



Stratigraphic Evidence of Holocene Thermo-erosional Valley Development in Ice-rich Permafrost on Kurungnakh Island (Lena River Delta, Siberia)

GEO 511 Master's Thesis

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Abstract

As arctic regions experience rapid climate warming, degradation processes such as thermal erosion increasingly affect the geomorphology and hydrology of vast permafrost landscapes. These processes involve the formation of distinct landforms, such as thermo-erosional valleys (TEVs). Especially in the ice-rich Yedoma permafrost of Siberia, thermal erosion has created widespread TEV networks during the Holocene, but our understanding of their impact on affected deposits remains very limited.

Here, I investigated the Holocene development of a TEV in late Pleistocene Yedoma deposits on Kurung-nakh Island (Lena River Delta, Siberia), based on near-surface stratigraphy and surface characteristics. I analysed five permafrost cores up to 184 cm long, taken on the floor, slope and adjacent Yedoma upland of a TEV. The stratigraphic analysis included computed tomography (CT) scans, cryostratigraphic classification, radiocarbon dating, volumetric ice content, grain size distribution, robust endmember analysis, total organic carbon (TOC) content, TOC/TN (total nitrogen), stable water isotopes ($\delta^{18}\text{O}$, $\delta^2\text{H}$, d-excess), pH and electrical conductivity. Additionally, I included distributed measurements of active layer thickness (ALT) and water table depth (WTD) and GNSS transects, taken within the study area in September 2017 during the core drilling campaign.

The deposits were mainly of Holocene age, reaching 11399 calibrated years before present (cal yr BP) on the Yedoma upland and about 753.5 cal yr BP at the valley floor. For a sediment lens within the wedge ice contained in one slope core I specified an age of up to 22220 cal yr BP. All samples exhibited high volumetric ice contents (>52 vol%), and except for the valley floor, all cores contained massive wedge ice. Stable isotope data showed more depleted values in wedge ice and a negative relationship between $\delta^{18}\text{O}$ and d-excess at specific depths of the valley floor, suggesting freezing fractionation. Polymodal grain size distributions and five robust endmembers indicated multiple sediment transport processes. Fluctuating grain size distributions, TOC contents, and TOC/TN ratios in the near-surface stratigraphy of the valley floor core reflected alternating high- to lower-energy depositional phases.

Together with ALT and WTD measurements, these findings suggest a stable, well-drained Yedoma upland, shaped by postdepositional processes which reworked the Yedoma deposits during the Holocene. The slopes show evidence of little soil accumulation while reflecting a stable TEV system from the Mid Holocene up to present. The young valley-floor deposits, along with high ALT and low WTD measurements, indicate ongoing TEV development in a variable energy environment with alternating processes and magnitudes of sediment accumulation during the Late Holocene. Although the permafrost core dataset does not allow a precise determination of the onset of TEV development, initiation around the Late Pleistocene–Holocene transition appears likely. Overall, the valley, particularly its slopes with shallow sediment covers above wedge ice, remain highly vulnerable to continued warming and extreme climatic events. Multi-year data and deeper cores could provide further valuable insights into the temporal evolution of this fragile permafrost landscape.

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List of Abbreviations

ALT	Active Layer Thickness
a.r.l.	Above River Level
a.s.l.	Above Sea Level
AWI	Alfred Wegener Institute Helmholtz Centre for Polar and Marine Research, Germany
BP	Before Present
cal yr BP	Calibrated Years Before Present
CT	Computed Tomography
d-excess	Deuterium-excess
EC	Electrical Conductivity
GMWL	Global Meteoric Water Line
GNSS	Global Navigation Satellite System
GSD	Grain-size Distribution
H1-4	Hypotheses 1-4
IC	Ice Complex
LMWL	Local Meteoric Water Line
TOC	Total Organic Carbon
TN	Total Nitrogen
rEM	Robust Endmember
rEMMA	Robust Endmember Analysis
RQ1-2	Research Questions 1-2
$\delta^2\text{H}$, δD	Stable Water Hydrogen Isotope
$\delta^{18}\text{O}$	Stable water oxygen isotope
SD	Standard Deviation
TEG	Thermo-erosional Gully
TEV	Thermo-erosional Valley
UFZ	Helmholtz Centre for Environmental Research
VIC	Volumetric Ice Content
VSMOW	Vienna Standard Mean Ocean Water

WTD	Water Table Depth
yr BP	Years Before Present

1 Introduction

The Arctic is warming approximately four times faster than the global average, making it one of the most climate-sensitive regions on Earth (Rantanen et al., 2022). This trend of increasing temperatures leads to the thawing of permafrost. Permafrost is defined as ground that remains at a temperature below 0 °C during at least two consecutive years (Boike et al., 2012). Permafrost covers substantial geographical areas of the Earth. On the northern hemisphere about 15 % of the total land surface have a mean annual ground temperature below 0 °C and thus account as permafrost area (Obu et al., 2019). In continuous regions such as the Lena River Delta in Siberia its thickness reaches 400 m to 1000 m (Zaplavnova et al., 2024). The thawing of permafrost has been shown to liberate substantial quantities of greenhouse gases stored in frozen soils, while also altering the geomorphology, hydrology, and ecology of Arctic landscapes (Briones et al., 2024; Schuur et al., 2015). These highly dynamic responses to climate forcing are, related to its feedback mechanisms, a highly sensitive climate archive that preserve information on past environments while simultaneously shaping present-day ecosystems (Meyer et al., 2015). Against this backdrop, understanding permafrost dynamics in specific Arctic environments becomes critical for assessing future landscape responses.

At study sites across the Arctic, permafrost and its archive of interactions with the environment are being studied for many decades. Especially, permafrost deposits with high ice and organic content are sensitive indicators of the past environment and climate. At the same time these deposits form a potential amplifier of future change. Thaw processes, such as thermokarst and thermal erosion can induce rapid degradation in lowlands, wetlands and on hillslopes (Strauss et al., 2021). A geological formation of high interest are the ice- and organic-rich Pleistocene deposits, named Yedoma, which cover extensive areas in Siberia, Alaska, and Canada (Strauss et al., 2017). This geological unit represents syngenetically frozen permafrost. Yedoma is ice- and organic-rich and consists of poorly-sorted silt to fine sand (Grosse et al., 2007; Schirrmeister et al., 2011b). The high ice content makes the deposits especially prone to rapid thaw processes and highly susceptible to permafrost degradation. One of the most prominent regions where the interactions of Yedoma deposits can be studied is the Lena River Delta in northeastern Siberia. Beyond its sensitivity to thaw, Yedoma constitutes an exceptional sedimentary and cryostratigraphic archive, preserving information on Late Pleistocene to Holocene sedimentation, permafrost formation, and palaeoenvironmental conditions in Arctic lowlands (Schirrmeister et al., 2011b).

The Lena River Delta and its Yedoma deposits combine permafrost, hydrology, and sedimentation interactions, making it an ideal region to study permafrost dynamics and to access well-preserved stratigraphic archives. Here, the polygenetic permafrost is very thick with underlying epigenetic permafrost and upper syngenetic deposition horizons such as the Yedoma and overlaying Holocene deposits. Insular remnants of a late-Pleistocene accumulation plain with Ice Complex (IC), Yedoma deposits feature the third out of the three geological river terraces of the delta (Grigoriev, 1993). Kurungnakh, an Island located in the southern part of the delta, is mainly shaped by this third river terrace. The sediments of this island compose of a sandy deposit formation, covered by the IC as the second formation which is overlain by Holocene deposits (Schirrmeister et al., 2003, 2011a; Schwamborn et al., 2002; Wetterich et al., 2008a). Thus, study sites on Kurungnakh provide key settings to investigate permafrost dynamics and their responses to climatic and geomorphic processes.

The modern surface geomorphology of Kurungnakh Island is strongly shaped by permafrost degrada-

tion, such as thermokarst and thermal erosion (Morgenstern et al., 2021b). Both are major processes in Arctic ice-rich periglacial landscapes and form landscape features that interrupt the ice-rich Yedoma uplands on Kurungnakh Island (Schirrmeister et al., 2011b). Thermokarst results from the thawing of ice-rich permafrost or the melting of massive ice (e.g. thermal mean only) (Van Everdingen, 2005). This in-situ process of thawing permafrost is followed by surface subsidence, creating thermokarst ponds, lakes, basins and alases (Morgenstern et al., 2011; Schwamborn et al., 2002), which allow the accumulation of Holocene material in the depressions (Wetterich et al., 2008a). On the other hand thermo-erosion results from the coupled action of thermal thaw and mechanical erosion by flowing water, forming landforms such as thermo-erosional gullies (TEGs) and thermo-erosional valleys (TEVs) (Gagnon et al., 2024; Morgenstern et al., 2021b; Van Everdingen, 2005). So far different thermokarst stages have been studied on Kurungnakh and other Yedoma landscapes in the Lena River Delta based on a spatial approach, using remote sensing and geoinformation systems (Morgenstern et al., 2011). To focus on thermo-erosional landscape features, Morgenstern et al. (2021b) inventoried TEVs in ice-rich coastal lowlands adjacent to the Laptev Sea. They found that TEVs are a prominent, hitherto overlooked permafrost degradation landform. While these spatial inventories provide important insights into the distribution and surface morphology of TEVs, they offer limited information on the subsurface processes and stratigraphic records that govern their formation and long-term evolution.

TEVs have been recognised as a widespread and significant permafrost degradation landforms in ice-rich Arctic lowlands, yet their internal structure and developmental history remain largely unknown. In particular, it is unclear how sequences of erosion and subsequent sediment infilling are recorded within the stratigraphy (French & Shur, 2010; Schirrmeister et al., 2011b). While thermokarst activity starts around the Late Pleistocene to Holocene transition (Morgenstern et al., 2013b), the timing of valley initiation has not been specified so far. Furthermore, rates of activity, and mechanisms leading to partial or complete stabilisation during the Holocene are also poorly constrained, limiting the ability to distinguish transient erosional features from rather stable landforms (Strauss et al., 2017; Wetterich et al., 2008a). Understanding these aspects is critical for reconstructing the evolution of TEVs and linking surface geomorphology to subsurface processes. A broader knowledge on these coherences will allow interpretations on the resilience and sensitivity of Yedoma landscapes to ongoing permafrost degradation. Consequently, filling this knowledge gap requires detailed stratigraphic evidence with records of erosion, sediment redistribution, and post-depositional processes.

Previous studies on ice-rich permafrost landscapes document a range of processes relevant to the formation of Yedoma deposits, the degradation of ice-rich permafrost, and the accumulation of Holocene covers and infilling of depressions. Owing to the geological and geomorphological setting of Kurungnakh Island, this thesis considers these process groups as a conceptual basis for interpreting TEV development. Primary material input during Yedoma formation broadly attributes to the aeolian- and hydraulic sedimentary system (Schirrmeister et al., 2011b, 2020). In contrast, the degradation of Yedoma terrain and the formation of TEVs are governed by interacting erosional, depositional, and postdepositional processes. These include thermal erosion driven by meltwater streams, overland flow, and permafrost thaw (Gagnon et al., 2024; Morgenstern et al., 2013a, 2021b), while cryogenic weathering, cryoturbation, ice-wedge growth, and soil formation shape postdepositional modification (French & Shur, 2010; Opel et al., 2018; Schirrmeister et al., 2011b, 2020; Schwamborn et al., 2012; St-Jean et al., 2011). Downstream transport of mobilised sediment by hydraulic processes represents a key system output (Lafrenière & Lamoureux, 2019; Morgenstern, 2012). Together, this published process understanding provides the

framework for interpreting TEV stratigraphy in terms of erosional removal, sediment reworking and sorting, and postdepositional modification.

To obtain such stratigraphic evidence and address the limited understanding of the subsurface structure and temporal development of TEVs, this study is based on four permafrost cores recovered from three geomorphic positions within a TEV on Kurungnakh Island. Since, compared to TEGs (Gagnon et al., 2024), the initial phases of TEV development can not be assessed fully on a time-span of several years to decades this thesis analyses permafrost cores from a single point in time (September 2017) and links the datasets from various analysed parameters to processes shaping the valley development. By analysing cores from the Yedoma upland, valley slope, and valley floor, the study captures spatial contrasts in erosion, thaw and sediment accumulation that reflect the internal dynamics of valley development in ice-rich permafrost landscapes (French & Shur, 2010; Strauss et al., 2017). Focusing on a single valley allows the linkage of stratigraphic records to local geomorphic processes while minimizing regional climatic variability, thereby providing a detailed reconstruction of thermo-erosional processes that cannot be resolved through surface observations or remote sensing alone (Morgenstern et al., 2021b). This approach forms the conceptual basis for integrating stratigraphic, cryostratigraphic, sedimentological and hydrochemical evidence to investigate the development and current state of a TEV within Yedoma terrain.

The datasets obtained from the four permafrost cores allows the reconstruction of spatial gradients in erosion, thaw, and sediment deposition within the TEV. Specifically, an in depth analysis of grain-size distributions (GSDs) provides insights into sediment provenance, transport mechanisms, and depositional environments (Weltje & Prins, 2007), while robust endmember modelling analysis (rEMMA) facilitates the identification of distinct sedimentary components and their process-based signatures (Dietze & Dietze, 2019; Schirrmeister et al., 2020). Furthermore, the analysis reveals valuable insights, since endmember unmixing has rarely been applied to thermokarst, thermo-erosional and Holocene grain-size datasets (Ulrich et al., 2019). Here, I interpret core locations relative to existing stratigraphic frameworks from Kurungnakh, the Lena River Delta and Siberia (Khazin et al., 2019; Schirrmeister et al., 2003, 2011b, 2013; Ulrich et al., 2019; Wetterich et al., 2008a, 2020) as well as Arctic TEGs stratigraphy (Gagnon et al., 2024). Integrating these datasets with measurements of active layer thickness (ALT) and water table depth (WTD) measurements from the three study sites and topographic information from GNSS transects and the ArcticDEM Mosaic (Porter et al., 2023) provides a comprehensive basis to link observed stratigraphy with geomorphic processes. Building on this integrated approach, I address the following research questions (RQ1-2) and hypotheses (H1-4):

RQ1 Which processes dominate the development of the TEV and characterise the stratigraphy at the Yedoma upland (core C), the valley slope (cores A001 and A003) and the valley floor (core B)?

H1 I hypothesise that the stratigraphy at the Yedoma upland (core C) remains largely undisturbed, preserving Yedoma deposits.

H2 I hypothesise that predominantly erosional processes characterise the stratigraphy at the slope (cores A001 and A003), causing erosional gaps and little soil formation.

H3 I hypothesise that the stratigraphy at the valley floor (core B) reflects sequences of erosion and subsequent sediment infilling.

RQ2 Does the connection between processes and geomorphic sites in the TEV create a stable Holocene

landform?

H4 I hypothesise that the TEV represents a Holocene landform resilient to minor disturbances but sensitive to changing hydrological and permafrost conditions.

To address the research questions outlined above, Chapter 2 provides a detailed overview of the study area, including geological and geomorphological characteristics of Kurungnakh Island and the third terrace and environmental conditions. Chapter 3 presents the data collection strategy and laboratory analyses that underpin the study, including cryostratigraphic, sedimentological and hydrochemical measurements. The results of these analyses are presented in Chapter 4, forming the basis for interpretations in the discussion (Chapter 5). The discussion is structured to first address RQ1, starting with the interpretation of robust endmembers (rEMs) derived from the rEMMA, followed by three sections concerning hypotheses H1, H2 and H3 corresponding to the Yedoma upland, valley slope, and valley floor cores, respectively. The final section of the discussion evaluates RQ2 concerning the overall development and stability of the TEV, before concluding with a synthesis of findings and limitations.

2 Study Area

The study site of this thesis lies in the southern part of the Lena River Delta at the coast of the Laptev Sea, Northeast Siberia (Figure 1) (Schirrmeister et al., 2003; Wetterich et al., 2008a). The data collection focused on a TEV on Kurungnakh Island (72°36'N; 126°25'E). Kurungnakh Island provides a well-studied framework for interpreting TEV deposits within ice-rich Yedoma terrain. Kurungnakh is a remnant island of a foreland plain of the delta that became fragmented by several outlets of the lower Lena River. The island covers an area of 350 km² (Morgenstern et al., 2013b). The Olenyoksky Channel separates Kurungnakh from the mainland and shapes its southwestern coastline. Along this channel, coastal cliffs expose the island's stratigraphy (Figure 2) and provide key outcrops for stratigraphic investigations (Schirrmeister et al., 2003, 2011b; Wetterich et al., 2008a). The delta is located in a continuous permafrost zone, which strongly influences the geomorphology of the landscape. Chapter 9.1.1 provides further information on permafrost characteristics and its distribution across Arctic regions.

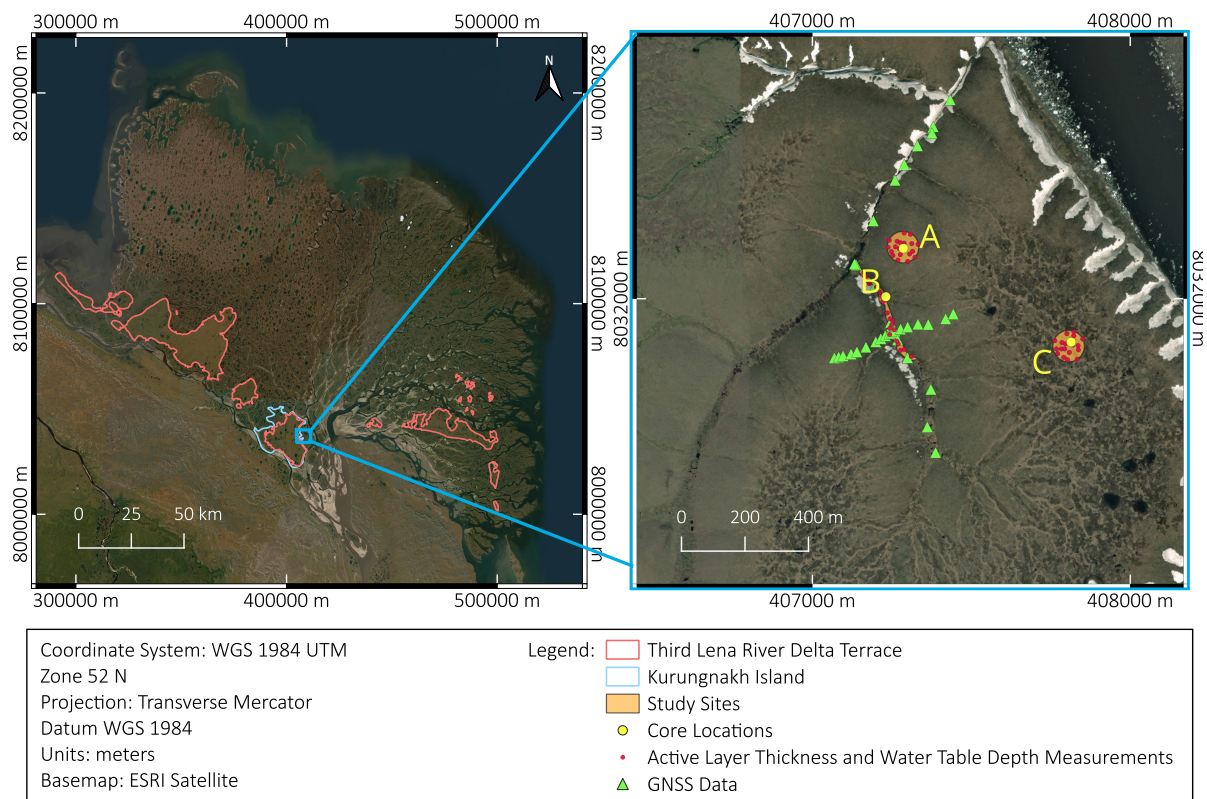


Figure 1: The left map depicts the Lena River Delta in Northeast Siberia and the area of the third Lena Delta Terrace (red) (according to Morgenstern et al. (2021b)) and Kurungnakh Island (blue). The map on the right shows the extent of the TEV on Kurungnakh Island (Esri, 2023). The locations of the retrieved permafrost cores A-C (dots) and study sites A-C (letters) are labelled yellow and the orange areas the study sites. Additionally, red dots mark the locations of distributed active-layer thickness and water table depth measurements. Green triangles represent Global Navigation Satellite System locations.

2.1 Geology of Kurungnakh Island

As described in Chapter 1, three Lena River Delta terraces shape the geological setting. While Kurungnakh Island belongs to the third, all three terraces remain relevant, since the first and second terrace depict potential sources of aeolian transportation and a deposition zone of suspended and dissolved

load exported out of the TEV. The first out of three Lena River Delta terraces is the youngest with a still active floodplain. It consists mainly of Holocene deposits, stretching across the central and eastern parts of the delta with elevations from 0 m to 10 m above sea level (a.s.l.). Sandy, fluvial Late Pleistocene deposits dominate the second river terrace, often referred to as the Arga Complex, which covers the northwestern part of the delta and reaches altitudes from 20 m to 30 m a.s.l.. The third terrace, previously mentioned as the now fragmented foreland plain, consists of the Late Pleistocene accumulation plain north of the Chekanovsky Ridge. This terrace extends across the southern and southwestern delta areas and reaches altitudes of 30 m to 55 m a.s.l. (Schirrmeister et al., 2011b). Regarding the stratigraphy of Kurungnakh Island, Yedoma deposits characterise most of the island's geology, belonging to the third Lena River Delta terrace (Figure 2). Due to the persistently cold and arid continental climate of northeastern Siberia, the Lena River Delta remained unglaciated throughout the Late Quaternary, forming part of the ice-free Beringian landscape. These periglacial environmental conditions promoted the syngenetic accumulation of ice-rich Yedoma deposits that now compose the Late Pleistocene strata of the delta and its elevated third river terrace (Schirrmeister et al., 2003, 2013; Wetterich et al., 2008a). The separate Chapter 9.1.2 further elaborates the specific characteristics and theoretical background of Yedoma deposits.

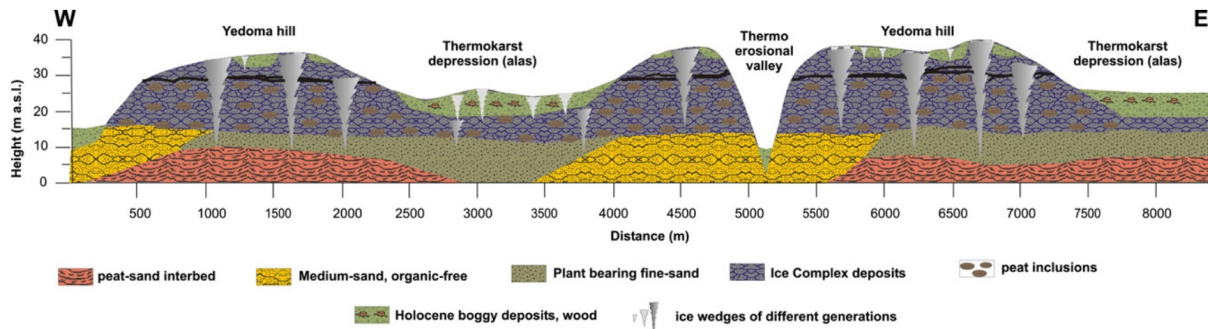


Figure 2: Cross section of Kurungnakh Island based on exposures along the Olenyokskaya Channel (Figure in Schirrmeister et al. (2011b)).

On Kurungnakh Island, within the third Lena River Delta terrace, two main sediment formations are of particular significance: sand deposits underlie ice-rich peaty and silty Yedoma deposits (Wetterich et al., 2008a). Holocene deposits form the upper part of the third Lena River Delta terrace. They are widely distributed and fill small- and large-scale thermokarst depressions (alases) (Wetterich et al., 2008a) as well as other depressions, such as valley bottoms (Figure 2). These sequences of sandy deposits, Ice Complex deposits, and interruptions by thermokarst depressions and valleys are exposed along the Olenyoksky Channel.

Wetterich et al. (2008a) classify the geologic sequences on Kurungnakh into five units (units I-V), defining the corresponding altitude, age, stratigraphy, and dominant sediments for each unit (Table 1). According to the analysis on Kurungnakh by Schirrmeister et al. (2003) and Wetterich et al. (2008a), the average boundary between the Ice Complex deposits and underlying fluvial sands lies between 15 m to 20 m a.r.l.. The elevation of the analysed cores in this thesis corresponds with units III, IVa, IVb, and V, while Chapter 5 evaluates whether the TEV stratigraphy matches these units. Nevertheless, the cores were collected within a TEV, where Holocene deposits cover the Yedoma upland and accumulate in the valley floor. Since the stratigraphy of Kurungnakh Island dips gently from northeast to southwest, the elevation of the units differs slightly between sites and is used as an orientation.

Table 1: Depth, age and important characteristics of five deposition units on Kurungnakh Island (Table in Wetterich et al. (2008a)).

Unit	Altitude (m a.r.l.)	AMS-ages (kyr BP)	Stratigraphy	Sediments
V	33.5–37	6–3	Middle and Late Holocene	Grey peaty silt with peat lenses and the uppermost modern active layer
IVb	32–33.5	About 8	Early Holocene	Grey sandy silt with rare plant detritus
IVa	29.5–32	About 17	Gap Late Weichselian Stadial (Sartan) Gap	Grey sandy silt with rare plant detritus
III	17–29.5	42–32	Middle Weichselian Interstadial (Kargin)	Grey peaty silt as continuous Ice Complex deposits with peat lenses and layers, including cryoturbated palaeosol horizons
II	12–17	>50	Early Weichselian Stadial (Zyryan)	Yellow medium-grained homogeneous finely laminated pure sand
I	Up to 12	>50	Early Weichselian Stadial (Zyryan)	Interbedded yellow medium-grained sand and grey silty sand with plant detritus, roots and single silt layers

According to the study by Wetterich et al. (2008a), the ground ice content on Kurungnakh varies with the corresponding five geological units, where overall a minimum of 23 wt% and a maximum of 150 wt% was measured. In coarser sediments, such as sand deposits the ice content is generally lower (about 25 wt%) (Wetterich et al., 2008a). A high ice content in soils does not only account for Kurungnakh Island, but most frozen soils have an ice content of 30 vol% to 60 vol% in near-surface layers (French & Shur, 2010). With freezing, the volume of water expands by about 9 vol% (French & Shur, 2010), causing freezing soils to expand and highlighting a seasonal relevance. Furthermore, thawed permafrost can only retain porewater (former pore ice), but not the meltwater originating from thawed excess ice (French & Shur, 2010). This results in the subsidence of thawing permafrost soils and makes soil ice content highly relevant in the context of climate warming (Kokelj & Jorgenson, 2013; Murton, 2009; Murton & French, 1994).

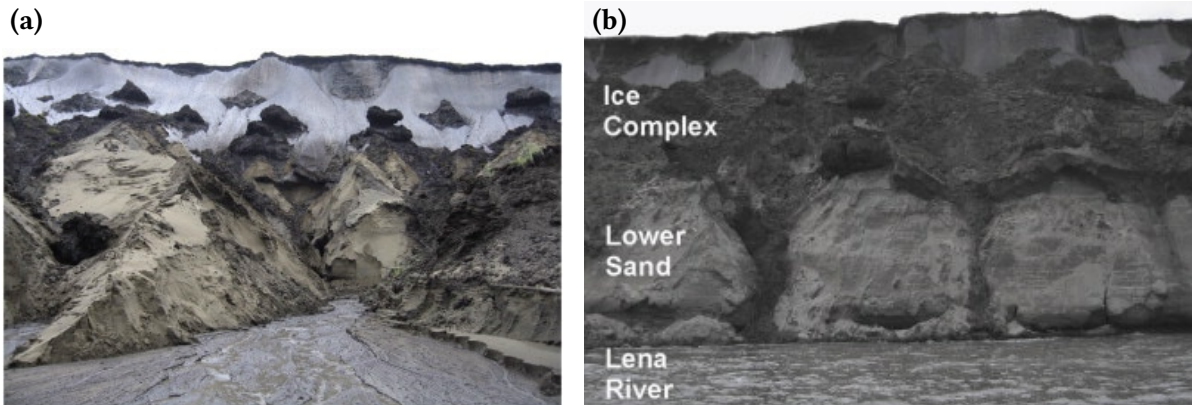


Figure 3: Coastal exposures of Kurungnakh Island along the Olenyoksky Channel: (a) bluff exposure on the eastern coast showing Holocene deposits overlying Ice Complex deposits and underlying sands at the modern Lena River shore (Image: M. Ulrich, Summer 2008 in Morgenstern et al. (2013b)); (b) coastal exposure illustrating the stratigraphy along the Olenyoksky Channel (Image: n.d., Summer 2000 in Wetterich et al. (2008a)).

