Glacial lakes in the Torneträsk region, northern Sweden, are key to understanding regional deglaciation patterns and dynamics

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Preface

This Master’s thesis is Karlijn Ploeg’s degree project in Physical Geography and Quaternary Geology at the Department of Physical Geography, Stockholm University. The Master’s thesis comprises 60 credits (two terms of full-time studies).

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Abstract

The prospect of sea level rise due to melting ice sheets affirms the urgency of gaining knowledge on ice sheet dynamics during deglaciation. The Fennoscandian Ice Sheet serves as an analogue, whose retreat can be reconstructed from the geomorphological record. The recent development of a high-resolution LiDAR-derived elevation model can reveal new relationships between landforms, even for well-studied areas such as the Torneträsk region in northwestern Sweden. Therefore, this study aims to refine the reconstruction of the deglaciation in this region based on an updated glacial geomorphological map. A range of glacial landforms were mapped, which by means of an inversion model were utilized to form swarms representing spatially and temporally coherent ice sheet flow systems. Additionally, glacial lake traces allowed for the identification of ice margins that dammed lakes in Torneträsk, Rautasjaure, and other (former) lake basins. Eight glacial lake stages were identified for the Torneträsk basin, where final drainage occurred through Tornedalen. Over 20 glacial lake stages were identified for the Rautasjaure basin, where drainage occurred along the margins of a thinning ice lobe. The disparity between the glacial lake systems results from different damming mechanisms in relation to the contrasting topography of the basins. A strong topographic control on the retreat pattern is evident, as the ice sheet retreated southward in an orderly fashion in the premontane region, but disintegrated into ice lobes in the montane region. The temporal resolution of current dating techniques is insufficient to constrain the timing of ice retreat at the spatial scale of this study. Precise dating of the Pärvie fault would pinpoint the age of the ice margin which at the time of rupture was located between two glacial lake stages of Torneträsk. Collectively, this study provides data for better understanding the final retreat of the ice sheet and associated processes, such as interactions between glacial lakes and ice dynamics.
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1 Introduction

In recent decades, anthropogenic climate warming has caused substantial ice mass loss from ice sheets and mountain glaciers (IPCC, 2021). Together, ice sheets and glaciers were the dominant contributors to global mean sea level rise between 2006 and 2018 (IPCC, 2021). Moreover, ice sheets are most capable of causing abrupt change in sea level (Lenton et al., 2019; Golledge et al., 2019). Sea level rise is already challenging low-elevation coastal zones with increased risks of flooding, saltwater intrusion, and coastal erosion (Nicholls and Cazenave, 2010), but there are more global environmental consequences predicted for the 21st century, such as changes in ocean mixing and atmospheric circulation (Golledge et al., 2019).

The prospect of accelerated ice sheet melt due to predicted future warming affirms the urgency of gaining more knowledge on ice sheet dynamics, especially during deglaciation. However, monitoring and research on the Greenland Ice Sheet (GrIS) and the Antarctic Ice Sheet (AIS) has only been conducted over a short time period. Information on ice sheet evolution, which is critical for understanding and predicting the response of the current ice sheets to global warming, must therefore be attained from past glaciations.

Large ice sheets have repeatedly covered the northern hemisphere, particularly during the last 2.6 million years (Ma) (Batchelor et al., 2019). The Laurentide Ice Sheet (LIS) and the Eurasian Ice Sheet (EIS) were under similar topographic conditions at their last maximum extents as the current ice sheets are (Stroeven et al., 2016), and therefore serve as a convenient analogue to the GrIS and AIS. Gaining knowledge of past glaciations can be achieved through paleoglaciology, which is the study of reconstructing the extent and dynamics of former glaciers and ice sheets from the geomorphological record (Stroeven et al., 2021).

Initially, paleoglaciological reconstructions of northern hemisphere ice sheets were predominantly based on the geomorphological and sedimentological record (e.g., Boulton and Clark, 1990; Kleman et al., 1997). The development of a coherent glacial inversion model (Kleman and Borgström, 1996) was crucial for deriving ice flow patterns and ice sheet configurations from landforms. The field experienced major advances when new remote sensing techniques, such as Light Detection and Ranging (LiDAR), and geographical information systems (GIS) were developed, which allowed for systematic mapping of landforms (Stokes et al., 2015). Needless to say, landforms are crucial for detailing the glaciation history. However, the generalization of landforms, which is necessary to create an ice sheet-wide reconstruction, might obscure details at a level of interest to researchers.

Paleoglaciological reconstructions are increasingly achieved through ice sheet-modeling. Ice sheet model outputs appear to be highly sensitive to the choice of climate forcing (Seguinot et al., 2014), which necessitates model performance assessments against geomorphological reconstructions (Seguinot et al., 2014; Patton et al., 2017; Ely et al., 2021). Empirical evidence of the extent and behavior of former ice sheets is steadily growing.
to contribute to a database that can be used for model validation. As such, empirical reconstructions are still highly relevant, as they continuously refine the level of detail and enhance knowledge on ice dynamics, as shown by recent publications on retreat of the LIS (e.g., Margold et al., 2018), the Cordilleran Ice Sheet (CIS) (e.g., Dulfer et al., 2022), and the Fennoscandian Ice Sheet (FIS) (e.g., Stroeven et al., 2016; Hughes et al., 2016).

The Torneträsk region, previously covered by the FIS, has interested researchers from a range of disciplines, for which the Abisko Scientific Research Station is especially valuable as a logistical base (Jonasson et al., 2012). The region has been one of Sweden’s most important areas for studying geomorphology (Jonasson et al., 2002). In the early 20th century the geomorphology and glacial geology in the area were investigated by Sjögren (1909), with a special focus on the glacial lake shorelines of Torneträsk (Sjögren, 1908). In the 1950s multiple studies on the deglaciation in the region were conducted by Holdar (1952, 1957, 1959). In the late 1970s a geomorphological mapping project started with the aim of identifying valuable areas across the Swedish mountains for conservation and scientific purpose. The maps covering the Torneträsk region were produced by Melander (1977a,b), who continued research on the deglaciation and glacial lakes (Melander, 1977c, 1980). In 2011 the Swedish Geological Survey (SGU) finished a map on soil types, which includes Quaternary deposits, but is limited in mapped landforms (SGU, 2020). As of now, the geomorphological maps by Melander (1977a,b) are the most recent detailed maps.

In this study the deglaciation dynamics of the FIS are revisited for the Torneträsk region in northern Sweden. Using the new LiDAR-based elevation model provided by Lantmäteriet, landforms are mapped that detail the retreat history of the FIS. The following sections will introduce the history of the FIS, the importance of glacial lakes in reconstructions, and the use of LiDAR for mapping landforms. The final sections of the introduction will elaborate on which research gaps are required to be addressed and the objectives of this study.

1.1 Fennoscandian Ice Sheet

The glacial history of Fennoscandia started 2.6 Ma ago following a shift in climate, which initiated a period characterized by approximately 40 glacial cycles driven by the earth’s orbital parameters (Mangerud et al., 1996; Kleman and Stroeven, 1997; Kleman et al., 2008). The Quaternary deposits and geomorphology, which reflect spatial patterns of erosion and deposition, indicate there were two modes of ice sheet configurations during this period: elongated west-centered mountain ice sheets (MIS), and large east-centered ice sheets (similar to the FIS) (Kleman and Stroeven, 1997; Kleman et al., 2008). The latter configuration covered Fennoscandia during the late Weichselian, also known as the last glaciation (Kleman et al., 2008). FIS-style ice sheets were the dominant configurations during the late Quaternary, while MIS-style ice sheets were dominant during the early and middle Quaternary (Kleman et al., 2008).
Figure 1.1: a) Map from Hughes et al. (2016) showing the maximum extent of the Eurasian Ice Sheet (EIS) constituting the British-Irish Ice Sheet (BIIS), Scandinavian Ice Sheet (SIS; same as FIS), and the Svalbard-Barents-Kara Ice Sheet (SBKIS). b) Map from Stroeven et al. (2016) showing the deglaciation pattern and chronology of the FIS.

The FIS was the largest sector of the EIS during the last glaciation, with its maximum extent connecting the British-Irish Ice Sheet (BIIS) to the Svalbard-Barents-Kara Ice Sheet (SBKIS) (Figure 1.1, Hughes et al., 2016). The time when the ice sheets reached their greatest extent is referred to as the Last Glacial Maximum (LGM), although the precise timing of the LGM is likely to have varied across different sectors of the individual ice sheets (Böse et al., 2012; Hughes et al., 2016; Stroeven et al., 2016). The three ice sheets were coalesced for a relatively short period considering the total glacial cycle, as they disintegrated into separate ice masses following the LGM (Stroeven et al., 2016). The most recent reconstructions of the deglaciation of the FIS and EIS are conducted by Stroeven et al. (2016) based on geomorphological and geochronological data, and by Patton et al. (2017) based on an ice sheet model validated against empirical data, although multiple attempts to establish the retreat patterns of the FIS precede their work (Lundqvist, 1986; Lundqvist and Saarnisto, 1995; Kleman et al., 1997; Boulton et al., 2001).

Different sectors of the FIS display varying patterns of retreat, revealing variations in climatic conditions, oceanic influence, topographical settings, basal thermal regime, and marine versus terrestrial termination (Figure 1.1b, Stroeven et al., 2016; Patton et al., 2017). From its LGM extent retreat occurred mainly along the southeastern margin of the FIS, while initial eastward-retreating ice along the western margin pinned itself at the Norwegian coast (Stroeven et al., 2016; Patton et al., 2017).

The southeastern margin showed initial slow, step-wise retreat, yet accelerated once the terrestrial margin transitioned into a calving margin in the Baltic Basin (Stroeven et al., 2016; Patton et al., 2017). Retreat was occasionally interrupted by stand-stills and minor re-advances, especially along the dynamic western margin (Hughes et al., 2016; Stroeven...
et al., 2016; Patton et al., 2017). A major re-advance occurred during the Younger Dryas, mainly along the southern margin (Stroeven et al., 2016; Patton et al., 2017). After the eastern ice margin retreated onshore, continuous terrestrial retreat resulted in an elongated ice ridge along the Scandinavian mountains from south-central Norway to northern Sweden (Stroeven et al., 2016; Patton et al., 2017). By now the ice margin was probably highly lobate, while the thinning ice sheet melted out mountain peaks as nunataks (Hughes et al., 2016). During the early Holocene the remaining FIS split into two separate ice masses over the southern Norwegian mountains and the northern Swedish mountains (Stroeven et al., 2016; Patton et al., 2017). The final ice remnant is believed to have disappeared in the eastern Sarek Mountains in northern Sweden (Stroeven et al., 2016).

The study area in northwestern Sweden is known for its palimpsest landform systems, which reflect ice sheet dynamics during the last deglaciation and prior glaciations (Kleman, 1992; Kleman and Stroeven, 1997). The existence of relict areas is attributed to the patchy erosional impact of the ice sheets owed to spatially varying basal thermal regime conditions (Stroeven et al., 2002; Goodfellow et al., 2008; Stroeven et al., 2013). The regime was largely cold-based in the region, except for the through valleys where ice flow accelerated (Stroeven et al., 2002). During MIS-style glaciations ice flow was directed eastward along the Torneträsk basin (Figure 1.2a), while during FIS-style glaciation ice flow was directed westward, where Torneträsk valley was probably one of the largest outlets of the FIS (Figure 1.2b, Stroeven et al., 2002). Deglacial flow is characterized by a shift from western ice flow, to northern ice flow, to northeastern ice flow (Figure 1.2c, Stroeven et al., 2002). The final deglaciation of the Torneträsk area occurred around 9.5 cal ka (Stroeven et al., 2002). The shoulders of the Torneträsk valley were most likely deglaciated earlier than the valley floor (Melander, 1980; Stroeven et al., 2002).

### 1.2 Glacial lakes

The existence of glacial lakes has been extremely valuable for ice sheet reconstructions. First and foremost, glacial lakes have been critical for reconstructing ice-marginal positions,
such as for the LIS (e.g., Jansson, 2003), the CIS (e.g., Perkins and Brennand, 2015), and the FIS (e.g., Lundqvist, 1972; Høgaas and Longva, 2018; Regnéll et al., 2019). Moreover, they were crucial for identifying the final ice remnant of the FIS, as glacial lake traces along the Scandinavian Mountains must have formed between the water divide in the west (running along the Scandinavian Mountains) and a retreating ice sheet to the east (e.g., Svenonius, 1898; Lundqvist, 1972; Regnéll et al., 2019). Furthermore, differential glacio-isostatic uplift patterns have been inferred from tilted raised shorelines from former glacial lakes (e.g., Pässe, 1998; Dawson et al., 2002; Rayburn and Teller, 2007). Differential uplift changed lake hydrology, for example by shifting outlets and influencing the size and outline of lakes, and continues to do so nowadays (Seppä et al., 2012). Torneträsk has its length axis approximately parallel to the tilting direction, which renders it especially sensitive to the impacts of tilting.

Glacial lake shorelines of Torneträsk have been described as early as the 19th century (Gumælius, 1876; Svenonius, 1898). Sjögren (1908) proposed three stages of glacial lakes in the Torneträsk valley as a consequence of east-ward ice retreat: smaller ice-marginal lakes between the ice and valley walls, an open glacial lake that drained partly through Sördalen, and an even larger glacial lake before final drainage in the east. Holdar (1952) disputed the idea of an open glacial lake, instead suggesting a dead ice body remained in Torneträsk valley with merely ice-marginal lakes along the sides. Melander (1977c) recognized five stages of glacial lakes, although both open lakes (Melander, 1977c) and smaller ice-marginal lakes (Melander, 1980) remained possible explanations for the shoreline distribution. Several potential outlets of the glacial lake were proposed. Namely, an outlet west of Vassijaure with drainage towards Rombaksfjorden, an outlet at Pålnoviken with drainage via Sördalen into Norway, and a final outlet somewhere to the east of Torneträsk (Svenonius, 1898; Sjögren, 1908; Melander, 1977c). Additionally, Melander (1977c) proposed that subglacial drainage also occurred.

Besides its value for ice sheet reconstructions, glacial lakes hold information on the effect of ice-marginal lake formation on ice sheet dynamics. Carrivick and Quincey (2014) observed an increase in ice-marginal lakes along the western ice margin of the GrIS over the last decades, which is alarming considering lacustrine-terminating ice margins appear to retreat faster than land-terminating margins (Mallalieu et al., 2021). The knowledge of interactions between ice-marginal lakes and ice sheet dynamics is limited, thus necessitating research on former glacial lakes to accurately predict the response of current ice sheets.

1.3 The advent of LiDAR

LiDAR is an active remote sensing technology based on the principle of measuring the round-trip time of an emitted laser pulse between the sensor and the target surface (Liu, 2008; Höfle and Rutzinger, 2011). Airborne LiDAR scanning is done from a laser scanning system on-board of a fixed wing aircraft or a helicopter. The round-trip time of the laser
pulse is used to calculate the distance between the sensor and the target surface, while a combination of a Global Positioning System (GPS) and an inertial measurement unit (IMU) provide the absolute position and orientation of the system (Vosselman and Maas, 2010). The technique produces a 3D digital representation of the surface of objects and sites in the form of a point cloud with each point having an x, y, and z-coordinate (Vosselman and Maas, 2010).

Airborne LiDAR is one of the most effective and accurate methods for acquiring elevation data (Liu, 2008). The standard accuracy of the acquired elevation data is 0.05–0.20 m for height and 0.2–1.0 m for position (Vosselman and Maas, 2010). A considerable advantage of LiDAR is the ability to retrieve point measurements under forest cover, such that Digital Elevation Models (DEMs) can be created that only represent the ground surface (Höfle and Rutzinger, 2011; Dowling et al., 2013). In order to generate a DEM without objects or vegetation, the ‘bare ground’ points have to be extracted from the raw LiDAR data (Liu, 2008). This is especially useful in Sweden, where 68.7% of the land area is covered by forest (The World Bank, 2020).

