; Lohrberg et al., 2022). Consequently, these chronospatial relationships indicate that TV0 may reflect this early-Elsterian or pre-Elsterian ice advance into the southeastern North Sea. Considering recent modelling results for global ice sheets (Fig. 4; Batchelor et al., 2019), we hypothesize that TV0 may be a relic of the Cromer glaciation (MIS 16). These considerations show that the distributions and the depth of TVs are major factors when trying to attribute their formation to specific glaciations in specific regions. Furthermore, the anomalous fill of TVs dictates that detailed knowledge of their distribution and depth as well as sedimentological composition is critical for offshore operations, such as construction sites for wind turbines or infrastructure for carbon capture and storage (CCS).

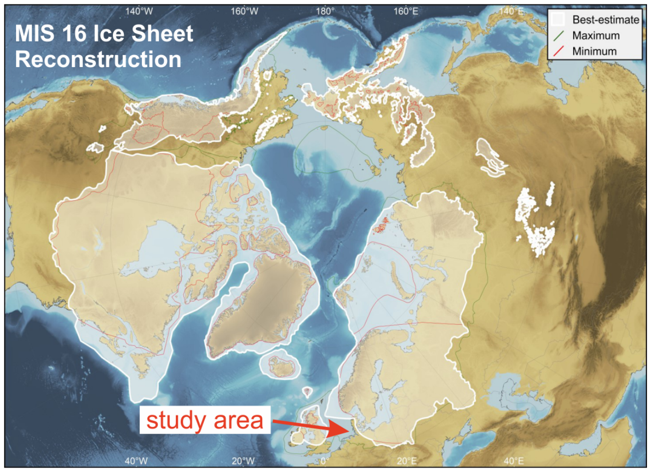

Figure 4. Reconstruction of MIS 16 ice sheets for the Northern Hemisphere (adapted from Batchelor et al., 2019).

It is relevant to understand the mechanisms of erosion and filling of TVs to understand their impact on the subsurface during future glaciations. In particular, the maximum incision depth of TVs is the most important number for radioactive waste repositories, as a potential storage location needs to be protected from erosion during climate extremes in the upcoming 1 Myr. To find this number, as many TVs as possible have to be examined for their incision depth. Our examples are representative of a large area of the North Sea where TVs have incised Neogene sands and clays, thereby providing an estimate of the maximum incision depth in Cenozoic sediments.

Equally important is the distribution of TVs as their fill may act as aquifers and thus lead to hydraulic connections between otherwise isolated aquifers in greater depths (BURVAL Working Group, 2009). Our results show that a very dense profile spacing and high resolution of reflection seismic data are essential to decipher the complexity and distribution of TVs for a specific region. In this regard, the possibility of acquiring such data in the marine realm is unmatched by land-based measurements. Therefore, marine seismic surveys offer a unique opportunity to increase our understanding of the formation, filling and distribution pattern of TVs. This opportunity should be further exploited to answer open questions, such as a potential spatial correlation of TVs with widespread salt tectonics, and to improve our understanding of these highly erosive features to ensure the long-term protection of radioactive waste repositories.

Data availability. The data that support the findings of this study are available from the corresponding author upon request. The data will also be made publicly available through the PANGAEA data repository.

Author contributions. AL wrote the article, guided data acquisition and processed the data used for the article. JSvD and SK led the data acquisition, which was carried out by AL, HG and KFL. HG and SK provided valuable input for the streamlining of the article. The project administration was handled by JSvD and SK. All authors contributed valuable comments, discussions and guidance for the improvement of this article.

Competing interests. The contact author has declared that none of the authors has any competing interests.

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