2.2 Geomorphology of the studied thermo-erosional valley

The detailed geomorphology of the studied TEV has not been described in previous publications. Site characterisation is therefore based on remote sensing products and Global Navigation Satellite System (GNSS) measurements. These data provide the basis for classifying the valley type following the distinction proposed by Morgenstern (2012) (Figure 4) and for interpreting spatial patterns within the

study site (Chapters 3.2 and 4.2).

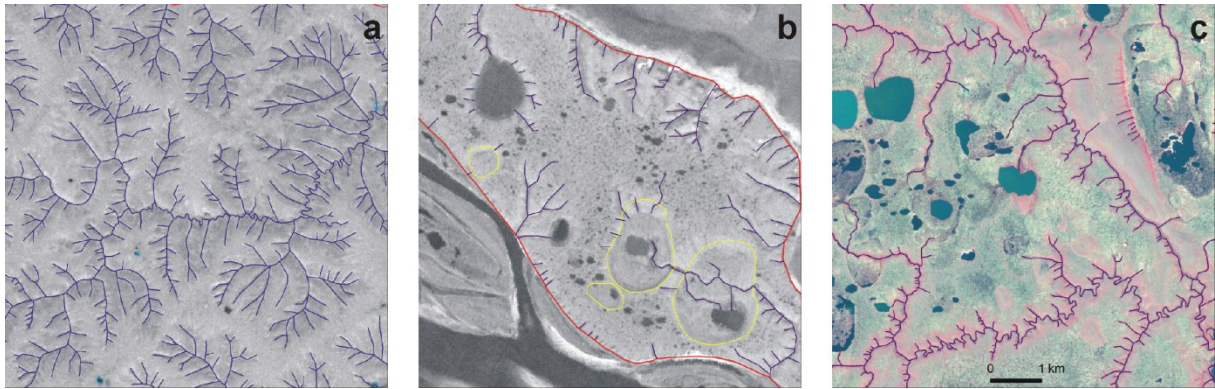


Figure 4: Different Arctic valley types: (a) extensive dendritic valleys; (b) short parallel and radial valleys and drainage pathways; (c) extensive longitudinal valleys with sharp meanders and short contributing valleys (Figure in Morgenstern (2012)).

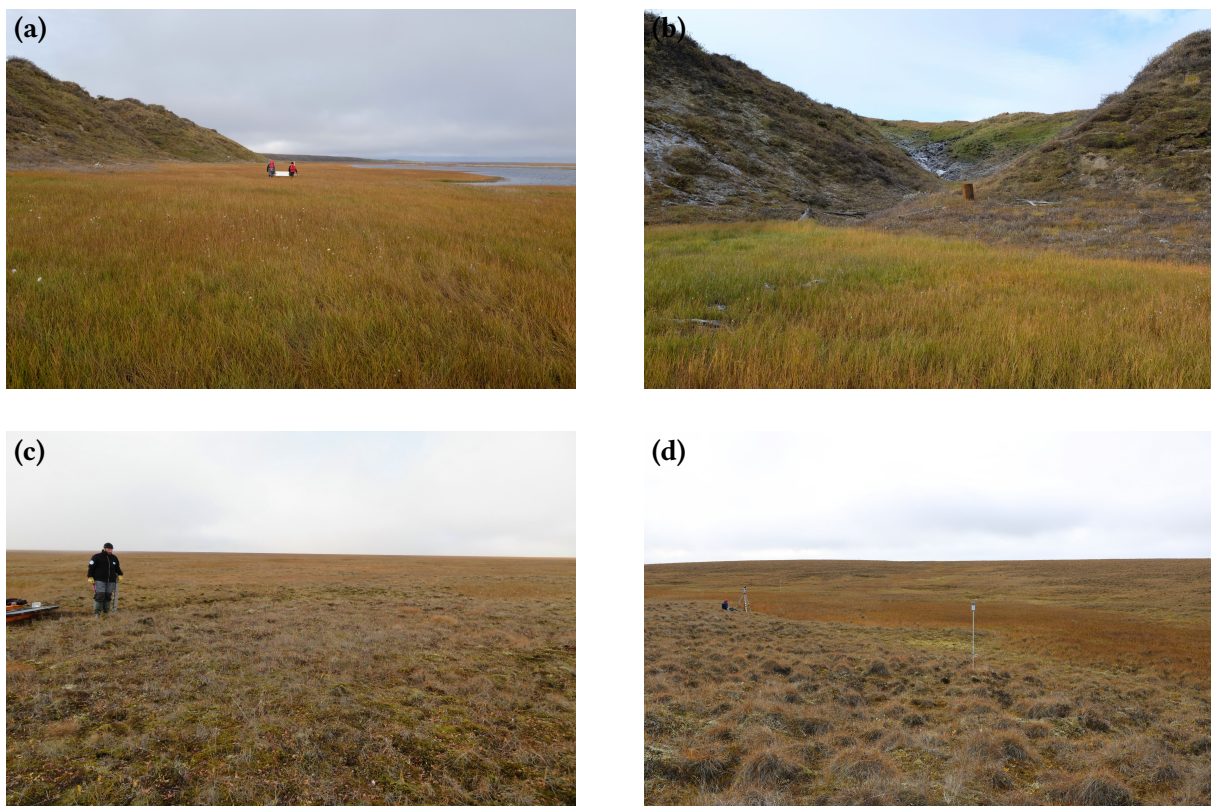
Data from a meteorological station on Samoylov Island (72°N, 126°E) serve as a reference for present meteorological conditions in the central Lena River Delta (Boike et al., 2013, 2019). From 1998 to 2017, the mean annual air temperature is -12.3°C . July is the warmest month and February the coldest, with mean monthly temperatures of 9.5°C and -32.7°C , respectively (Boike et al., 2019). Minimum winter air temperatures fall below -45°C , and summer air temperatures can exceed 25°C (Boike et al., 2019). Most rainfall occurs between mid-May and the end of September (Boike et al., 2013). Mean annual precipitation amounts to about 125 mm for 1999 to 2011 (Boike et al., 2013) and 169 mm for 1998 to 2017 (Boike et al., 2019). Average winter snow cover is 0.3 m for 2002 to 2017, although snow depth shows high spatial variability due to strong winds.

Holocene temperature trends in the Arctic and the Lena River Delta remain uncertain, including the timing and magnitude of potential warming or cooling phases and the existence of a millennial-scale thermal maximum during the Mid Holocene (Kaufman & Broadman, 2023). However, at the transition from the Late Pleistocene to the Holocene (11 700 yr BP), the Lena River Delta region experienced a major climatic reorganization with short-term fluctuations, shifting from cold, arid steppe-tundra conditions characteristic of the Late Glacial to a warmer and more humid Early Holocene climate (Andreev et al., 2011; Bischoff et al., 2013; Schwamborn et al., 2002). Based on stable water isotope analyses of ice wedges from the Lena River Delta, Meyer et al. (2015) identify a winter warming trend during the Mid to Late Holocene in the Siberian Arctic. In contrast, other proxy reconstructions suggest a warming trend over this interval but are often biased towards summer conditions (Vinther et al., 2009; Wanner et al., 2008). Recent decades show rapid warming that likely overrides Holocene-scale trends (Consortium, 2013; Kaufmann, 2018; Meyer et al., 2015). Ground surface temperatures increased by 5° to 9° between 1855 to 2012 (Groenke et al., 2024).

Climate warming affects permafrost hydrology (Lafrenière & Lamoureux, 2019) by controlling flow-paths within the active layer and supra-permafrost water flow (Woo, 2012). Hydrological processes in permafrost terrain are closely linked to seasonal conditions, precipitation availability, and snow redistribution (Young et al., 2010). Permafrost degradation can alter surface drainage by causing ponding and may contribute to rapid landscape disturbances such as active-layer detachments, retrogressive thaw slumps, and thermo-erosional gullying driven by the erosion of ice wedges (Kokelj et al., 2009;

Kokelj & Jorgenson, 2013; Lafrenière & Lamoureux, 2019). TEVs can also modify downstream hydrology (Morgenstern et al., 2021b). The dry, continental and non-glaciated environment leads to a snowmelt-dominated system on Kurungnakh Island. Time-lapse imagery (Appendix 9.2.3) illustrates seasonal snow distribution within the TEV. In snowmelt-dominated systems, the snow water equivalent is a key control on suspended sediment transfer (Cockburn & Lamoureux, 2008). Lower air temperatures at the onset of snowmelt can lead to higher-energy sediment delivery regimes, while melt energy, snow distribution, and sediment availability further influence sediment transport (Cockburn & Lamoureux, 2008; Macumber et al., 2018). To get a concrete idea about hydrology and thaw depth in the studied TEV, the following chapters include active layer thickness (ALT) and water table depth (WTD) measurements.

Seasonal hydrology in the Lena River Delta is also influenced by the annual spring ice-breakup flood, a common phenomenon in Arctic rivers that remain frozen throughout winter (Ma & Fukushima, 2002). Flood heights measured at the Khabarova gauging station (200 km at similar latitudes as Kurungnakh Island) did not exceed 12 m during the period 2013 to 2019 (Pertiwi et al., 2021; The Arctic Great Rivers Observatory ArcticGRO, 2024). Further downstream, water levels flatten across the delta, suggesting that flooding does not reach the studied TEV branch on Kurungnakh Island. Field observations from the 2017 expedition likewise show no evidence of flood impact within the valley, while driftwood deposits occur along the coastline and adjacent steep valleys (Figure 5b). In addition to seasonal flooding, recent studies show that the Lena catchment has experienced catastrophic outburst floods during the Pleistocene, which provide important context for stratigraphic and hydrological disturbances in northeast Siberia (Arzhannikova et al., 2025; Margold et al., 2011, 2018; Wetterich et al., 2020).



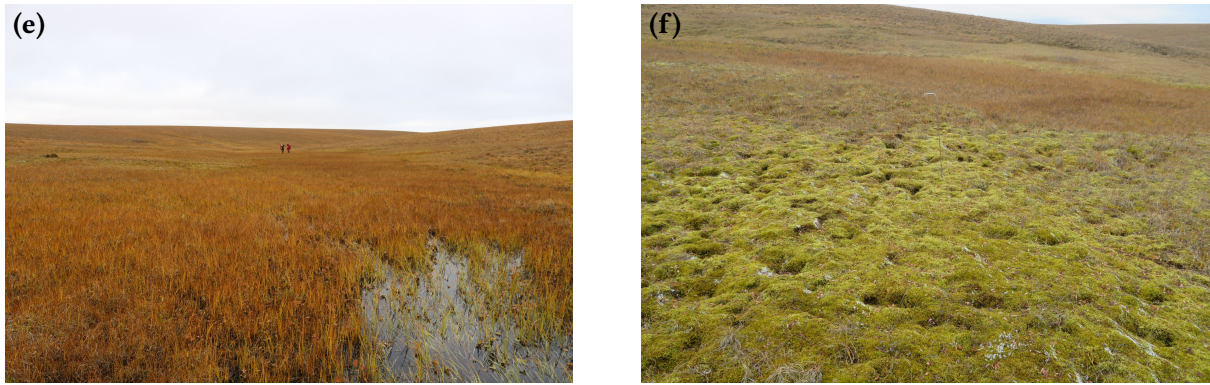


Figure 5: Field images from Kurungnakh Island: (a) northeast coast with river channel to the right and two expedition participants carrying freight; (b) relatively short and steep thermo-erosional valley with a steel barrel for scale; (c) study site C on the Yedoma upland; (d) study site A at the west slope; (e) valley floor with standing water in the thermo-erosional valley and two people for scale; (f) study site B at the valley floor (Images: Jan Nitzbon; September 2017).

Surface cover represents another important factor influencing TEV processes. Raschke and Savelieva (2017) describe the dominant surface cover on Kurungnakh Island as subarctic swamp hummock-hollow tundra complexes. A recent land cover map of the Lena River Delta identifies polygonal tundra with up to 10% surface water on the Yedoma upland (Lisovski et al., 2025), but observations during the 2017 expedition show that the ice-wedge polygonal surface topography on the Yedoma upland of the analysed TEV is only weakly developed. The upland surface depicts locally wet depressions (Figure 1 & 11), but no evidence of young Holocene ice-wedges. The drained slopes mainly correspond to dry tundra tussock type (Figure 5d), while the valley bottom is characterised by moist to wet sedge communities (Figure 5e) (Lisovski et al., 2025). Furthermore, Figure 5e and 5f show that the surface cover at the valley floor is locally heterogeneous. Based on pollen records, Wetterich et al. (2008a) describe a dry shrub tundra during the early Holocene, followed by middle to Late Holocene conditions characterised by wet tundra with willow shrubs.

3 Methods

This thesis combines field, remote sensing, and laboratory methods to reconstruct near-surface stratigraphy and development processes within a TEV on Kurungnakh Island. Figure 6 summarises the complete workflow, from site characterisation with remote sensing, GNSS, and distributed field measurements, to permafrost core recovery, subsampling, and laboratory analyses. The following sections describe these methods in the same order, starting with geomorphological and near-surface site characterisation and followed by the core-based laboratory procedures and analytical approaches.

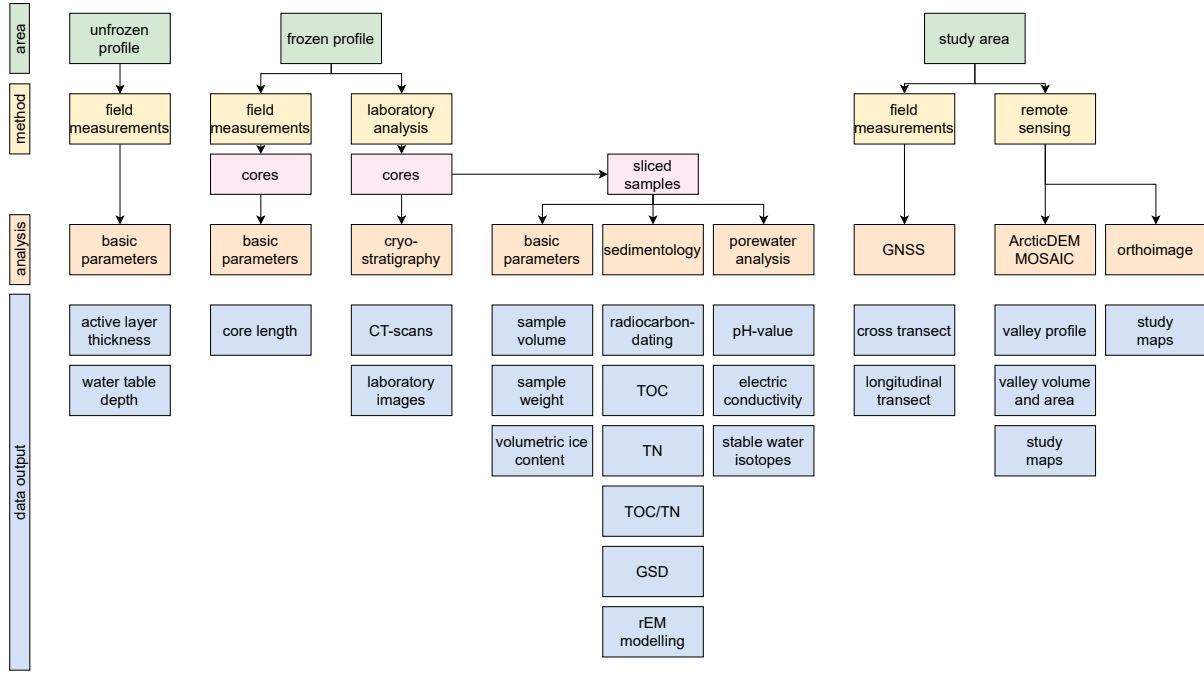


Figure 6: Flowchart depicting the workflow, all relevant datasets and parameters for this thesis.

3.1 Valley mapping

Remote sensing and elevation data provided the basis for the study area maps and supported the geomorphological characterisation of the TEV as well as the spatial interpretation of the core locations. Spatial data processing and map design were carried out in QGIS Version 3.44 using EPSG:3413 (WGS 84 / NSIDC Sea Ice Polar Stereographic North).

Remote sensing data from three main products provided the basis for maps and visualisations: (i) the ESRI World Imagery basemap (Esri, 2023), (ii) a GeoEye Mosaic acquired on 05.08.2010, and (iii) the ArcticDEM Mosaic (Porter et al., 2023). ESRI World Imagery served as the basemap for the regional overview (Figure 1) with a maximum spatial resolution of 2 m. The GeoEye Mosaic also had a spatial resolution of 2 m and included four spectral bands (red, green, blue, and near-infrared). A false-colour composite (red = band 4, green = band 3, blue = band 2) highlighted surface features of the TEV, including streams, depressions, and ponds (Figure 11). The ArcticDEM Mosaic provided elevation and terrain information at 2 m spatial resolution and represents a processed and smoothed product compared to ArcticDEM strip data (Porter et al., 2023).

Core locations and distributed measurement points were added as point datasets. The TEV catchment polygon was delineated by expedition members in 2017 and is shown in Figure 12. Catchment metrics were derived by clipping the ArcticDEM to this polygon and calculating elevation statistics using the *Zonal statistics* and *Zonal minimum/maximum* tools. A depth raster was then calculated by subtracting DEM elevations from the maximum catchment elevation. This raster layer formed the basis for calculating catchment area and volume using the QGIS *Raster surface volume* tool.

In addition, two GNSS transects collected during the 2017 expedition provided information on valley morphology, including one longitudinal profile along the valley floor and one cross section. Since the GNSS cross section did not reach the valley floor depth of core B and did not extend upslope to the elevation of core C on the Yedoma upland, I derived an supplementary elevation profile from the ArcticDEM to capture the full elevation range of all cores (Figure 10).

3.2 Distributed measurements

To characterise near-surface permafrost and hydrological conditions within the studied TEV, I included ALT and WTD measurements collected during the core-drilling expedition in September 2017. Around each core location (A–C), twenty distributed field measurements were carried out (Figure 1). At the valley slope and Yedoma upland, these study sites depicted circular areas with a diameter of 50 m, while at the valley bottom, the study site extended along the valley floor and included measurements in small streams to capture local hydrological variability. For both parameters, positive values indicated greater depths below the ground surface, whereas WTD included positive values indicating a water table above the surface due to ponding water or streams. Finally, I summarised the results of each study site as average values and their standard deviation (SD).

3.3 Permafrost cores

Members of the Permafrost Research Section at the Alfred Wegener Institute Helmholtz Centre for Polar and Marine Research (AWI) drilled the permafrost cores during the Lena River Delta Expedition in September 2017. Core lengths varied, whereas the diameter was approximately 7.5 cm. As shown in Figure 11, study site C is located on the Yedoma upland, site A on the valley slope, and site B at the valley floor. An additional core (core D) originated from the source region of the TEV, representing the transition zone between the Yedoma upland and the slope. Due to limited availability of laboratory data, I excluded core D from the main interpretation and present it in the Appendix (Chapter 9.2.1).

The core sampling strategy aimed to obtain one core per geomorphological position. To account for potential small-scale stratigraphic heterogeneity at the valley slope, expedition members collected two cores at site A (cores A001 and A003). Figure 7 shows field photos from the near surface parts of all retrieved cores and for core C and A003 from the entire core.

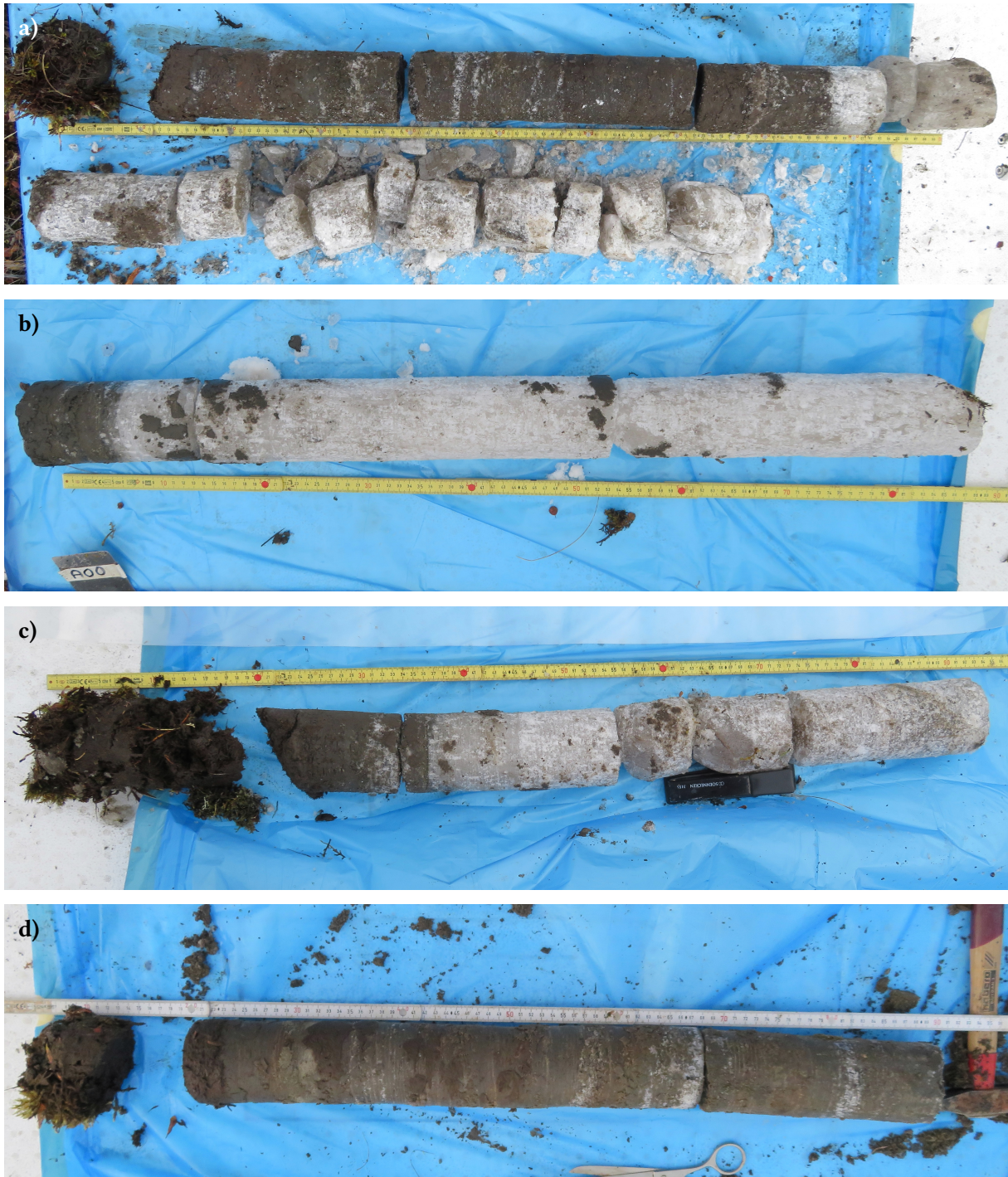


Figure 7: Field images of permafrost cores from the studied thermo-erosional valley on Kurungnakh Island: (a) complete core C from the Yedoma upland; (b) upper section of core A001 from the valley slope; (c) complete core A003 from the valley slope; (d) upper section of core B from the valley floor (Image: Jan Nitzbon, September 2017).

To support the interpretation of parameter changes with depth and to structure the results, I subdivided the cores into core sections representing active-layer, sediment, ice-wedge, and ice-lens intervals. I identified section boundaries primarily from CT scans and images of the halved cores, while cryostratigraphic classification and laboratory results provided additional support. Sediment section boundaries therefore represented an interpretive subdivision rather than sharp stratigraphic limits.

To prepare cores and samples for the laboratory analysis, I split cores A001, A003, and B longitudinally

into two halves (Figure 8). While I archived one half of the cores, I subsampled the second halves into a total of 83 slices (including both slope cores and core B). Core C showed circular slices since they did not provide a one half for the archive back in 2019 (Nitzbon et al., 2022). Overall, slice thickness varied depending on core characteristics such as mineral, ice, and organic content, as well as fractures and breaks. The combined evaluation of field notes, photographs of the halved cores, and CT scans defined the sampling positions and resolution. In particular, CT scans provided key guidance by revealing the internal distribution of ice and sediment prior to subsampling.

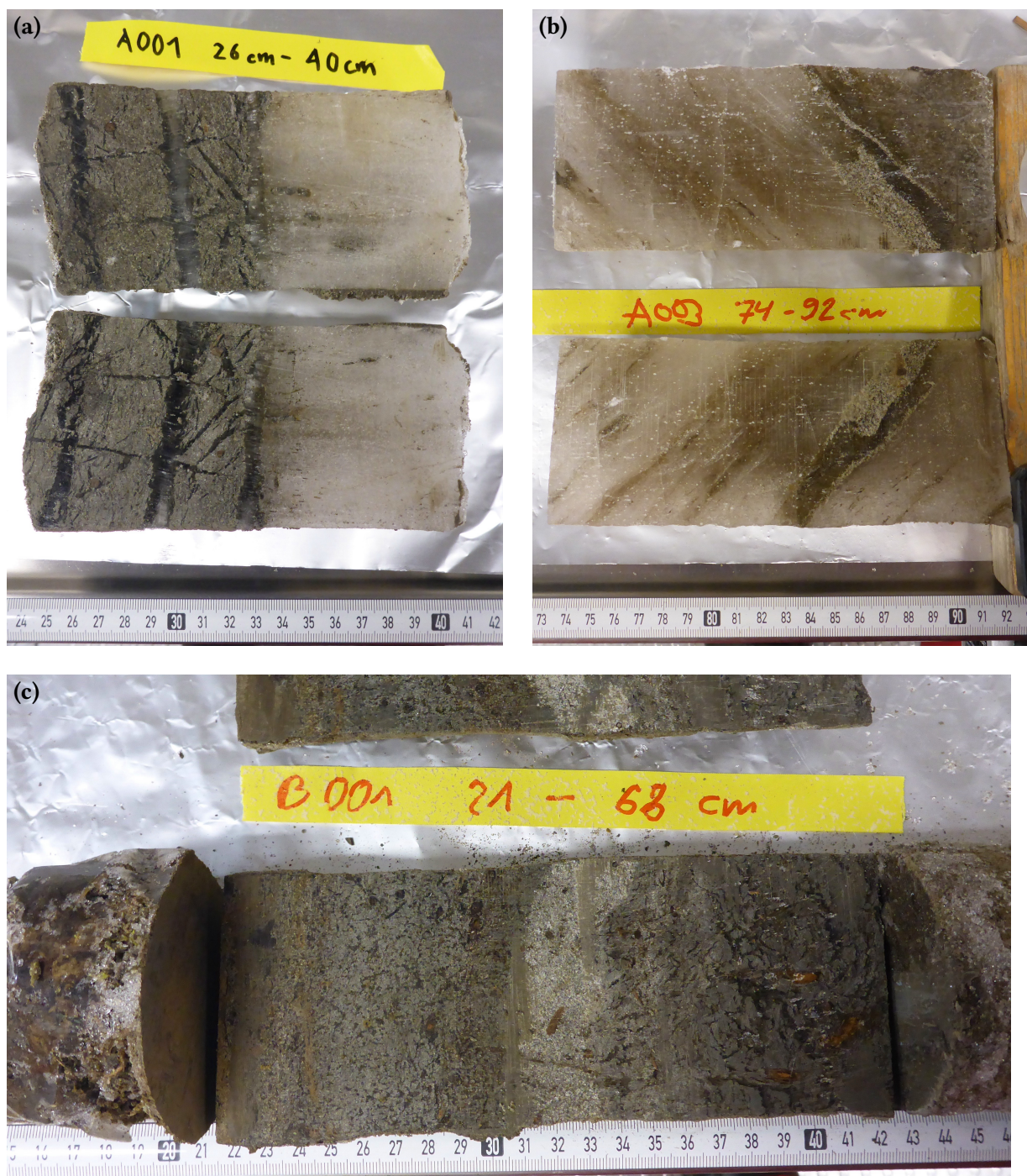


Figure 8: Images of halved valley slope and valley floor cores: (a) valley slope core A001 shows the sediment section and the top of the wedge ice; (b) valley slope core A003 shows a section of wedge ice including the sediment lens; (c) valley floor core B shows an extent of the first and second sediment section (Images: Jael Mencke, January 2025).