Prior to the development of LiDAR, geomorphological mapping was mostly based on field investigations and aerial imagery. DEMs derived from aerial imagery were only available at a horizontal resolution of 50 m, which was even coarser or interpolated in forested areas (Smith et al., 2006). Since 2019 the LiDAR-based National Height Model, with a 2 m horizontal resolution, has been made available for the Torneträsk region by Lantmäteriet (the Swedish mapping, cadastral, and land registration authority). The high-resolution DEMs can reveal previously unidentified landforms or expose new relationships with already identified landforms, both due to the higher resolution and forest canopy penetration, such that even well-studied areas are worth exploring by the new data (Dowling et al., 2013; Johnson et al., 2015). This is already demonstrated by several recent LiDAR-based studies on glacial geomorphology (Ojala et al., 2015; Goodship and Alexanderson, 2020; Öhrling et al., 2020; Ojala et al., 2021), and glacial lake traces (Thorndycraft et al., 2016; Regnéll et al., 2019).

### 1.4 Research gaps

A number of research gaps can be identified from the literature discussed in the previous sections. The most recent detailed geomorphological maps of the Torneträsk area were produced by Melander (1977a,b). His mapping was based on interpretations from aerial photographs following extensive field verification. Because high-resolution LiDAR-based elevation data now circumvents problems inherent in the use of aerial photographs (e.g., forest cover), a reinterpretation has the promise to alter previous interpretations, clarify previous mapping uncertainties, and introduce new research gaps. GIS technology was not yet adopted in the 1970s, so it undoubtedly has the potential to improve mapping and analysis.
Reconstructions of the deglaciation of the FIS were recently updated (Hughes et al., 2016; Stroeven et al., 2016; Patton et al., 2017). However, these are largely based on interpretations based on geomorphological and geochronological data, precede the advent of LiDAR data, and address ice retreat on a continental scale. Moreover, no attempts were made to realistically reconstruct the deglaciation in the mountains due to the chaotic nature of geomorphological evidence (Kleman, 1992). Hence, the dynamics of ice sheet demise in topographically challenging terrain (e.g., Dulfer et al., 2022) remains an important research gap in Scandinavia. It is therefore important to address this issue of outdated maps by re-mapping the glacial geomorphology in the region based on the high-resolution LiDAR-based elevation data.

There has been disagreement whether the interpretation of raised shorelines along Torneträsk indicate the presence of a large open glacial lake (Svenonius, 1898; Sjögren, 1908) or smaller ice-marginal lakes (Holdar, 1952; Melander, 1980). Authors proposing that raised shorelines were formed by ice-marginal lakes impounded by dead-ice bodies adopt the assumption that all lateral meltwater channels were formed during the last deglaciation. However, Kleman (1992) argues that some landforms from previous glaciations have been preserved underneath subsequent cold-based ice sheets. Lateral meltwater channels sloping east, for example, indicate the presence of a west-centered ice sheet, the last occurrence of which predates the Late Weichselian glaciation (Kleman, 1992). These new insights contradict the previous assumptions by Holdar (1952) and Melander (1980). A re-evaluation of the landform record, which includes an evaluation of the complexities imposed by landform preservation during cold-based glaciations, will shed light on the nature of the glacial lakes.

A detailed reconstruction of glacial lakes in the Swedish Mountains by Regnéll et al. (2019) stops short of the Torneträsk basin. Although Rautasjaure, immediately south of Torneträsk, is formally included in the study by Regnéll et al. (2019), the reconstruction of the damming of this lake remains incomplete (Edward, 2020). Additionally, shorelines of Torneträsk were last mapped in large scale (1:250 000) in the 1970s by Melander (1977b), while the current resolution of LiDAR enables identification of additional smaller and less apparent shorelines. This alters the number of lake stages and the extents of the glacial lakes as previously interpreted and identified. Moreover, GIS allows for more thorough analysis of the glacial lakes, rendering it easier to identify outlets. Ultimately, refining the glacial lake reconstruction impacts the precision of reconstructed patterns of ice sheet retreat (e.g., Regnéll et al., 2019).

1.5 Research questions

This study is driven by the aforementioned research gaps concerning the deglaciation history of the Torneträsk region and the reconstruction of its glacial lakes. The aim of this study is therefore twofold: (i) to refine the reconstruction of the deglaciation of the
Torneträsk region, and (ii) to improve the reconstruction of the glacial lake systems in the Torneträsk and Rautasjaure basins. This aim results in the following objectives:

- To map glacial landforms in unprecedented detail based on a high-resolution LiDAR-derived DEM
- To identify glacial lake systems and their outlets
- To reconstruct the retreat pattern of the FIS during the last deglaciation based on the updated glacial geomorphological map and the refined reconstruction of the glacial lake systems
- To propose a time frame for the deglaciation of the Torneträsk region by evaluating literature

In meeting this aim and these objectives, the project is structured around the following key research questions:

- How did glacial lake systems of Torneträsk and Rautasjaure develop in terms of configuration and drainage routes?
- How did spatial and temporal retreat patterns of the ice sheet evolve during the last deglaciation?

The following chapters attempt to systematically address the objectives of this study and to provide answers to the key research questions. Chapter 2 provides a description of the study area in terms of its geology, physical geography, and climate. The methodology is introduced in Chapter 3, with a particular focus on the mapping strategy and the adopted inversion model. Chapter 4 presents the landform classification, which details the landforms that are crucial for the reconstruction and highlights their characteristics. The following Chapter 5 presents the results. This chapter contains the map of the glacial geomorphology of the Torneträsk region and offers spatial patterns of each essential landform. Chapter 7 constitutes an analysis of the glacial lakes and discusses the spatial and temporal retreat patterns of the ice sheet. Finally, Chapter 8 provides answers to the research questions posed in this chapter.
2 Study area

The study area is located in Kiruna municipality within the county Norrbottens län, in northernmost Sweden (Figure 2.1). The research is centered around the WNW-ESE trending valleys Torneträsk and Rautasjaure. The Torneträsk basin cuts through the Scandinavian mountain range, also known as the Scandes. The mountain range runs along the length axis of the Scandinavian Peninsula and straddles the border between Sweden and Norway. The study area is approximately 5980 km² and its outline is following the border with Norway in the northwest.

The Scandes was formed during the Caledonian orogeny through the collision of the continents Baltica and Laurentia in the mid Silurian to early Devonian after closure of the Iapetus Ocean, the ancestor of the Atlantic Ocean (Gee et al., 2008). The Caledonian rocks, essentially composed of the outer margin of the continent Baltica (Seve Nappes) and overlying sedimentary rocks deposited at the bottom of the Iapetus Ocean (Köli Nappes), were thrust over the older Baltic (or Fennoscandian) Shield in the form of nappes (Gee et al., 2010). The structure of the mountains is characterized by an east-directed thrust system, predominantly made up of amphibolites and gneisses (Figure 2.2; Kulling, 1964; Gee et al., 2008). The nappes have been subjected to high-grade metamorphism during the Caledonian orogeny (Williams and Claesson, 1987; Gee et al., 2010). The Baltic Shield, which is a segment of the East European Craton, is composed of Precambrium crystalline rocks formed during the Archean–Proterozoic Eras (Gee et al., 2010). Part of the Precambrian basement rocks have been incorporated into the nappes during the collision. The basement rocks are exposed in the low relief areas in the eastern part of the

Figure 2.1: Overview of the study area. (a) Outline of the study area in Sweden. (b) Study area of the Torneträsk region, including the border of the premontane and montane regions based on the glacial geomorphological regions of Hättestrand (1998).
area, and are dominated by respectively granite gneisses and granites (Figure 2.2; Kulling, 1964; Kleman and Stroeven, 1997; Ebert et al., 2011). The exposed basement in the study area is formed during the Svecofennian (or Svecokarelian) orogeny, which took place during the Paleoproterozoic Era. Based on relief, the study area can be divided in a montane region in the west and a premontane region in the east (Figure 2.1). The montane region has several peaks above 1000 m above sea level (a.s.l.), with the highest peak Kåtotjåkka at 1986 m a.s.l. according to the latest National Elevation Model of Lantmäteriet. The lowest elevation at 312 m a.s.l. is found along Sevujoki, draining Sevujärvi towards southeast.

The landscape has experienced extensive glacial erosion, as demonstrated by deep glacial valleys, cirques, and scoured bedrock (Kleman and Stroeven, 1997; Ebert et al., 2011; Jansson et al., 2011). In contrast, significant upland areas have been preserved, which are typically characterized by gentle slopes, round summits with occasional tors, wide shallow fluvial valleys, and open passes (Kleman and Stroeven, 1997). Although there seems to be no significant recent glacial erosion of these upland areas, considerable erosion occurred by means of subsurface weathering processes throughout the Quaternary (Goodfellow et al., 2014; Andersen et al., 2018). As such, referring to these upland areas as "preglacial" remnants with a pre-Quaternary origin is not correct (Goodfellow et al., 2008).
The premontane region is a low relief area ranging between 300 and 600 m a.s.l., which consists of extensive plains with residual hills (Lidmar-Bergström, 1995). The plains are formed through the exhumation of the Baltic shield and denudation of its surface (Lidmar-Bergström, 1995). Furthermore, the region is characterized by long and narrow lakes formed by glacially deepening and widening of river valleys (Hättestrand, 1998). The plains are found in stepped sequences along the major valleys, which is thought to relate to uplift events during the Cenozoic and consequent fluvial erosion (Lidmar-Bergström et al., 2007).

The majority of the area has exposed bedrock at the surface or only a thin till cover (<1 m), especially at the higher elevations and naturally in the scoured area in the west (Figure 2.3). The soil cover increases to depths of 20 m towards the east, where it is predominantly composed of till and to a lesser extent of glaciofluvial sediments and peat. The Torneträsk area is located within the discontinuous or sporadic permafrost zones (Brown et al., 1997). Generally, mountain permafrost is found approximately above 880 m, while permafrost at lower elevations is restricted to peat mires (Johansson et al., 2011; Sannel et al., 2016).

There is a steep precipitation gradient across the mountains with a relatively warm oceanic climate with high precepitation rates on the Norwegian coast and mountain stoss side, and a dry and cold continental climate in the east, which is intensified by the Scandes
mountains through its orographic effect on precipitation (Barnekow, 2000; Meyer-Jacob et al., 2017). The present drainage divide is located close to the border between Sweden and Norway (Kleman and Stroeven, 1997). The largest inlet at present of Torneträsk is the Abiskojäkka river, that enters at the western end of the lake. The current outlet is Tarrakoski fors (68°11’37.7”N 19°59’17.1”E), which is a rapid that drains Torneträsk to the 3 m lower water level of Alajärvi, which eventually connects to Torneälven. The current lake level of Torneträsk is at 342 m a.s.l., although the lake level varies over a meter throughout the year (SMHI, 2020). The lake has a maximum depth of 168 m according to bathymetric measurements from 1920/1921 (SMHI, 2020), but recent sonar measurements by the boat of Abisko Research Station show that its depth, in places, exceeds 190 m (A. Kristoffersson, personal communication, 26 January 2022).
3 Methods

3.1 Fieldwork

Fieldwork was conducted from 13 August to 2 September 2021. The aim of the fieldwork was to ground-truth landforms that were already identified from the LiDAR data, to develop a general understanding for the character of the landscape, and to "calibrate" the scale and appearance of the landforms as reflected by the LiDAR data through comparison to these actual landforms in the field. The field sites are concentrated around the southern shore of lake Torneträsk, given the focus on glacial lake traces, such as raised shorelines and perched deltas, and the extend of roads limiting the inspection of sites with easy access (Figure 3.1).

Figure 3.1: Overview of the locations visited during field work. a) Field stops along the routes taken by foot, car, and boat (see corresponding list of locations). b) Location of the DGPS measurements (see Table A.1 in Appendix A.)

Hiking routes were recorded on a Garmin GPS, including waypoints for locations with important observations. A hillshade relief model of the area was available on a tablet, which allowed for the identification of landforms in the field and for the posting of notes directly on the hillshade relief model. At every stop the morphology and sedimentological properties of the landform and surrounding landscape were assessed.

At location 11 the elevations of multiple shorelines were measured with a differential GPS (DGPS), as series of shorelines were found that were not or hardly identifiable on
the hillshade relief model (Figure 3.1b). The DGPS device was a Trimble Geo7X with an external antenna, for which the Trimble Business Center software was used to calculate the positions in cm-precision. Each measurement was taken at the base of the shoreline, directly below the wave-washed boulders, at the break in slope (Figure 3.2). All measurements were taken while descending a ridge at points where the raised shorelines wrapped around the ridge. The measurements can be found in Table A.1 in Appendix A.

3.2 Datasets

Different types of data were used in this study to map the glacial geomorphology of the area (Table 1). The mapping relied heavily on a DEM derived from LiDAR-data. The terrain was visualized through different techniques, such as a hillshade relief model, contours, and a slope model. Additionally, aerial imagery in the form of orthophotos and satellite imagery as visualized in Google Earth were utilized.

Lantmäteriet provides a high-resolution LiDAR-based DEM, also known as the National Height Model, ready for the end user. The horizontal resolution of the elevation data as published in 2019 is 2 m. The DEM has an absolute vertical accuracy of <0.1 m and an absolute horizontal accuracy of <0.3 m, but these can vary depending on point density of the laser scanning, time of scanning and survey technique (Lantmäteriet, 2019). The tiles are provided in 2.5 × 2.5 km and are adjusted to the SWEREF 99 TM reference system.

As of 2021 the elevation model has been upgraded to a horizontal resolution of 1 m (Lantmäteriet, 2021b), but has the same horizontal and vertical accuracy as it is based on

<table>
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<tr>
<th>Dataset</th>
<th>Temporal coverage</th>
<th>Spatial resolution</th>
<th>Source</th>
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<tr>
<td>DEM</td>
<td>2010–2018</td>
<td>2 m</td>
<td>Lantmäteriet</td>
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<tr>
<td>Orthophotos (RGB &amp; IRF)</td>
<td>2008–2018</td>
<td>0.5 m</td>
<td>Lantmäteriet</td>
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the same LiDAR data. The newer elevation model was not utilized in this project due to the larger file sizes and thus longer processing times. A comparison between both resolutions yielded that differences were imperceptible at the scale of the mapping. However, the most recent elevation model with the spatial resolution of 1 m is available through an e-service called My Map (https://minkarta.lantmateriet.se/), which was used as an additional source. Shorelines can be especially narrow on steep slopes, for which the 1 m horizontal resolution DEM can increase the certainty of correct identification.

The first mapping cycle was performed in QGIS 3.16, while the second mapping cycle was done in ArcGIS Pro 2.8.0. The DEM was processed to create a hillshade relief model. According to Chandler et al. (2018), an illumination angle of $30^\circ$ and azimuths of $45^\circ$ and $315^\circ$ are the optimal values for the visualization of hillshade relief models for the purpose of glacial geomorphological mapping. Hence, these azimuths were predominantly used for the mapping, both with and without 3 times vertical exaggeration. Smith and Clark (2005) suggest using two illumination azimuths, orthogonal to one another, and parallel and perpendicular to the dominant orientation of lineations. The dominant lineation orientation in this study is north-south, and therefore additional azimuths of $90^\circ$ and $180^\circ$ were applied. Using multiple illumination azimuths reduces the ‘azimuth bias’, which refers to linear landforms possibly being poorly visible when illuminated from certain azimuths (Smith and Clark, 2005; Chandler et al., 2018). Additionally, a slope model was derived from the DEM and contour lines were created with intervals of 10 m, 20 m, and 100 m.

Orthophotos with a resolution of 0.5 m were also provided by Lantmäteriet. Orthophotos are aerial photographs which are geometrically corrected to an orthogonal map projection, such that the scale is uniform and the orthophoto can be used to measure true distances (Lantmäteriet, 2021a). The tiles are provided in $2.5 \times 2.5$ km and are adjusted to the SWEREF 99 TM reference system. Only the natural colour (RGB 4, 3, 2) and colour infrared (RGB 5, 4, 3) orthophotos were used in the mapping project. Although the mapping was primarily DEM-based, the aerial imagery could in certain cases of doubt enhance landform detectability. Additionally, it showed whether valleys had running water, and where the terrain is affected by anthropogenic activities.

The freely available Google Earth Pro enabled visualization of the terrain in 3D, which was mostly used to cross-check the mapping based on the other imagery. The imagery of Google Earth Pro is primarily based on satellite imagery with varying resolutions. The ability to import or export Keyhole Markup Language (KML) files, which can be used in GIS software, is extra convenient.