3.3.1 CT scans

As previously described, CT scans from each permafrost core highlighted the cryostratigraphy and enabled the stratigraphic visualisation. Core C was scanned at the Helmholtz Centre for Environmental Research (Helmholtz-Zentrum für Umweltforschung, UFZ) in 2019 by Dr. habil. John Maximilian Köhne, as described by Nitzbon et al. (2022). In 2024, the same laboratory performed CT scans on cores A, B, and D. The UFZ provided TIFF files of two-dimensional longitudinal cross sections, which were processed in ImageJ (Schneider et al., 2012). The processing resulted in grayscale cross-section images, saved as TIFF files.

To normalise grayscale values between cores and improve the comparability of ice, mineral, and organic components, I reopened each image in ImageJ and selected the largest possible connected area depicting massive ice. I then saved the selection as a *.roi* file in the same directory as the corresponding TIFF image. Using an adjusted macro script originally developed by Dr. Steffen Schlüter (UFZ), I normalised the grayscale values from the overview scans by shifting the histogram peak of the selected ice area to a value of 120. Figures 15–18 show the according to their depth position assembled scans per permafrost core.

3.3.2 Basic parameters - Sample volume and volumetric ice content

Some basic parameters formed the foundation for analytical results from the laboratory analysis. One such basic parameter was the sample volume enabling the determination of the volumetric ice content (VIC). For core C, AWI members determined sample volumes exclusively from manual measurements. I therefore used geometric volumes for all volume-based calculations of core C. On the other hand, I determined sample volumes for cores A and B using three-dimensional scanning. This approach replaced the commonly used water displacement method (Strauss et al., 2013) to improve measurement accuracy by scanning the samples without its packaging. I performed all scans in a climate chamber, which minimised the risk of thawing during volume determination.

To validate the three-dimensional scans, I measured each sample manually immediately after cutting it in the climate chamber. The manual volume calculation used the height of the half-cylinder sample at three positions (left, centre, right), the radius perpendicular to the halving cut, and the diameter along the cut. I approximated the sample geometry as a circular segment and calculated the geometric volume using:

$$V_{\text{sample}} = \left(\frac{h}{2}\right) \pi \left(\frac{d}{2}\right)^2 100$$

V_{sample} = geometric volume (cm³)

h = mean sample thickness (mm)

d = chord diameter of the circular segment (mm)

I compared scanned and manual volumes using the absolute difference between the scanned and geometric results. When deviations exceeded 1 cm³, I assessed scan quality by comparing the three-

dimensional model to two-dimensional images of the frozen sample. Cryostructures and ice lenses locally produced bulges or cavities, which caused over- or underestimation in some scans. For samples with a reliable scan representation despite deviations, I used the scanned volume in further calculations. For samples with an imprecise scan model, I used the geometric volume instead.

To calculate VIC, I weighed wet samples before and after freeze-drying. I combined wet and dry masses with the sample volume and an ice density of 0.9 g cm^3 to calculate VIC for each sample.

3.3.3 Cryostratigraphy – Classification and ice-wedge structures

In addition to quantified ice contents, the occurrence and arrangement of ground-ice types provide key information on permafrost genesis and post-depositional processes. Therefore, I used the cryostratigraphy to describe ground-ice conditions in the perennially frozen sediment and to support interpretations of permafrost conditions.

Cryostructures and ground-ice types adapted from the North American cryostructure classification compiled by French and Shur (2010) formed the classification basis in this thesis. Additional categories were included following Gagnon et al. (2024) to enable comparison with TEG stratigraphy. Cryostructure classes not present in the analysed cores (e.g. *crustal*) were excluded. Figure 9 summarises the resulting cryostratigraphy classes, which I consistently applied across all cores. For visualisation of cryostratigraphy plots (Figures 15–18 and 26), I converted the designed symbols from the classification (Figure 9) into a simplified colour scheme.

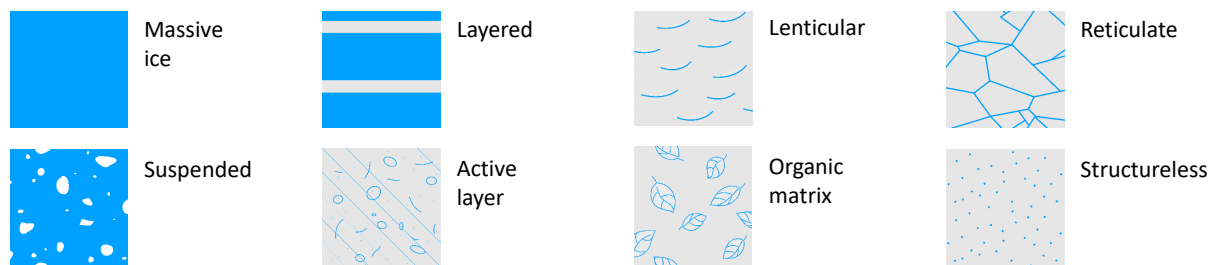


Figure 9: Relevant cryostratigraphy classes for the thermo-erosional valley data from Kurunghakh Island.

Cryostructure types supported the interpretation of permafrost formation, as layered cryostructures typically indicate syngenetic permafrost, whereas reticulate and irregular cryostructures indicate epigenetic freezing (French & Shur, 2010). However, the cryostructure classification did not capture structures within massive ice bodies. Ice wedges were identified by vertically oriented foliations of the ice body (Meyer et al., 2015). Characteristic bubble morphologies, clarity and layering delivered further information about the age and development of the ice wedges. While the ice wedges reflected winter temperatures, they helped distinguish between Pleistocene and Holocene ice (Wetterich et al., 2008a). Therefore, I described massive ice characteristics separately in the results section and interpreted them together with the available stable water isotope data.

3.3.4 Hydrochemistry – pH, EC and stable water isotopes

For the analysis of hydrochemical parameters, I extracted porewater from the thawed samples using Rhizon samplers. Depending on the available porewater volume, I divided each porewater sample into subsamples for pH, electrical conductivity (EC), and stable water isotope measurements. Due to limited porewater availability in some samples, I prioritised the allocation of subsamples accordingly.

I measured pH-values using a Lab 870 pH meter and calibrated the device with standard buffer solutions of pH 4.01, 7.00, and 10.01. I measured EC with an Orion Versa Star Pro device and calibrated it using a 1413 $\mu\text{S}/\text{cm}$ standard. All samples contained sufficient porewater volumes for pH and EC measurements.

For study sites A and B, 76 samples contained sufficient porewater volumes for stable water isotope measurements of hydrogen ($\delta^2\text{H}$ or δD) and oxygen ($\delta^{18}\text{O}$). Limited availability of ice-wedge samples from the valley slope prevented hydrochemical analysis of these ice-wedge sections. Therefore, measurements of wedge ice were only available from the Yedoma upland. The AWI in Potsdam, Germany, measured all samples using Finnigan MAT Delta-S mass spectrometers. They reported the isotope ratios per mil difference relative to VSMOW (‰, Vienna Standard Mean Ocean Water) in ‰, with internal 1σ uncertainties better than 0.8‰ for $\delta^{18}\text{O}$ and 0.1‰ for $\delta^2\text{H}$.

Since stable hydrogen and oxygen isotopes and deuterium-excess (d-excess) provide insights into palaeotemperatures and moisture sources (Meyer et al., 2000), I plotted $\delta^2\text{H}$ against $\delta^{18}\text{O}$ and compared the dataset to the global meteoric water line (GMWL). Furthermore, I calculated regression slopes ($S_{\text{D-18O}}$) for isotopic trends within individual core sections (Figures 19a and 19b) to test whether trends in the isotope data (Figure 18) indicated equilibrium freezing effects. According to Lacelle (2011), slope values below 7.3 can suggest freezing under equilibrium conditions, while earlier studies interpreted slope values below 6 as an indicator of ice formed by freezing of liquid water. To further evaluate this relationship, I plotted d-excess against $\delta^2\text{H}$, where no relation reflects meteoric precipitation and a negative relation indicates freezing-related fractionation (Lacelle, 2011). Additional theoretical background on stable water isotopes is provided in Appendix Chapter 9.1.3.

3.3.5 Biogeochemistry – Total organic carbon and total nitrogen

To analyse biogeochemical parameters, I subsampled the freeze-dried sediment and milled the required amount of material for elemental analysis of total organic carbon (TOC) and total nitrogen (TN). I milled the samples for at least 200 s using a Retsch MM200 Mixer Mill. In most cases, sufficient material remained for subsequent grain-size and magnetic susceptibility measurements.

For carbon and nitrogen measurements, I prepared 50 mg per sample in two replicates. Where material availability was limited, I reduced the sample mass to 20 mg. I measured TOC using a soli TOC cube, which differentiated total carbon into TOC400, residual organic carbon (ROC), and TIC900 (total inorganic carbon) based on temperature-dependent combustion in an oxygen-containing gas flow. I calibrated the instrument following EN15936 and DIN19539. I measured TN using a rapid MAX N exceed and calibrated the device with blank samples and standards iva 2150 and iva 2156. I calculated TOC/TN ratios by dividing TOC by TN.

3.3.6 Sedimentology – Grain-size distributions and endmember modelling

Grain-size analysis is a long-established and widely used approach to reconstruct past Earth surface dynamics (Dietze & Dietze, 2019; Folk & Ward, 1957; Krumbein, 1932). Transport energy controls the GSD of deposited sediments, creating a distinct distribution upon deposition. Thus, GSDs provide essential information about source-to-sink relations, transport modes, sorting, and depositional processes (Dietze et al., 2014; Folk & Ward, 1957; Sun et al., 2002; Visher, 1969; Weltje & Prins, 2007). As a first-order description, the GSD characterises the average grain size and sorting of the mineral fraction.

To prepare samples for grain-size analysis, I transferred the freeze-dried sediment into glass beakers and documented the material by photographing each beaker to capture the amount and visual composition. From the 83 samples collected at sites A and B, I selected 63 samples with sufficient sediment content and material availability for grain-size analysis. To remove organic matter, I added 100 mL of 30 % hydrogen peroxide (H_2O_2) to each sample and placed the beakers on a shaker with a heating plate at 60 °C. Every day, except for weekends, I replenished the oxidant by adding 10 mL of 30 % H_2O_2 per sample and cleaned the beaker walls using 3 % H_2O_2 . To prevent drying, I added distilled water when necessary. I maintained a pH range of 6.5–8.0 by adding ammonia solution (NH_3) to increase pH or acetic acid ($\text{C}_2\text{H}_4\text{O}_2$) to decrease pH. After approximately three to four weeks, organic matter was fully dissolved, and no visible reaction (bubbling) occurred when adding oxidant.

Next, I separated the mineral fraction by centrifugation in several rounds using a Cryofuge 8500i and a Heraeus Megafuge 40. After reducing the supernatant to a minimum, I freeze-dried the mineral residue to remove remaining moisture. Depending on the expected grain-size range, I weighed between 0.6 g and 3.9 g of freeze-dried mineral material into 250 mL beakers, added sodium hexametaphosphate as dispersant, and mixed the suspension for at least 24 hours. I then sieved the samples through 2 mm and 1 mm sieves and split the fraction < 1 mm into eight subsamples using a Rotary Sample Divider. After drying and weighing the subsamples, I analysed them using a Mastersizer 3000+ (Malvern) laser diffraction particle size analyser. While the sieved proportion determined the upper limit of the measured grain-size range, the Mastersizer 3000+ has a lower limit at 0.1 μm (Panalytical, 2025). Each subsample was measured three times. To obtain homogeneous results, I analysed a minimum of three and a maximum of eight subsamples per sample, resulting in 9–24 GSDs. The average of these measurements formed one representative GSD per sample.

I processed the grain-size data using GRADISTAT 8.0 (Blott & Pye, 2010). The software provided the relative proportions of standard grain-size classes, including gravel (> 2 mm), sand ($> 63 \mu\text{m}$ to ≤ 2 mm), silt ($> 2 \mu\text{m}$ to $\leq 63 \mu\text{m}$), and clay ($\leq 2 \mu\text{m}$) (Blott & Pye, 2001). I further applied the GRADISTAT naming convention (Blott & Pye, 2001) to classify each sample (Figure 4). The relevant categories for this dataset included very coarse silt, coarse silt, medium silt, fine silt, medium sand, fine sand, and very fine sand. Although GRADISTAT also calculated arithmetic mean grain size and sorting parameters, I did not include these metrics in this thesis because they provide limited interpretive value for and non-Gaussian sediments.

To better resolve mixed grain-size populations, I applied an endmember modelling analysis (EMMA). EMMA is a multivariate statistical unmixing approach that decomposes compositional grain-size datasets into a limited number of independent components and their relative contributions to each sample (Weltje, 1997; Weltje & Prins, 2007). There are two main modelling approaches: conceptual, quali-

tative inverse modelling and numerical inverse modelling. While conceptual inverse modelling was established early in the history of sedimentary geology, numerical inverse modelling is much younger and still developing (Dietze & Dietze, 2019; Weltje, 1997; Weltje & Prins, 2007).

Endmember modelling was performed using the open-source EMMAgeo package implemented in R (Dietze & Dietze, 2019). EMMAgeo extends the classical EMMA framework by combining eigenspace rotation and non-negative matrix factorisation (Dietze & Dietze, 2019; Weltje, 1997). For datasets with unknown mixing conditions, Dietze and Dietze (2019) propose the robust EMMA (rEMMA) workflow, which includes a compact and an extended protocol. Here, I selected the compact protocol since it provides an automated and reproducible approach for extracting robust endmembers (rEMs). Furthermore, the compact protocol suited the process-oriented paleoenvironmental interpretation well.

The input dataset combined all available GSDs from the Yedoma upland (core C), valley slope (cores A001 and A003), and valley floor (core B). EMMAgeo required the removal of all grain-size classes containing only zero values across all samples, resulting in a numeric matrix with 74 grain-size bins ranging from 0.276 μm to 3080 μm . The rEMMA workflow tested ten weight-transformation values l , with $l_{\min}$ set to 0 and $l_{\max}$ set to 95% of the maximum allowable value. These transformations did not alter the grain-size bins but adjusted the weighting of the distribution prior to unmixing. Based on the l -space exploration, rEMMA determined a plausible range of robust endmember numbers (q) and selected the final solution using goodness-of-fit criteria (Figure 27). The resulting robust endmembers represent distinct and well-separated grain-size mode clusters (Dietze & Dietze, 2019). Endmember loadings and scores were used to interpret dominant sedimentary processes and their variability within and between core sections, supported by the explained variance of the model output (Figure 28).

3.3.7 Chronology – Radiocarbon dating

For radiocarbon dating, I hand-picked organic material from sediment samples whenever TOC values were sufficiently high and plant remains were clearly identifiable. The extracted material varied depending on sample composition and availability and included twigs, grass fragments, and roots.

The MICADAS laboratory at the AWI in Bremerhaven, Germany, performed the radiocarbon dating. Due to limited capacity at the MICADAS laboratory, I selected only a subset of samples for dating. The sampling strategy aimed to capture the overall stratigraphy and key processes reflected in the cores. In general, the average sampling resolution was around 15 cm. Where horizontal cracks or breaks occurred, I selected samples directly above and below the respective structures to assess potential stratigraphic disturbances and depositional or erosional events. Due to the thin sediment intervals in both slope cores, I dated only seven samples from site A. The sediment lens in core A003 contained insufficient identifiable organic material for organic matter picking. Therefore, the MICADAS laboratory analysed the bulk dried sediment for these samples. According to Mollenhauer et al. (2021), both hand-picked material and bulk sediment measurements provide reliable radiocarbon ages for stratigraphic interpretation. The stratigraphy of core B allowed continuous sampling in depth steps of 10 cm to 15 cm, resulting in 13 dated samples.

I calibrated all positive radiocarbon ages (before present, BP) using CALIB Version 8 (Stuiver & Reimer, 2020), based on the 2020 international calibration dataset and a 2σ confidence range. Negative radiocarbon values, representing post-1950 ages, were calibrated using CALIBomb (Reimer & Reimer, 2025).

I retained all results to avoid subjective exclusion and evaluated them in relation to stratigraphy, TOC, stable water isotopes, and cryostratigraphic evidence, even if some indicated potential contamination.

4 Results

Here Chapters 4.1 and 4.2 summarise the results of valley mapping and distributed measurements, followed by Chapter 4.3, which presents the detailed results of the analysed cores from the Yedoma upland, valley slope, and valley floor. Finally, Chapter 4.3.5 portrays the results of the combined GSD and rEMMA dataset.

4.1 Valley mapping

Valley mapping products provided the spatial framework for the study site. The GeoEye Mosaic highlighted surface and topographic features of the TEV, including streams, depressions, ponds, and characteristic plateau structures (Figure 11). The ArcticDEM Mosaic provided elevation and terrain reference and formed the basis for the catchment delineation and topographic metrics (Figure 12). I summarised core coordinates and elevations extracted from the ArcticDEM in Table 2 and visualised them along the elevation profile in Figure 10. The mapped TEV catchment covered an area of 1.08 km², with elevations ranging from 12 m to 32 m a.s.l., and a total catchment volume of approximately 4 309 000 m³.

Table 2: Study site details from the permafrost cores including coordinates and elevation.

Site Name	Latitude [°N]	Longitude [°E]	Elevation Arctic DEM [m a.s.l.]	Core length [cm]	Number of analysed samples
Yedoma Upland - Core C	72.366132	126.272724	33.9	184	72
Valley Slope - Core A001	72.368553	126.256731	26.4	164	8
Valley Slope - Core A003	72.368553	126.256731	26.4	92	13
Valley Floor - Core B	72.367170	126.255306	17.5	169	62

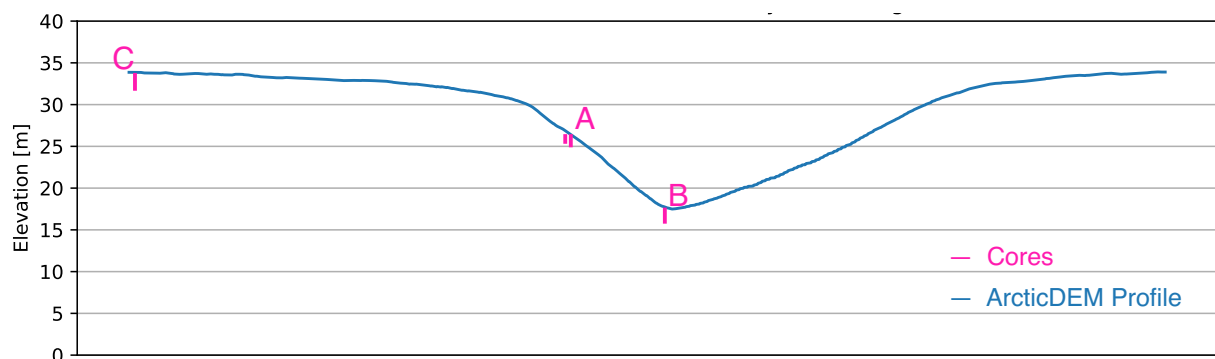


Figure 10: Elevation profile through the thermo-erosional valley on Kurungnakh Island derived from the ArcticDEM (Porter et al., 2023). The blue line in Figure 11 indicates the location of the profile within the valley. The profile starts at study area C on the Yedoma upland, descends to site B at the valley floor, and continues uphill on the opposite valley slope. Core labels indicate core positions relative to elevation, including slope cores that are not located directly on the profile line.

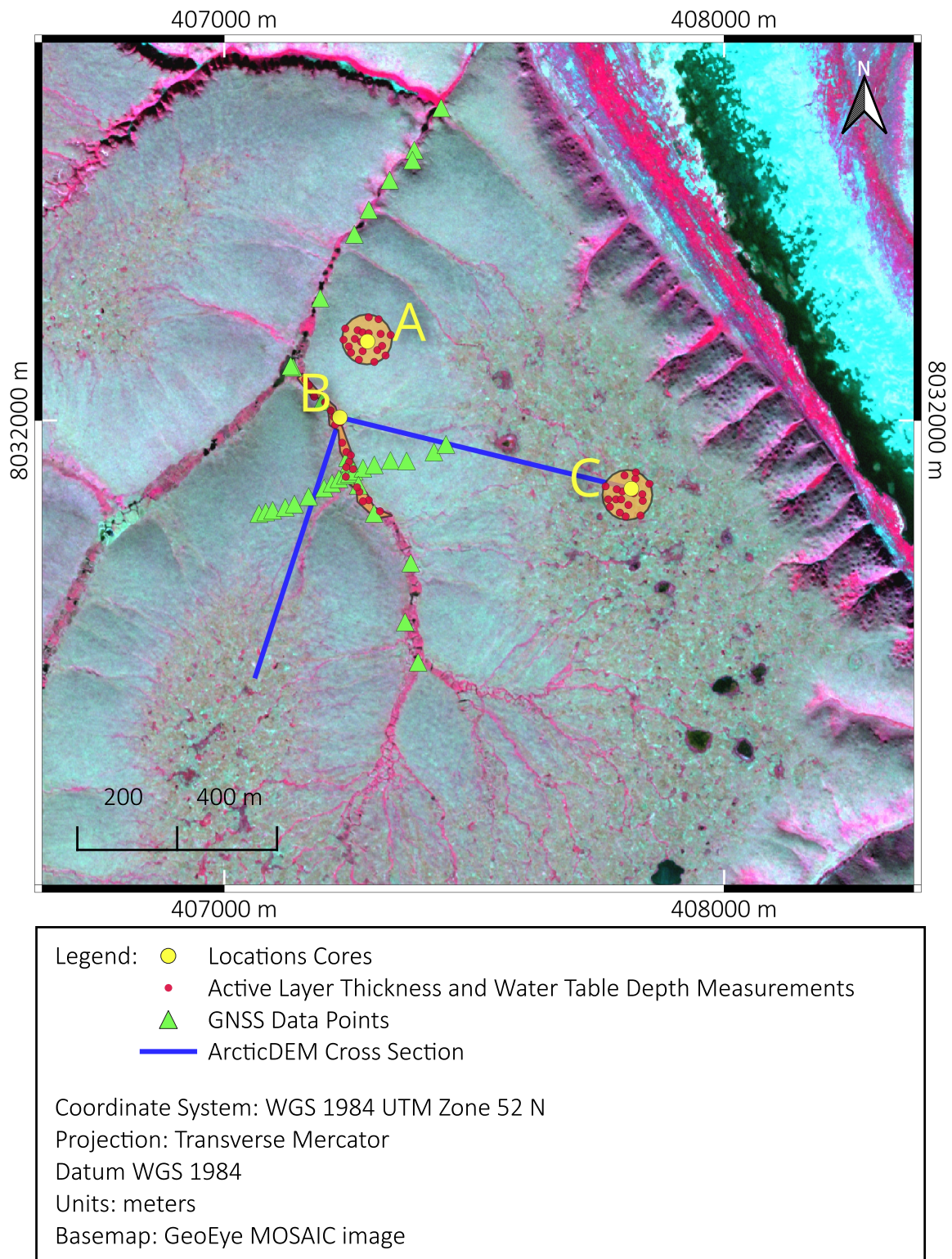


Figure 11: Thermo-erosional valley on Kurungnakh Island, Lena River Delta, Northeast Siberia. The locations of the retrieved permafrost cores A-C and study sites A-D are labelled in yellow. The locations of distributed measurements within the study areas A-C are marked with red dots. The green triangles show the GNSS data points and the blue line depicts a cross section through the Arctic DEM (Porter et al., 2023).

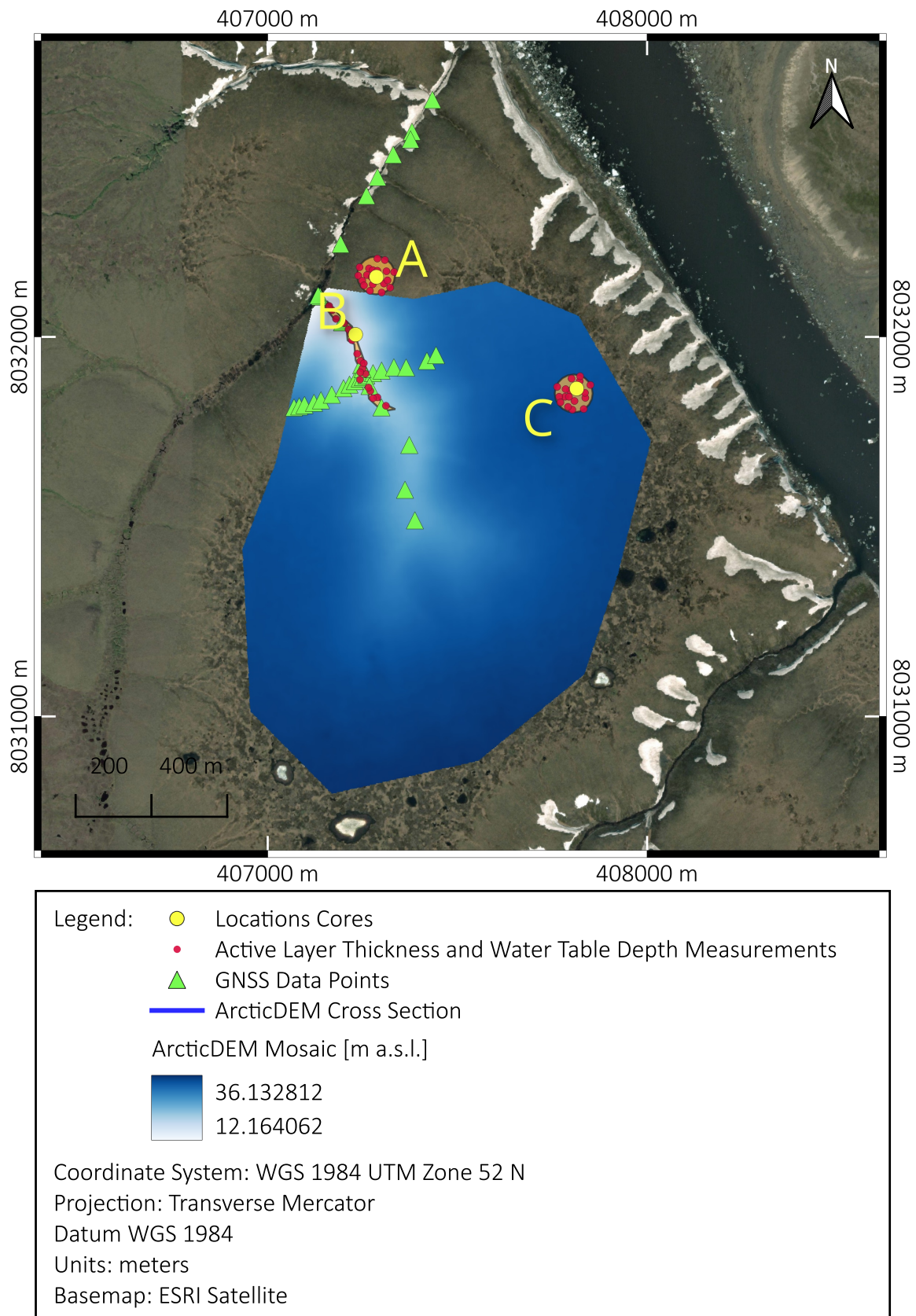


Figure 12: Thermo-erosional valley on Kurungnakh Island, Lena River Delta, Northeast Siberia. The locations of the retrieved permafrost cores A-C (dots) and study sites A-C (letters) are labelled yellow and the orange areas the study sites. Additionally, red dots mark the locations of distributed active layer thickness and water table depth measurements within the study areas. Green triangles represent Global Navigation Satellite System locations and the blue line depicts a cross section through the ArcticDEM (Porter et al., 2023).

The GNSS cross section indicated a U-shaped rather than V-shaped valley, reflecting sediment infilling at the valley bottom (Figure 13a). The longitudinal GNSS profile showed a distance of 1.3 km from the main valley outlet to the Yedoma upland within the analysed TEV branch, while the distance from the studied catchment outlet to the Yedoma upland amounted to approximately 0.7 km. Based on these metrics and the remote sensing visualisations in Figures 11 and 12, I classified the TEV on Kurungnakh Island as a side branch of an extensive dendritic valley system, comparable to extensive dendritic TEVs described for Cape Mamontov Klyk (Morgenstern, 2012). The valley clearly differed from short parallel valleys along Ice Complex margins and short radial valleys around thermokarst lakes and on alas slopes, which are also typical for the Lena River Delta (Morgenstern, 2012). The erosional base of the analysed valley was located at its outlet into the main dendritic TEV.