Several geomorphological maps were used as a background map to refer to while mapping. The geomorphological maps by Melander (1977a,b) were provided by the Agency for Digital Government’s Sveriges dataportal (DIGG, 2020), who georectified the scans to RT90 2,5gV. However, the maps were not yet georeferenced, which was accomplished in QGIS 3.16 by using points on the map with known coordinates. Additionally, scanned and
georeferenced maps were available by Hättestrand (1998) of the glacial geomorphology of central and northern Sweden and by Sjögren (1908) of the ice-dammed lakes in Torneträsk and Vassijaure. Stroeven et al. (2016) provide GIS layers of the extent of the Fennoscandian ice sheet during final deglaciation, for every 100 years, the pattern of which is primarily based on glacial geomorphology, and the timing of which is constrained by radiocarbon, Optically Stimulated Luminescence (OSL), and cosmogenic nuclide exposure ages. Similarly, the DATED-1 database by Hughes et al. (2016) contains published deglaciation ages constraining the extent of the Fennoscandian ice sheet for every 1000 years. Lake polygons were retrieved from the HydroLAKES database (Messager et al., 2016) and rivers from OpenStreetMap data of Sweden (Geofabrik, 2018).

### 3.3 Mapping and analysis

Remote mapping on the scale of a regional sector of a paleo-ice sheet requires a systematic mapping approach, in order to map a large area in a time-efficient, yet accurate manner. Relevant for the mapping process itself is the construction of a landform identification table, which describes the morphology, identification criteria, paleoglaciological significance, and provides an example location of each glacial landform in the study area. Developing clear criteria for identification and mapping of the landforms is essential for reducing the uncertainties related to the skill, experience, or approach of the mapper (Chandler et al., 2018).

As the scale of the final map is 1:300 000, landforms that were visible at a scale of 1:30 000 were mapped. Depending on the landform, zooming in to a larger scale was required to outline the landforms correctly. For reliable mapping results, the area should be inspected multiple times and at different scales (Chandler et al., 2018). The study area was divided into 15 sectors, such that every sector was examined for all landforms at multiple scales for a few consecutive days before inspection of the next sector. There were two iterations for the total area, thus passing all 15 sectors in the manner described twice.

Landforms were mapped either as polyline or polygon. Shorelines were mapped as polylines that were marked midway between the toe and inner break of the shoreline. The polylines were converted to vertices, for which the elevation in meters above sea level (m a.s.l.) was extracted from the DEM by means of the Add Surface Information tool in ArcGIS Pro 2.8. Perched deltas were mapped as polygons that demarcated the flat top surface. The minimum elevation along the polygon, assumed to represent the delta front, was extracted from the DEM by means of the Add Surface Information tool in ArcGIS Pro 2.8.

The elevations of the shorelines and perched deltas were plotted against the distance along a reference plane (Figure 3.3), of which the distance was calculated by means of the Near tool in ArcGIS Pro 2.8. The reference plane had an orientation of 325°N, which is oriented perpendicular to the isobases of raised shorelines as inferred from literature (Møller, 1987; Regnéll et al., 2019). In this way, different stages of glacial lakes and the
effect of the glacio-isostatic uplift could be distinguished. The lines representing glacial lake stages were determined through visual interpretation, as linear regression would have required separation of the shoreline data points by re-labeling them in ArcGIS Pro, which was not feasible given the time frame of this study.

The shorelines corresponding to the specific stages were isolated and the outlines of the glacial lakes were drawn by tracing the corresponding contour line of the stage. Interpolation between the contour lines that corresponded to the elevation range of the stage was necessary to account for the tilting. The ice dam of each glacial lake stage was placed to fit the distribution of glacial lake traces, including the drainage channels and spillways.

### 3.4 The inversion model

The application of an inversion model is required for extracting ice-sheet properties, such as its thermal regime, subglacial hydrology, or the presence of ice streams, from mapped glacial geomorphology (Kleman et al., 2006). In using their conceptual framework, the aim is to deduce the ice sheet evolution through time. The inversion model is composed of a set of assumptions, a classification system for glacial landform assemblages, and a procedure for managing the landform data and incorporating absolute chronological data (Kleman et al., 2006).
3.4.1 Assumptions

The inversion model is based on a set of genetic assumptions, which explain the linkage between the formation of landforms and ice sheet properties. The assumptions therefore dictate how the existence of individual landforms is interpreted in terms of ice sheet properties, which eventually will lead to ice sheet-wide glaciologically-consistent patterns by aggregation of the individual landforms into swarms. The following assumptions underpin the inversion model of Kleman et al. (2006, p. 196), which is based on previous inversion models of Kleman and Borgström (1996) and Kleman et al. (1997):

1. the basic control on landform creation, preservation, and destruction is the location of the phase boundary between water and ice, separating frozen from thawed material, at or under the ice-sheet base (i.e., basal temperature);
2. basal sliding requires a thawed bed;
3. glacial lineations can form only if basal sliding occurs;
4. glacial lineations are created aligned parallel to local ice-flow directions and perpendicular to the ice-sheet surface contours at the time of creation;
5. frozen-bed conditions inhibit the reshaping of the subglacial landscape;
6. regional deglaciation is always accompanied by the creation of a spatially coherent but metachronous system of meltwater features, such as meltwater channels, eskers and glacial lake traces;
7. eskers are formed in an inward-transgressive fashion inside a retreating ice front;
8. meltwater channels will form the major landform record during frozen-bed deglaciations, whereas eskers are typically lacking under these conditions.

3.4.2 Data reduction

Data reduction is essential for the analysis of the distribution of the mapped landforms, due to their amount and complexity of disparate landforms occurring adjacent or on top of each other. The glacial geomorphological map is composed of over 6000 individual landforms, which through the data reduction procedure are aggregated into a number of glacial landform systems with particular genetic conditions. These landform systems are referred to as 'swarms', with each swarm representing a spatially and temporally coherent ice sheet flow system, reflecting a glaciologically plausible flow pattern (Kleman et al., 2006). Naturally, the swarms are integrative representations of the individual landforms, but this allows for the spatial and temporal analysis that is required.

Kleman et al. (2006) recognizes three basic swarm types: the deglacial envelope, the event swarms and the ice-stream swarms. Based on subglacial temperature distributions, the
The deglacial envelope is further divided into two components, namely the wet-bed deglaciation swarm and the frozen-bed deglaciation swarm.

1. The wet-bed deglaciation swarm is composed of lineations with aligned eskers, which are both formed parallel to the ice surface slope direction and perpendicular to the ice margin, and can therefore be used to determine the overall outline of the ice margin (Stroeven et al., 2021). These swarms form time-transgressively as the margin retreats across the landscape. Additionally, ribbed moraine can indicate a transition from frozen to wet bed conditions (Hättestrand, 1997). Its individual ridges are formed perpendicular to the surface slope direction, and can provide insight into the configuration of the ice sheet as well (Kleman et al., 2006; Stroeven et al., 2021).

2. The frozen-bed deglaciation swarm is composed of meltwater channels, shorelines and perched deltas, and a notable lack of subglacial landforms due to the frozen bed conditions (Kleman et al., 2006). The meltwater traces are imprinted on a relict surface, which can be non-glacial or glacial, thus demonstrating the preservation of landforms for a previous (phase of) glaciation (Kleman et al., 2006; Stroeven et al., 2021). Meltwater traces that form during wet-bed deglaciation can usually be followed across the boundary to the frozen-bed deglaciation, although eskers tend to disappear or become smaller (Stroeven et al., 2021).

3. The event swarm is composed solely of lineations, while aligned meltwater traces are lacking (Kleman et al., 2006). This indicates that the swarm did not form close to the ice margin, but within the ice sheet at a time when wet-bed conditions existed (Stroeven et al., 2021). Therefore, the lineations might have formed synchronously, hence the term event swarm (Stroeven et al., 2021).

4. The ice stream swarm is composed of highly elongated lineations, which typically show a convergent pattern towards the main trunk of the ice stream and a divergent pattern towards the margin of a land-terminating ice stream (Kleman et al., 2006). Although they are comparable to the event swarms in terms of landform composition, their distinct "hourglass" pattern indicates that the ice stream swarms are reflecting events of short-lived, enhanced ice flow within sheet-flow regions (Stroeven et al., 2021).

The spatial distribution of the landforms and their morphological characteristics are used to classify and spatially delineate swarms (Kleman et al., 2006). Given the aim of this study, primarily deglaciation swarms are used for the reconstruction. A set of landforms with a coherent system of flow traces is spatially defined by longitudinal continuity lines that reflect the ice flow direction (Stroeven et al., 2021). These patterns are then converted to swarms, by drawing the upstream and downstream boundaries of a swarm perpendicular to the continuity lines. Most importantly, the swarms and their patterns need to be a realistic representation of the ice flow of an ice sheet. In practice, this means that most swarms are
composed of landforms indicating parallel, slightly divergent or slightly convergent ice flow directions (Stroeven et al., 2021). However, as deglaciation is usually characterized by the migration of ice sheet divides, and by pronounced ice thinning, ice flow directions during deglaciation can strongly differ over time (Stroeven et al., 2021).

3.4.3 Age determination

Relative chronologies of the swarms can be established based on cross-cutting relationships of landforms within the swarms, such as smaller lineations on top of larger lineations, eskers on top of other landforms, and meltwater channels cutting into underlying landforms (Kleman et al., 2006; Stroeven et al., 2021). The deglacial envelope is the youngest and usually the most complete landform system, while older swarms increasingly become incomplete with age as evidence is removed by the imprinting of landforms that are part of the younger swarms. The deglacial envelope is transformed into time slices reflecting the changing ice sheet configurations during deglaciation by using the near-margin ice flow traces as a guide to drawing the contours. The configurations can be placed in time by means of chronological constraints, such as dated records from radiocarbon, cosmogenic nuclide dating, and Optically Stimulated Luminescence (OSL) dating, tephrochronology, and Swedish varve chronology (Stroeven et al., 2021). In certain cases, glaciofluvial deposits or till beds with till fabric directions can be correlated to dated stratigraphical sequences.
4 Landform classification

The glacial landforms can be subdivided in subglacial, ice-marginal and proglacial landforms. For each category, the landforms are described in terms of their morphology, size, and composition. Furthermore, the state-of-the-art knowledge on the formation process is explained, which leads to the paleoglaciological significance of the landform. An overview is provided in Table 2, a landform identification table that details how landforms can be identified from the utilized data sets (Table 1), and which possible identification errors can be expected.

4.1 Subglacial landforms

4.1.1 Lineations

The term glacial lineations is an umbrella term that covers flutes, crag-and-tails, and drumlins. Lineations are elongated subglacial landforms that are aligned to the direction of ice flow (Figure 4.1; Stroeven et al., 2016). There is a large diversity in terms of shape, size and composition, which makes it difficult to explain their formation (Clark et al., 2009; Stokes et al., 2011). The shapes vary from oval-shaped hills with a steep stoss-side and a gentle sloping, tapering lee-side to long and narrow symmetrical hills; these end members are in fact part of a continuum of lineation shapes (Benn and Evans, 2010). Their sizes vary from metres long and tens of centimetres high (flutes) to kilometres long and tens of metres high (drumlins) (Clark et al., 2009; Stroeven et al., 2021). As lineations are formed through both depositional and erosional processes, the features may consist of bedrock, sorted sediments, and diamict (Stokes et al., 2011). They commonly occur in swarms, rather than as individual features (Clark et al., 2009; Stroeven et al., 2021).

Figure 4.1: Lineations southwest of Käärävaa (67°57’25.7”N 20°2’30.7”E) as visible on the hillshade (a) and outlined with 10 m contours as background (b).
The paleoglaciological significance of lineations is related to the observation that lineations often occur together with deglacial landforms, such as eskers, and occur parallel to the youngest set of striae on bedrock outcrops, thereby implying that the lineations reflect the ice flow direction close to the former ice sheet margin prior to deglaciation (Stroeven et al., 2016). However, larger lineations, such as drumlins, may have formed during multiple glaciations with the same ice flow direction, and survived glaciations with another ice flow direction (Hättestrand et al., 2004). Lineations are only formed underneath an ice sheet with a warm-based thermal regime (Kleman et al., 2006).

4.1.2 Eskers

Eskers are straight or winding ridges composed of sand and gravel (Figure 4.2; Brennand, 2000). They occur as single ridges or as a network of several parallel ridges (Stroeven et al., 2016). In addition, there are, for example, variations in esker width (Dewald et al., 2021), number of crests (Livingstone et al., 2020), and sinuosity (Storrar et al., 2014). In fact, the morphology of a ridge can vary substantially along its course (Dewald et al., 2021). Eskers often occur in valleys or depressions and are commonly flanked by dead-ice depressions (Stroeven et al., 2021). The majority of eskers are less than 10 km long, although the esker systems extend much longer if the gaps between the segments (or esker beads) are interpolated (Storrar et al., 2014). The ridges are typically tens to hundreds of meters wide and meters to tens of meters high (Livingstone et al., 2020).

Eskers are typically formed inside subglacial tunnels at the base of ice sheets (Stroeven et al., 2016), although field research has shown eskers may form in englacial conduits and supraglacial channels as well (Price, 1966). Deposition of sediments occurs near the mouth of the water-filled conduits, either within the tunnel or as a fan beyond the margin in a

Figure 4.2: Esker east of Matalajärvi (68°18′47.1″N 19°59′16″E) as visible on the hillshade (a) and outlined with 10 m contours as background (b).
proglacial channel or lake (Brennand, 2000; Hewitt and Creyts, 2019). Most eskers are aligned more or less parallel to the ice flow direction, as the channels drain towards the ice margin (Benn and Evans, 2010). It remains slightly uncertain whether the eskers reflect a synchronous (Brennand, 2000) or a time-transgressive drainage system (Livingstone et al., 2015), although it is argued that the temporal variability in meltwater supply indicates formation of the esker in a time-transgressive manner close to the margin (Greenwood et al., 2016; Livingstone et al., 2020). As the eskers are formed close and perpendicular to the margin during deglaciation, their distribution can be used to reconstruct the ice retreat during deglaciation.

An exception in terms of formation processes and relationship to ice flow direction is the subglacially-engorged esker, which are the depositional versions of subglacial chutes (Sissons, 1961). Subglacially-engorged eskers likely reflect sedimentation in subglacial tunnels running downslope valley sides (Johansson, 1994; Goodship and Alexanderson, 2020). Subglacially-engorged eskers are therefore often oriented perpendicular to major eskers aligning valleys.

### 4.1.3 Ribbed moraine

Ribbed moraine is usually described as fields of curved or concave ridges, which form transverse to former ice flow with their outer limbs pointing down-ice (Figure 4.3, Benn and Evans, 2010). The ridges often have asymmetric cross sections, are anastomosing and occasionally separated by gaps (Hättestrand and Kleman, 1999). They are typically 10–20 m high, 300–1200 m long and 150–300 m wide (Hättestrand and Kleman, 1999). However, an assessment by Dunlop and Clark (2006) showed that ribbed moraine exists in highly variably forms and sizes. The composition of the ribbed moraine varies widely,

**Figure 4.3:** Ribbed moraine southwest of Bogivárdu (68°7’51.8”N 20°30’26.8”E) as visible on the hillshade (a) and outlined with 10 m contours as background (b).
ranging from silts and sands to gravel and diamict, both massive and laminated, but most commonly consist of till (Benn and Evans, 2010; Hättestrand and Kleman, 1999). They frequently seem to be composed of the same material as the surrounding terrain (Hättestrand and Kleman, 1999).

Ribbed moraine ridges are seldom used in ice sheet reconstructions because consensus is lacking as to the formation of the landform. Boulton (1987) suggests that ribbed moraine is part of a continuum of streamlined landforms, formed through altering of a previously existing landform by a later ice flow direction and differential sediment transport rates. Lindén et al. (2008) suggest that compressive flow due to a change in topography or thermal boundaries causes a stacking and folding of basal ice or till, although Hättestrand and Kleman (1999) argue that several observations contradict this "shear and stack theory". Instead, Hättestrand (1997) argues that the ribbed moraine ridges form sheets of fractured frozen substrate due to extensional flow caused by the transition from cold-based to warm-based conditions. According to Hättestrand (1997), ribbed moraine is formed closely to the ice margin during deglaciation. Such uncertainties concerning its formation process have resulted in ribbed moraine being primarily used as ice flow direction indicator in previous reconstructions (Greenwood and Clark, 2009), while only few studies have used the ribbed moraine as indicator of the spatial distribution of frozen and thawed bed conditions (Kleman and Hättestrand, 1999; Kleman et al., 2006, 2010).