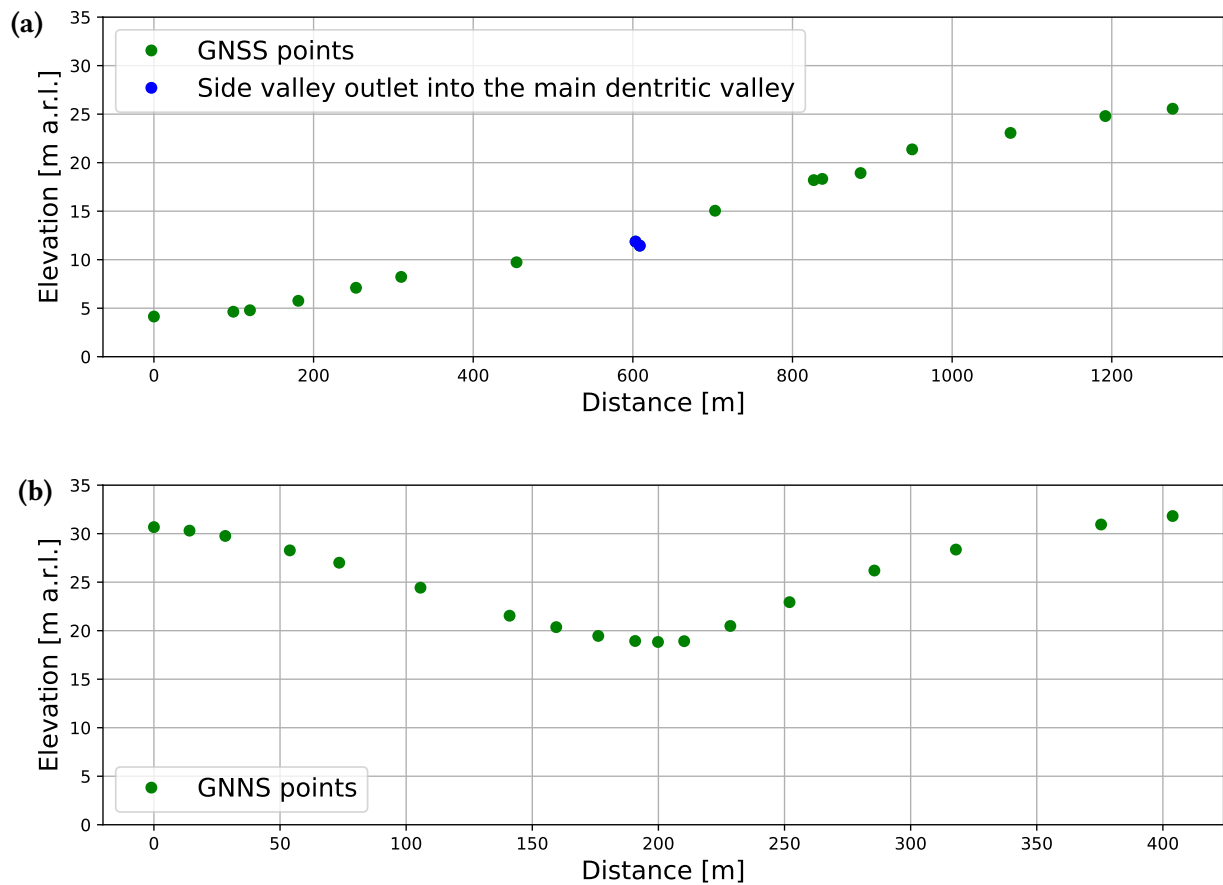


Figure 13: GNSS (Global Navigation Satellite System) elevation profiles of the analysed thermo-erosional valley: (a) longitudinal elevation profile starting close to the dendritic thermo-erosional valley outlet and leading up to the source region of the analysed thermo-erosional valley branch; (b) cross section elevation profile through the studied thermo-erosional valley, starting at the eastern source region, crossing the valley and ending on the western Yedoma upland.

4.2 Distributed measurements

Figure 14 summarised the distributed ALT and WTD measurements and revealed contrasts in hydrology and permafrost conditions across the geomorphological positions. The Yedoma upland (site C) showed the shallowest thaw depths, with a mean ALT of 34 cm (± 5 cm SD), and a comparatively deep mean WTD of -20 cm (± 3 cm SD). In contrast, the valley floor (site B) showed the deepest thaw depths, with a mean ALT of -51 cm (± 2 cm SD). The mean WTD at site B was closest to the surface (-3 cm ± 4 cm

SD) and included negative values associated with small streams and ponding water along the valley bottom. Measurements from the valley slope (site A) fell between upland and valley-floor conditions, with a mean ALT of -36 cm and a mean WTD of -24 cm (± 6 cm SD). These distributed measurements provided additional context for later interpretations of hydrological and permafrost-related development processes (Chapter 5).

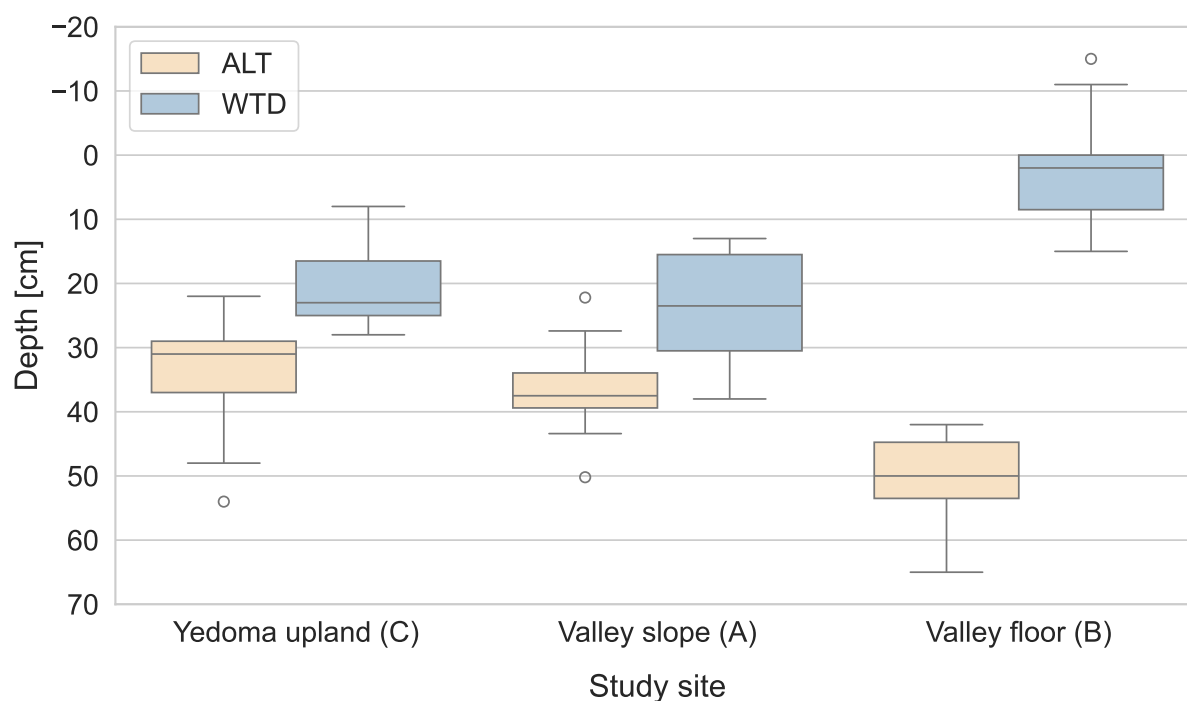


Figure 14: Distributed active layer thickness and water table depth measurements from the thermo-erosional valley, Kurung-nakh Island. Members from the Alfred-Wegener-Institute (AWI) took the measurement during the 2017 Lena Delta Expedition. The data has not been published so far.

4.3 Permafrost cores

The following chapters present the results from the permafrost core analysis. While Chapters 4.3.1–4.3.4 focus on the four cores they exclude the GSD and results from the rEMMA, on which Chapter 4.3.5 presents.

4.3.1 Yedoma upland – core C

This section presents the results from the Yedoma upland (core C), illustrated in Figure 15. Core C had a total length of 184 cm (Table 2), which I subdivided into an active layer, an upper sediment section between 20 cm to 57 cm, a lower sediment section between 57 cm to 85 cm, and an underlying massive ground-ice section extending to the base of the core. Cryostratigraphic classification indicated that massive ground ice accounted for approximately 54 % of the core length. The sediment sections consisted predominantly of organic-matrix cryostructures, reticulate structures, and minor occurrences of layered and lenticular cryostructures. VIC varied with cryostructure type and showed greater variability in the upper sediment section than in the lower section. VIC was not determined for samples consisting of massive ground ice.

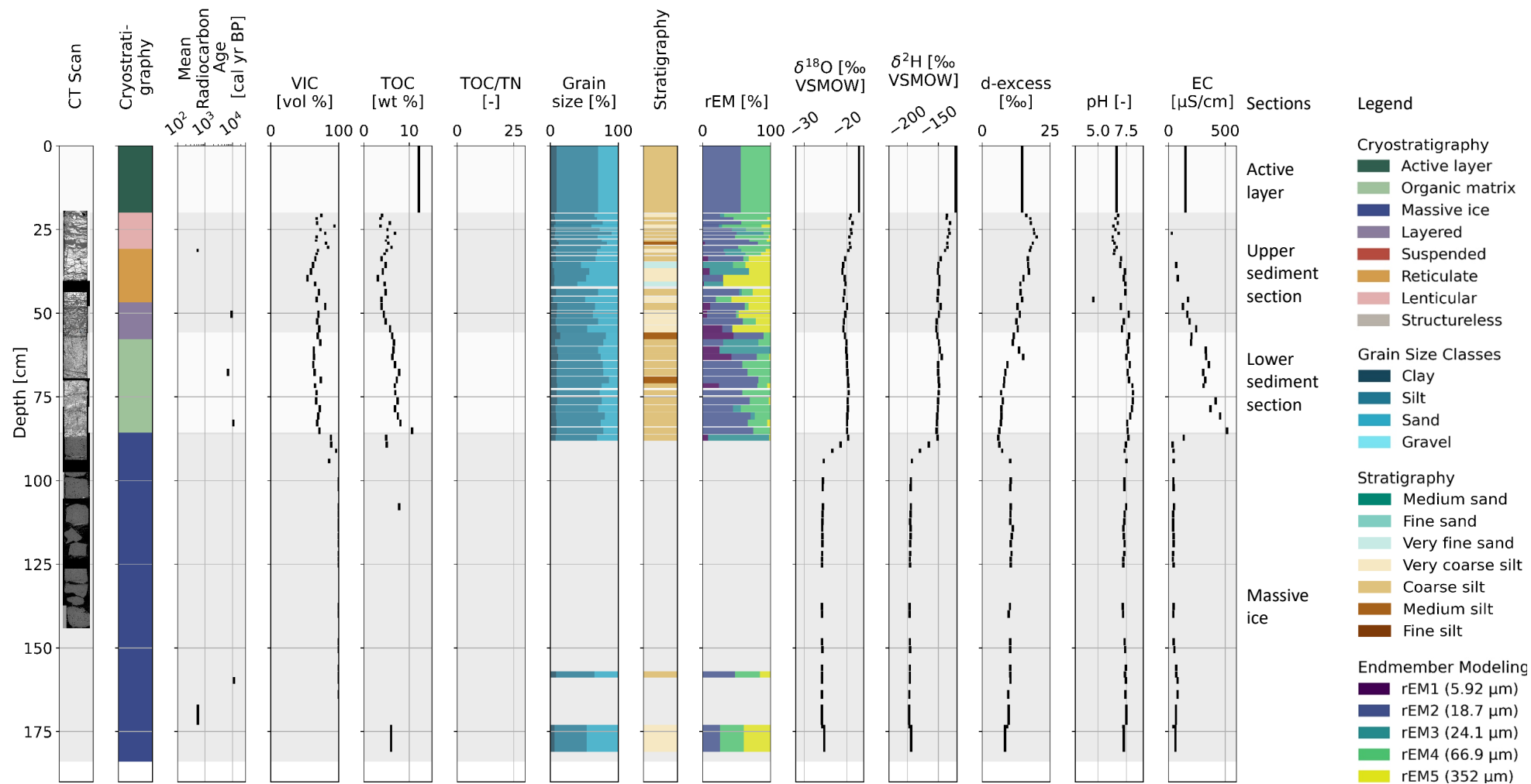


Figure 15: Data from the Yedoma upland core C, plotted by depth: CT scan, cryostratigraphy classification, Calibrated radiocarbon age, volumetric ice content (VIC), total organic carbon (TOC), total organic carbon to total nitrogen ratio (TOC/TN), grain-size contributions, stratigraphy classification, rEM contributions, $\delta^{18}\text{O}$, $\delta^2\text{H}$, d-excess, pH-values and electric conductivity (EC).

The sample selection for the radiocarbon dating contained a total of seven samples (Figure 3). The youngest near-surface sample from the active layer yielded an age of 86.1 cal yr BP. Within the sediment section, an inverse age relationship occurred at the transition between the upper and lower sediment units, where an older sample overlaid a younger one (Figure 3). The oldest age from core C (11 398.5 cal yr BP) was obtained from ice-wedge material at a depth of 160 cm. In contrast, a sample taken from the lowermost part of the core within massive ground ice indicated a younger age of –22.4 cal yr BP at depths around 183 cm. The logarithmic x-axis of the radiocarbon results in Figure 15 excluded the negative results. As later discussed in Chapter 5.2, these anomalously young ages most likely reflect contamination.

Table 3: All results of radiocarbon dating and calibrated ages per sample and core.

Core	Sample	Mean depth [cm]	14C-age [yr BP]	±	Calibrated age min (2 sig) [yr BP]	Calibrated age max (2 sig) [yr BP]	Calibrated age mean (2 sig) [cal yr BP]
C001	KUR17-C001_2	20.8	93	21	32.5	139.6	86.1
	KUR17-C001_12	31.2	536	24	516	555	535.5
	KUR17-C001_22	50.3	8232	96	9001	9454	9227.5
	KUR17-C001_30	67.7	6065	27	6846	6995	6920.5
	KUR17-C001_37	82.8	9488	56	10576	10878	10727
	KUR17-C002_61	159.7	9876	96	11143	11654	11398.5
	KUR17-C002_64	170	512	70	445	653	549
	KUR17-C002_66	177	-58	55	-22.1	-22.8	-22.4
A001	KU17_A001_002	28.6	926	65	721	936	828.5
	KU17_A001_005	32.2	1002	66	771	1007	889
A003	KU17_A003_004	25.6	248	25	277	318	297.5
	KU17_A003_013	34.7	6294	28	7162	7268	7215
	KU17_A003_010	84.8	18276	93	22036	22404	22220
	KU17_A003_011	87.7	17872	87	21415	21995	21705
	KU17_A003_012	90.4	9638	30	11066	11179	11122.5
B001	KU17_B001_001	22.8	178	24	139	225	182
	KU17_B001_008	34.5	4873	27	5580	5610	5595
	KU17_B001_015	49.9	181	24	139	225	182
	KU17_B001_020	64.4	860	70	677	912	794.5
	KU17_B001_025	77.9	602	65	524	668	596
B002	KU17_B002_002	95.2	315	57	284	497	390.5
	KU17_B002_006	106.0	517	58	475	571	523
	KU17_B002_010	115.4	633	58	540	672	606
	KU17_B002_014	124.1	846	62	673	834	753.5
	KU17_B002_018	134.5	885	24	729	800	764.5
	KU17_B002_021	143.1	762	31	666	728	697
	KU17_B002_027	156.0	748	56	666	728	697
	KU17_B002_032	166.4	637	56	542	672	607

Nitzbon et al. (2022) did not measure TN during laboratory analyses in 2019 for core C. Therefore, I report only TOC values and no TOC/TN ratios. TOC contents showed strong variability in the upper sediment section, ranging from 3.1 to 6.8. Towards the base of this section, the values stabilised. In the

lower sediment section, TOC values increased overall and ranged between 5.8 to 8.1. The massive ice section offered only sporadic sediment samples and the low sample size generally prevented further TOC analysis.

Stable water isotope compositions varied markedly with depth in core C and broadly reflected the subdivision into the active layer, sediment sections, and massive ice. Overall, $\delta^{18}\text{O}$ values ranged from -25.88‰ to -17.11‰ and $\delta^2\text{H}$ values from -197.25‰ to -122.24‰ , while d-excess values ranged between 5.93‰ to 20.04‰ . The single sample from the active layer showed the most isotopically enriched $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values and a d-excess of 14.64‰ . Below the active layer, isotope values became progressively more depleted down to a depth of approximately 38 cm. Within the remainder of the upper sediment section, $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values showed no consistent trend, whereas d-excess values initially increased to a maximum at 27 cm before decreasing towards the section base.

In the lower sediment section, d-excess values continued to decrease, apart from a limited increase directly below the sediment section boundary. In contrast, $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values showed comparatively little variability throughout most of this section. At approximately 88 cm, both isotope values began to decrease more strongly, coinciding with the onset of massive ground ice. Below 94 cm, isotope values exhibited low variability, consistent with the homogeneous cryostratigraphy of the massive ice.

pH-values and EC showed a wide range throughout core C, with EC reaching the highest values observed within the entire TEV dataset. In the upper sediment section, pH-values varied more than in the lower sediment section, where values increased up to 8.09 before decreasing to approximately 7.5. Within the massive ice section, pH-values remained relatively stable slightly below 7.5. EC increased strongly towards the base of the lower sediment section, reaching a maximum of $518\text{ }\mu\text{S}/\text{cm}$. The EC record did not reflect a pH outlier, suggesting a measurement artefact. With the onset of massive ground ice, EC values decreased sharply to approximately $50\text{ }\mu\text{S}/\text{cm}$.

4.3.2 Valley slope – core A001

Core A001 had a total length of 164 cm (Table 2). Based on cryostructural characteristics, I subdivided the core into the active layer, a sediment section extending from 27 cm to 34 cm, and underlying massive ice reaching the base of the core. Furthermore, the cryostratigraphic classification indicated predominantly layered cryostructures within the sediment section.

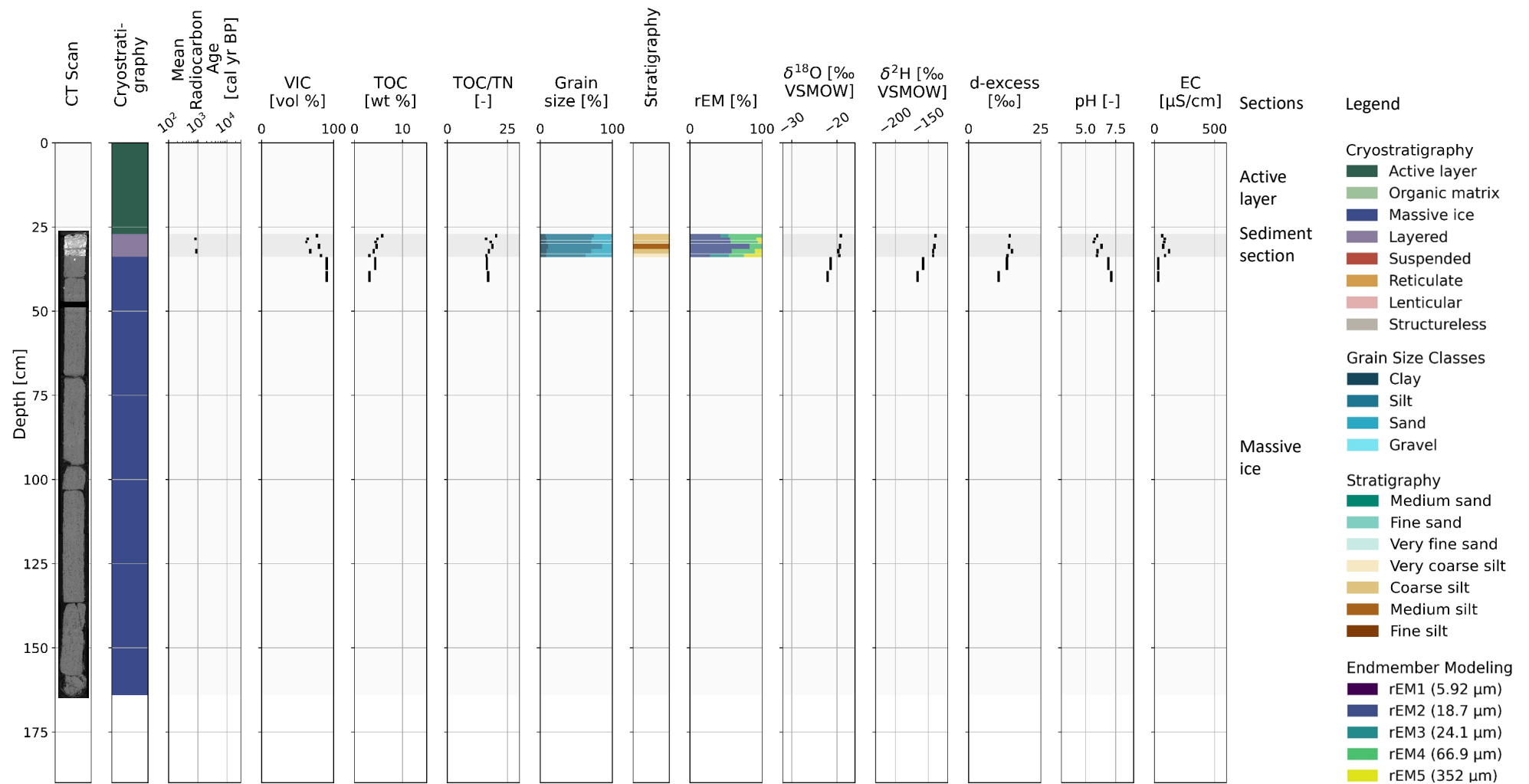


Figure 16: Data from the valley slope core A001, plotted by depth: CT scan, cryostratigraphy classification, Calibrated radiocarbon age, volumetric ice content (VIC), total organic carbon (TOC), total organic carbon to total nitrogen ratio (TOC/TN), grain-size contributions, stratigraphy classification, rEM contributions, $\delta^{18}\text{O}$, $\delta^2\text{H}$, d-excess, pH-values and electric conductivity (EC).

For radiocarbon dating, I selected two samples from core A001, both originating from the sediment section (Figure 3). The upper sample at the top of the sediment section yielded a calibrated ages of 828.5 cal yr BP. A second sample, taken 3 cm below, dated to 889 cal yr BP.

TOC values showed a decreasing trend with depth within the sediment section, whereas TOC/TN ratios fluctuated near the top of the section and became more stable towards its base. As shown in Figure 16, the spatial sampling resolution increased with depth, which likely contributed to the reduced variability of the TOC/TN values.

Due to the limited thickness of the sediment section, several samples did not provide sufficient material for stable water isotope analysis. In addition, I did not sample the massive ice-wedge section. As a result, isotope data were available from four sediment samples and two samples from the uppermost ice section. Despite the limited dataset, $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values from the sediment and upper ice section showed a clear depletion trend with depth. The ice-section samples still contained minor sediment traces and internal structures. Overall, $\delta^2\text{H}$ values ranged between -166.15‰ to -138.72‰ , while $\delta^{18}\text{O}$ values ranged from -22.07‰ to -19.12‰ . The d-excess values also decreased with depth and varied between 10.38‰ to 15.01‰ .

pH-values followed a similar pattern to the stable water isotope data. The highest pH-values occurred at the top of the sediment section, while the lowest values appeared near the transition into the massive ice section. Over short depth intervals, pH-values changed rapidly and ranged overall between 5.68 to 6.9. The EC showed no consistent trend within the sediment section and ranged from $65\text{ }\mu\text{S/cm}$ to $122\text{ }\mu\text{S/cm}$. In contrast, EC values dropped sharply to approximately $32\text{ }\mu\text{S/cm}$ in the two samples from the massive ice.

4.3.3 Valley slope – core A003

Core A003 had a total length of 92 cm (Table 2). Cryostructural characteristics allowed a clear subdivision into a sediment section extending from the surface to 37 cm depth and an underlying massive ice section reaching to the base of the core. Within the massive ice, a sediment lens occurred between 83 cm and 91 cm. This resulted in the following top-down stratigraphic sequence of active layer, sediment section, massive ice, and sediment lens. Massive ice accounted for approximately 51 % of the core length, while I classified the sediment sections as layered, organic-matrix, and reticulate cryostructures. The VIC varied with the characteristics in cryostructures.

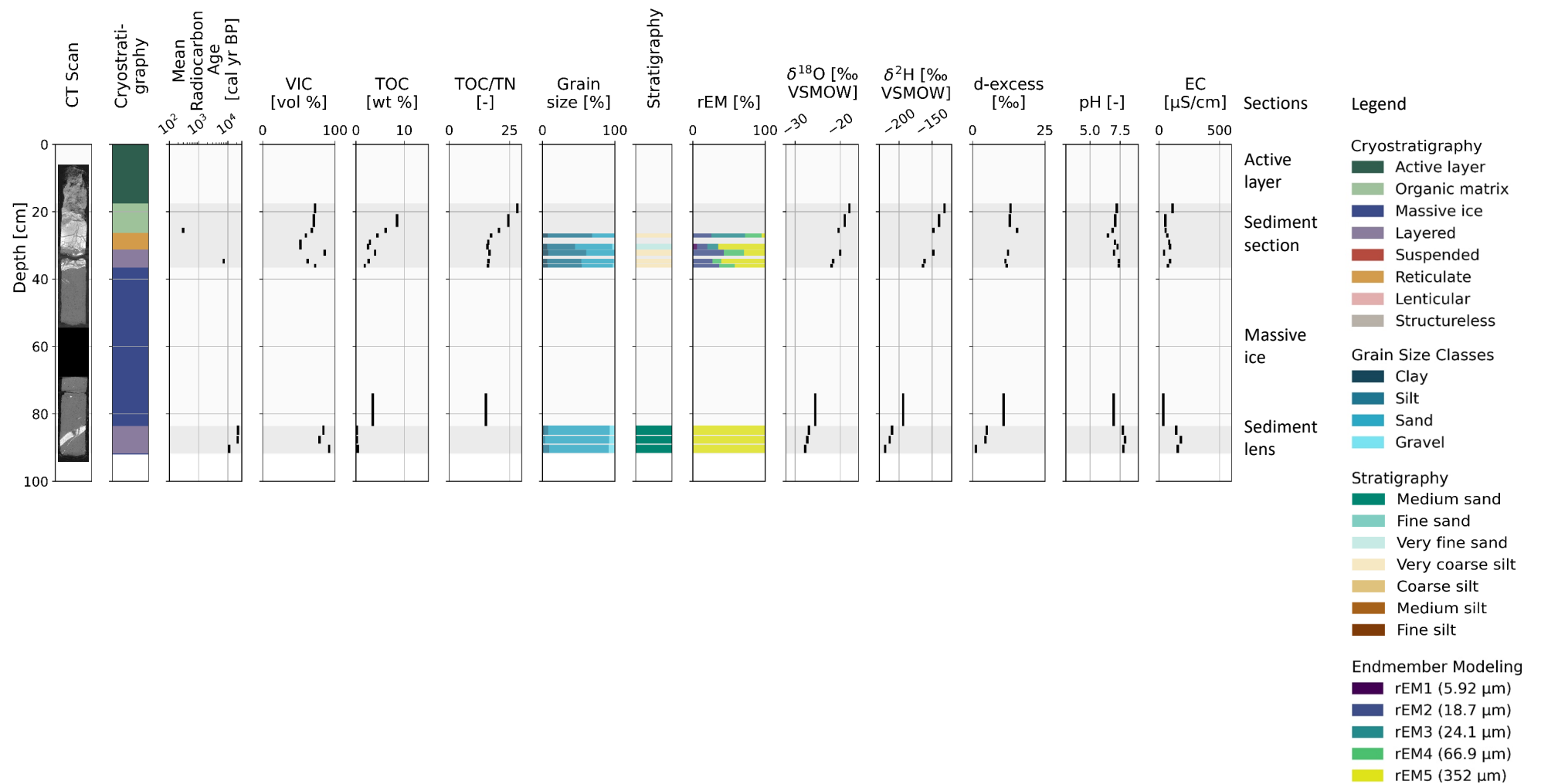


Figure 17: Data from the valley slope core A003, plotted by depth: CT scan, cryostratigraphy classification, Calibrated radiocarbon age, volumetric ice content (VIC), total organic carbon (TOC), total organic carbon to total nitrogen ratio (TOC/TN), grain-size contributions, stratigraphy classification, rEM contributions, $\delta^{18}\text{O}$, $\delta^2\text{H}$, d-excess, pH-values and electric conductivity (EC).