4.1.4 Subglacial meltwater channels

Meltwater channels are erosional features, which are expressed as incisions into underlying sediments or bedrock (Kleman, 1992; Kleman et al., 1997; Greenwood et al., 2007; Margold et al., 2013). Different channel types that are recognized include subglacial

![Figure 4.4](image-url) Subglacial meltwater channels at Guoskkaláhku (68°8’15.9”N 19°31’53.2”E) as visible on the hillshade (a) and outlined with 10 m contours as background (b).
channels, lateral channels (see 4.2.1), proglacial channels (see 4.3.1), and spillway channels (see 4.3.1); (Jansson, 2003; Greenwood et al., 2007). These channel types differ in their morphology and relationships with the parent ice body and topography (Greenwood et al., 2007), although these criteria cannot always be used to exclusively identify a type of meltwater channel.

The direction of subglacial channels is primarily determined by the ice surface slope, but also by underlying topography (Sugden et al., 1991). Hence, channels that are found oblique to the slope can confidently be characterized as subglacial meltwater channels (Figure 4.4). Additionally, channels connecting esker segments can be identified as of subglacial origin (Kleman and Borgström, 1996). Subglacial meltwater channels indicate a warm-based deglaciation, where the channels reflect ice sheet flow direction close to the ice margin (Greenwood et al., 2007).

4.2 Ice-marginal landforms

4.2.1 Lateral meltwater channels

Lateral meltwater channels are series of channels formed parallel to the contemporary contours of a mountain side into which they are eroded (Figure 4.5; Greenwood et al., 2007; Stroeven et al., 2016). Because the channels formed perpendicular to the current drainage system, they are often dry. The channels are commonly tens of meters deep, meters wide, and hundreds of meters long (Stroeven et al., 2016).

Lateral meltwater channels are formed by erosion of a meltwater stream wedged between an ice margin and the adjacent ice-free mountain side. Here, meltwater from the wider convex ice surface collects and is guided by the surface slope of the ice margin.

Figure 4.5: Lateral meltwater channels along southeastern slope of Čoragás (68°1’36.3”N 19°21’36.4”E) as visible on the hillshade (a) and outlined with 10 m contours as background (b).
(Mannerfelt, 1949; Greenwood et al., 2007; Margold et al., 2013). Series of lateral meltwater channels therefore record the lowering of the ice surface (Margold et al., 2013), and typically indicate a steepening of the ice surface through time. The lateral meltwater channels are especially useful for reconstructing cold-based deglaciations, as the channels also form during frozen-bed conditions, in contrast to landforms such as lineations and eskers (Kleman, 1992; Greenwood et al., 2007; Hättestrand and Stroeven, 2002; Stroeven et al., 2016). However, Syverson and Mickelson (2009) observed lateral meltwater channels to have formed along a temperate ice margin, and the channels are therefore not an exclusive indication of a cold-based regime.

4.2.2 Marginal moraine

Marginal moraines are straight or arcuate ridges of sediment, which are deposited or deformed at the margins of glaciers or ice sheets (Figure 4.6; Benn and Evans, 2010). End moraines are formed at the foot of a glacier or ice lobe, while lateral moraines are formed at the sides (Benn and Evans, 2010). The morphology and composition of moraines can vary significantly, which is partly related to the various processes that may contribute to the formation of a moraine, such as dumping, pushing, shearing and squeezing (Benn and Evans, 2010). They range from small ridges, few meters long and high, to entire complexes which extend for hundreds of kilometers (Heyman and Hättestrand, 2006). The moraines are usually composed of till, but can also be composed of overridden and deformed sediments, such as glaciofluvial deposits. Moraines are occasionally interrupted by gaps of non-deposition or meltwater erosion, but usually the moraines can be extrapolated to either side assuming lateral ice sheet continuity (Stroeven et al., 2016).

![Figure 4.6: Marginal moraine southeast of Nagirvári (68° 10' 1.6" N 19° 44' 50.4" E) as visible on the hillshade (a) and outlined with 10 m contours as background (b).](image)
Moraines reflect the outline of a glacier or an ice sheet margin at the time of formation (Heyman and Hättestrand, 2006). As such, they indicate a maximum advance (or re-advance) or an interruption in ice sheet retreat (Stroeven et al., 2016). The use of the moraine for reconstruction purposes strongly depends on a correct paleoglaciological interpretation of the location of the parent glacier on either side of the ridge. The topographical setting and the plausibility of certain ice behavior generally points to the location of the parent glacier, but this can be complicated when both local valley glaciers and ice lobes from ice sheets have covered the location in the past, such as in the Scandinavian mountains (Heyman and Hättestrand, 2006) or in the Cordillera of North America (Dulfer and Margold, 2021; Dulfer et al., 2022).

Heyman and Hättestrand (2006) classified non-contemporary marginal moraines in the Swedish mountains into four classes based on morphology and topographical setting: cirque-and-valley moraines, valley-side moraines, complex moraines, and cross-valley moraines. Cirque-and-valley moraines are predominantly situated as arched ridges in or below ice-free cirques, valley-side moraines are subhorizontal straight ridges situated on valley sides or mountain slopes, and cross-valley moraines are usually situated across the valley floor as straight ridges (Heyman and Hättestrand, 2006). Complex moraines are also usually situated along valley sides or mountain slopes, but show an irregular morphology and their outer ends usually point up slope (Heyman and Hättestrand, 2006). Although all classes were adopted while mapping, the final map only displays the complex moraines and contemporary-glacier moraines separately, while the three remaining classes were incorporated into one marginal moraine class.

4.3 Proglacial landforms

4.3.1 Proglacial meltwater channels

Proglacial meltwater channels align to the local bed slope, as they are not directed by the ice margin (Figure 4.7, Stroeven et al., 2021. Even though these can be difficult to distinguish from contemporary channels, two possible identifications are as dry-bed or having underfitted streams (Stroeven et al., 2021). They can reach enormous dimensions in length and width, and up to several hundred meters in depth (Benn and Evans, 2010).

Channels associated with glacial lakes can be difficult to distinguish from proglacial meltwater channels. There are two types of channels that are related to glacial lake systems. Spillway channels drain the excess water of a dammed glacial lake, controlling the maximum lake level. Spillways occur at the lowest point of water divides (cols) (Jansson, 2003; Benn and Evans, 2010). Drainage channels are associated with the (stepped) drainage of glacial lakes due to ice margin retreat or breaking of the threshold of the ice dam. Note that the spillway and drainage channels could be one and the same, for example when the lowest point of the terrain surrounding a glacial lake is found along the ice dam. Drainage channels can be difficult to distinguish from regular lateral meltwater channels (see 4.2.1),
as they can also occur on valley slopes parallel to the ice margin (Jansson, 2003). However, drainage channels are generally larger and are often associated with washed bedrock zones (Jansson, 2003). The channels, together with other glacial lake traces (see 4.3.2), can be used to identify the outline of former glacial lakes and the threshold, which is useful for deglaciation reconstructions (Kleman, 1992; Jansson, 2003; Regnéll et al., 2019). The spillway and drainage channels are not outlined on the map of the glacial geomorphology, as their identification strongly depended on the inferred glacial lake stages.

4.3.2 Raised shorelines

Raised shorelines are (roughly) horizontal zones of wave-washed substrate (commonly till) or consist of an accumulation of sediment (Figure 4.8). The shorelines often occur in series (Jansson, 2003), and are characterized by a break in slope (Regnéll et al., 2019). The shorelines can commonly be traced over large distances, and occasionally onto perched deltas (see 4.3.3; Regnéll et al., 2019). The width of the shoreline is depending on the initial topography, where thin shorelines correlate with steep slopes and wider shorelines with gentler slopes. The shorelines are not exactly horizontal due to uneven glacio-isostatic uplift along the former glacial lakes (Pässé, 1998; Berglund, 2012), with a slope of approximately 0.5 m/km in the Scandinavian mountains in northwestern Sweden (Regnéll et al., 2019).

The almost horizontal landforms are representing former shorelines of glacial lakes, which are lakes that are formed by trapping of meltwater due to a damming ice margin (Stroeven et al., 2016). The series of shorelines represent several different lake levels, which indicate the (partial) drainage of the glacial lake through either an ice dam failure, overtopping of the ice dam or the exposure of a new outlet by the retreating ice margin (Stroeven et al., 2016). The glacial lake shorelines are especially useful for reconstructions
of cold-based glaciations, where the position of the ice-margin (as an ice-dam) is estimated based on the distribution of the glacial lake shorelines (Jansson, 2003; Stroeven et al., 2016; Regnéll et al., 2019).

### 4.3.3 Perched deltas

Perched deltas are characterized by a steep delta front and a flat top surface, of which the surface often shows signs of erosion by braided or meandering streams (Figure 4.9; Jansson, 2003; Margold et al., 2013; Goodship and Alexanderson, 2020). The deposits are situated on the valley slopes above present lake levels, often directly in front of the valley.
mouth. The perched deltas are composed of sand and gravel, although this is influenced by the gradient of the feeding river (Benn and Evans, 2010; Peterson and Smith, 2013).

The perched deltas are deposits formed by meltwater streams originating from ice sheets and glaciers, which transport sediments into a standing body of water, such as a glacial lake (Benn and Evans, 2010). The perched deltas are situated higher in the landscape due to lake-level lowering since the formation of the perched deltas (Peterson and Smith, 2013). Hence, the perched deltas can be used as an indication of former glacial lake-levels, and can be used to reconstruct the position of an ice margin towards the valley mouth.
Table 2: Landform classification table describing the morphology, dimensions, possible identification errors and paleoglaciological significance of the landforms mapped in this study. References can be found in the corresponding sections in Chapter 4.

<table>
<thead>
<tr>
<th>Landform</th>
<th>Morphology</th>
<th>Dimensions</th>
<th>Possible identification error</th>
<th>Significance</th>
<th>Example</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>SUBGLACIAL</strong></td>
<td></td>
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</tr>
<tr>
<td>Lineations</td>
<td>Elongated ridges, both depositional and erosional. Tend to occur in swarms</td>
<td>Meters long and tens of centimeters high to kilometers long and tens of meters high</td>
<td>May be confused with bedrock structures, although hillshades with multiple illumination angles may clarify</td>
<td>Formed parallel to ice flow, reflects ice flow prior to deglaciation when occurring together with eskers</td>
<td>Southwest of Käyrävaara (67° 57’25.7”N 20° 2’30.7”E)</td>
</tr>
<tr>
<td>Esker</td>
<td>Single ridges or networks of parallel ridges. Typically sharp-crested, long and winding</td>
<td>Size up to hundreds of kilometers long and tens of meters high</td>
<td>Misinterpretation as type of moraine, although esker is usually more sinuous</td>
<td>Formed parallel to ice flow and close to a retreating ice margin</td>
<td>East of Matalajärvi (68° 18’47.1”N 19° 59’16”E)</td>
</tr>
<tr>
<td>Subglacial meltwater channels</td>
<td>Channels incised into bedrock or sediment, descent oblique to slope. Occasionally connect esker segments</td>
<td>Highly variable dimensions, up to several hundred of meters</td>
<td>May be confused with sub-marginal lateral meltwater channels</td>
<td>Reflect ice sheet flow direction close to the ice margin</td>
<td>At Guoskaláhku (68° 15’9.0”N 19° 31’53.2”E)</td>
</tr>
<tr>
<td>Ribbed moraine</td>
<td>Fields of curved ridges, regularly and closely spaced</td>
<td>Hundreds of meters long and tens of meters high</td>
<td>May be confused with solifluction lobes, although ribbed moraine are usually found in depressions</td>
<td>Formed transverse to ice flow, with outer limbs pointing down-ice. Indicative of frozen bed conditions</td>
<td>Southwest of Bogivárdu (68° 7’51.8”N 20° 30’26.8”E)</td>
</tr>
<tr>
<td><strong>ICE-MARGINAL</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Marginal moraine</td>
<td>Straight or arcuate ridges, can occur in series. Potentially continuous, but often interrupted by gaps of non-deposition or erosion</td>
<td>From few meters long and high to complexes of hundreds of kilometers long</td>
<td>Difficult to distinguish from a protalus rampart, but the location might suggest unfavourable conditions for glaciers</td>
<td>Formation at the ice margin, outline indicates shape and position of the former ice margin</td>
<td>Southeastern slope of Nagirvárári (68° 10’1.6”N 19° 44’50.4”E)</td>
</tr>
<tr>
<td><strong>Lateral meltwater channels</strong></td>
<td>Series of straight or winding channels cut into valley walls, subparallel to the contours</td>
<td>Tens of meters deep, hundreds of meters long, meters wide</td>
<td>Misinterpreting as bedrock structures, steplike solifluction lobes or shorelines, although the latter is strictly horizontal</td>
<td>Formation along ice margin, possible to infer ice surface slope and ice thickness</td>
<td>Along southeastern slope of Čorugåš (68°1’36.3”N 19°21’36.4”E)</td>
</tr>
<tr>
<td>-------------------------------</td>
<td>-------------------------------------------------------------------------------------------------</td>
<td>-------------------------------------------------</td>
<td>---------------------------------------------------------------------------------</td>
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</tr>
<tr>
<td><strong>Proglacial meltwater channels</strong></td>
<td>Channels incised into bedrock or sediment, aligned to the local bed slope</td>
<td>Tens to over hundreds of meters long to tens of meters wide</td>
<td>May be confused with contemporary river incisions, but identifiable by dry-bed or underfitted stream. May be misinterpreted as glacial lake drainage channels</td>
<td>Formation at terminus of the ice margin</td>
<td>At the mouth of Kärkevagge (68°25’1.6”N 18°17’53.8”E)</td>
</tr>
<tr>
<td><strong>Shorelines</strong></td>
<td>Zones of (nearly) horizontal wave washed till, eroded rock terrace or an accumulation of sediment. Slope of 0.5 m/km. Charaterized by a break in slope. Often occurs in series.</td>
<td>Few meters wide, can extend hundreds of meters in length</td>
<td>Misinterpreting as bedrock structures (although those usually have a more extensive spatial distribution and are seldom strictly horizontal), steplike solifluction lobes or meltwater channels</td>
<td>Indicative of former lake levels of ice-dammed lakes, possible to infer the location of the ice margin</td>
<td>Valley side to the north-east of Giepmåšluokta (68°19’25”N 19°30’54.8”E)</td>
</tr>
<tr>
<td><strong>Perched deltas</strong></td>
<td>Flat top surface and steep delta front, situated above present lake levels. Signs of erosion by streams</td>
<td>Hundreds of meters wide</td>
<td>May be confused with an ice-contact delta, although these often have kettle holes</td>
<td>Indicative of former lake levels of ice-dammed lakes, possible to infer the location of the ice margin</td>
<td>Lowermost part of Bessešvággi valley (68°17’56”N 19°13’21.5”E)</td>
</tr>
</tbody>
</table>
5 Results

The glacial geomorphology of the Torneträsk basin is presented in Figure 5.1. In total, there are 6369 features mapped, of which there are 2629 lineations, 671 eskers, 39 ribbed moraine, 1232 meltwater channels, 153 marginal moraines, 510 minor ridges, 859 raised shorelines, 192 perched deltas, 38 veiki moraines, 15 outwash plains, 15 pitted outwash deposits, and 16 undifferentiated glaciofluvial deposits. Note that the count includes all segments of a landform, so it represents a feature count instead of a landform count. To consider the relevance of each feature in a paleoglaciological reconstruction, this chapter offers a description of the landform distribution, relation to topography, ice flow direction, and relation to other landforms.

5.1 Subglacial landforms

5.1.1 Lineations

Lineations are found across the whole area, but are most common in the premontane region in the east (Figures 5.2a and 2.1). Here the lineations occur in fields, where their morphology often represents the characteristic streamlined tear-drop shape (Table 2). The glacial lineations are typically hundreds of meters in length, while large-scale drumlins (thousands of meters in length) are rare. Fluting is widespread in the region as well, especially in the forefield of contemporary glaciers. This small-scale fluting has not been included on the map, as they are not useful for reconstructing the paleoglaciology of the FIS. The lineations in the montane region are most commonly found on valley floors and even uplands. These lineations are often in the form of drumlinized thin till cover. The lineations in the scoured bedrock zone in the west are mostly composed of rock drumlins and whale backs.

There are two dominant ice flow directions that can be inferred from the lineations, namely towards the northeast (NE) and southeast (SE). Ice flow towards NE is the most frequent orientation, where the lineations are orientated predominantly NNE in the north, but gradually change towards a NE flow direction in the south. These lineations are exclusively in the typical size-range. The SE ice flow direction is mostly represented by large-scale drumlins. The streamlined bedrock in the western scoured zone has an elongated direction E-W, without a clear indication of either westerly or easterly flow because sediment tails are usually absent. At a few locations these two sets of lineations are crosscutting each other (Figure 5.2b), where it can be inferred that the NE lineations are superimposed on the larger SE drumlins and are younger. A third set of NNW lineations is crosscutting the SE lineations (Figure 5.2b), but it cannot be inferred which of the NE and NNW lineations are the youngest.

The lineations occasionally occur jointly with eskers and ribbed moraine. For example, the majority of the larger eskers align with lineations with a NE ice flow direction (Figure
Figure 5.1: The map of the glacial geomorphology of the Torneträsk basin, which is designed for a scale of 1:300,000 (fit for A0).
5.2c). However, seemingly degraded esker fragments are sporadically overprinted by these younger lineations. Additionally, there are a few ribbed moraines that are fluted.

### 5.1.2 Eskers

Eskers occur across the area, but are most frequent in the montane region (Figures 5.3a and 2.1). There, the eskers mainly trace the valley floors parallel to the valley axis, while subglacially-engorged eskers extend transverse to the valley axis, down the valley slopes. Eskers are almost completely lacking in the scoured bedrock zone in the west, and short and narrow if present. Eskers are relatively rare in the premontane region as well. The valleys in the premontane region are often occupied by lakes, which may therefore hide eskers underneath the water surface.

The longest continuous esker is almost 20 km long, and is clearly connected to an upstream eroded meltwater channel (Figure 4.2). The majority of the eskers are, however, fragmented pieces of tens to hundreds of meters in length, but they can be traced over distances of several kilometers. The fragments are mostly ridge-like, but can also be short and almost circular, in which case they represent esker beads (Livingstone et al., 2020; Figure 5.3b).
In addition to larger eskers frequently aligning with the orientation of NE lineations, most esker fragments, often also traceable over the longest distances, parallel a NE direction. There exist a few eskers, frequently degraded, that display a SE orientation. Two relatively long eskers are curved (Figure 5.3c), which complicates assigning it to a certain retreat direction. Occasionally, esker ridges widen into sections that consist of multiple ridges, after which they converge into single ridges again. These esker enlargements are only observed in the premontane region in the east.

Esker fragments commonly grade into meltwater channels, thereby confirming the type of meltwater channels as subglacial (Figure 5.5b). Furthermore, eskers are often found in valleys where perched deltas or outwash plains are observed further downstream. It was especially difficult to distinguish eskers from certain minor ridges west of Alajärvi (Figure 5.7c), as both consist of glaciofluvial material (Melander, 1977b), and their morphologies are rather similar.

5.1.3 Ribbed moraine

Ribbed moraine occurs predominantly north of Torneträsk and in the premontane region (Figures 5.4a and 2.1). Ribbed moraine is located on relatively flat surfaces, either as part of valley bottoms or on uplands. They are often situated in long and narrow zones next to lake basins, which leads to a suspicion that ribbed moraine also exists on the bottom of the
Figure 5.4: a) Distribution of ribbed moraine in the study area. b) Location where ribbed moraine is found at higher elevation. c) Location where ribbed moraine is fluted and eroded by lateral meltwater channels.

adjacent lakes. The montane region is generally devoid of ribbed moraine, although there are exceptions north of Rautasjaure (Figure 5.4b) and Torneträsk where ribbed moraine is found on flat uplands.

Ribbed moraine ridges are generally oriented perpendicular to lineations, such that the ridges approximately have a NW-SE orientation and correspond to NE flow. In contrast, the northwesternmost occurrences of ribbed moraine show an orientation of N-S, which corresponds to an ice flow direction roughly parallel to the long axis of Torneträsk, although it is unclear whether it concerns westerly or easterly flow.

Sporadically, the ribbed moraine has been fluted, usually perpendicular to the ribbed moraine ridges themselves (Figure 5.4c). Additionally, meltwater channels occasionally cut into ribbed moraine (Figure 5.4c).

5.1.4 Subglacial meltwater channels

Meltwater channels are the second-most abundant landform in the study area (Figure 5.5a). They occur most densely in the eastern part of the montane region, while meltwater channels are almost completely lacking in the scoured bedrock zone in the west (Figures 5.5a and 2.1). Subglacial meltwater channels are prevalent in the entire study area, although most of them occur in the premontane region. Subglacial channels are frequently situated on valley floors or valley sides. The aspect of the channels varies with local topography, but
the majority of the channels indicate water flow towards the north(east). The dimensions of the channels vary, but they are often relatively short (few hundreds of meters), considerably shorter than other types of channels (up to kilometers). There are locations where meltwater channels with different flow directions cross-cut each other. Subglacial channels often occur in relation to esker fragments, which indicates fluctuations in conditions that lead to erosion and deposition along the subglacial drainage path (Figure 5.5b).

A few meltwater corridors are mapped where individual meltwater channels were difficult to distinguish. The corridors were either characterized by an exceptionally wide but shallow channel or multiple crosscutting channels with adjacent pronounced escarpments. The profiles along the corridors are often undulating, which is interpreted as a sign of a subglacial drainage system. The corridors are observed in both the premontane and montane region, and can be connected to eskers as well.

5.2 Ice-marginal landforms

5.2.1 Lateral meltwater channels

Whereas subglacial channels are abundant in the premontane region, lateral meltwater channels, on the contrary, are relatively rare, even though multiple sets of channels occur in
the northeastern corner of the study area (Figures 5.5 and 2.1). Lateral meltwater channels are usually occurring on steep valley slopes, and relatively often on east-facing slopes. The channels differ in size, depth, and degree of sinuosity. The majority of the lateral meltwater channels are cut by water flow towards north or east, which suggests an ice retreat towards the south or west. However, in the mountains the channel slope directions are varying considerably, which could have been caused by increasing importance of the topography of the mountains during downwasting of the ice sheet. Lateral channels on the steepest valley slopes are associated with subglacially engorged eskers (Figure 5.5c). The lateral channels are cut into the subglacially engorged eskers, indicating the channels are younger. Additionally, lateral meltwater channels occasionally transition into subglacial meltwater channels further downstream. Sporadically, the orientation and elevation of valley-side moraines correspond to the same ice margin that formed lateral channels downslope.

5.2.2 Marginal moraine

There are relatively few moraines in the study area (Figure 5.6a). Most moraines are found close to the highest mountain peaks south of Torneträsk, close to the margins of contemporary glaciers (as contemporary-glacier moraines) or in ice-free valleys (as cirque-valley moraines). There are only few marginal moraines that are unambiguously associated with ice sheet configurations. These latter predominantly occur as either valley-
side moraines or complex moraines, and occasionally as cross-valley moraines (Figure 5.6b). The orientation of the moraines in the montane region is variable. This is according to expectation if they are formed during thinning of the ice sheet, which caused the flow to be predominantly determined by the local topography. The majority of complex moraines occur on south-facing slopes. Marginal moraines are virtually lacking in the premontane region (Figures 5.6a and 2.1). Those few marginal moraines that do exist have an approximately E-W orientation, which is at an almost perfect perpendicular angle to lineations and eskers in the vicinity (Figure 5.6c).

5.2.3 Minor ridges

Minor ridges are scarce in the study area, but they always occur in groups of parallel ridges (Figure 5.7a). Crevasse-squeeze ridges are exclusively observed in the montane region (Figures 5.7a) and 2.1). The remaining ridges, categorized as undifferentiated because they demonstrate a different morphology and setting than the crevasse-squeeze ridges, are all in the premontane region. All ridges, except for one group on an even upland, are found on valley bottoms, and occur close to lakes. Their thin and generally sharp-crested ridges appear fragile and susceptible to erosion.

Although all are categorized as undifferentiated ridges, they vary in morphology and setting. Crevasse-squeeze ridges usually form perpendicular to ice flow due to crevasse

Figure 5.7: a) Distribution of minor ridges in the study area. b) Location where undifferentiated ridges occur in parallel, lobate sets. c) Location where undifferentiated ridges are found adjacent to eskers.
formation through fracturing perpendicular to the principle tensile stress (Rea and Evans, 2011). If these undifferentiated ridges are similar to crevasse-squeeze ridges, or moraines, the ridges are oriented perpendicular to ice flow. However, landforms in the vicinity of the ridges are occasionally contradictory to this assumption.

The ridges downstream of Rautasjaure are particularly striking due to their lobate shape (Figure 5.7b). Their orientation suggests NNE ice flow, aligning with lineations at an perpendicular angle to the ridges. Another set of straight ridges further northeast seemingly connects to the Rautasjaure group of lobate ridges. Field work showed that these ridges are composed of large, subrounded boulders. Their orientation implies ice flow towards NW, which is in agreement with a small patch of lineations (Figure 5.3b). However, esker beads in close proximity, interpreted to be part of an esker aligning with the ridges, are incompatible with NW ice flow (Figure 5.3b).

A set of undifferentiated ridges west of Alajärvi are oriented parallel to the current shoreline of the lake (Figure 5.7c). These ridges are unusually straight, sharp-crested, and closely-spaced. They are situated close to eskers that look rather similar. The orientation of the ridges represents NE ice flow, although lineations in the vicinity occur at an oblique angle to the ridges. Furthermore, eskers parallel to the ridges would imply an ice flow direction towards NW.

5.3 Proglacial landforms

5.3.1 Proglacial meltwater channels

Only very few proglacial meltwater channels were confidently mapped. It is the question whether the identification criteria might be insufficient (Greenwood et al., 2007), for example due to the difficulty of distinguishing them from contemporary channels as both flow directly downslope. The proglacial channels that were mapped originated from a glacier tongue (Figure 4.7). Often other types of meltwater channels were found further upstream of the proglacial channels. Additionally, outwash plains are found in close association.

5.3.2 Raised shorelines

Raised shorelines are widespread along the Torneträsk basin, Rautasjaure basin, and several smaller (former) lake basins (Figure 5.8a). The individual shoreline segments are usually hundreds of meters in length, but are ultimately traceable over distances of tens of kilometers. The shoreline width seems to vary depending on the slope, usually rather narrow on steep slopes (meters in width), but much wider on gentle slopes (tens of meters in width). The shorelines appear more clearly developed on the south- and west-facing slopes of Torneträsk, which is mostly visible for the middle part of the basin. The number of raised shorelines along Torneträsk decreases towards the southeast, with the southernmost occurrence along Torneälven, a mere 10 km northeast of Kiruna (Figure
5.8b). The shorelines at Rautasjäure are relatively narrow and vertically closely-spaced. They are generally less clearly developed than the shorelines of Torneträsk. The shorelines are often cut by chutes or covered by slope deposits (Figure 5.8c).

Several lake stages can be distinguished for Torneträsk and Rautasjäure (Figure 5.9 and Table 3). Figure 5.9 shows how the elevations of the shoreline segments plot in straight lines with slopes reflecting post-formation differential glacio-isostatic uplift. Table 3 provides the lake stages that were identified from Figure 5.9, including the elevation range of the stages and the shoreline elevation gradient along the reference plane. The computation of the latter was complicated by post-glacial movement of the Pärvie fault (Figure 5.8), differentially displacing shoreline elements on either side of the fault.

For Torneträsk eight lake stages are identified (T-0 to T-7), with the highest lake stage at least 143 m above the current lake level of 342 m a.s.l. and the lowest lake stage extending below the current lake level due to lake-tilting by differential post-glacial rebound (Figure 5.9a). Lake stages T-2 to T-5 appear to be affected by the fault, as gradually-changing shoreline elevations suddenly jump in elevation by 5–8 m at the location of the fault line (Figures 5.8 and 5.9). Remarkably, glacial lakes T-0 and T-1 do not appear to have a step in elevation, although this is perhaps related to the few shoreline segments that were identified for this lake stage. The shorelines of glacial lakes T-6 and T-7 are clearly unaffected by the Pärvie fault displacement, suggesting the fault ruptured between glacial lakes T-5 and T-6.
Figure 5.9: Lake stages identified from the shoreline and perched delta elevations of the Torneträsk glacial lakes (a) and the Rautasjaure glacial lakes (b), including the location of the Pärvie fault. The distance is calculated along an axis perpendicular to the isobases of the shorelines (see Figure 3.3).
Table 3: Overview of the glacial lakes that have been identified based on the raised shorelines and perched deltas. The elevation range is inferred from the identified glacial lake stages (see Figure 5.9). The gradient for several lake stages are missing due to an insufficient length over which the gradient could be calculated. Gradients were calculated separately for shorelines east of the cross-cutting Pärvie fault (*). The highlighted rows are visualized in Figure 6.1.

<table>
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<th>Distance</th>
<th>Gradient</th>
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</table>
The gradients of the lake stages range from 0.5 m/km to 0.7 m/km, which are corresponding to gradients of shoreline elevations of the glacial lakes of Akkajaure and Sitojaure, situated 140 km south from Torneträsk, between Kebnekaise and Sarek (Regnél et al., 2019). Shorelines with relatively high elevations (i.e., above 500 m a.s.l.) that are situated along the northern side of Torneträsk, do not appear to have been part of a large open lake as the shorelines are spatially limited. However, DGPS measurements of shorelines at Varrégaska (Location 11, Figure 3.1b), show that minor shorelines have developed at elevations above 500 m a.s.l. along the southern shore of Torneträsk as well (A.1 in Appendix A).

For Rautasjaure 22 lake stages were identified (R-1 to R-22), with the highest lake stage around 274 m above the current lake level of 560 m a.s.l. (Figure 5.9b). The lowest lake stage was situated downstream of the current Rautasjaure, occupying the basin where nowadays the river Rautasälven is draining Rautasjaure. Glacial lake R-22 was situated around 26 m above the current level of Rautasälven at 511 m a.s.l. (Figure 5.9b). The shorelines of glacial lakes R-12 to R-19 were most continuous, while the other stages are based on fragmented occurrences of shorelines. The lowest stages are mainly identified based on the perched deltas. The fault is not crossing the lake basin, so the elevation ranges are unaffected by steps. The gradients of the stages show higher variation, ranging from 0.3 m/km to 0.8 m/km. This high variation might be related to the relatively few mapped shorelines the gradients were inferred from.

Fewer lake stages are recognized for glacial lakes Vittankijärvi, Nakerijärvi, Sautasjärvi, Vassijaure, Kaitasjärvi and Sevujärvi (Table 3). The distribution of shorelines belonging to these lakes is spatially limited and therefore no gradients were calculated.

### 5.3.3 Perched deltas

Perched deltas are predominantly found in the Torneträsk and Rautasjaure basins (Figure 5.10a). The perched deltas usually occur inset at successive elevations at the end of valleys or chutes. The majority of perched deltas are formed on the southern shores of the glacial lakes by meltwater draining north. This is consistent with ice retreat towards the south, and the ice margin producing meltwater and an abundance of sediment. Given the active retreat of the ice margin, deltas are only formed for a short duration.

The perched deltas are often laterally traceable to raised shorelines (Figure 5.10b). Although the elevations of the deltas do not plot as coherent on one line as the shorelines, Figure 5.9 demonstrates multiple perched deltas that correspond to certain lake stages as identified by their shorelines. Occasionally, an esker connects to the apex of a delta, strengthening their glacial meltwater origin (Figure 5.10c).
Figure 5.10: a) Distribution of perched deltas in the study area. b) Location where shorelines and perched deltas occur at same elevations. c) Location where eskers are connected to the apex of perched deltas.
6 Discussion

6.1 Reconstruction of glacial lakes

Several glacial lake systems were in operation during deglaciation in basins currently occupied by Torneträsk, Vassijaure, Rautasjaure, Nakerijärvi, Kaitasjärvi, Vittankijärvi, and Luovijärvi (Figure 6.1). The identified ice dams are consistent with the distribution of shorelines, perched deltas, drainage channels, and spillways, impounded by surrounding topography, and the orientation of meltwater channels, lineations, and eskers. Those stages for which a plausible drainage and/or spillway location could be identified are outlined in Figure 6.1 and further elaborated on in this section.