For radiocarbon dating, I selected samples from both the sediment section and the sediment lens of core A003 (Figure 3). The uppermost dated sample at a depth of 26 cm displayed a relatively young age of 297.5 cal yr BP. In contrast, the sample taken at 35 cm dated to 7215 cal yr BP. Samples derived from the sediment lens showed a wide range of ages, with the oldest dating to 22 220 cal yr BP and the youngest to 11 122.5 cal yr BP.

At the top of the sediment section, TOC values decreased rapidly with depth and became slightly more stable towards the base of the section (Figure 17). Within the sediment lens, TOC values were very low and ranged between 0.2 wt% to 0.4 wt%, except for one sample directly above the lens that showed a comparatively high TOC value of 3.5 wt%. This elevated value coincided with a very small sedimentary sample caused by high ice content. TOC/TN ratios followed a similar pattern, with high values at the top of the sediment section that decreased with depth. Due to the limited sample size, TN could not be measured for all samples.

Stable water isotope data from core A003 did not include samples from the active layer or the main ice section (Figure 17). Nevertheless, $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values from the sediment section and sediment lens showed a pronounced depletion trend with depth with a steep gradient. Overall, $\delta^2\text{H}$ values ranged between -221.14‰ to -130.94‰ , while $\delta^{18}\text{O}$ values ranged from -27.79‰ to -18.01‰ . The three samples from the sediment lens exhibited particularly depleted isotope values, representing the lowest hydrogen and oxygen isotope values measured across all sites. The d-excess values also decreased with depth and overall ranged between 1.15 ‰ to 15.34 ‰, although the trend showed greater scatter and did not follow a well-defined linear relationship.

pH-values and EC from core A003 showed no consistent depth-dependent trend within the sediment section and ranged between 6.48 to 7.4 and 35 $\mu\text{S}/\text{cm}$ to 111 $\mu\text{S}/\text{cm}$, respectively (Figure 17). In contrast, samples from the sediment lens deviated markedly from this pattern, with pH-values reaching up to 7.9 and EC values increasing to 181 $\mu\text{S}/\text{cm}$.

4.3.4 Valley floor – core B

Core B had a total length of 169 cm (Table 2). In contrast to the other analysed cores from the TEV, this core did not show a clear separation between sediment and massive ice. Based on CT scans and subsequent laboratory analyses, I subdivided the core into the active layer (0 cm to 22 cm) and four sediment sections arranged from top to bottom. Section boundaries occurred at depths of 41 cm, 70 cm, and 91 cm (Figure 18). Physical breaks separated the active layer from sediment section one, sediment section two from section three, and section three from section four, with an additional break occurring within sediment section four.

Cryostratigraphic classification showed organic-matrix cryostructures as the most abundant type, followed by reticulate, layered, lenticular, suspended, and structureless cryostructures. Among all analysed cores, core B displayed the highest cryostructural diversity, comprising seven distinct cryostructure types. The CT scan further revealed differences in cryostructure orientation between sediment sections. In sediment section two, cryostructures, particularly layered structures, were slightly inclined, whereas sediment section three showed predominantly horizontal alignment. Sediment section four was characterised by overall homogeneous, mainly horizontally aligned cryostructures, with alternating sequences of organic-matrix and fine reticulate structures. Overall, core B exhibited high

VICs, even in sections dominated by fine cryostructures without prominent ice lenses. Nevertheless, ice-rich features showed locally elevated VIC values (e.g. sediment section three displayed in Figure 18), as sampling targeted such structures specifically, to enable the validation of CT scan segregation methods.

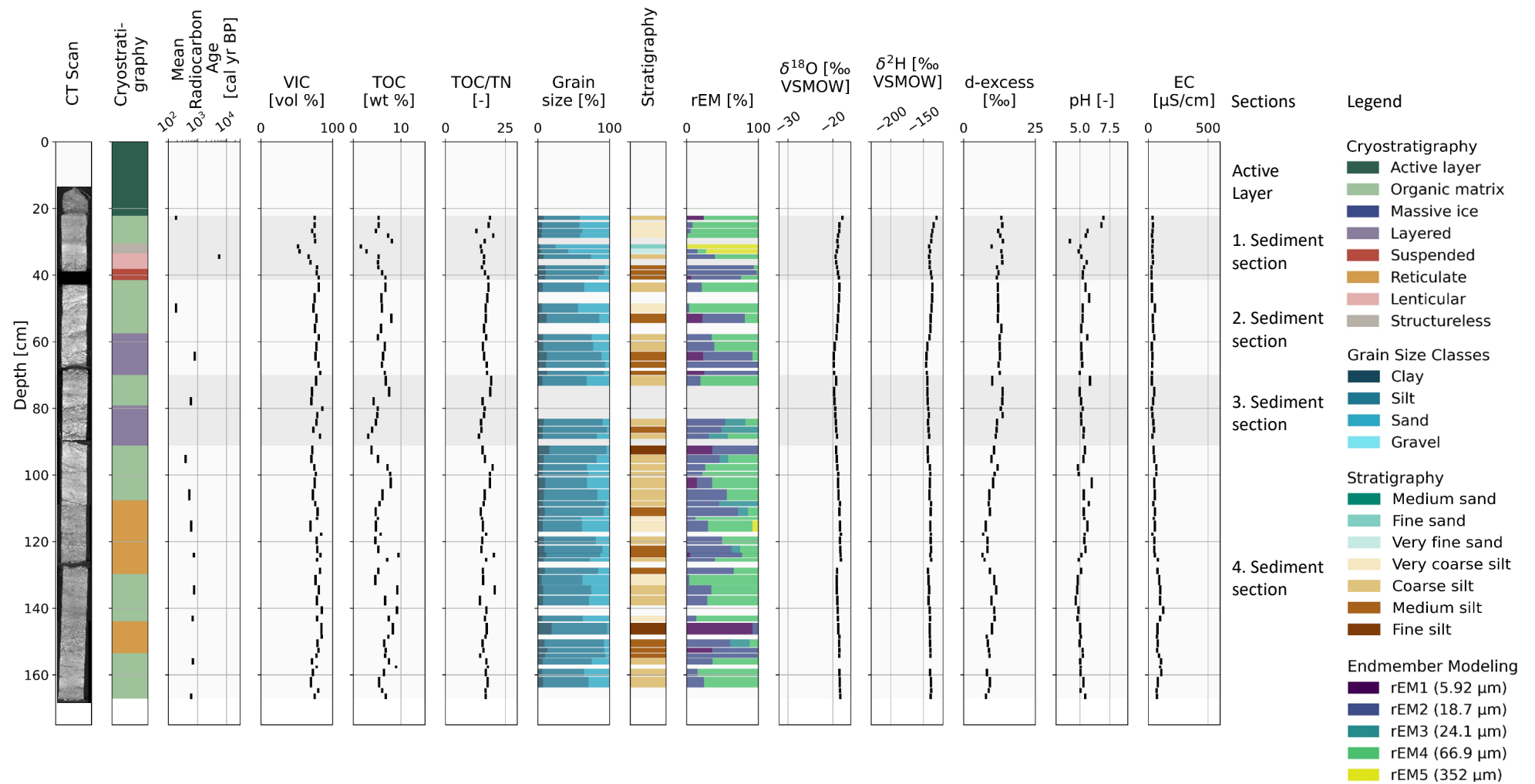


Figure 18: Data from the valley slope core B, plotted by depth: CT scan, cryostratigraphy classification, Calibrated radiocarbon age, volumetric ice content (VIC), total organic carbon (TOC), total organic carbon to total nitrogen ratio (TOC/TN), grain-size contributions, stratigraphy classification, rEM contributions, $\delta^{18}\text{O}$, $\delta^2\text{H}$, d-excess, pH-values and electric conductivity (EC).

Radiocarbon ages from core B ranged between 182 cal yr BP to 794.5 cal yr BP, except for one distinct outlier (Figure 3). Within this narrow range, ages fluctuated with depth and locally showed inverse stratigraphy, with only limited depth-dependent trends. The outlier strongly characterised the first sediment section. The second sediment section contained a sample with the same age as the one directly above the outlier (182 cal yr BP), followed by an increase to 794.5 cal yr BP. In sediment section three, ages decreased again to 390.5 cal yr BP, before increasing in the fourth sediment section, although without reaching the maximum age observed in section two.

TOC values varied strongly throughout core B and generally remained high, ranging between 1.6 wt% to 9.5 wt%. The first sediment section showed pronounced variability with abrupt changes in TOC. In sediment section two, TOC values varied more smoothly, whereas sediment section three marked a rapid decrease followed by noisy fluctuations. These fluctuations continued into sediment section four, before TOC values stabilised again towards the base of the core. TOC/TN ratios followed a comparable pattern and ranged between 13 to 21.

Stable water isotope compositions showed relatively low overall variability and broadly reflected the subdivision into the active layer and four sediment sections. The isotope data displayed several distinct features that corresponded to cryostratigraphic characteristics and grain-size variations. The first sediment sample exhibited the most enriched isotope values, with $\delta^2\text{H}$ of -129.78‰ , $\delta^{18}\text{O}$ of -17.87‰ , and a d-excess of 13. From this depth, $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values became progressively more depleted down to approximately 34 cm, before increasing again towards the base of first sediment section.

Sediment section two began with more enriched isotope values which from this point to the base of the section smoothly decreased, while d-excess values remained comparatively stable. Sediment section three showed slowly increasing isotope values accompanied by a noisy decrease in d-excess. In the fourth sediment section, d-excess values became more variable, whereas $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values showed a gradual enrichment with depth. A abrupt jump towards more depleted isotope values marked a physical break at approximately 125 cm, interrupting the general enrichment trend.

To evaluate whether observed isotope trends reflected freezing processes, I plotted $\delta^2\text{H}$ against $\delta^{18}\text{O}$ and calculated regression slopes ($S_{2\text{H}-18\text{O}}$) for selected depth intervals (Figures 19a,b). Regression lines in Figure 19a represent intervals with decreasing isotope values with depth, whereas Figure 19b shows regressions for intervals with increasing isotope values.

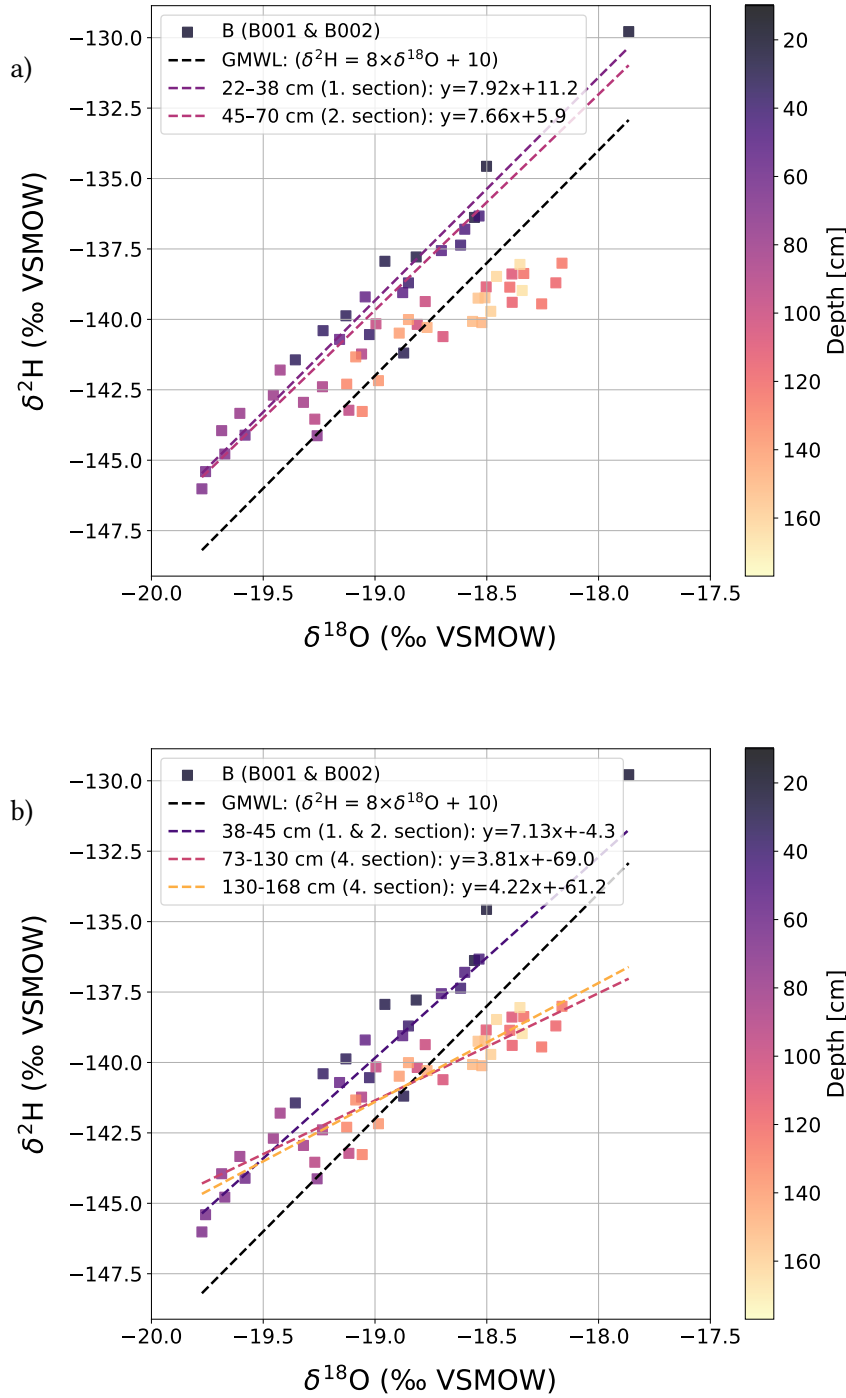


Figure 19: Stable isotope relationships for core B. Plot a and b show $\delta^{18}\text{O}$ – $\delta^2\text{H}$ plots with the Global Meteoric Water Line (GMWL) and regression lines for top-down and bottom-up trends, respectively.

Regression slopes associated with potential top-down freezing exceeded 7.3 and therefore did not indicate freezing processes (Figure 19a). A regression slope of 7.13 calculated for data from sediment sections one and two, where isotope patterns suggested possible bottom-up freezing between 38 cm to 45 cm, fell only marginally below this threshold (Figure 19b). Therefore, I did not consider it as an indicator of freezing fractionation. In contrast, regression slopes for sediment section 4 were considerable below 7.3, with values of 3.81 and 4.22, indicating bottom-up freezing (Figure 19b).

To further support these interpretations, I plotted d-excess against $\delta^2\text{H}$ for the same depth intervals

(Figures 20a,b). Slopes in Figure 20a showed no evidence for freezing, nor did the slope derived from sediment sections one and two in Figure 20b. In contrast, strongly negative trends for data from the third and fourth sediment section supported the interpretation of freezing fractionation.

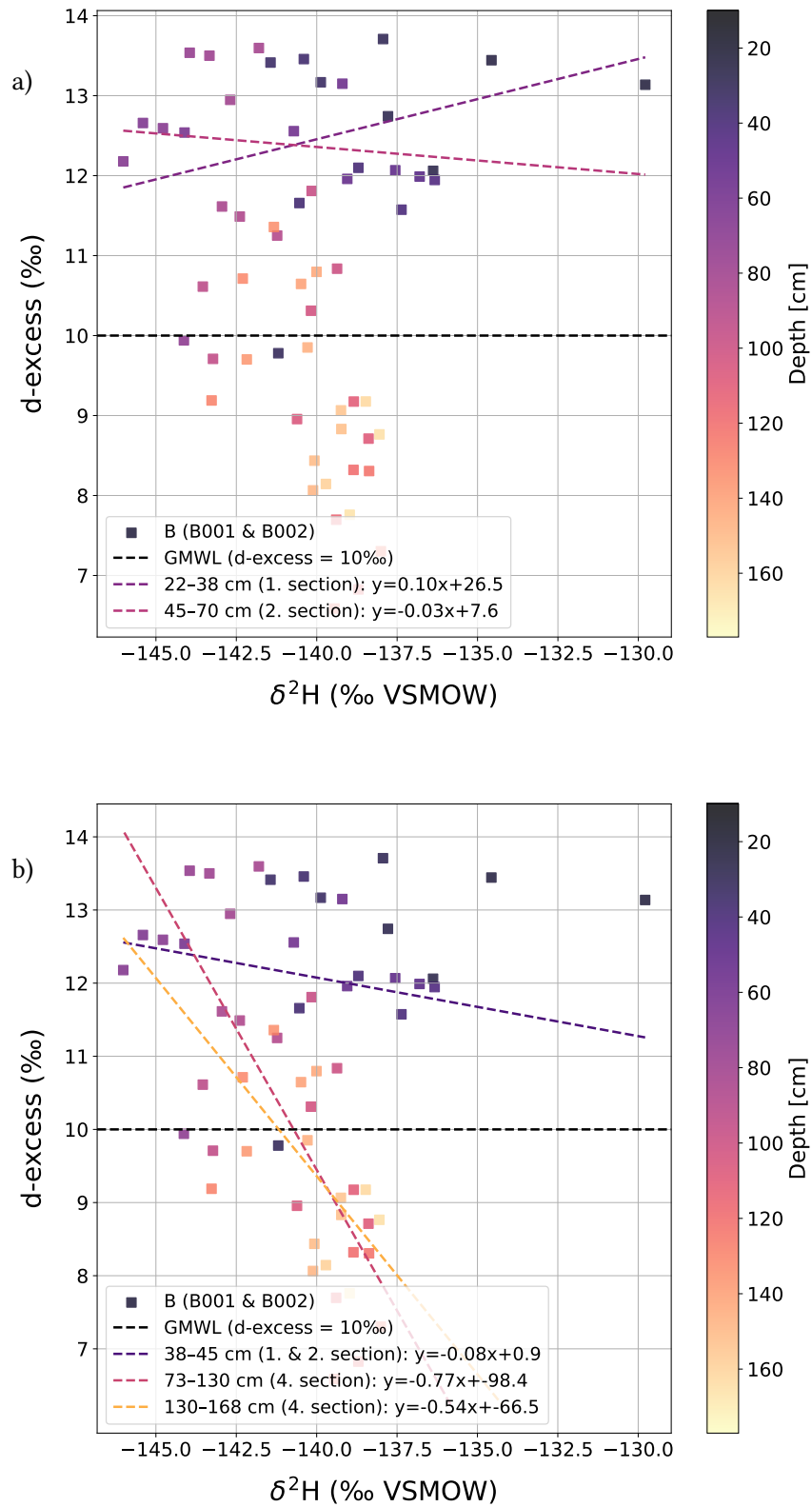


Figure 20: Plot a and b show the relationship between d-excess and $\delta^2\text{H}$ for the same datasets as in Figure 19.

pH values in core B did not follow isotope trends and instead displayed independent variability (Figure 18). Sediment section one showed the largest fluctuations, including the minimum (4.16) and maximum (6.97) pH-values observed in the core. In the second and third sediment sections, as well as the upper half of sediment section four, pH-values fluctuated less and remained relatively stable. The lower half of sediment section four exhibited the smoothest pH trend. In contrast to pH, EC values showed little overall variability, with a gradual increase within the lowest sediment sections. The minimum EC value (28 $\mu\text{S}/\text{cm}$) occurred near the boundary between sediment sections one and two, whereas the maximum (125 $\mu\text{S}/\text{cm}$) was recorded in the fourth sediment section.

Several parameters showed distinct responses in and around the sample with the radiocarbon age outlier at 34 cm. The dated sample yielded an age of 5580 cal yr BP to 5610 cal yr BP. Two samples directly above this outlier showed a pronounced shift in GSD, classifying as *very fine sandy very coarse silt* by GRADISTAT (Blott & Pye, 2010), whereas samples below classified as *very coarse silty medium sand*. The outlier sample itself again classified as *very fine sandy very coarse silt*. These two overlying samples also exhibited the lowest TOC and TOC/TN ratios within core B. This change in material composition was visually evident in CT scans (Figure 18) and documented in laboratory photographs (Figure 8c). Cryostratigraphic classification identified these samples as structureless, and additional datasets showed coincident shifts, including a decrease in VIC and excursions in stable water isotope values.

4.3.5 Grain-size distributions and robust endmembers

The GSDs of the TEV dataset varied strongly between sites and with depth and ranged from mud, with an arithmetic mean of 17.7 μm in core B, to medium sand, with an arithmetic mean of 470.3 μm in core A003 (Figures 21c–d). The output of the GRADISTAT analysis (Blott & Pye, 2001) showed that, out of 103 analysed samples, 76 classified as *very poorly sorted*, 26 as *poorly sorted*, and only 1 as *moderately sorted* (Table 4). Overall, the GSDs displayed unimodal, bimodal, trimodal, and polymodal distributions (Figure 21). In some samples from cores A001, A003, and B, the sieved grain fractions greater than 1 mm and greater than 2 mm produced abrupt peaks in the GSD curves (Figures 21b–d).

Table 4: Summary of sorting categories per analysed core.

Site name	Sorting	Count	Percentage
Yedoma Upland - Core C	Very Poorly Sorted	40	98
	Poorly Sorted	1	2
Valley Slope - Core A001	Very Poorly Sorted	6	100
Valley Slope - Core A003	Very Poorly Sorted	5	62
	Poorly Sorted	2	25
	Moderately Sorted	1	13
Valley Floor - Core B	Very Poorly Sorted	25	52
	Poorly Sorted	23	48

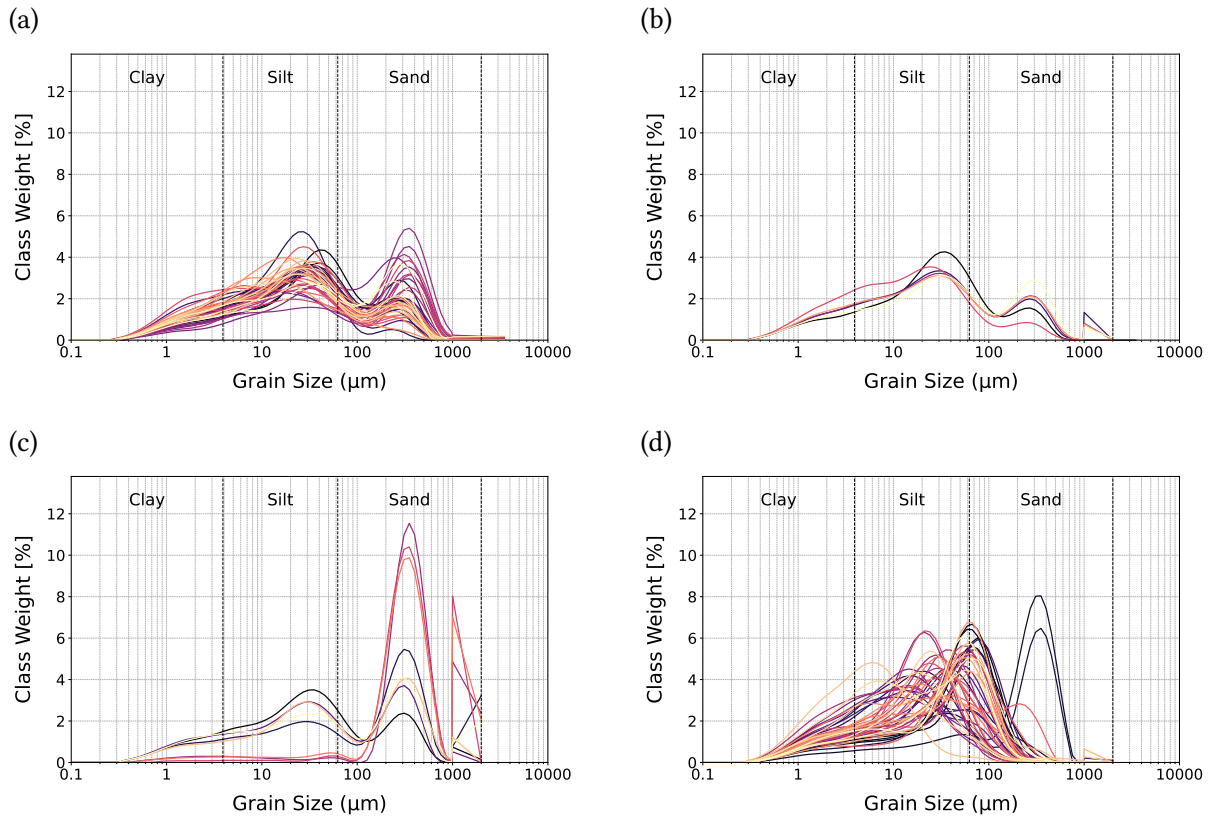


Figure 21: Grain-size distribution curves for all analysed samples from: the Yedoma upland core C (a); valley slope core A001 (b); valley slope core A003 (c) and valley floor core B (d). The included sediment nomenclature by Blott and Pye (2001) serves as a reference. The magenta colour scheme depicts the full range of samples for each core, where darker samples are located deeper below the surface.

The rEMMA with the EMMAgeo R package (Dietze & Dietze, 2019) identified five rEMs (Table 5). The dominant modes ranged from fine silt to medium sand. All rEMs exhibited polymodal distributions (Figure 22 and Appendix Figure 28). Among the five endmembers, rEM4 explained the highest proportion of variance (32 %), followed by rEM2 and rEM5, each accounting for 25 %. On a sample-wise basis, the five rEMs explained an average of approximately 90 % of the total variance (Appendix Figure 28). Table 5 lists the range, main mode and explained variance of all five rEMs.

Table 5: This table lists the size range, main mode and explained variance of each rEM from the data of this thesis. Furthermore, it lists the corresponding best matching rEMs published by Schirrmeister et al. (2020) on Arctic Yedoma samples.

	rEM5	rEM4	rEM3	rEM2	rEM1
Size range [μm]	(310–352 μm)	(59–76 μm)	(21–27 μm)	(15–21 μm)	(5–8 μm)
Main mode [μm]	310	66.9	24.1	18.7	5.92
Explained variance [%]	6.5	25	12	32	25

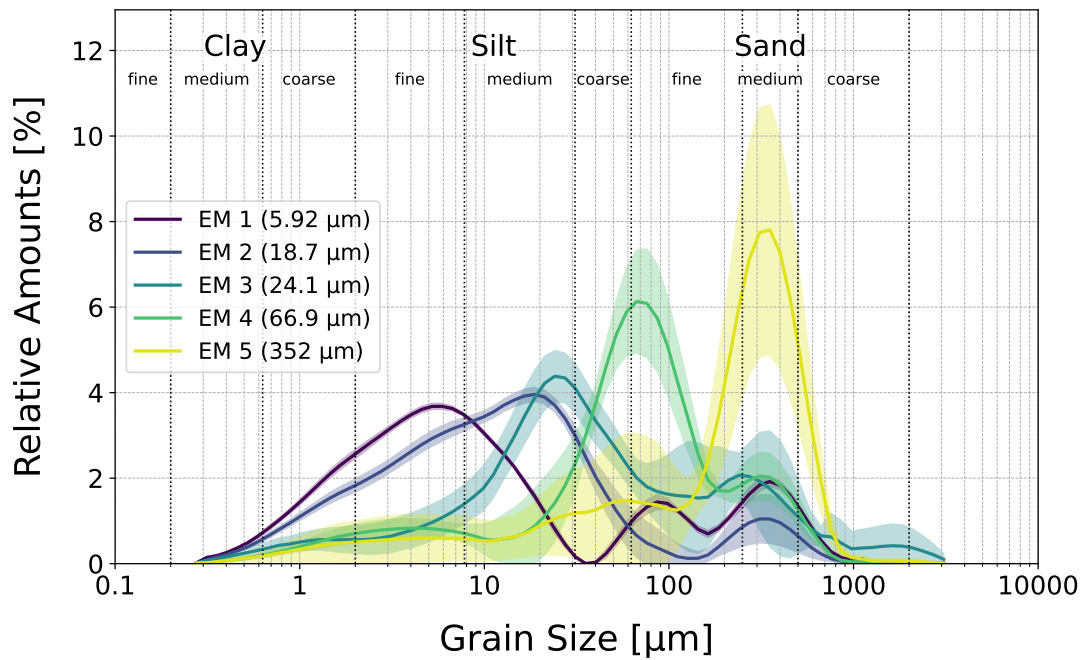


Figure 22: Endmember loadings for all samples from the sites in the thermo-erosional valley on Kurungnakh Island.

The coarsest endmember, rEM5, showed a main mode at 352 μm (Table 5). In addition to this dominant coarse mode, rEM5 exhibited minor secondary modes, including a fine component around 4 μm to 6 μm (Figure 22). rEM5 contributed to all four analysed cores and explained up to 100 % of the grain-size variance in the sediment lens of core A003 (Figure 17) and in the previously described outlier samples from the valley floor (Figure 18). These rEM5 dominated samples were moderately sorted and displayed a dominant coarse-grained peak in the laser-derived distributions, accompanied by sieved proportions of grains exceeding 1000 μm and 2000 μm (Figure 21). As EMMAgeo modelling excluded the sieved coarse fractions, the main mode of rEM5 likely underestimated the coarsest grain-size component present in these samples. At the Yedomia upland, rEM5 mainly contributed to the upper sediment section of the core (Figure 15), where individual sample curves revealed distinct coarse-grained peaks (Figure 21a) aligned with the rEM5 distribution.

The second coarsest endmember, rEM4, represented very fine sand to coarse silt and showed a main mode at 66.9 μm (Table 5). rEM4 explained the highest proportion of variance among all five endmembers, accounting for 32 % of the total dataset variance (Figure 28). It contributed to all analysed cores (Figure 15–18), although its relative contribution in the slope core A003 was low (Figure 17). GSD curves indicated spatial variability in the representation of rEM4. Samples from the valley floor exhibited dominant peaks closely matching the rEM4 main mode (Figure 21d and Table 5), whereas samples from the Yedomia upland and valley slope showed peaks predominantly within the coarse silt range rather than the fine sand range. Nevertheless, rEM4 contributed to these samples. rEM4 also displayed secondary modes, including a fine component comparable to the finest mode of rEM5 and a weak mode within the grain-size range of the rEM5 main mode (Figure 22). Similar secondary modes occurred in rEM1 and rEM2.

rEM3 showed a main mode at 24.1 μm , classifying as coarse silt Blott and Pye (2001). This endmember explained 12 % of the total dataset variance (Figure 28) and occurred in all analysed cores (Figures 15–

18). In addition to the dominant coarse-silt mode, rEM3 exhibited a weak secondary mode at grain sizes slightly below 200 μm (Figure 22). The relative contribution of rEM3 varied between study sites but remained consistently present throughout the dataset (Appendix Figure 28).

rEM2 showed a main mode at 18.7 μm , classifying as silt Blott and Pye (2001). This endmember explained 25 % of the total GSD variance and occurred in all analysed cores from the TEV. In contrast to the coarser endmembers, rEM2 did not display a smooth fine-grained tail of the main mode but showed an irregular distribution curve, indicating the combination of multiple grain-size components (Figure 22).

The finest robust endmember, rEM1, showed a main mode at 5.92 μm , classifying as fine silt Blott and Pye (2001) (Figure 22). This endmember explained 6.5 % of the total GSD variance and occurred in samples from the Yedoma upland, the valley slope (core A003), and the valley floor. Compared to the coarser endmembers, rEM1 displayed a narrow GSD concentrated in the fine-silt and clay range.

5 Discussion

This chapter integrates the site-specific observations from Kurungnakh Island into an overarching interpretation of TEV development in ice-rich permafrost terrain. Based on the stratigraphic, cryostratigraphic, and geochemical results, the discussion evaluates the processes shaping the Yedoma upland, valley slopes, and valley floor during the Holocene, while also considering potential LatePleistocene influences.

The discussion first provides a process-based interpretation of sedimentary signals and then addresses the three geomorphological study sites: the Yedoma upland, valley slope, and valley floor. Finally, I synthesise the site-specific findings to assess the overall development of the TEV and its resilience to climatic, hydrological, and permafrost dynamics.

5.1 Grain-size distribution and robust endmember interpretations

The interpretation of GSDs and rEMs derived from the EMMAgeo modelling (Dietze & Dietze, 2019) follows the general assumption that coarser sediment fractions require higher transport energy than finer material. Accordingly, I interpret the rEMs in terms of sedimentary processes acting within hydraulic and aeolian systems under Arctic permafrost conditions. In the context of the Lena River Delta and Kurungnakh Island, these processes include fluvial transportation and depositions related to snowmelt-driven runoff and precipitation, a thermo-erosional component related to permafrost thaw and overland flow as well as aeolian sediment transport (Morgenstern, 2012; Morgenstern et al., 2021b; Wetterich et al., 2020). In addition, postdepositional processes such as gelifluction, cryoturbation, and ground-ice formation may modify primary depositional signals (Schirrmeister et al., 2020).

The predominantly polymodal GSDs of the TEV dataset reflect the superposition of multiple transport and depositional processes (Tanner, 1964). In the following sections, I interpret the shapes and dominant modes of the five rEMs of the studied dataset (rEM_{TEV}) and compare them with grain-size characteristics of known sedimentary systems Sun et al. (2002) and with published rEMs from Yedoma and thermokarst environments (Schirrmeister et al., 2020; Ulrich et al., 2019; Wetterich et al., 2020). Differences between the $rEMs_{TEV}$ and published Yedoma endmembers (rEM_{TEV}) are expected, as the TEV integrates reworked Yedoma material and Holocene deposits that have undergone erosion, transport, and redeposition. In addition, Schirrmeister et al. (2020) analysed a much larger dataset of 771 samples from heterogeneous study sites distributed across the Arctic, which further limits direct comparisons. Nevertheless, comparing the main modes of both datasets, including $rEMs_{Yedoma}$ not represented on Kurungnakh Island, provides valuable insights into the links between sedimentary processes and GSDs.

5.1.1 $rEM5_{TEV}$

The coarse main mode of $rEM5_{TEV}$ at $352\mu m$ requires high transport energies to mobilise sand-sized grains. As summarised by Dietze et al. (2014), endmembers with grain sizes larger than $250\mu m$ indicate fluvial transport within hydraulic sedimentary systems. Consistent with this, the main mode of $rEM5_{TEV}$ lies within the saltation load range of $200\mu m$ to $400\mu m$ reported for fluvial environments (Sun et al., 2002), supporting a fluvial transport origin.

The presence of grains exceeding 1 mm and 2 mm in several samples indicates that the rEM5_{TEV} distribution likely represents a sand-to-gravel fraction. As these coarse fractions were removed prior to rEMMA, the main mode of rEM5_{TEV} is probably shifted towards lower grain sizes. In addition, reworking and fractionation during Holocene pedogenic and cryogenic processes may further contribute to the slightly lower peak position compared to published saltation modes (Sun et al., 2002).

Secondary fine-grained components, such as the mode around 4 μm to 6 μm (Figure 22), commonly occur in aeolian sedimentary systems, including desert sand and loess deposits (Sun et al., 2002). Here, this fine component likely represents long-term suspended dust. Additional weak modes between 20 μm to 80 μm most likely reflect overlap with the dominant modes of rEM1_{TEV} to rEM4_{TEV} and therefore represent spurious secondary signals (Dietze & Dietze, 2019). Consequently, the interpretation of rEM5_{TEV} focuses primarily on its dominant coarse-grained mode.

The shape and main mode of rEM5_{TEV} closely match rEM2_{Yedoma} identified by Schirrmeister et al. (2020), which they associate with high-energy fluvial processes such as saltation and traction during strong meltwater runoff. Consistent with this interpretation, I assign rEM5_{TEV} to fluvially transported and hydraulically sorted material derived from Yedoma deposits that is relocated or reworked during the Holocene. This material may include sand originating from MIS 4 deposits (Figure 2 and Table 1) (Schirrmeister et al., 2003, 2011b; Wetterich et al., 2008a) and incorporated into Yedoma deposits during MIS 2 and MIS 3. Results from core B (Figure 18) indicate that high-energy transport mechanism within the TEV system occur episodically. In addition, according to (Ulrich et al., 2019) subsidence can further enhance the accumulation of such a coarse sediment component.

5.1.2 rEM4_{TEV}

The main mode of rEM4_{TEV} at 66.9 μm represents very fine sand to coarse silt and does not indicate a single, distinct sedimentary transport process. Instead, this grain-size range suggests a combination of transport and depositional mechanisms operating within the fluvial sedimentary system. Comparable grain-size modes occur in mixed sediments derived from reworked loess deposits, where aeolian material is subsequently transported and redeposited by fluvial processes (Sun et al., 2002). In particular, Sun et al. (2002) describe mixed sediments with a mode at around 60 μm , which closely matches the main mode of rEM4_{TEV}. Such mixed signals are consistent with environments affected by variable discharge, temporary flooding, and sediment reworking.

The main mode of rEM4_{TEV} also closely matches rEM4_{Yedoma} identified by Schirrmeister et al. (2020). These authors interpret this endmember as representing fine-grained overbank deposits and settled suspended load in temporarily flooded areas with low flow velocities. They further highlight the role of repeated frost weathering in producing and enriching fine-grained sediment fractions (Schirrmeister et al., 2020; Viran & Binal, 2018). Experimental studies demonstrate that seasonal freeze-thaw cycles promote grain disintegration and enrichment of fines in quartz-rich sediments (Schwamborn et al., 2012), a process likely to operate in ice-rich Yedoma and TEV environments.

The pronounced contribution of rEM4_{TEV} to valley floor sediments, combined with its weaker representation in Yedoma upland and slope samples, suggests that this endmember reflects relocated and sorted material during downslope and downstream transport. Different transport mechanisms selectively redistribute fine sand to coarse silt fractions within the fluvial sedimentary system (Middleton, 1976; Sun

et al., 2002). As the main mode and overall shape of $rEM4_{TEV}$ do not correspond to a single fluvial transport process, the signal likely integrates multiple mechanisms related to thermo-erosional reworking and overland flow under permafrost thaw conditions. Although reference GSDs for thermo-erosional transport are not well constrained, such processes provide a plausible geomorphological context for the observed grain-size characteristics. The shape and main mode of $rEM4_{TEV}$ also match a widely distributed robust endmember associated with Holocene thermokarst infilling in Central Yakutia, Siberia (Ulrich et al., 2019). In addition, postdepositional cryogenic processes likely enhance the relative abundance of fine-grained material at the valley floor, as indicated by high ALT and shallow WTD (Figure 14). Together, these observations support an interpretation of $rEM4_{TEV}$ as a product of fluvial and thermo-erosional reworking and cryogenic modification of relocated Yedoma-derived material.

5.1.3 $rEM3_{TEV}$

The coarse-silt main mode of $rEM3_{TEV}$ at $24.1\ \mu m$ indicate low-energy sediment transport and deposition. Nevertheless, the main mode and GSD curve of the rEM do not perfectly resemble one of the exemplary samples published by Sun et al. (2002), but depict a similar main mode as a suspended load component in lacustrine sediments with a range of 25 T to 0 The comparison to the Yedoma samples (Schirrmeister et al., 2020) showed that $rEM3_{TEV}$ falls within the main mode range of $rEM6_{Yedoma}$, which they interpret as a fluvial suspension component. Accordingly, $rEM3_{TEV}$ can reflect deposition from suspension under hydraulic conditions characterised by reduced flow velocities. The influence of storms might explain the secondary main mode greater than $200\ \mu m$ in the GSD, just like in lacustrine sediments (Sun et al., 2002). Nevertheless, since this coarse secondary mode is placed directly below the main mode of the $rEM5_{TEV}$ secondary mode, it is more likely that it represents a spurious secondary mode, than an independent transport process such as a storm signal.

I conclude that $rEM3_{TEV}$ represents a fluvial or alluvial suspension component that settles when shields stress falls below a critical threshold, for example in ponded water or during waning flow stages (Dietrich, 1982). During the Holocene, I consider fluvial snowmelt- and overland-flow environments as the dominant settings for deposition from suspension. In addition, settling suspension load in a Late Pleistocene alluvial environment (Minprirody, n.d.; Pertiwi et al., 2021) may be reflected in reworked Yedoma deposits. Furthermore, $rEM3_{TEV}$ may include a contribution from coarse regional dust that can remain in atmospheric suspension for several days during storm events (Schirrmeister et al., 2020). This interpretation is most consistent with the Yedoma upland, where samples extend back to the Early Holocene.

5.1.4 $rEM2_{TEV}$

The irregular shape of the distribution curve and the absence of a smooth fine-grained tail suggest that $rEM2_{TEV}$ integrates multiple fine-grained components rather than representing a single transport process. The silt-sized main mode of $rEM2_{TEV}$ at $18.7\ \mu m$ indicates sediment transport and deposition dominated by suspension. Comparable GSDs occur in lacustrine sediments composed of overlapping suspended-load components. For example, Sun et al. (2002) describe lake sediment samples collected at shallow water depths that consist of two superimposed silt fractions, both deposited from suspension. A similar superposition of suspended-load components with varying proportions likely contributes

to the composite shape of $rEM2_{TEV}$. Although the main mode of $rEM2_{TEV}$ falls within the grain-size range of $rEM5_{Yedomas}$ (Schirrmeister et al., 2020), differences in curve shape preclude a direct comparison between the TEV and Yedomas endmembers.

I therefore interpret $rEM2_{TEV}$ as representing sediment deposited from suspension under low-energy hydraulic conditions, such as in ponded water or during phases of reduced flow velocity (Dietrich, 1982; Schirrmeister et al., 2020). Particularly at the Yedomas upland, this rEM may additionally include an aeolian component that contributes fine-grained material subsequently reworked and redistributed downslope and along the valley floor during the Holocene.

5.1.5 $rEM1_{TEV}$

The fine-silt main mode of $rEM1_{TEV}$ at $5.92\text{ }\mu\text{m}$ indicates deposition from suspension under very low-energy conditions. This grain-size range matches $rEM8_{Yedomas}$ identified by Schirrmeister et al. (2020), which they associate with several potential sedimentary origins. These include long-term fluvial suspension in larger streams or repeated meltwater runoff with sufficient still-water conditions for fine particles to settle, as well as aeolian or mixed aeolian–fluvial processes involving background dust. Within the geomorphological and environmental setting of the studied TEV, rather overland flow and meltwater runoff linked to permafrost thaw provide a relevant context for such suspension-driven deposition (Figures 11 and 13; Appendix Figure 29).

In addition, Schirrmeister et al. (2020) discuss a possible contribution from pedogenic clay deposited in ice-wedge polygon ponds and subsequently reworked. Based on the available grain-size data alone, these potential origins cannot be uniquely distinguished for $rEM1_{TEV}$. Accordingly, this endmember likely integrates multiple fine-grained suspension components rather than representing a single, distinct depositional process.

The regional setting of the Lena River Delta and Kurungnakh Island does not indicate the presence of nearby desert environments during the Holocene or Late Pleistocene. Therefore, an origin from regionally transported desert dust appears unlikely. Instead, $rEM1_{TEV}$ most plausibly reflects locally sourced aeolian material derived from riverbanks and floodplains of the Lena River Delta and subsequently redistributed within the TEV by fluvial suspension. This interpretation is consistent with the widespread occurrence of $rEM1_{TEV}$ across geomorphological positions within the valley.

5.2 Yedomas upland – core C

To test H1, I evaluate whether the stratigraphy at the Yedomas upland preserves largely undisturbed Yedomas deposits with minimal postdepositional modification. This assessment is based on (i) sedimentological characteristics and GSDs as indicators of primary Yedomas deposition, (ii) cryostratigraphic features, radiocarbon ages, and geochemical proxies constraining the timing of deposition and the influence of postdepositional processes, and (iii) distributed measurements indicating current permafrost conditions.

5.2.1 Stratigraphic resemblance to Yedoma deposits

Typical Yedoma sediments consist of poorly sorted silt to fine sand (Grosse et al., 2007; Schirrmeister et al., 2011b), with occasional contributions of medium sand and, more rarely, coarse sand (Schirrmeister et al., 2020). GSDs of core C show generally poor sorting with a dominant silt fraction and a subordinate fine-grained component (Figure 21a). In addition, rEMMA results indicate a high contribution of rEM5_{TEV} within the upper sediment section of core C (Figures 22), with its main mode in the medium-sand fraction. Together, the GSD and rEMMA results match the expected sedimentological characteristics of Yedoma deposits on Kurungnakh Island (Grosse et al., 2007; Schirrmeister et al., 2011b, 2020).

High TOC contents further support the presence of Yedoma-derived material (Grosse et al., 2007; Schirrmeister et al., 2011b; Strauss et al., 2013, 2017). Although TOC values fluctuate close to the active layer, core C is overall organic-rich, particularly at greater depth within the lower sediment section. This division into two sediment sections (Figure 15) is consistent with the stratigraphic boundary identified by Nitzbon et al. (2022) based on CT scan analysis. The observed TOC pattern resembles the transition between unit IV (8000 yr BP) and the underlying Yedoma Ice Complex (unit III, 17 000 yr BP) described by Wetterich et al. (2008a) at approximately 29.5 m a.r.l.. However, radiocarbon ages indicate that the analysed deposits do not extend into the Late Pleistocene.

A study of TOC in near-surface TEV stratigraphy in the Lena River Delta, including sites on Kurungnakh Island, reported decreasing TOC values with depth on the Yedoma upland (Kuhl, 2018). In contrast, the cores analysed in this thesis exclude active-layer material. The differing TOC trends likely reflect the investigated depth intervals and the influence of postdepositional processes. The following Chapter 5.2.2 deepens the interpretation on such processes based on cryostratigraphy, radiocarbon and hydrochemical data. The absence of TOC/TN data for the Yedoma upland prevents further assessment of organic-matter decomposition. Generally, I would expect high TOC/TN ratios at the Yedoma upland, where vascular-plant organic matter accumulates and cold, oxygen-limited conditions promoted carbon preservation and slow decomposition (Haugk et al., 2022; Lin et al., 2025; Treat et al., 2023; Wetterich et al., 2008a).

Yedoma deposits are typically ice-rich and characterised by massive ground ice and diverse cryostructures (Grosse et al., 2007; Schirrmeister et al., 2011b). Core C reflects these characteristics through the presence of a massive ice body interpreted as wedge ice and various ice-rich microstructures such as layered, reticulate, and lenticular cryostructures. Although halved core images are unavailable for massive ice of core C, limiting detailed structural interpretation, the observed cryostructures are consistent with Yedoma Ice Complex characteristics. The isotope values of the wedge ice from the Yedoma upland and within and surrounding the sediment lens at the slope (A003) match the wedge ice of the MIS 2 and MIS 3 on Sobo-Sise (Wetterich et al., 2020), the predominant range of MISs for Yedoma deposits (Schirrmeister et al., 2011a, 2011b).

5.2.2 Holocene development and processes

Despite sedimentological, cryostratigraphic, and carbon characteristics consistent with Yedoma deposits, radiocarbon ages indicate Holocene deposition. Only the deepest dated sample of the lower sediment section, coinciding with the shallowest dated ice-wedge sample, almost dates back to 11 700 yr BP. Accordingly, core C corresponds to units IV and V (Wetterich et al., 2008a). Holocene deposits com-

monly overlie the Yedoma Ice Complex, and their thickness varies with local relief (Figure 2), preferentially accumulating in topographic depressions (Morgenstern et al., 2013b; Schirrmeister et al., 2011b; Wetterich et al., 2008a). The term *Yedoma upland* therefore refers to the geomorphological position of core C, whereas the recovered sediments predominantly represent the Holocene cover at this location.

Radiocarbon dating results and stratigraphy of core C provide insight not only into the timing of deposition but also into postdepositional processes. A pronounced age inversion, together with ice-rich samples and reticulate, layered, and lenticular cryostructures in the upper sediment section and organic-matrix cryostructures in the lower section (Figure 15), indicates substantial postdepositional reworking, most likely by cryoturbation. This interpretation is supported by the clear distinction between the two sediment sections in terms of GSDs and rEM contributions. Neither active fluvial processes nor major aeolian sediment sources were present at the Yedoma upland during the Holocene that could explain the representation of the coarse fluvial rEM5 within the upper sediment section. As discussed in Chapter 5.1.1, Yedoma deposits can incorporate reworked material, including fluvial sands reworked into Yedoma deposits and Yedoma material subsequently reworked into Holocene sediments. In addition to the inverse age records between the upper and lower sediment sections, coarse cryostructures coincide with pronounced cryoturbation processes during the Mid Holocene, when the Arctic experienced warming (Bockheim, 2007). Thus, postdepositional mixing modified the original sedimentary signal. Although solifluction can influence grain-size composition in periglacial environments (Schirrmeister et al., 2020) and would match the clear distinction between two sections, the very low slope angles and weakly developed ice-wedge polygonal networks at the Yedoma upland make solifluction an unlikely process.

EC data further constrain postdepositional processes. Low EC values near the surface reflect infiltration by meltwater and precipitation, whereas EC values increase with depth toward groundwater characteristics identified in the Lena River Delta (Juhls et al., 2020). Maximum EC values in core C exceed $500\ \mu\text{S}/\text{cm}$, approaching the groundwater endmember of approximately $600\ \mu\text{S}/\text{cm}$ (Juhls et al., 2020). Near-surface and specifically active-layer stable water isotope values correspond to summer (June, July and August) and autumn (September, October and November) precipitation values in the Lena River Delta (Bonne et al., 2020), indicating exchange between pore, excess and surface water. Together with the comparatively high ALT and high WTD measurements at the well-drained Yedoma upland (Appendix Figure 14) (Lisovski et al., 2025), these observations suggest that the lower sediment section remained frozen over long periods, while the upper sediment and active layer experienced a more recent thawing and hydrological exchange. This confines the Holocene postdepositional processes to mainly cryoturbation and cryogenic weathering.

5.2.3 Conclusion on the Yedoma upland stratigraphy and processes

I therefore reject H1 that core C represents largely undisturbed Yedoma deposits. Instead, Holocene deposits affected by postdepositional processes overlie the Yedoma Ice Complex. Yedoma material is locally reworked into the Holocene cover, which is itself strongly modified by cryoturbation. Beneath this cover, syngenetic ice wedges and ice-rich sediments indicate that the underlying Yedoma Ice Complex remains largely intact. The well-drained upland position and the presence of ice wedges minimise an Holocene impact of thermal erosion.

5.3 Valley slope – cores A001 & A003

According to H2, predominantly erosional processes characterise the stratigraphy of the valley slope, resulting in erosional gaps and limited soil formation. To test this hypothesis, I analysed the slope cores for (i) erosional features such as gaps or hiatuses in radiocarbon ages and in the stratigraphy of GSDs and rEM contributions, and (ii) indicators of depositional environments and soil-forming processes.

5.3.1 Erosional stratigraphic evidence

Both valley slope cores A001 (Figure 16) and A003 (Figure 17) exhibit a thin sediment cover overlying massive wedge ice. In particular, the CT scan of core A001 (Figure 16) highlights the sharp contrast between the thin sediment section and the underlying ice. The thickness of the sediment cover differs between the two cores, which likely reflects the uneven hummocky tundra surface at the valley slope (Figure 5d).

Radiocarbon dating further illustrates differences between the two slope cores. While core A001 contains two samples of similar age within the sediment section, core A003 shows a strong increase in age with depth. Despite a vertical distance of only 9 cm between the dated samples in core A003, the age difference amounts to 6917.5 cal yr BP (Table 3). Although the radiocarbon ages are discontinuously sampled, most sedimentological parameters vary smoothly without abrupt shifts, the presence of Late Holocene sediments directly above massive wedge ice in core A001 indicates localised thermo-erosional removal of the sediment cover. On the other hand there are not specific indicators for an erosional event between 7215 cal yr BP to 297.5 cal yr BP in core A003. This suggests that erosion acts at a very local scale on the valley slope, rather than producing laterally extensive erosional surfaces.

5.3.2 Depositional environment and soil formation

In core A003, TOC values and TOC/TN ratios decrease rapidly with depth. Very high TOC values close to the active layer indicate high organic productivity near the surface, whereas low TOC values near the ice wedge, combined with much older radiocarbon ages, suggest less favourable conditions for organic matter storage at the valley slope during the mid-Holocene around 7162 cal yr BP. The with depth decreasing TOC/TN ration support this interpretation. Erosion and relocation of vegetation can additionally influence TOC values at the slope. Exposure of deeper sediment layers through thermo-erosion promotes aeration and microbial activity, leading to organic matter degradation (Obu et al., 2017). Enhanced microbial decomposition in thawed sediments further supports this interpretation (Walz et al., 2017). This process results in lower TOC and TOC/TN values at depth, which are consistent with earlier observations from TEV slopes on Kurungnakh Island (Kuhl, 2018) and indicate limited soil formation under predominantly erosional conditions.

ALT at the slope is only slightly higher, and WTD only slightly lower, than at the Yedoma upland (Figure 14), indicating well-drained conditions consistent with the local geomorphology (Figure 5d and 13b). According to Gagnon et al. (2024), increased thawing and ALT at slopes typically occur during the initial stages of TEG development, whereas subsequent vegetation colonisation reduces thermal disturbance by limiting air convection and solar radiation. A thicker active layer increases the volume of thawed, mobile sediment susceptible to erosion, supporting the interpretation of ongoing slope instability.

GSDs and rEM contributions at the slope closely resemble those of the Yedoma upland. Both settings are characterised by poorly sorted sediments and include all five rEMs_{TEV}, indicating relocation of material from the Yedoma source region. Compared to the valley floor, rEM5_{TEV} contributes a relatively high proportion at the slope, whereas rEM1_{TEV} occurs only in a single sample from core A003. These contrasts among upland, slope, and valley floor sediments highlight the role of sorting processes during transport, likely reflecting different fluvial and gravitational transport modes that produce hydraulically polymodal sediments (Ashley, 1978; Middleton, 1976; Sun et al., 2002; Visser, 1969). Coarser grains transported predominantly by saltation tend to be deposited at the slope, indicating shorter transport pathways (Visser, 1969), whereas overland flow or fluvial processes transported finer material further downslope or into valley-floor and channel environments. Field measurements in the Lena River Delta show increased release of suspended and dissolved material from Ice Complex deposits during summer (Fedorova et al., 2015), reflecting efficient transfer of thawed material into deltaic channels and ultimately into the Laptev Sea.

Cryogenic processes further shape the slope stratigraphy. Layered cryostructures with high visible ice content in core A001 (Figure 16) and reticulate cryostructures near the ice wedge in core A003 (Figure 17) indicate freeze-thaw dynamics. Moreover, I interpret the massive ice of both cores as wedge ice. Yellowish-grey ice wedges with well-defined subvertical elementary ice veins and thin mineral layers are typical for the Yedoma Ice Complex (Meyer et al., 2002). However, whereas Late Pleistocene ice wedges typically preserve rounded gas bubbles (Boereboom et al., 2013; St-Jean et al., 2011), the ice wedges at the valley slope contain elongated bubbles, some exceeding 1 mm in length (Figure 8b), which are characteristic of Holocene ice wedges. Ice crystallography alone therefore does not allow a definitive age assignment.

Stable water isotope compositions provide additional context for ice-wedge formation. Late Pleistocene ice wedges are typically isotopically depleted, reflecting colder winter temperatures, whereas Holocene ice wedges show more enriched values associated with warmer conditions (Boereboom et al., 2013; Meyer et al., 2015; Opel et al., 2018; St-Jean et al., 2011). Relatively enriched isotope values in core A001, sampled near the top of the ice wedge directly beneath the Holocene cover, correspond to MIS 1 ice-wedge signatures reported from Sobo-Sise Island (Wetterich et al., 2020). Similarly, isotope values from the sediment lens in core A003 resemble enriched Holocene ice-wedge signatures from TEVs on Bol'shoi Lyakhovsky Island (Meyer et al., 2002). Furthermore, the still active ice wedges on Bol'shoi Lyakhovsky Island display elongated gas bubbles and subvertically oriented ice veins. These findings suggest that ice-wedge growth at the valley slope persisted into the Holocene.

5.3.3 Conclusion on the valley slope stratigraphy and processes

Overall, the thin sediment cover and high radiocarbon ages at shallow depths indicate limited deposition and soil formation at the valley slope, consistent with predominantly erosional conditions. While discontinuous radiocarbon sampling prevents high-resolution reconstruction of slope stability throughout the Holocene, dated samples at 7162 cal yr BP and 297.5 cal yr BP point to phases of relative stability from the mid-Holocene to the present. At the same time, evidence from core A001 (Figure 16) demonstrates that thermo-erosion remains an active process at the valley slope, capable of thawing, detaching, and removing Holocene sediment cover above wedge ice.

The comparison of cores A001 and A003, both of which contain shallow ice wedges, indicates a broadly similar near-surface stratigraphy along the west–southwest-facing valley slope, with varying locally varying age records in the Holocene cover. Although the stratigraphy of the opposing slope remains uncertain, the regional geological setting of Yedoma deposits on the third terrace of Kurungnakh Island and the ArcticDEM Mosaic topography (Figure 12) suggest a largely symmetric valley geometry without pronounced lateral contrasts.