6.1.1 Glacial lakes of Torneträsk

Because Vassijaure is located at the western end of the Torneträsk basin, glacial lake Vassijaure (Va-1), with perched delta elevations between 510 and 513 m a.s.l., is included in this subsection (Figure 6.2a). Its spillway, located at 510 m a.s.l., directed meltwater through the valley Norddalen towards Rombaks-fjorden in Norway (Figure 6.3a). A set of perched deltas at the head of the fjord are likely related to the incision of the spillway, and indicates relative sea level lowering due to glacio-isostatic uplift of the land. Va-1 required an ice dam just west of Sördalen (Figures 6.2b and 6.3b), and was in operation until further dam failure opened a much lower passage towards Norway through Sördalen at 412 m a.s.l. Multiple channel incisions and a general lack of sediment along the western valley side demonstrate the erosive power of the drainage of glacial lake Va-1, as the lake level dropped around 100 m (Figure 6.3b).

Glacial lake Torneträsk stage T-1, with shoreline elevations between 471–495 m a.s.l., presumably had its spillway through Sördalen at 412 m a.s.l. Because the lowest shoreline elevations are meters above the current elevation of the spillway (i.e., in close vicinity to the

54
Figure 6.1: Glacial lake systems associated with deglaciation of the FIS across the Torneträsk and Rautas valleys, including the position of the damming ice margin. Note that T-1, R-1, and R-4 are dammed from multiple sides.
Figure 6.2: The extent, spillways, and drainage channels of the glacial lake stages of Vassijaure (a) and Torneträsk (b-f).
Figure 6.3: Locations of spillways and drainage channels of the glacial lakes of Torneträsk as visible on the hillshade. a) Valley Norddalen that served as a spillway for Vassijaure (Va-1) towards Rombaks-fjorden. b) Sördalen that served as a spillway (T-2) and drainage channel for Torneträsk (T-1) and Vassijaure (Va-1). c) Potential drainage path of T-2 towards glacial lake Vittankijärvi. Contours at 460 m and 470 m a.s.l. are included to aid visualization. d) Drainage channel of T-4. e) Spillway and drainage channel of T-6 and spillway of T-7. f) Drainage channels of T-7.

The spillway at 417 m a.s.l; Figures 6.2c and 6.3b), this potentially reflects deepening of the spillway by c. 5 m through continuous spilling. A coherent ice dam in the southeast must have blocked two lower cols in the topography, corresponding to an ice dam with an E-W orientation. An E-W ice dam appears also justified from nearby eskers and lineations with orientations perpendicular to the ice margin. There are no distinct morphological traces of a drainage event that lowered the lake level to the next stage, hinting it might have been continuous subglacial leakage in conjunction with shifting ice marginal positions. Another
possibility is drainage via the outlet of glacial lake Vittankijärvi (Figure 6.8f), although the ice margin must have retreated further south from its position where it was damming glacial lake Vi-1 to expose terrain between glacial lake Torneträsk T-2 and a lower stage of glacial lake Vittankijärvi (Figures 6.1 and 6.3c).

The ice front that dammed glacial lake Torneträsk stage T-4, with current shoreline elevations between 391 and 432 m a.s.l., probably had a rather similar configuration compared to the ice margin that dammed glacial lake stage T-2 (Figure 6.2d). The ice sheet retreated around 5 km, although the eastern end of the dam curved more towards the southeast. The latter is related to the overall shrinkage of the ice sheet towards the mountains. Glacial lake T-4 had its drainage channel at 424 m a.s.l. with fluvial deposits downstream of the channel over a distance of 1.5 km (Figure 6.3d). It is unclear whether there existed a spillway at all. Perhaps subglacial leakage controlled the highest glacial lake level, until the retreating ice margin exposed the lower terrain through which glacial lake T-4 drained.

Glacial lake Torneträsk stage T-6, with shoreline elevations between 351 and 382 m a.s.l., was dammed approximately 30 km further south (Figures 6.1 and 6.2e). The level was determined by a spillway channel at 383 m a.s.l., with a width of approximately 750 m, the position of which occurs only 2.5 km from its drainage channel (Figure 6.3e). The drainage channel at 365 m a.s.l. has an internal topography of c. 5 m. Erosion of at least 5 m appears to have occurred due to the drainage of glacial lake stage T-6, thus establishing a new spillway altitude for glacial lake Torneträsk stage T-7 at 365 m a.s.l.

The final stage of glacial lake Torneträsk (T-7), with shorelines situated between 347 and 369 m a.s.l., was probably controlled by the drainage channel of T-6 (Figures 6.2f and 6.3e). Drainage of glacial lake T-7, the final drainage of glacial lake Torneträsk, occurred through Torneälven. Two drainage channels paralleling the current river are found at 365 m a.s.l. and 357 m a.s.l. (Figure 6.3f), which appears to demonstrate that the drainage occurred in two steps: an initial drainage along the ice margin, and a final drainage as the low point of Tornedalen opened up.

The drainage of glacial lake Torneträsk T-7 presumably caused an outburst flood of significant volume, namely in the order of magnitude of 19 km³. This was calculated by multiplying the height difference between the lake surface elevation and the drainage channel elevation with the glacial lake area. In comparison, glacial lake Missoula, dammed by the CIS, had a maximum flood volume of 2200–2600 km³ (Hanson et al., 2012). Glacial lake outburst floods (GLOFs), even low magnitude floods, are known to modify the landscape due to their high-magnitude stream power (Thorndycraft et al., 2016; Wells et al., 2022), and, therefore, Tornedalen was inspected for geomorphological traces such as trimlines, channel incision, scoured bedrock, boulder bars, and sediment fans. Indeed, 25 km downstream of the southernmost glacial lake shoreline there are trimlines and deposits in the form of sediment bars, which could be interpreted as erosional and depositional
flood landforms, respectively (Figure 6.4). However, they are difficult to distinguish from ice-marginal glaciofluvial deposits. These deposits remain a topic for future investigation.

The shoreline gradients of Torneträsk have a tilting direction towards the northwest and decrease from 0.7 m/km for the oldest glacial lake stages to 0.5 m/km for the youngest glacial lake stages. The gradients clearly reflect the glacio-isostatic uplift pattern following the deglaciation of Fennoscandia (Steffen and Wu, 2011; Berglund, 2012). The increase of gradients with age demonstrate the longer exposure to post-glacial uplift. Accordingly, the gradients of Torneträsk are slightly higher than the gradients of 0.4–0.5 m/km that Regnéll et al. (2019) calculated for glacial lakes Akkajaure and Sitojaure, located approximately 140 km further south between Kebnekaise and Sarek Mountains. It is, however, questionable whether the gradients can be confidently calculated at this precision in the first place. The calculations strongly depend on the direction of the tilt along which they were calculated, the resolution and accuracy of the DEM, the precision of the mapping of shorelines in ArcGIS, and are complicated by the crosscutting Pärvie Fault.

Extrapolation of the available shoreline and perched delta data (Figure 5.9) to determine the western extent of glacial lake Torneträsk T-7, shows that an absence of shorelines here is most likely due to their submergence as a result of lake-tilting. Glacial lake T-7 shorelines would be predicted at elevations of approximately 325 m a.s.l., which is 17 m below the current lake level of Torneträsk. Bathymetric data is therefore required to faithfully outline the perimeter of T-7, but the most recent open-source bathymetric map is based on measurements from 1920–1921 (SMHI, 2020). Nevertheless, the map shows that
the lake bathymetry contour of 20 m occurs at a relatively short distance inboard from the current lake level, indicating a rather steep basin. The current shoreline of Torneträsk is therefore used as perimeter for the western part of T-7.

The reconstruction of the glacial lakes in the Torneträsk basin resulted in eight glacial lake stages, of which the two highest and lowest stages were the least clear in their expression. Similarly, Melander (1977c) identified at least five glacial lake stages, which overlap in elevation with the stages T-2, T-3, T-4, T-5 and T-6. The notion of Sördalen and its canyon as an outlet for glacial lakes in Torneträsk finds strong support in literature (Sjögren, 1908; Holdar, 1952; Melander, 1977c). Melander (1977c) additionally suspected two potential outlets at the southeastern end of glacial lake Torneträsk, but as the shorelines could not be traced to these proposed outlets, subglacial drainage is mentioned as alternative. Neither Sjögren (1908) nor Melander (1977c) mapped shorelines further south than Jiekajärvi, while the southernmost shoreline in this study is found 25 km further south at Alanen Kallovaara along Torneälven. Hence, this reconstruction shows that the glacial lakes of Torneträsk were more extensive than previously thought. Furthermore, an additional three drainage channels and four spillways could be identified and connected to the stages. It has been debated whether the shorelines reflect the presence of open lakes or ice-marginal lakes (Sjögren, 1908; Holdar, 1952; Melander, 1977c). The distribution of shorelines, with consistent elevations along the lake for the different glacial lake stages, points to open lake systems.

6.1.2 Glacial lakes of Rautasjaure

Many glacial lakes existed at Rautasjaure, yet drainage channels for only a few lake stages could be identified (Figure 6.1, Figure 6.5). Glacial lake Rautasjaure R-1 had shoreline elevations between 834 and 843 m a.s.l., whose high lake levels required dams from both the western and eastern end of the valley (Figure 6.5a). A western ice dam is supported by a lack of shorelines in the valleys west of Rautasjaure and could be realized by an ice tongue in Bessešvágg (see place name in Figure 6.10). This is in agreement with lateral meltwater channels along the valley sides. The eastern dam was realized by a northward flowing ice lobe in the flatter terrain between Bergfors and Rautas (see place names in Figure 6.10), along which the drainage occurred as demonstrated by lateral channels along the slopes of Roahpi (Figure 6.6b). A spillway at 835 m a.s.l. controlled the lake level (Figure 6.6a). The spillway is connected to a deeply eroded canyon, although it is arguable whether the erosion occurred solely due to spilling. It is likely that deepening of the canyon occurred over multiple deglaciations, and not solely due to spilling of glacial lakes, but mainly through drainage of meltwater.

Glacial lake Rautasjaure R-4 had shoreline elevations between 733 and 743 m a.s.l. The ice dams of glacial lake R-4 had a seemingly similar configuration to glacial lake R-1. The western end of the lake was still dammed by the same ice tongue (Figure 6.5b), while the ice lobe between Bergfors and Rautas seemingly pinned itself at Roahpi and continued
to act as the eastern ice dam. Although the frontal margin of the ice lobe probably retreated further south compared to its position during glacial lake Rautasjaure R-1, it was the ice margin in contact with the slopes of Roahpi that thinned and exposed lower terrain. Presumably, drainage occurred repeatedly at this location for multiple glacial lake stages of Rautasjaure, coinciding with thinning of the ice lobe. Each lake level stabilized to the elevation of the drainage channel of the previous glacial lake stage, of which the channel then operated as the new spillway. Deeply eroded lateral channels at multiple elevations support this mechanism (Figure 6.6b).

Glacial lake Rautasjaure R-12 with shoreline elevations between 653 and 668 m a.s.l. was no longer dammed by the ice tongue in the west (Figure 6.5c), as perched deltas demonstrate the lake extended into the adjacent valleys. However, the eastern dam was
Figure 6.6: Locations of spillways and drainage channels of the glacial lakes of Rautasjaure as visible on the hillshade. a) Canyon that served as a spillway of R-1. b) Drainage channels south of Roahpi draining several stages of Rautasjaure (R-1, R-4, R-12). c) Spillway of R-22. d) Drainage channel of R-22.

still maintained by the ice lobe between Bergfors and Rautas. Again, drainage occurred at the same location south of the Roahpi along the ice lobe (Figure 6.6b).

The final glacial lake stage R-22 with a shoreline elevation of 543 m a.s.l. was considerably smaller and only covered the current Rautasälven (Figure 6.5d). The ice lobe at Rautas must have unpinned itself from Roahpi and retreated further south for this glacial lake to have existed. The spillway channel is found at 546 m a.s.l. with fluvial deposits in the form of bars further downstream (Figure 6.6c). This channel probably also drained the previous stages, until glacial lake R-22 established its lake level. The final drainage of glacial lake R-22 probably occurred through Rautasälven towards the northeast. Along the northern shores of the river traces of flowing water have been found, perhaps indicating drainage along an ice front that was partly covering Rautasälven (Figure 6.6d).

Surprisingly, there are no distinct morphological traces of a large drainage event, while there must have been a considerable volume of water that drained at the moment that the ice lobe between Bergfors and Rautas unpinned itself from Roahpi. The only hint seems to be washed bedrock over an area of around 20 km² with a few channels cut into bedrock draining into Njuohčamjávri (Figure 6.6c). If these were indeed traces from a drainage event, there must have been a smaller ice lobe still present at Rautas to direct the water towards north.
For glacial lake Rautasjaure an exceptional number of glacial lake stages were identified compared to Torneträsk (Figure 5.9). In the first place, there were less shoreline data points to infer the glacial lake stages from, which set the bar lower for the amount of data points that were ‘sufficient’ to identify a glacial lake stage. Nonetheless, the shorelines of the glacial lakes of Rautasjaure were generally poorly developed compared to the shorelines of the glacial lakes of Torneträsk, implying each lake level of Rautasjaure perhaps existed for a shorter duration. Hence, the total duration of the Rautasjaure glacial lakes was probably not much shorter than the Torneträsk glacial lakes, even though there were more stages of Rautasjaure. The disparity between the glacial lake systems could be the result of the dissimilar damming mechanisms due to contrasting topography. Glacial lakes Rautasjaure were dammed by an ice lobe that was pinned at a mountain slope, while glacial lakes Torneträsk were dammed by an ice margin that retreated across a lower relief terrain and halted at different locations in the basin.

6.1.3 Other glacial lakes

There are a few other glacial lakes of much smaller size compared to Torneträsk and Rautasjaure (Figure 6.7). Nakerijärvi had two glacial lake stages, although only glacial lake Nakerijärvi N-1 is outlined in Figure 6.1. Glacial lake N-1 with a shoreline elevation around 552 m a.s.l. was dammed in the south (Figure 6.7a). A spillway channel at 555 m a.s.l. controlled the highest lake level by spilling towards the north into Torneträsk, while the adjacent spillway to the west at 544 m a.s.l. controlled the lake level of glacial lake Nakerijärvi N-2 (Figure 6.8a). The drainage of glacial lake N-1 occurred through a deeply eroded lateral channel of which the threshold is at 539 m a.s.l. (Figure 6.8b), which suggests there was an ice lobe present to the east of Nagirvärr.

For Kaitasjärvi only a single glacial lake was identified with shoreline elevations between 497 and 500 m a.s.l., which could fit a northwest-southeast orientated ice dam (Figure 6.7b). Glacial lake Kaitasjärvi K-1 had its spillway spilling towards the north at 497 m a.s.l. (Figure 6.8c). Further downstream the spillway channel is flanked by fluvial deposits. Final drainage occurred south of Kaitasvaara, where a lateral channel starts at 485 m a.s.l. (Figure 6.8d). This is considerably lower than shoreline elevations of glacial lake K-1, but perhaps there was an initial lake level lowering through subglacial leakage, before the final drainage occurred.