5.3.4 Sediment lens in core A003

As described in Chapter 4.3.3, the sediment lens embedded within the ice wedge of core A003 exhibits characteristics that clearly distinguish it from the surrounding slope sediments and from all other analysed sediment samples. Stratigraphic records from the Lena River Delta frequently contain hiatuses during the Late Pleistocene and early Holocene. Based on cryostratigraphic analyses from Sobo-Sise Island, Wetterich et al. (2020) identified hiatuses between 36 kyr BP to 29 kyr BP, 20 kyr BP to 17 kyr BP, and 15 kyr BP to 7 kyr BP, which they relate to shifts between depositional and erosional regimes. Against this background, the sediment lens in core A003 potentially represents a rare archive of sedimentary processes otherwise poorly preserved in the regional stratigraphic record. This archive spans a wide age range between 22 220 cal yr BP to 11 122.5 cal yr BP within a thin sediment lens composed of coarse, comparatively well-sorted material. An outlier age caused by sample contamination is unlikely, as I scraped off the outer surfaces of the samples prior to laboratory analyses. This raises the question of the depositional origin and environmental context of the sediment lens. In the following, I discuss three possible interpretations.

A first approach compares the sediment lens with established stratigraphic units of the Lena River Delta. Based on its mean grain size and very low TOC values, the sediment lens resembles unit II as described by Wetterich et al. (2008a), which consists of yellow, medium-grained, homogeneous sand exposed north of the Olenyoksky Channel. Both the analysed outcrops on Kurungnakh Island and the sediment lens in core A003 consist of clean, organic-poor sand (Figure 17). However, several discrepancies argue against this interpretation. Radiocarbon ages from the sediment lens are substantially younger than the age range assigned to unit II ($>50\,000$ yr BP) (Wetterich et al., 2008a). In addition, the surface elevation of core A003 lies more than 10 m above the upper boundary of unit II. Although geological units on Kurungnakh Island are slightly tilted from northeast to southwest (Grigoriev, 1993; Morgenstern, 2012) and stratigraphic boundaries vary spatially across the third Lena River Delta terrace (Grosse et al., 2007; Khazin et al., 2019; Schirrmeister et al., 2003, 2011a; Wetterich et al., 2008a, 2020), and although the position of the sediment within wedge ice suggests potential vertical displacement during ice-wedge growth, the radiocarbon age mismatch alone is sufficient to exclude unit II as the source of the sediment lens.

A second interpretation considers the sediment lens as material that entered an open ice-wedge crack and was subsequently preserved. Sediment inclusions above and below the lens within the ice wedge (Figure 8b) support the possibility that sediment infiltrated the crack during periods of ice-wedge opening. Ice-wedge cracking is episodic and does not necessarily occur annually (Opel et al., 2018). However, given the widespread formation of syngenetic ice wedges during Late Pleistocene Yedoma accumulation (Grosse et al., 2007; Schirrmeister et al., 2011b), it is unlikely that cracking ceased for more than 10 000 cal yr BP, which could explain the wide age range recorded by the sediment lens. Instead, sed-

iment may have entered the crack during a single or a few short-lived events. Nevertheless, the well-sorted GSD and the dominance of a single coarse endmember argue against sediment infiltration from a mixed sediment cover, which would more likely produce mixed grain-size signatures.

A third interpretation links the sediment lens to high-magnitude flood events associated with glacial lake outbursts in the Vitim River basin. Arzhannikova et al. (2025) identified three major Vitim megafloods, with the youngest event occurring at approximately 21 kyr BP to 22.3 kyr BP (MIS 2). Except for the youngest of the three dated samples, radiocarbon ages from the sediment lens overlap with the timing of this third megaflood. While direct megaflood evidence in the Lena River Delta has so far been inferred mainly from erosional hiatuses rather than preserved deposits (Arzhannikova et al., 2025; Wetterich et al., 2008a, 2020), the coarse, fluvial rEM5 characterising the sediment lens may reflect deposition within the broader megaflood environmental system (Dietze et al., 2014; Schirrmeister et al., 2020). Mineralogical analyses currently underway aim to test this hypothesis further. The youngest radiocarbon age does not fit this interpretation and may reflect reworking or flushing and dissolved organic carbon input linked to ice-wedge activity. Repeat dating is planned to clarify this discrepancy. Even if the sediment lens does not represent a direct megaflood deposit, it likely records sedimentation under high-energy flood-related conditions that were exceptionally preserved within the ice wedge.

Overall, neither core A003 nor any previously published sedimentological record from Kurungnakh Island or the Lena River Delta describes a comparable sediment lens embedded within wedge ice. The absence of this feature in core A001, drilled only a few metres away, is most plausibly explained by an inclined geometry of the sediment lens within the wedge ice. A more horizontal feature would likely have been intersected by both cores. At present, the available data do not allow a definitive conclusion in favour of one interpretation, and further analyses are ongoing.

5.4 Valley floor – core B

5.4.1 Depositional continuity and postdepositional disturbance

To test H3, I evaluate whether the valley-floor stratigraphy records alternating phases of erosion and subsequent sediment infilling. I base this assessment on (i) the radiocarbon age and cryostratigraphy over the depth as an indicator of depositional continuity and reworking of the deposits due to postdepositional processes, (ii) sedimentological evidence for depositional events and environments, and (iii) hydrochemical and stable isotope signatures that constrain freezing and water influence during and after deposition. Together, these proxies allow me to assess depositional continuity, reworking, and the preservation of erosion-infilling cycles at the valley floor.

Radiocarbon ages indicate that the valley floor consists predominantly of Late Holocene deposits down to 185 cm depth, with the exception of a single outlier in the upper sediment section. In contrast to the Yedoma upland and slope sites, which preserve early Holocene and Late Pleistocene material, this young age range suggests that the valley floor mainly records recent sediment infilling, while older material is present only as reworked components.

Age inversions in sediment sections two and three indicate postdepositional disturbance, whereas sections one and four show comparatively consistent age-depth relationships, apart from the discrete

outlier. Cryostratigraphic observations support this interpretation, since coarser layered and organic-matrix cryostructures, as well as locally inclined bedding, occur mainly in sections two and three (Figure 18) (French & Shur, 2010). The Inclined layers (Figure 18) cannot be linked to specific movement directions because the CT scans represent random longitudinal cuts through the core. Furthermore, the stratification does not indicate cross-bedding, which would suggested a clear idea of relevant fluvial processes and the depositional environment (Wetterich et al., 2008a). The association of age inversions with coarse and tilted cryostructures points to cryoturbation as a key mechanism overprinting the depositional sequence in the middle part of the core. Consequently, discrete erosion–infilling phases cannot be resolved unambiguously in the middle part of the core. In contrast, intact sediment layers in section one and fine cryostructures in section four indicate a comparatively minor postdepositional overprint.

Distinct temporal hiatuses are not clearly identifiable in the valley-floor record. Age inversions and postdepositional processes further complicate their detection. While specific erosional phases cannot be dated or isolated, erosion and reworking remain plausible components of valley-floor stratigraphy formation, consistent with H3.

5.4.2 Sediment infilling processes and depositional environments

As described in chapter 4.3.4 and listed in table 3, a pronounced radiocarbon age outlier of 5595 cal yr BP occurs within the first sediment section, while the surrounding samples date back to 182 cal yr BP. This pattern indicates redeposition of older material within an otherwise young stratigraphic context and highlights the strong influence of depositional processes at the valley floor. The absence of age offsets in the samples above and below the outlier suggests that this event did not involve erosion of the surrounding valley-floor deposits.

Additional stratigraphic parameters support the interpretation of a discrete depositional event. Around the depth of the outlier, GSDs shift toward *very fine to fine sand*, and rEM5_{TEV} appears exclusively in samples between 30 cm to 35 cm of core B (Figure 18). For the interpretation of this depositional event, I considered two scenarios. The first scenario links the deposition of this coarse, fluvial GSDs and rEM to a flood, reaching the valley floor of the TEV. Nevertheless, a flood-driven origin is unlikely, as modelling of Lena River spring floods indicates maximum flood heights of 12 m at the Khabarova gauging (The Arctic Great Rivers Observatory ArcticGRO, 2024) station, with further flattening downstream (Pertiwi et al., 2021). Moreover, no sedimentological evidence for flood-induced erosion at the valley floor is present. A more plausible scenario involves short-lived meltwater stream, an event strong enough to mobilised coarse sediment and organic matter from upslope source areas, adjacent to the Yedomia upland. The fluvial transport then deposited coarse grains at the valley floor without eroding the underlying deposits. This interpretation is consistent with episodic sediment delivery associated with thermo-erosional activity (Morgenstern et al., 2021b).

Throughout core B, high TOC values and elevated TOC/TN ratios indicate substantial organic matter input and comparatively well-preserved plant detritus. Fluctuations along the profile (Figure 4.3.4) likely reflect shifting flow paths at the valley floor and associated changes in vegetation growth conditions and sediment supply, similar to interpretations by Wetterich et al. (2008a). TOC/TN ratios fall within the documented range for Lena River Delta deposits (Lin et al., 2025; Wetterich et al., 2008b) but

exceed values reported from lake and pond sediments on Samoylov and Kurungnakh Island (Wetterich et al., 2008b). Comparable high ratios in MIS 1 deposits from Sobo-Sise support this interpretation (Haugk et al., 2022). Furthermore, perennial snowfields and subsequent snowmelt can add to the accumulated of not only mineral debris but also plant detritus (Schirrmeister et al., 2011b). The generally high TOC/TN ratios indicate cold conditions with limited organic-matter degradation, which I attribute to rapid burial and subsequent freezing at the valley floor, shortening microbial processing times and promoting organic-matter preservation (Herbst et al., 2024).

GSDs and EMMAgeo results provide independent evidence for alternating depositional energy regimes at the valley floor. Variations in rEM contributions reflect changing runoff intensity and transport pathways (Wetterich et al., 2008a). Intervals dominated by fine-silt rEM1_{TEV} alternate with phases characterised by medium-sand rEM5_{TEV}, indicating strongly variable transport mechanisms (Figure 18) (Dietze & Dietze, 2019; Schirrmeister et al., 2020; Sun et al., 2002). In High Arctic non-glacial catchments, snow water equivalent and melt-season air temperatures exert primary control on suspended sediment transfer (Cockburn & Lamoureux, 2008; Macumber et al., 2018). Thus the varying sedimentary environments link to changing climate conditions and seasons. Fluctuating runoff conditions and the meandering of small streams at the valley floor likely contributed to variable depositional environments and changing WTD and ALT conditions.

Hydrochemical parameters provide further context for water influence during and after deposition. EC increases slightly with depth, while pH remains stable around 5.0. Overall, the EC values fall within the range of precipitation and snowmelt endmembers in the Lena River Delta (Juhls et al., 2020). Thus, the depth-related EC increase likely reflects longer water-sediment contact times in deeper deposits (Drever, 1997) combined with limited flushing by recent meteoric water cryogenic concentration of solutes in persistently frozen layers (Schirrmeister et al., 2011b). These trends are consistent with solute enrichment in older frozen sediments and limited flushing, supporting the interpretation of long-term frozen conditions after deposition.

Stable water isotope values in core B show low variability and broadly match present-day precipitation, indicating a comparatively homogeneous pore-water signal. The continuous availability of isotope data over the entire depth of core B and the absence of wedge ice with its characteristic isotope signatures (Opel et al., 2018) further enables these smooth profile. Overall, the isotope values correspond well to mean autumn precipitation measured at Samoylov Island between 2015 and 2017, with mean values of -19.1 and -147.4 (Bonne et al., 2020).

Based on this smooth isotope distribution, freezing-related fractionation patterns, such as top-down freezing reflected by decreasing isotope values with depth, could be expected in sediment sections one and two. However, isotope values from the upper three sediment sections (Figure 18) do not show significant freezing fractionation, as neither slopes exceeding the threshold of 7.3 nor a strong negative correlation between d-excess and δ^2 are observed (Lacelle, 2011). This absence of a clear freezing signal is likely related to the proximity of these samples to the active layer, supported by high ALT and low WTD measurements along the valley floor (Figure ??). Seasonal infiltration of precipitation and snowmelt likely caused repeated mixing of porewaters, thereby overprinting any primary isotope signals. Consequently, distinct depositional water signatures are not preserved in the upper parts of the core.

On the other hand, the results presented in Figures 19 and 20 indicate freezing fractionation due to bottom-up freezing within sediment section core, specifically between 73 cm to 130 cm and 130 cm to 168 cm. These patterns indicate two separate events of deepening of the ALT and subsequent bottom-up freezing within one season. Since soil commonly freezes top-down in permafrost soils, this trend seems unintuitive. Nevertheless, deviations from typical top-down freezing patterns could relate to conditions of ponding water or lateral water supply at the surface linked to changes in the WTD and ALT, creating complex freezing geometries.

In contrast, the results from sediment section four (Figures 19 and 20) suggest freezing-related isotope fractionation consistent with bottom-up freezing, particularly between 73 cm to 130 cm and 130 cm to 168 cm (Lacelle, 2011). These patterns are consistent with at least two phases of active-layer deepening followed by freezing. Because permafrost soils typically freeze from the surface downward, such bottom-up freezing behaviour appears unusual. However, deviations from idealised top-down freezing can occur under conditions of sustained surface saturation, ponding water, or lateral water supply linked to changes in water-table depth and active-layer thickness, resulting in complex freezing geometries.

5.4.3 Conclusion on the valley floor stratigraphy and processes

The limited length of core B restricts the reconstruction window and prevents a clear identification of the onset of TEV development. Core B does not contain basal coarse accumulation material overlain by finer alluvial sediments, as described by Gagnon et al. (2024) for the initial stages of TEG development. Consequently, the timing of valley initiation cannot be determined from the analysed profile. Borehole data from a 17 m long core on Kurungnakh Island suggest that thick Holocene deposits extend below the recovered depth (Khazin et al., 2019), indicating that deeper coring would be required to assess whether Early to Mid Holocene sediment infilling is preserved beneath core B.

Based on the combined evidence from radiocarbon ages, cryostratigraphy, sedimentology, and geochemical proxies, I conclude that core B records Late Holocene valley-floor sedimentation that is locally modified by postdepositional reworking. Fluvial sedimentation dominates deposition, primarily reworking material eroded from upslope Yedoma deposits. Since approximately 800 cal yr BP, pedogenic and cryogenic processes have increasingly overprinted the sedimentary record. Variations in GSDs, TOC values, and TOC/TN ratios reflect changing transport energies and environmental conditions within a dynamic valley-floor setting. Overall, core B partly supports H3 by documenting sediment infilling of the valley floor with reworked Yedoma material, while distinct erosional hiatuses cannot be robustly resolved.

5.5 Overall thermo-erosional valley development and system resilience

RQ2 and H4 address whether the TEV represents a Holocene landform that is resilient to minor disturbances but sensitive to hydrological and permafrost thresholds. To evaluate H4, I synthesise results from the Yedoma upland, valley slope, and valley floor cores to assess how erosion, sediment transfer, deposition, and postdepositional processes interact within the TEV. The following sections (i) conceptualise the TEV as an geomorphic system, (ii) discuss three plausible scenarios for its developmental

onset, (iii) evaluate system resilience and threshold sensitivity in a broader permafrost context, and (iv) outline limitations of the dataset and interpretations.

5.5.1 Thermo-erosional valley as an interconnected system

I conceptualise the TEV as an open geomorphic system controlled by external climatic forcing and internal boundary conditions imposed by Yedoma and underlying sand deposits (Schirrmeister et al., 2011b; Wetterich et al., 2008a). Figure 23 summarises the system structure and dominant processes, integrating stratigraphic evidence and interpretations from Chapters 5.1–5.4 with field observations and published stratigraphic frameworks (Khazin et al., 2019; Schirrmeister et al., 2003; Wetterich et al., 2008a, 2020). Within this framework, sediment redistribution is primarily driven by spring snowmelt, permafrost thaw, and precipitation leading to overland flow. Although aeolian input may contribute fine-grained material (Dietze et al., 2014; Schirrmeister et al., 2011b; Sun et al., 2002), the sedimentary signal at the study sites predominantly reflects reworked Yedoma material, making it difficult to distinguish Holocene aeolian additions from fine particles generated by cryogenic weathering and reworking (Schwamborn et al., 2012).

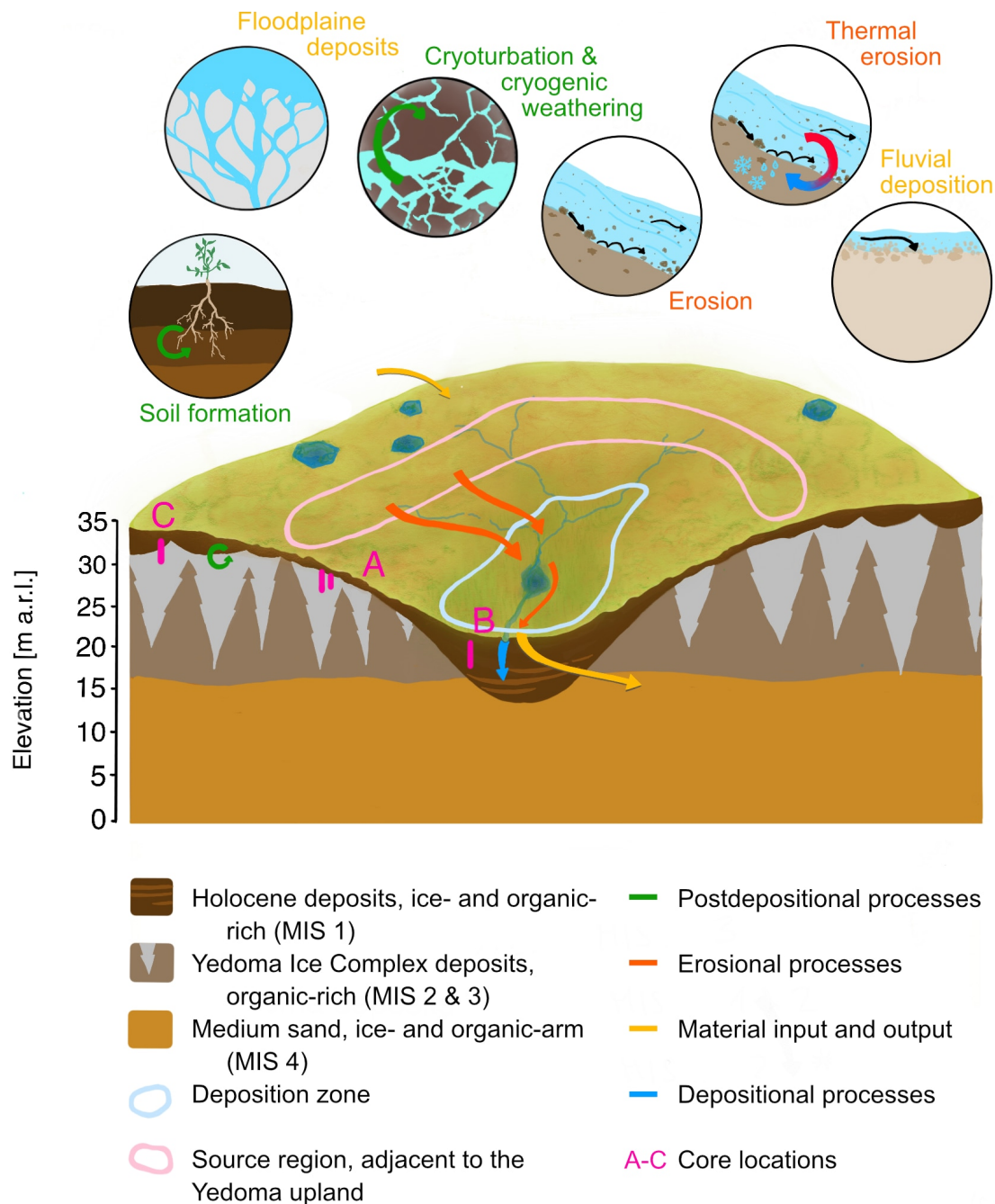


Figure 23: Simplified illustration of the thermo-erosional valley on Kurungnakh Island. The stratigraphy includes Holocene deposits (MIS 1), underlying Yedoma deposits with its syngenetic Ice wedges (MIS 2 & 3) and underlying sand deposits (MIS 4). Floodplaine deposits depict the origin of aeolian sedimentary input into the system. Pedogenic processes and cryoturbation and cryogenic weathering represent major postdepositional processes (green) while erosion, thermal erosion (dark orange) and fluvial deposition (blue) depict further highly relevant processes for the near surface stratigraphy and TEV development. The source zone is marked in light pink and the depositional zone in light blue.

Observed variability in sedimentological and biogeochemical parameters (GSD, rEM contributions, TOC, and TOC/TN) highlights the importance of fluvial and thermo-erosional processes within the TEV. Spatial differences in endmember contributions constrain sediment transfer pathways. Higher rEM_{5TEV} contributions at the Yedoma upland and valley slope compared to the valley-floor sites indicate limited transport capacity for coarse material, whereas low clay proportions suggest efficient export of fine sediment out of the system. This spatial organisation reflects a coupling between sediment mobilisation in

the source region and selective deposition at the valley floor, while fine material and dissolved components are transported further downstream. In this way, TEVs link ice-rich permafrost terrain to surface drainage networks and downstream sediment, hydrological, and biogeochemical fluxes (Lafrenière & Lamoureux, 2019; Morgenstern et al., 2021b; Zolkos et al., 2020).

5.5.2 Thermo-erosional valley development timeline

The analysed core data do not constrain the onset or full development period of the TEV (Chapter 5.4). I therefore consult published evidence on thermokarst and thermo-erosional features in the Lena River Delta and discuss three plausible onset scenarios in combination with the near-surface stratigraphic evidence from the TEV cores. The scenarios start an initiation (i) during the Late Pleistocene, (ii) around the Late Pleistocene to Holocene transition, and (iii) during the Mid to Late Holocene.

The first scenario assumes TEV initiation during the Late Pleistocene. In this case, thermal erosion and sediment relocation would have begun while Yedoma formation and ice-wedge aggradation continued within MIS 2 and into the onset of MIS 1. This scenario is broadly consistent with the wedge-ice evidence, since stable isotope values from the Yedoma upland core C do not reach the strongly depleted signatures typical for Late Pleistocene ice wedges (Meyer et al., 2002, 2015; Opel et al., 2018).

The second scenario links TEV development to the Late Pleistocene to Early Holocene transition, analogous to the onset of thermokarst activity. On Kurungnakh Island, Morgenstern et al. (2013b) dated initial thermokarst lake and basin development to 13 kyr BP to 12 kyr BP and documented drainage of a thermokarst lake at around 5.7 kyr BP due to the development of a TEV (Morgenstern et al., 2013b), indicating that thermo-erosional incision shaped the island's geomorphology by at least this time. In the analysed TEV, the oldest dated sediment above the ice wedge at the valley slope (core A003) is 7215 cal yr BP, which suggests that slope conditions were already established by the Mid Holocene. However, the thermokarst-lake sequence described by Morgenstern et al. (2013b) does not necessarily represent the development history of the analysed TEV.

The third scenario proposes a Mid to Late Holocene onset, after 7215 cal yr BP. This would imply that an extensive Yedoma upland remained stable through the Early Holocene in the area of the present TEV, with continued ice-wedge growth producing isotopic signatures compatible with Early Holocene conditions (Opel et al., 2018; St-Jean et al., 2011). In this case, the 7215 cal yr BP slope deposits would either have to represent reworked material deposited with and age signal from its origin, or initial valley formation would have been dominated by subsidence rather than thermo-erosional sediment transfer.

While, the goal of these discussed scenarios remains to determine the timeframe for the development onset, the TEV evidence does not reveal one verified scenario. The third scenario can only be explained by the impact of subsidence and weak thermal erosion at the valley slopes. While a weak effect of thermal erosion remains plausible based on the thin sediment cover at the slopes, a strong impact of subsidence without relocation of material from Source region to the slope and from the slope to the valley floor does not fit with the previously conclusions on the stratigraphic parameters and related processes. Further, the magnitude of the cored sediment section is several times larger at the valley floor, highlighting the impact of spatial reorganisation within the valley and sediment infilling. Finally, I discard the third scenario. Both scenarios one and two match the in this thesis presented core data and radiocarbon dating records. Nevertheless, the first scenario with its Late Pleistocene initial devel-

opment is not able to specify the driver of the initial development, while the second scenario links the development to the climatic changes between MIS 2 and MIS 1. Thus, a development onset at the transition between Late Pleistocene to Holocene remains most likely and matches findings on thermokarst activity on Kurungnakh Island (Morgenstern et al., 2013b).

Overall, the available evidence does not uniquely identify one onset scenario. However, a Mid to Late Holocene initiation would require substantial subsidence with limited sediment relocation from the source region to the slopes and valley floor, which is inconsistent with the pronounced valley-floor infilling and spatial reorganisation inferred from the cores. I therefore discard the third scenario. Both scenarios (i) and (ii) remain plausible, but scenario (ii) provides a clearer driver by linking initiation to climatic and hydrological change at the MIS 2 to MIS 1 transition. Consequently, a TEV onset near the Late Pleistocene to Holocene transition appears most consistent with the core data and with regional evidence for thermokarst activity on Kurungnakh Island (Morgenstern et al., 2013b).

5.5.3 System resilience and broader implications

The young valley-floor deposits of core B reflect ongoing valley development and directly inform the resilience of the TEV to disturbance events. Discrete depositional events recorded within the last 800 cal yr BP document short-lived phases of enhanced sediment mobilisation. Such events relocate material, most likely originating from the source region adjacent to the Yedomu upland, where erosion removes records of up to at least 5595 cal yr BP. The eroded material deposits the coarse, medium-sand fractions at the valley floor, while finer sediments are transported further downstream. GSDs and elevated $rEM5_{TEV}$ contributions indicate high-energy fluvial deposition (Chapter 5.1.1) (Dietze et al., 2014; Schirrmeister et al., 2020), consistent with extreme runoff driven by snowmelt, precipitation, or permafrost thaw. Despite these disturbances, the Holocene stratigraphy at the Yedomu upland extends back to the Early Holocene (Figure 15), and GNSS and ArcticDEM profiles show a smooth valley morphology, indicating no major recent geomorphological changes.

A key mechanism promoting this resilience is the presence of ice-rich transient layers. According to Shur et al. (2005), such layers buffer permafrost against short-term climatic variability and extreme weather events by retarding degradation through the latent heat of fusion of ice (Jorgenson et al., 2010), thereby enhancing system resilience. All investigated geomorphic positions within the TEV contain wedge ice at shallow depth, providing the potential to slow degradation processes. However, surface runoff can still infiltrate open thermal contraction cracks and cavities, leading to annual thermal erosion of permafrost and ice wedges without causing sustained gullying (Gagnon et al., 2024). At the valley floor, where a distinct transient layer is absent, deeper active-layer penetration is plausible, particularly under conditions of ponding water or meandering streams. Such conditions likely contributed to the freezing patterns inferred from stable water isotope data (Figures 19 and 20).

Despite this buffering capacity of ice-rich permafrost and a transient layer, continued climate warming poses clear limits to TEV resilience. Progressive permafrost thaw leads to degradation of transient layers, loss of wedge ice, and the development of more consolidated soils (Shur et al., 2005). In ice-rich terrain, thaw-induced subsidence destabilises flat surfaces and increases susceptibility of slopes to mass wasting (Murton, 2009). Such destabilisation can trigger thermo-erosional disturbances and catastrophic slope failures (Kokelj et al., 2009; Pizano et al., 2014). However, high ground-ice content alone

does not directly determine morphological stability, as recently evolved TEGs may exhibit ground-ice contents comparable to undisturbed terrain (Gagnon et al., 2024).

Overall, the analysed TEV has remained morphologically stable since at least 7215 cal yr BP, without evidence for catastrophic mass wasting. Nevertheless, the system is sensitive to threshold behaviour. Recent observations indicate that the timing of snowmelt onset is shifting, altering hydrological regimes and runoff patterns in Arctic permafrost landscapes (Heim et al., 2022). In parallel, numerical modelling from northeastern Siberia suggests that ground surface temperatures have increased by approximately 5 °C to 9 °C since the pre-industrial period. It remains uncertain at which point these changes may exceed critical thresholds that strongly affect thermo-erosional processes and system resilience. Consequently, the TEV represents a active, semi-stable Holocene landform that has persisted under past climatic variability but remains fragile under ongoing and future warming.