For Vittankijärvi three glacial lake stages were identified, but only for the oldest lake stage at shoreline elevations between 498 and 504 m a.s.l. the spillway and drainage channels could be recognized (Figure 6.7c). In order to impound glacial lake Vittankijärvi Vi-1, the glacial lake had to be dammed at three outlets in the south, which required a curved ice margin. The ice sheet probably retreated faster on the low relief terrain around the current lake Vittankijärvi, while its recession was halted in the east due to higher terrain acting as obstacles. The spillway of glacial lake Vi-1 spilled northward at 501 m a.s.l. through the current Rienakjohka (Figure 6.8e). The drainage occurred in the southeastern
The extent, spillways, and drainage channels of the glacial lake stages of Nakerijärvi (a), Kaitasjärvi (b), Vittankijärvi (c), and Luovijärvi (d). The only stage of glacial lake Luovijärvi at an elevation of 442 m a.s.l. covered both the current Luovijärvi and Vittankijärvi, and had to be dammed in a similar way as Vi-1 (Figure 6.7d). A curved ice margin blocked meltwater at three potential outlets in the south and east. Again, ice retreat was probably faster in the flat lake basin than the higher hills around it. The spillway is located at 441 m a.s.l. between two hills in the southwestern end of the glacial lake (Figure 6.8g). The drainage channel at 441 m a.s.l. opened up in the east
Figure 6.8: Locations of spillways and drainage channels of several glacial lakes as visible on the hillshade. a) Spillways of N-1 and N-2. b) Drainage channel of N-1. c) Spillway of K-1. d) Drainage channel of K-1. e) Spillway of Vi-1. f) Drainage channels of Vi-1. g) Spillway of L-1. h) Drainage channels of L-1.
between Taavinunnanen and Koijulaki (Figure 6.8h). Another wider and deeper drainage channel at 433 m is found parallel to it, also draining towards the east, which may indicate an even lower glacial lake Luovijärvi existed.

6.2 Spatial reconstruction of the deglaciation

The reconstruction of the deglaciation of Torneträsk is accomplished by using multiple landform assemblages, each of which contribute to the reconstruction in various levels of detail. Lineations and eskers enable constructing swarms that represent ice flow directions, which express the coarsest level of detail that is accomplished (e.g., isochrons of reconstruction by Stroeven et al., 2016). The ice dams from the glacial lake reconstruction offer a relatively precise indication of successive ice-marginal positions, but only over shorter distances (section 6.1). The finest level of detail is achieved through tracing lateral meltwater channels and marginal moraines, which delineate the ice margins directly, occasionally even in a sequence portraying thinning and retreat of ice tongues. Marginal moraines formed by ice sheet configurations are scarce (Kleman, 1992; Heyman and Hättestrand, 2006), so ice-marginal positions at this level of detail were mostly inferred from lateral meltwater channels.

6.2.1 Swarms

The inversion model from Kleman et al. (2006) is developed to infer ice sheet-wide ice flow patterns, and is thus well-suited for constructing swarms in the premontane region, where the low relief was not a large factor in influencing ice sheet flow. However, in the montane region, ice flow became topographically steered as ice thinned. Ice flow directions, therefore, varied considerably at a local spatial scale, which rendered it impossible to apply the swarm approach in the montane region. For the same reasons, Kleman et al. (2010) treated the CIS different from the LIS in their reconstruction of the North American Ice Sheet. Five different swarms associated with at least two different ice sheet configurations were recognized based on the landforms in the area (Figure 6.9). The swarms are exclusively located in the lower-lying areas, namely the premontane region and the Torneträsk basin. The recognized swarms are aligning with the fans that were identified by Kleman et al. (1997).

The deglacial envelope is confidently identified in the premontane region (Figure 6.9). There are numerous cases where lineations align with eskers, indicative of a wet-bed deglaciation swarm. Lateral meltwater channels in the northern part of the premontane region indicate frozen surfaces despite the wet-bed conditions. The ice sheet initially retreated towards south, but curved towards southwest in the southern part of the study area. It is well-known that the ice sheet retreated faster in the east than west (Boulton et al., 1985; Kleman, 1992; Stroeven et al., 2016). It is thought that the final remnant disappeared in the eastern Sarek mountains (Kleman et al., 1997; Boulton et al., 2001; Stroeven et al., 2016).
The distribution of glacial lakes indicates, more precisely, that the final deglaciation occurred to the southeast of Sarek (Regnéll et al., 2019). The change in retreat direction supports the suggested location of the final remnant.

The deglacial envelope is overprinting an event swarm that is solely composed of lineations, mostly in the form of large drumlins (event swarm 1, Figure 6.9). These drumlins are interpreted to have formed during multiple glaciations of similar ice sheet configurations (Hättestrand et al., 1999, 2004). The increase in lineation size with age appears to be merely because larger drumlins were more likely to survive overriding than smaller lineations (Hättestrand et al., 1999). The orientation of the drumlins indicate ice flow towards southeast, which corresponds to an ice sheet configuration with its ice divide parallel to the mountain divide in the west (Figure 1.2a; Kleman, 1992; Kleman et al., 2008).

Several event swarms with other flow directions were identified along the Torneträsk basin in the west (event swarm 2, Figure 6.9). These lineations are mostly formed in bedrock, while only few crag-and-tails indicate ice flow towards (north)west. The scoured bedrock is part of a western scouring zone, which is interpreted as demonstrative of rapid ice flow towards marine-terminating ice streams on the Norwegian shelf, compatible with west- and east-centered ice sheet configurations (Kleman et al., 2008). Hence, it cannot be ruled out that these lineations were in fact part of the deglacial envelope, as the ice flow

Figure 6.9: Distribution of swarms in the study area. Note that the deglacial envelope is overprinting event swarm 1.
direction corresponds to the ice sheet configuration during the last deglaciation (Figures 1.2c). The observed lack of eskers and lineations with sediment tails might be due to a lack of sediment availability, as the bed was already scoured before the onset of the last deglaciation due to the cumulative erosive power of previous glaciations (Kleman et al., 2008). The lack of meltwater channels and eskers precludes a conclusive explanation regarding the significance of these swarms in reconstructing the deglaciation history.

### 6.2.2 Ice-marginal positions

Ice-marginal positions are largely based on the inferred locations of ice margins damming glacial lakes. Their positions are guided by topography and extrapolated to construct a glaciologically plausible ice sheet margin. The lateral meltwater channels and marginal moraines intermittently assisted in this process. The inferred ice sheet margins are shown in Figure 6.10. The successive ice margin positions, from older to younger, are indicated by numbers 1–9. Factors influencing the spatial patterns of retreat are first discussed on a general level, after which the ice-marginal positions are discussed separately in a more detailed manner for the premontane and montane regions.

There is a clear distinction between the retreat pattern in the premontane region and the montane region in terms of the configuration of the ice sheet margins. In the premontane region the ice sheet maintains its shape as one more or less coherent ice body, while the ice sheet disintegrates into several ice lobes in the montane region. The strong control of topography on ice retreat is evident, as studies for the FIS (Stroeven et al., 2016), the BIIS (Greenwood et al., 2007; Hughes et al., 2014), and CIS (Kleman et al., 2010; Dulfer et al., 2022) support as well. The underlying topography becomes increasingly dominant as the ice thins (Hughes et al., 2014), hence topographic control is especially significant during ice expansion and final deglaciation.

Besides topography the presence of glacial lakes presumably influenced ice dynamics during deglaciation. Proglacial lakes amplify glacier mass loss and glacier velocity (Carrivick and Tweed, 2013; Sutherland et al., 2020), so an effect on regional-scale ice sheet patterns is likely. Numerical modeling of the North American Ice Sheet demonstrate that large proglacial lakes can accelerate grounding line retreat (Quiquet et al., 2021). Reconstructions of retreat of the LIS show that large, long-lived proglacial lakes promoted surging (Stokes and Clark, 2004; Utting and Atkinson, 2019). It is evident that lacustrine-terminating ice and terrestrial-terminating ice demonstrate different behavior (e.g., Stroeven et al., 2016), but the effect is probably dependent on whether calving occurs. Calving rates increase with water depth, but flotation in the first place is controlled by the ice thickness relative to the lake depth (Benn et al., 2007).

A question remains as to whether glacial lakes in the study area reached a sufficient water depth for the ice sheet to start floating and calving. De Geer moraines, which are interpreted to have formed at the grounding line of water-terminating ice margins (Lindén and Möller, 2005; Bouvier et al., 2015; Ojala et al., 2015), have occasionally been found
Figure 6.10: The inferred ice margins based on ice dams, lateral meltwater channels, and marginal moraines. The numbers indicate successive ice margin positions from old to young (1-9).
in relation to glacial lakes (Johansson and Kujansuu, 2005, as cited in Ojala et al., 2015; Öhrling et al., 2020). Lindén and Möller (2005) presumed De Geer moraine formation, indicative of potential calving, required a water depth of at least 70 m. There were no De Geer moraines identified in the study area, but based on the water depths of the glacial lakes, it appears likely that at least glacial lakes Torneträsk T-0, T-1, and T-2 had sufficient water depths for calving (Table 3). Thus, ice retreat across the Torneträsk basin along the lacustrine-terminating margin might have been accelerated through calving.

6.2.2.1 Premontane region

The ice sheet exhibits a relatively orderly retreat in the premontane region (Figure 6.10), which at large fits the retreat pattern of the deglacial envelope (Figure 6.9). The northernmost ice marginal positions (1) are outlined by lateral meltwater channels that document ice sheet thinning. Ice sheet flow is predominantly directed towards the north. The orientation of the ice dam of glacial lake Vassijaure Va-1 indicates that ice probably retreated eastward in the western part of the Torneträsk basin. The ice sheet must have had a radial flow pattern at that time.

Ice marginal positions 2–7 are approximately parallel, but slightly undulating as local relief and glacial lakes influenced ice retreat rates (Figure 6.10). The ice margin initially retreated southward, but eventually converged towards southwest. The ice lobe at the Bergfors-Rautas corridor retained its configuration during retreat. The ice sheet pinned itself at Roahpi (positions 7–8), which acted as a pivot point as the ice margin to the east of it retreated in a "swinging" manner towards the southwest. The ice margin to the west of Roahpi simultaneously blocked glacial lakes Rautasjaure. Glacial lakes Rautasjaure R-1, Kaitasjärvi K-1, and Torneträsk T-7 possibly existed around the same time (position 4 in Figure 6.10). The drainage of the final glacial lake stage of Torneträsk probably occurred while glacial lake Rautasjaure was still dammed (position 7 in Figure 6.10). The latest, also southernmost, ice margins (9) were mainly outlined based on lateral meltwater channels. By now the previously merely undulating ice sheet margin appears to have transformed into more distinct lobes. The final stage of glacial lake Rautasjaure R-22 had drained by now.

Two smaller lobes emerging from both sides of Sálvvučohkka are inferred from a series of lobate-shaped minor ridges (between positions 8–9 in Figure 6.10; description in section 5.2.3). Although their exact formation process is elusive, the morphology implies ice-marginal formation. Annual deposition of the ridges would indicate a retreat rate of around 50–90 m/yr, which is low compared to reported retreat rates of 200–1600 m/yr (Stroeven et al., 2016). Therefore, the ridges are more likely to have formed sub-annually. The rarity of these minor ridges is perhaps explained by formation through a small re-advance in stead of ordinary retreat.
6.2.2.2 Montane region

It is noteworthy to mention the complexity of reconstructing the deglaciation in the montane region. Disentangling the traces in the mountains and determining during which glaciations the landforms were formed is challenging, as it is unclear whether moraines were formed by advancing valley glaciers situated up-valley of the moraine or by outlet glaciers from a thinning ice sheet situated down-valley of the moraine (Heyman and Hättestrand, 2006). It is known, however, that the moraines in the mountains are not exclusively from the last deglaciation (Kleman, 1992; Fredin and Hättestrand, 2002; Heyman and Hättestrand, 2006). Melander (1980) attributed all meltwater traces to the last deglaciation, but some of these could just as well originate from previous deglaciations (Kleman, 1992).

As the ice sheet thinned, the higher peaks of the mountains emerged as nunataks, which in all probability influenced the surrounding ice surface elevations (Mas e Braga et al., 2021), and thus impacted local flow patterns. As thinning continued, the ice sheet disintegrated into several ice tongues, most likely with various retreat directions and pace. The large ice lobe originating from Abiskodalen must have retreated quickly southward, as indicated by perched deltas formed in glacial lake Torneträsk T-2 at the mouth of Abiskodalen (see difference in configuration between the piedmont lobe from position 2 to 4 in Figure 6.10). The ice lobe probably separated into a larger ice tongue in Abiskodalen and a smaller valley glacier in Gorsavággí.

The lack of evidence for further glacial lakes in mountain valleys might be due to active slope processes that obscured the shorelines or due to the short-lived nature of the glacial lakes. For example, the E-W trending Gorsavággí valley must have been dammed by the ice lobe in Abiskodalen (Figure 6.10). However, if the valley was filled by ice at this time, no glacial lake traces should be expected.

The ice tongue in Bessešvággí retreated southward as well, simultaneously damming the western end of Rautasjaure (Figure 6.10). It is possible that the ice tongue became pinched off from the ice sheet, as the N-S trending Bessešvággí continued into the E-W trending valleys of Rávtasvággí and Alisvággí. This is supported by hummocky moraine in the latter valley, which could be indicative of a dead-ice body (Boone and Eyles, 2001).

Present glaciation is restricted to the higher massifs, such as Kebnekaise and Sarek, in the form of cirque and valley glaciers. It seems plausible that after deglaciation ice remnants remained in the higher mountains, while the surrounding landscape became ice free. The mapped contemporary-glacier moraines, which were not utilized for demarcating the ice-marginal positions, indicate that variations in glacier extent occurred throughout the Holocene (Karlén and Denton, 1976; Heyman and Hättestrand, 2006).

6.3 Ice properties of the ice sheet

The lobate ice margins signify the dynamic nature of the ice sheet, although this merely portrays ice-marginal velocity variations. Further away from the retreating ice margin,
the distribution of lineations illustrates that basal sliding and sediment deformation was spatially varying. Additionally, preservation of non-glacial surfaces (Hättestrand and Stroeven, 2002; Goodfellow et al., 2008, 2014) and landforms of previous glaciations (Kleman, 1992; Hättestrand and Stroeven, 2002) demonstrate negligible glacial erosion occurred in places, which is confirmed by cosmogenic nuclide inheritance (Stroeven et al., 2002; Fabel et al., 2002). The distribution of those preserved landscapes is patchy, indicating the thermal regime was regionally varying (Stroeven et al., 2021). Moreover, the erosional effectiveness of the ice sheet declined with increasing elevations (Jansen et al., 2019), which is a logical consequence of the relationship between ice sheet thickness and pressure melting at the bed.

Several landform assemblages support spatial and temporal variation in the thermal regime at the base of the ice sheet. The aligned lineations and eskers demonstrate that the deglaciation was largely warm-based in the premontane region. These landforms were formed at the bed relatively close to the retreating margin (Stroeven et al., 2021), but the ice margin itself must have been impenetrable to act as ice dams of the glacial lakes. The same accounts for the formation of lateral meltwater channels, especially abundant in the montane region, as the surface of the ice must have been impenetrable to drain supraglacial meltwater towards the valley sides. The ice dams and lateral meltwater channels therefore suggest the ice margins were at least cold-based.

Sets of lateral meltwater channels are usually associated with cold-based conditions (Greenwood et al., 2016), but this relationship does not appear to be so clear-cut. Observations of contemporary polythermal glaciers show that meltwater drainage occurs both subglacially and supraglacially (Irvine-Fynn et al., 2011). Furthermore, lateral meltwater channels have been reported along the margin of a temperate glacier margin in Alaska (Syverson and Mickelson, 2009), and the supraglacial drainage network of a cold-based glacier reached the bed by means of downcutting (Naegeli et al., 2014). These examples illustrate that interpretations regarding the thermal regime as inferred from the presence of lateral meltwater channels should be treated with caution.

A relatively large proportion of the region in the east is covered by ribbed moraine (Figure 5.4), which is indicative of a transition from cold-based to warm-based conditions further towards the interior of the ice sheet (Kleman and Hättestrand, 1999). The ribbed moraine is often fluted, which supports formation of the ribbed moraine in the interior, and subsequent fluting by a retreating margin. Ribbed moraine is rare in the montane region, which might signify that a transition from cold-based to warm-based flow never occurred. Ice was likely funneled through trough valleys during the whole glaciation, maintaining warm-based flow throughout. Ice at higher elevations probably never reached a sufficient thickness for pressure melting at the base, hence maintaining cold-based ice flow.

There exists additional geomorphological evidence that is suspected to represent changes in thermal regime. For example, there is a zone that extends for over 20 km from the minor ridges west of Alajärvi to Kiruna that shows a transition between a smoother
surface texture in the west to a rougher surface texture to the east. This possibly reflects a thermal regime transition from warm-based to cold-based conditions. Additionally, minor ridges that are associated with crevasse formation are potentially created due to spatially varying ice velocities, which in turn could result from a thermal regime shift. Although their origin might be arguable, they appear to support the observation that the region was subjected to varying basal thermal conditions.

6.4 Timing of ice retreat

This study did not apply any absolute dating techniques, and therefore has to rely on previously published data, and the cross-cutting relationship between the Pärvie fault and raised shorelines.

6.4.1 Previously-published chronological data

Resources that can be utilized are the chronological databases underlying the reconstructions of the FIS (Stroeven et al., 2016), and the EIS (Hughes et al., 2016). The available dates for the study area are acquired through radiocarbon ($^{14}$C) and Terrestrial Cosmogenic Nuclide (TCN) dating. The samples for the $^{14}$C-dates were taken from macrofossils imbedded in minerogenic sediments, bulk sediment, and basal peat from several different lakes (Karlén, 1979; Shemesh et al., 2001; Rubensdotter, 2006; Rubensdotter and Rosqvist, 2009). The ages should therefore reflect a minimum deglaciation age, as these lake basins became available for accumulation of organic material after ice retreated (Karlén, 1979; Lundqvist and Mejdahl, 1995; Stroeven et al., 2016). The calibrated $^{14}$C-ages range from 9.5–10.4 cal ka BP, which is a substantial range of 1000 years, considering their uncertainties are in order of magnitude of 1%.

The TCN ages were sampled from different bedrock surfaces, glacial boulders, and rock falls assumed to be triggered by post-glacial pressure release (Fabel et al., 2002; Stroeven et al., 2002, 2016; Li et al., 2005; Goodfellow et al., 2014). The samples can accurately reflect deglacial ages as accumulation of cosmogenic nuclides first starts when ice cover is less than 10 m thick (Stroeven et al., 2002). However, samples from relict surfaces that barely eroded due to cold-based ice sheet conditions, inherited cosmogenic nuclides from previous ice-free periods and yield older apparent exposure ages (Fabel et al., 2002; Stroeven et al., 2002; Goodfellow et al., 2008).

The TCN ages that are statistically deviating from the assumed deglaciation age of 9.5 ka BP range from 13 to 128 ka BP, and are therefore assumed to indicate surfaces that were not eroded sufficiently to remove prior cosmogenic exposure (Stroeven et al., 2002). Their ages are significantly older than the expected deglaciation ages, and although very useful for deducing patterns of glacial erosion (Stroeven et al., 2002; Li et al., 2005; Jansen et al., 2019), they do not aid constraining the timing of deglaciation. The remaining ages are dated between 8.3 and 12.6 ka BP. The standard deviation of the TCN ages is
relatively large (around 10%) in comparison to the standard deviations of $^{14}$C-ages, and TCN-dating using $^{10}$Be, $^{26}$Al, or $^{36}$Cl, is, therefore, not precise enough to differentiate between deposits or surfaces with an age difference of maximum few hundred years.

The distribution of the ages is not consistent with the reconstructed pattern of retreat. The ice sheet retreated southward, with faster retreat on the eastern side than on the western side (this study; Boulton et al., 1985; Kleman, 1992; Stroeven et al., 2016). Hence, it would be expected that the oldest ages are located in the north(east) and the youngest ages in the south, but in reality the ages reflect no spatial pattern. Because the majority of the samples were taken in the mountains, their varying ages might reflect that the ice sheet did not retreat in orderly fashion in the montane region. It appears that the available dates are suitable for providing an approximate age of deglaciation, but can not resolve the temporal gradients associated with the smaller spatial scale at which this study was undertaken.

The time window that can be dated with the current techniques is simply too wide for the temporal resolution that is required to investigate the retreat pattern at the spatial scale of the study area. However, sampling across multiple landforms could enhance the reliability and temporal resolution. A cosmogenic study at a site in northern Norway yielded exposure ages across both erosional and depositional landforms, which allowed for an examination of the variation in age between the sampled locations (Stroeven et al., 2011). A multiple sampling environment provided a more reliable mean age with a lower standard deviation, which is an approach that should be adopted for more reliable deglaciation ages (Stroeven et al., 2011).

The isochrons of the deglaciating FIS, as reconstructed by Stroeven et al. (2016) and Hughes et al. (2016) based on the geomorphological record and the geochronological database, respectively, are presented in Figure 6.11. The retreating ice margin deglaciated the total study area in a time span of 500 yr (Figure 6.11a). Both studies agree on an approximate deglaciation age of 10 ka BP (Figure 6.11). The ice-marginal positions that dammed the successive Torneträsk glacial lakes fall approximately in between the positions of isochrons 10.1 and >9.9 ka (Figure 6.11a), which would suggest the glacial lake system of Torneträsk existed for a total duration of only 150 yr.

A duration of merely 150 yr appears short for the development of several generations of shorelines. Waves play a key role in the development of glacial lake shorelines (Lorang et al., 1993; Schuster and Nutz, 2018), although in periglacial environments lake ice potentially influences shoreline morphology as well (Matthews et al., 1986). Regardless, the development of shorelines is largely dependent on the duration of exposure to shoreline-building processes, which requires a stable lake level for a certain period. Shorelines are estimated to require at least a few decades to develop (Melander, 1977c; Thompson and Baedke, 1997), which would result in a minimum duration of approximately 240 yr when assuming eight glacial lake stages for Torneträsk and an average duration of shoreline formation of 30 yr.
Figure 6.11: Isochrons representing retreat of the FIS in the study area (assembled from supplementary datasets from Stroeven et al. (2016) and Hughes et al. (2016)). a) Isochrons representing timesteps of 100 years around 10 ka as reconstructed by Stroeven et al. (2016). b) Isochron representing the mean and maximum interpretation of the 10 ka isochron as reconstructed by Hughes et al. (2016).
6.4.2  Use of post-glacial faults

The 155 km long Pärvie fault is the longest fault line in Sweden, which is composed of a series of west-facing fault scarps (Lagerbäck and Sundh, 2008). The NNE trending fault line crosscuts Torneträsk and its glacial lake shorelines. The Pärvie fault cuts the shorelines of T-2, T-3, T-4, and T-5, while shorelines of T-6 and T-7 appear to postdate faulting (Figure 5.9a). The shorelines from glacial lake stages T-0 and T-1 also appear uninterrupted, but the jump in elevation is most likely poorly documented due to an insufficient number of data points. The cross-cutting shows that the Pärvie fault ruptured between the formation of glacial lake stages T-5 and T-6. This indicates the fault ruptured in close proximity to the ice margin, and according to the reconstruction in Figure 6.10, extended 80 km underneath the ice sheet. No spillways or drainage channels could be identified for glacial lake stage T-5, and therefore its configuration was not outlined. Perhaps the rupture of the fault aided drainage of glacial lake T-5, as dam failure could be initiated during seismic activity.

Previous research associated the formation of several faults in northern Sweden with the last deglaciation as indicated by cross-cutting relationships of the faults with landforms formed during the last deglaciation (Lagerbäck and Witschard, 1983; Lagerbäck, 1992; Lagerbäck and Sundh, 2008). Although Melander (1977c) does not mention a fault crosscutting the series of glacial lake Torneträsk shorelines, crosscutting of shorelines has been reported further south along the Pärvie fault (Lagerbäck and Witschard, 1983; Mörner, 1985). A vertical displacement of 5–6 m, as inferred from the offsets of glacial lake shorelines T-2 to T-5, fits field observations and photogrammetric measurements along the Pärvie fault scarp demonstrating displacements varying between 3 and 10 m (Lagerbäck and Sundh, 2008). Cross-cutting relationships do not always conclusively indicate whether rupturing of the faults happened before or after local deglaciation (Lagerbäck and Witschard, 1983), although it seems that most observations point towards a post-glacial age (Lagerbäck, 1992; Sutinen, 2005; Lagerbäck and Sundh, 2008; Smith et al., 2014). The postglacial faults are thought to have formed through a combination of tectonic and isostatic stresses caused by the glacio-isostatic rebound of Fennoscandia as the ice sheet vanished (Arvidsson, 1996; Lagerbäck and Sundh, 2008; Kukkonen et al., 2010; Smith et al., 2014).

It appears that the Pärvie fault was created through a single seismic event. Stratigraphical analysis of sediments at this fault and other fault scarps in northern Sweden yield that no (detectable) repeated movements occurred since activation around the last deglaciation (Lagerbäck and Sundh, 2008). A hydroacoustic survey of the lake floor of Torneträsk by Vogel et al. (2013) revealed that the deposition of (glacio)lacustrine sediments nearby the cross-cutting fault appear uninterrupted by seismic activity since deglaciation. The high stress drop (difference between stress across a fault before and after an earthquake) is similar to stress drops of other large earthquakes, which makes it likely that a single earthquake occurred (Arvidsson, 1996). The Pärvie fault was in that case formed by a thrust earthquake with a moment magnitude of $M_w \approx 8.2$ (Arvidsson, 1996). Furthermore,
a single seismic event is supported by the lack of vertical displacement for the series of shorelines of T-6 and T-7. On the other hand, equal vertical displacement for the interrupted series of shorelines would be expected in case of a single seismic event, but in reality the vertical displacement differs incoherently between the glacial lake stages for reasons that remain unclear.

Unfortunately, no absolute ages are available for the Pärvie fault. However, dating of secondary deposits is another method to estimate the timing of fault rupture. For example, the formation of landslides in northern Fennoscandia have been associated to earthquakes caused by post-glacial faulting (Sutinen, 2005; Lagerbäck and Sundh, 2008). There are two large rock slope failure (RSF) deposits in the study area that were potentially triggered by the Pärvie fault rupture. A boulder from the RSF deposit in Kärkevagge is TCN-dated at 9.1 ± 1.4 ka BP (Stroeven et al., 2002), while the RSF in Bessešvágggi is TCN-dated at 8.9 ± 0.7 ka BP and 10.9 ± 0.8 ka BP based on samples from the bedrock scar and a boulder, respectively (LSDn scaling method, supplementary dataset from Stroeven et al., 2016). The bedrock scar provides a more reliable age, as the boulder is more likely to have inherited previous exposure. However, slope activity after the main failure could result in a younger apparent age of the bedrock scar, but the age appears to strongly correlate to the deglaciation age of 8.8 ± 0.7 ka BP at the bottom of the valley of Alisvággi (Li et al., 2005). Alisvággi is the E-W trending valley extending from Bessešvággi, where the RSF is located. The approximate age of the Pärvie fault is thus around 9 ka BP, only slightly younger than the assumed deglacial age of 10 ka BP from the reconstructions by Stroeven et al. (2016) and Hughes et al. (2016).

6.5 Future research

This study managed to refine the glacial lake reconstruction of the Torneträsk area, but there are still elements in the reconstruction that could benefit from further exploration. First of all, the timeline of the formation of the glacial lakes is still uncertain. Dating the shorelines directly would be ideal, for example through radiocarbon dating of incorporated organic matter (e.g., Young et al., 2021) or OSL dating of shoreline sediments (e.g., Lee et al., 2009). During fieldwork it became apparent that obtaining fine-grained sediments of raised shorelines and perched deltas, which is required for OSL dating them (Jenkins et al., 2018), was problematic. In recent years luminescence dating developed to apply the technique to larger clasts (Freiesleben et al., 2015), which suddenly enables dating a wider range of sediments, and could further constrain the timing of the damming of the separate glacial lakes.

Another method to constrain the timing of the glacial lakes of Torneträsk is by dating the Pärvie fault. Dating the Pärvie fault itself would give a minimum age for stage T-5 and a maximum age for stage T-6. Usually the ages of fault scarps are determined by dating seismically deformed sediment by means of radiocarbon dating (e.g., Sutinen, 2005), cosmogenic nuclide dating and luminescence dating (e.g., Abbas et al., 2022). While fault
scarps themselves could technically be dated directly through TCN dating (Benedetti and van der Woerd, 2014), Lagerbäck and Sundh (2008) consider this method not accurate enough for more precise dating of the fault rupture. Perhaps future optimization of the technique allows for more precise dating.

Another issue is the outdated bathymetrical data of Torneträsk. The most recent freely-available bathymetry map is hand-drawn and based on measurements from 1920–1921 (SMHI, 2020). This data is inadequate for analysis in GIS software, and could nowadays be acquired in higher resolutions. Airborne LiDAR Bathymetry (ALB) has been used for data collection of shallow water environments, but maximum depths of up to 50 m can only be measured in clear water (Lague and Feldmann, 2020). Generation of deep lake bathymetry therefore relies on other techniques, such as single beam and multibeam echo-sounders, or satellite-derived bathymetry (Getirana et al., 2018; Wöfl et al., 2019). A high-resolution bathymetry map would allow for the lowest glacial lake stage of Torneträsk (T-7) to be demarcated, but would more importantly reveal lake floor morphology, which potentially contains morphological evidence of lacustrine-terminating ice margins. This is especially valuable for investigation the influence of glacial lakes on ice dynamics.

Last, there are potential traces of a GLOF along Tornedalen further downstream of the borders of the study area. Analysis of the GLOF erosion and deposition by use of LiDAR-based DEMs could give insight into the magnitude of the flood in terms of peak flow, duration and sediment yield (Thorndycraft et al., 2016). This is critical knowledge considering the observed and predicted increase in GLOF-related hazards due to climate change (Zheng et al., 2021)). As GLOFs are hardly ever witnessed during the event itself, the geomorphological and sedimentological modifications of the landscape is often the only way to reconstruct the flood (Wells et al., 2022). Further investigation is required to resolve the origin of these potential GLOF traces.
7 Conclusion

This study aimed to reconstruct the spatial and temporal retreat patterns of the FIS in the Torneträsk region in northwestern Sweden. The reconstruction was predominately based on the analysis of the glacial lake systems of Torneträsk and Rautasjäure in terms of lake configuration and final drainage routes. Geomorphological mapping based on a high-resolution LiDAR-derived elevation model enabled the identification of a range of glacial landforms, which were utilized through an inversion model to form swarms representing spatially and temporally coherent ice sheet flow systems. Additionally, glacial lake traces allowed for the identification of ice margins which dammed glacial lakes in Torneträsk, Rautasjäure and a few other (former) lake basins.

Eight distinct glacial lake stages were identified for the Torneträsk basin, of which the lowest stages demonstrate the lake covered a larger extent than previously thought. The multiple lake stages had different spillways and drainage routes, but the final drainage occurred through Tornedalen. Potential GLOF deposits were identified further downstream, but require closer inspection. Over 20 stages were identified for the glacial lakes of Rautasjäure, whose drainage occurred at the same outlet along the margins of a thinning ice lobe that pinned itself at a mountain slope. Geomorphological traces of a final drainage event are lacking, thereby remaining an issue to be addressed. The disparity between the glacial lake systems in terms of the number of lake stages is likely the result of different damming mechanisms related to the contrasting topography of the basins.

Collectively, the swarms and ice dams provided insight into potential successive ice-marginal positions during deglaciation. While the ice sheet retreated southward relatively orderly in the premontane region, it disintegrated into several ice lobes in the montane region. Evidently, the topographic control on ice sheet retreat was significant. Presumably, other factors played an important role, such as the interaction between the ice margin and (pro)glacial lakes. Several landform assemblages demonstrate spatial and temporal variation in the basal thermal regime of the ice sheet.

Chronological databases provided deglaciation ages ranging from 9.5 to 10.4 cal ka BP for radiocarbon dates and from 8.3 to 12.6 ka BP for TCN dates. Unfortunately, the temporal resolution of the current dating techniques is insufficient to constrain and differentiate the successive ice-marginal positions at the spatial scale of the study area. The total duration of the existence of the glacial lakes of Torneträsk is estimated to be around 200–300 yr as inferred from the required duration for shoreline formation. Rock slope failures assumed to be triggered by the Pärvie fault, which is cross-cutting a series of shorelines of glacial lake Torneträsk, are dated at 9 ka BP. Precise dating of the Pärvie fault itself would pinpoint the age of the ice margin that was located between two glacial lake stages of Torneträsk at the time of rupture.
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A Appendix

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Table A.1: DGPS elevation measurements of the shorelines at Varrégaska (see Figure 3.1b for locations). Coordinate system: SWEREF99.