5.5.4 Limitations

Several aspects limit the interpretation and upscaling of the TEV dataset presented in this study. These limitations mainly relate to the spatial representativeness of the data, methodological constraints, and chronological resolution.

First, the analysed TEV represents a specific geomorphological and geological setting on Kurungnakh Island, which constrains the direct transferability of the results to other Arctic permafrost landscapes. TEVs occur across a wide range of morphologies and hydrological contexts (Morgenstern, 2012), and the stratigraphic characteristics observed here do not necessarily represent valleys with different boundary conditions. This limitation is amplified by the restricted number of analysed cores. Although the four permafrost cores capture key geomorphological positions within the valley, they do not resolve small-scale spatial heterogeneity in sediment redistribution and postdepositional modification. Local variability is evident at the valley slope, where cores A001 and A003 differ in sediment thickness and structure while still sharing fundamental characteristics such as a shallow sediment cover. Additional and deeper cores within the studied TEV, as well as comparative datasets from parallel or radial valleys, would improve the assessment of spatial controls on TEV development. In this context, the supplementary data from core D at the source region displays valuable additional data.

Methodological limitations arise primarily from the grain-size analyses and the application of rEM modelling. Polymodal GSDs reflect the superposition of multiple transport, depositional, and postdepositional processes and therefore challenge synthetic unmixing approaches such as rEMMA (Dietze & Dietze, 2019). Although the EMMAgeo workflow identifies an optimal number of endmembers, the resulting rEMs do not explain the full variance of the dataset and represent simplified process proxies rather than unique sediment sources. In addition, postdepositional modification, including cryogenic disturbance and reworking, further complicates stratigraphic interpretations based on grain-size signals alone (Schirrmeister et al., 2020).

Finally, radiocarbon dating provides the chronological framework for this study but introduces additional uncertainty. The selected organic material represents only a subset of the sediment matrix and may include reworked components, affecting age–depth relationships. As a result, depositional and erosional phases can be resolved only within broad temporal frameworks rather than as a continuous, high-resolution developmental sequence. Nevertheless, the available age control is sufficient

to constrain the main phases of TEV development and to distinguish between Late Pleistocene and Holocene processes. Higher-resolution radiocarbon dating would further refine these interpretations and strengthen future reconstructions of TEV evolution.

6 Conclusion

This study addressed the limited understanding of the near-surface stratigraphy and Holocene development of a TEV incised into Yedoma deposits. By combining cryostratigraphic, sedimentological, geochemical, and hydrochemical analyses of permafrost cores from distinct geomorphic positions, this thesis investigated the processes and timescales shaping near-surface permafrost stratigraphy in an ice-rich Arctic landscape. In addressing RQ1, the analysed cores reveal pronounced spatial differences in stratigraphy and dominant processes that are closely linked to their geomorphic position within the valley.

At the Yedoma upland, stratigraphic records extend back to the Early Holocene and preserve syngenetic wedge ice beneath a Holocene cover. The sedimentary material shows strong postdepositional modification and reworking of Yedoma deposits, while clear indicators of active permafrost degradation are absent. Therefore, I reject H1, which proposed largely undisturbed Yedoma deposits at the upland site.

At the valley slopes, thin sediment covers and Mid Holocene ages at shallow depth indicate limited deposition and weak soil formation. The young deposits overlying wedge ice in core A001 suggests erosional removal, yet the stratigraphy does not allow a robust temporal constraint on individual erosional events. Consequently, I partly accept H2, suggested a stratigraphy, predominantly coined by erosional processes and little soil formation.

The valley floor records sediment infilling as hypothesised in H3. Deposits are constrained to the Late Holocene, and GSDs, rEM contributions, and organic-matter characteristics indicate alternating depositional conditions and episodic high-energy transport. While the stratigraphy does not preserve distinct erosional sequences, deposition and postdepositional modification dominate the valley-floor record. Due to the missing erosional evidence, I partly accept H3.

RQ2 and the corresponding H4 regard the TEV as an interconnected system with the various processes constrained in RQ1 and poses the question of its stability and resilience. Furthermore the H4 hypothesises an Holocene landform resilient to minor disturbances but sensitive to changing hydrological and permafrost conditions.

In addressing RQ2, I evaluated the TEV as an interconnected geomorphic system and tested H4, which proposed a Holocene landform that is resilient to minor disturbances but sensitive to hydrological and permafrost thresholds. Stratigraphic records from the valley slopes constrain the onset of TEV development to earlier than 7215 cal yr BP, although the data do not allow identification of a precise initiation phase. In agreement with previous studies from Kurungnakh Island, an initial development around the transition from the Late Pleistocene to the Early Holocene appears most likely. During the Mid to Late Holocene, the TEV evolved into a comparatively stable landform, characterised by limited modification of the Yedoma upland and valley slopes and continued sediment accumulation at the valley floor. Individual disturbance events mobilised sediment material uphill and redistributed it downslope and downstream, yet these events did not fundamentally alter valley morphology. I therefore accept H4.

Previous studies primarily focused on surface morphology, spatial distribution, and conceptual process models of thermokarst and thermo-erosional landscapes. This thesis adds stratigraphic evidence from the subsurface and refines the understanding of processes operating at distinct geomorphic positions

within TEVs. These findings improve conceptual models of TEV development and provide a basis for assessing how ice-rich permafrost landscapes may respond to ongoing climate warming and changing hydrological conditions.

7 References

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8 AI statement

Artificial intelligence (AI) was used during the preparation of this Master’s thesis exclusively in the form of ChatGPT (OpenAI, GPT-5.). The table on the specific AI statement includes exemplary prompts on the use of AI, covering all areas of AI involvement (Table 6).

All AI-assisted outputs were critically reviewed, verified against primary literature, and edited by the author. The author takes full responsibility for the scientific content, interpretations, and conclusions presented in this work.

Table 6: Documented use and exemplary prompts of AI in this thesis. All outputs were reviewed and validated by the author.

AI Tool	Version	Thematic Context	Purpose / Representative Prompts
ChatGPT, Python GPT (OpenAI)	GPT-5.2	Data visualisation (Chapter 4)	Data processing and plotting support in Python: “How can grain size distribution data stored in Excel, with four metadata rows and grain sizes in the first column, be processed in Python to plot logarithmic grain size distribution curves grouped by sample prefixes (A, B, C), with equal spacing on a logarithmic x-axis in micrometers and controlled plot layout (figure size, legend position)?”
ChatGPT (OpenAI)	GPT-5.2	MICADAS AMS radiocarbon dating (Chapter 3.3.7)	Literature identification: “Is there a publication from the MICADAS lab in Bremerhaven on dating accuracy?”
ChatGPT (OpenAI)	GPT-5.2	Solifluction & gelifluction (Chapter 5)	Core interpretation: “How can solifluction or gelifluction be recognised and distinguished in permafrost-core stratigraphy, preferably in Yedoma landscapes?”
ChatGPT, SciSpace (OpenAI)	GPT-5.2	Ice-wedge dynamics (Chapter 5)	Conceptual validation: “Is it unlikely for an ice wedge not to crack for thousands of years and then crack again? Provide citations.”
ChatGPT (OpenAI)	GPT-5.2	Paragraph structure (Chapter 1)	Writing structure: “What changes do you suggest for this paragraph, to make it fit into the flow of the introduction from the broad system towards the research question. The goal of this paragraph is to depict the research gaps on thermo-erosional valleys. Paragraph: [...]”
ChatGPT (OpenAI)	GPT-5.2	Grammar and repetitions	Writing: “Check the grammar (English UK) of this paragraph and make it more concise, while keeping the content, my style, active voice and [...] tense. Paragraph: [...]”

9 Appendix

9.1 Theoretical Background

9.1.1 Permafrost

Building on the brief definition and relevance of permafrost introduced in Chapter 1, this section summarises key theoretical concepts required for the interpretation of stratigraphic and cryostratigraphic observations. Regarding to its global distribution, permafrost is commonly classified into glaciers, continuous, discontinuous, sporadic, isolated, and subsea permafrost zones (Figure 24) (Schuur et al., 2022). The Lena River Delta is situated within the zone of continuous permafrost, characterised by thick and laterally extensive frozen ground (Schuur et al., 2022; Zaplavnova et al., 2024).

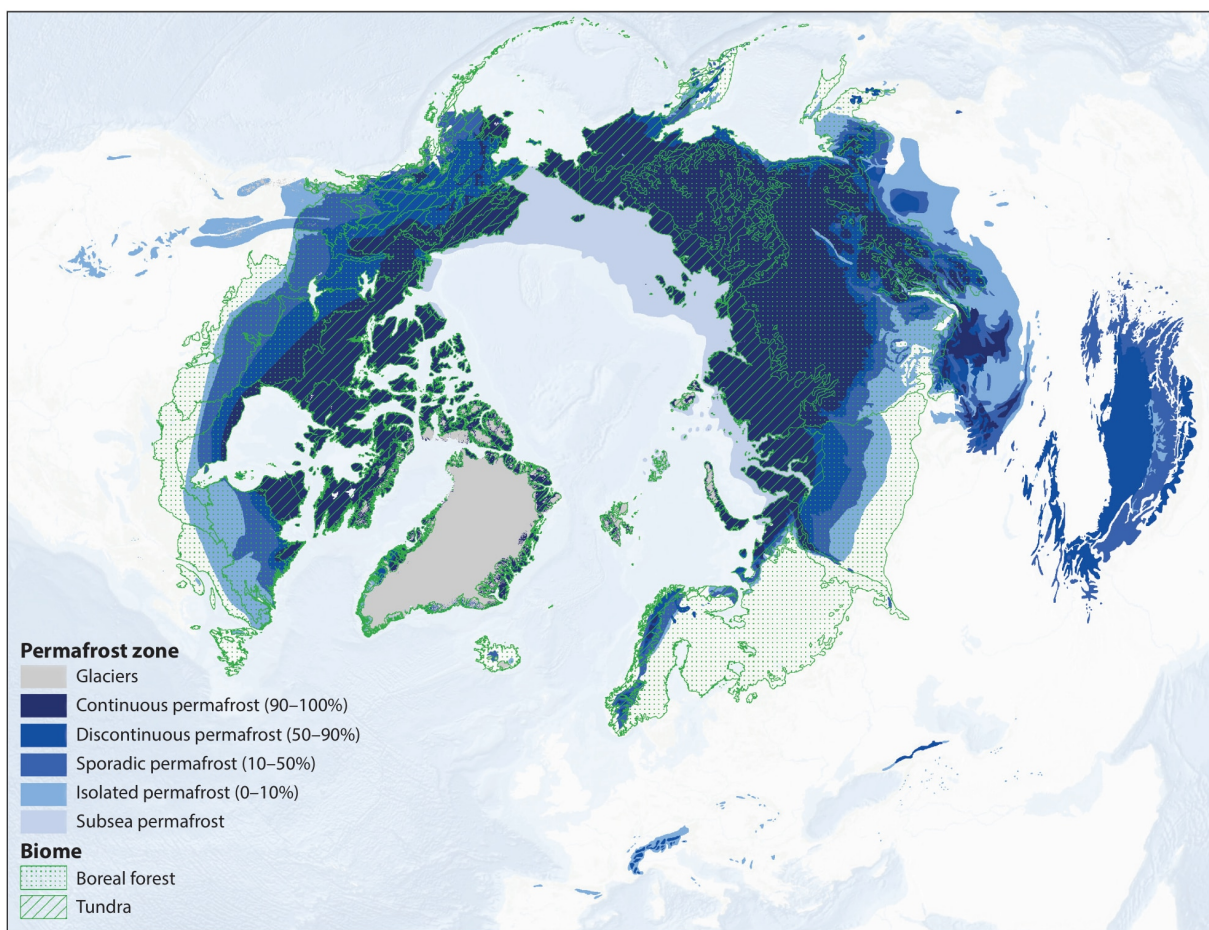


Figure 24: Map of Arctic permafrost regions classified into glaciers, continuous, discontinuous, sporadic, isolated, and subsea permafrost with respective ground cover percentages (Figure in Schuur et al. (2022)).

Permafrost literature generally distinguishes between the seasonally thawing active layer and the underlying perennially frozen permafrost body (Shur et al., 2005). ALT varies interannually and spatially as a function of air temperature, slope orientation and angle, vegetation, drainage, snow cover, soil properties, and water content (French & Shur, 2010). Mean annual ground temperatures reach their minimum at the base of the active layer, referred to as the permafrost table, and increase with depth following the geothermal gradient. In permafrost terrain, ground freezing can occur both from the

surface downward and from the permafrost table upward, in contrast to seasonally freezing ground in non-permafrost regions (French & Shur, 2010).

The depth of the permafrost table can vary on annual to millennial timescales. Therefore, while the distinction between active layer and permafrost body is sufficient for many applications, it becomes insufficient for analysing long-term permafrost evolution and palaeoenvironmental records (Shur et al., 2005). For such purposes, the concept of a transient layer is introduced, defined as the zone within which the active-layer thickness fluctuates over longer timescales. In this thesis, ice-rich stratigraphic sequences affected by long-term thaw and refreezing are therefore interpreted in the context of a transient layer.

Permafrost can further be classified according to its genesis into epigenetic, syngenetic, and polygenetic types (French & Shur, 2010). Epigenetic permafrost forms after deposition of the host material, whereas syngenetic permafrost develops contemporaneously with sediment accumulation, causing the permafrost table to migrate upward. Syngenetic permafrost formation is associated with a range of depositional environments, including aeolian, fluvial, colluvial, and lacustrine settings. Most thick permafrost bodies are polygenetic, combining syngenetic and epigenetic components (French & Shur, 2010).

Syngenetic and epigenetic permafrost both depict distinct characteristics. Syngenetically-frozen soil is typically ice rich with soil gravimetric ice contents possibly exceeding 100 % as well as high VICs. The sediment is silty or loess-like and can also consist of aggrading fluvio-aeolian sands and coarser floodplain deposits. Furthermore, the syngenetic soils often contain buried organic-rich horizons and exhibit a rhythmically-organised, layered appearance. Syngenetic permafrost typically develops in the Arctic North America and northern Siberia. Usually the permafrost in these regions is several hundred meters thick of which the contemporary syngenetic permafrost does not exceed a thickness of a few meters. The lower parts are often epigenetic. The epigenetic permafrost does not only appear in combination with syngenetic permafrost but also in high-latitude areas, previously covered by Late-Pleistocene ice sheets (French & Shur, 2010). Though, this is not of high importance in the Lena River Delta and for Kurungnakh Island, since the area was not glaciated during the late Pleistocene and Holocene (Hubberten et al., 2004).

Syngenetically frozen deposits are typically ice rich, with gravimetric ice contents that may exceed 100 % and high volumetric ice contents. They often consist of silty, loess-like material or aggrading fluvio-aeolian sands and floodplain deposits, and may contain buried organic-rich horizons and rhythmically layered structures. Syngenetic permafrost is widespread in northern Siberia and Arctic North America. Although permafrost thickness in these regions commonly reaches several hundred metres, the actively aggrading syngenetic portion typically comprises only the uppermost metres, underlain by epigenetic permafrost (French & Shur, 2010). In the Lena River Delta, this stratigraphic configuration is characteristic, as the region was not glaciated during the Late Pleistocene or Holocene (Hubberten et al., 2004).

Cryostratigraphy distinguishes between *pore ice*, *segregated ice*, and *vein ice* and documents their distribution within frozen deposits (French & Shur, 2010). Pore ice forms from moisture within capillary spaces and binds sediment particles. Segregated ice develops during freezing when unfrozen water migrates towards the freezing front by cryosuction and forms ice lenses ranging from millimetre-thick

layers to ice bodies several tens of metres thick (French & Shur, 2010). Vein ice forms when thermal-contraction cracks fill with wind-blown snow and snowmelt-derived water, which with repeated cracking can generate foliated ice and ultimately form ice wedges (French & Shur, 2010). Ground ice can further be described as *syngenetic* or *epigenetic*, depending on whether freezing occurs during sediment accumulation or after deposition (French & Shur, 2010). In addition, the conceptual framework of cryostratigraphy includes three permafrost horizons described by Popov (1973), which are summarised by French and Shur (2010): (i) the active layer, (ii) permafrost above the depth of zero annual amplitude, and (iii) permafrost below the depth of zero annual amplitude (French & Shur, 2010).

Cryostratigraphy provides the framework to describe the distribution and origin of ground ice within frozen deposits. It distinguishes between pore ice, segregated ice, and vein ice (French & Shur, 2010). Pore ice occupies capillary spaces between sediment grains, whereas segregated ice forms through cryosuction-driven migration of unfrozen water toward freezing fronts, producing ice lenses ranging from millimetres to several metres in thickness. Vein ice forms when thermal-contraction cracks fill with snow or meltwater and, through repeated cracking and infilling, may evolve into foliated ice and ice wedges (French & Shur, 2010). These different ground-ice types are expressed by contrasting VICs, cryostructural patterns observed in CT scans and core images, and characteristic stable water isotope signatures. This cryostratigraphic framework underpins the interpretation of ice-rich permafrost sequences throughout this thesis.

9.1.2 Yedoma

During the Late Pleistocene, global sea level was lowered by 120 ± 5 m, exposing the shallow shelves of the Laptev Sea and incorporating them into the Beringian landmass (Fairbanks, 1989; Grosse et al., 2007). The cold and dry continental climate prevented widespread glaciation of northeastern Siberia and the exposed shelf regions, with glaciation restricted mainly to the northwestern Taymyr Peninsula (Hubberten et al., 2004). Under these conditions, fine-grained, ice- and organic-rich sediments accumulated and froze syngenetically, forming the IC, while fluvial sediments were deposited within major river valleys (Grosse et al., 2007).

The term Yedoma refers not only to these late Pleistocene IC deposits but also to a distinct geological unit composed of syngenetically frozen permafrost. Yedoma deposits are characterised by high ice and organic contents and consist predominantly of poorly sorted silt to fine sand (Grosse et al., 2007; Schirrmeister et al., 2011b). Gravimetric ground-ice contents typically range between 40 wt% to 60 wt%, and together with syngenetic ice wedges the total ice volume can reach up to 80 vol% (Schirrmeister et al., 2011b). The origin of these deposits was long debated, but increasing paleoenvironmental evidence supports a predominantly nival to aeolian sedimentation under cold-climate conditions, with spatially heterogeneous depositional environments across periglacial landscapes (Schirrmeister et al., 2011b, 2020).

Following the onset of postglacial warming and marine transgression during the Holocene, beginning around 11 700 yr BP (Walker et al., 2008), thermokarst processes became increasingly important (Grosse et al., 2007; Morgenstern et al., 2013b). Thermokarst development and self-submergence led to the widespread degradation of the original Yedoma depositional environment. Nevertheless, extensive remnants of the Ice Complex persist today in the northeastern Siberian coastal lowlands and on

islands of the East Siberian Seas, where they remain subject to ongoing erosion (Grosse et al., 2007). Kurungnakh Island represents one such remnant as a part of the third Lena River Delta terrace.

Under ongoing Arctic warming, Yedoma deposits are increasingly viewed as vulnerable carbon reservoirs. The erosion and thaw of ice-rich permafrost, for example along fluvio-thermal-erosional cliffs such as Sobosise, release large amounts of previously frozen organic carbon and nitrogen (Fuchs et al., 2020; Schuur et al., 2008). This transition from long-term carbon storage to a potential carbon source has implications not only for the global carbon cycle but also for local riverine and nearshore ecosystems in the Arctic (Fuchs et al., 2020).

9.1.3 Stable water isotopes

Stable water isotopes $\delta^2\text{H}$ and $\delta^{18}\text{O}$ form the basis for investigating ice formation processes, stratigraphic correlation, and postcryogenic alteration in permafrost deposits (Schirmer et al., 2011a). In ice wedges, stable isotope compositions enable paleoenvironmental interpretations related to winter precipitation, which feeds ice-wedge growth. In contrast to many biological climate proxies, such as pollen or tree rings, which are biased toward summer conditions, ice-wedge isotopes primarily reflect cold-season climate signals (Meyer et al., 2015; Schirmer et al., 2011a). Consequently, ice wedges record local surface temperatures of the meteorological winter (December, January, February) and early spring (March, April, May). Although isotope signals can be influenced by changes in moisture source, transport pathways, snow-cover seasonality, and frost-cracking dynamics, air temperature remains the dominant control on $\delta^{18}\text{O}$ variability in Arctic ice wedges (Meyer et al., 2015).

At the global scale, $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of precipitation plot along the Global Meteoric Water Line (GMWL), defined by a slope close to 8, reflecting equilibrium fractionation during progressive cooling and condensation of atmospheric moisture (Craig, 1961). Deviations from the GMWL indicate non-equilibrium (kinetic) fractionation processes, such as evaporation from open water surfaces, snow formation, or sublimation (Bonne et al., 2020). During equilibrium freezing, heavier isotopes preferentially remain in the liquid phase, resulting in isotopically depleted ice. Kinetic effects during freezing or evaporation may cause isotope data to deviate from the GMWL and define a Local Meteoric Water Line (LMWL).

The influence of kinetic fractionation processes is commonly assessed using the deuterium excess (d-excess), defined as (Dansgaard, 1964):

$$\text{d-excess} = \delta^2\text{H} - 8 \times \delta^{18}\text{O}$$

The d-excess provides insight into moisture-source conditions and atmospheric processes affecting water vapour, reflecting variations in relative humidity, sea-surface temperature at the source, evaporation conditions, and the contribution of continental or recycled moisture (Dansgaard, 1964). In Arctic regions, including the Lena River Delta, pronounced seasonal variability in d-excess has been documented. Elevated vapour-phase d-excess values during late autumn and early winter are linked to low-humidity evaporation over open water during sea-ice formation, whereas increasing d-excess in spring and summer reflects enhanced snow sublimation and changing moisture recycling as inso-

lation rises and snow cover retreats (Bonne et al., 2020). Seasonally varying air-mass contributions with contrasting isotopic signatures further characterise the regional moisture budget. Rapid transport from the Atlantic delivers relatively warm and isotopically enriched moisture, whereas slower transport over northern Europe and continental environments is associated with more depleted isotopic signatures (Bonne et al., 2020).

9.1.4 Grain-size distributions of hydraulic and aeolian sedimentary systems

The methods, results, and interpretations presented in this thesis already address GSDs and their relation to hydraulic and aeolian sedimentary systems. Here, I step back and outline the fundamental composition of these systems and the associated transport and depositional mechanisms that are relevant for interpreting the GSDs of the analysed dataset.

In continental environments, the hydraulic sedimentary system can be subdivided into fluvial, lacustrine, alluvial, and deltaic facies. Sediment particles are transported in three principal modes depending on grain size and flow energy: suspension, saltation, and traction (bed load) (Middleton, 1976; Sun et al., 2002). When transport energy decreases, these fractions are commonly deposited together, resulting in polymodal GSDs. Sun et al. (2002) demonstrate how overlapping components of characteristic grain-size ranges combine to form the distinct GSD curves observed in sediment samples (Figure 25).

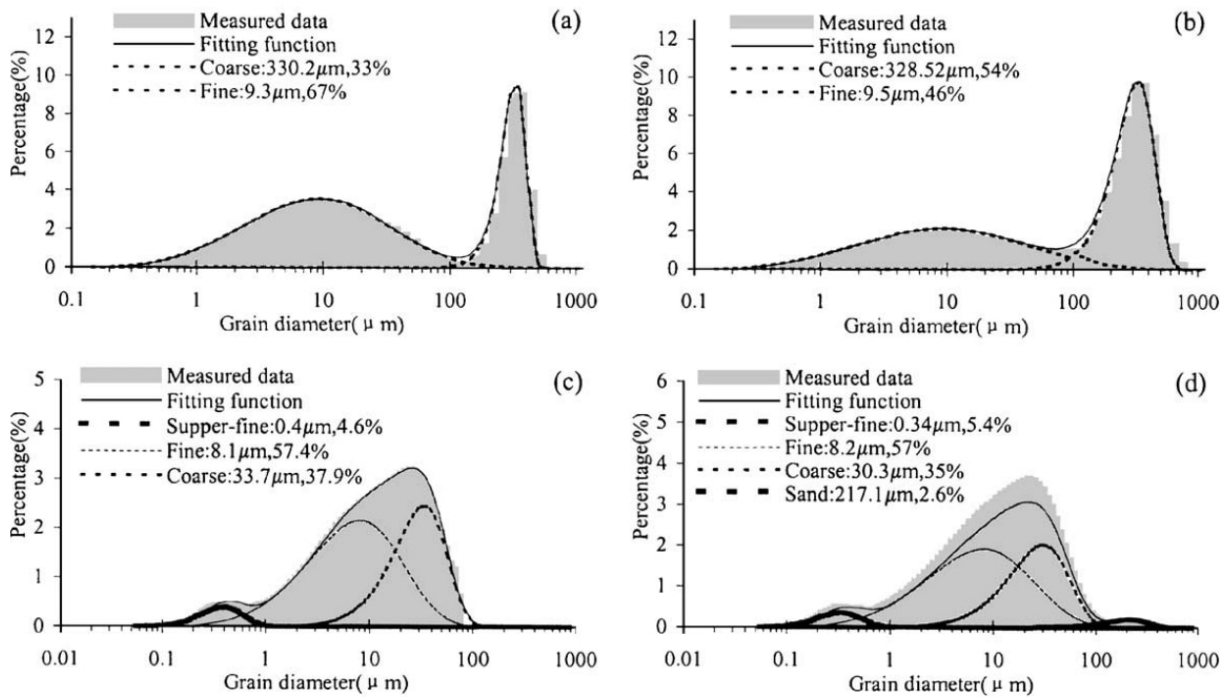


Figure 25: Grain-size distribution and sedimentary components of fluvial, lacustrine sediments of: (a) Modern riverbed sand from Xifeng in northern China; (b) Tertiary river sand from Xifeng in northern China; (c) Lake sediment from 0.2 m water depth in Taibei Lake, Qinlin Mountains and (d) Lake sediment from 0.4 m water depth in Taibei Lake (Figure in Sun et al. (2002)).

In fluvial environments, suspension and saltation loads are particularly relevant for environmental interpretation. Typical grain-size ranges for suspended load lie between 10 μm to 15 μm , whereas saltation loads usually range between 200 μm to 400 μm (Figure 25) (Sun et al., 2002). The saltation fraction

often accounts for a larger proportion of the sediment and is generally well sorted, producing a narrow and pronounced modal peak. In contrast, the suspended fraction is poorly sorted and forms a broader distribution curve.

Lacustrine sediments represent another subdivision of the hydraulic system. According to Sun et al. (2002), these sediments typically consist of two closely overlapping silt components with modes around 8 μm to 9 μm and 25 μm to 35 μm , as well as a finer component between 0.2 μm to 0.4 μm . Some lacustrine samples additionally contain an extremely coarse component with a modal size exceeding 200 μm , comparable to the saltation fraction in fluvial sediments and potentially related to episodic high-energy events such as storms. The silt-sized components are interpreted as suspended load originating from different surface-water processes, including seasonal runoff, whereas the very fine fraction may reflect chemical, biochemical, or pedogenic contributions (Sun et al., 2002).

Aeolian sedimentary systems rely on wind as the transporting medium. Particle movement is initiated by aerodynamic drag, lift forces, or impacts with other moving particles, including sediment grains, hail, or rain (Pye, 1987). Similar to hydraulic systems, aeolian transport occurs as traction, saltation, and suspension. The transport mode depends on the balance between particle settling velocity and wind shear velocity (Gillette et al., 1974). Generally, coarse grains remain close to the source area and form desert sand deposits, whereas finer silt and clay particles are transported in suspension and deposited either locally or over long distances, forming loess (Tsoar & Pye, 1987). Transport height plays a key role: coarse silt is carried in short-term, near-surface suspension, while finer clay particles are entrained into higher atmospheric levels, enabling long-distance transport.

Desert sand typically displays a bimodal GSD composed of an overlapping coarse sand and fine clay component. The coarse fraction dominates the distribution, is well sorted, nearly symmetrical, and has an average modal size between 100 μm to 200 μm , indicating transport by saltation near the surface. The finer component is poorly sorted, symmetrical, and exhibits a modal size between 2 μm to 6 μm (Sun et al., 2002).

Loess deposits represent another endmember of the aeolian system and primarily record deposition from suspension. It is important to distinguish aeolian suspension particles from those transported in hydraulic systems. Within aeolian suspension, coarse settling grains (>20 μm) occur in short-term suspension, whereas finer non-settling grains remain in long-term suspension (Pye, 1987; Tsoar & Pye, 1987). Loess GSDs are commonly bimodal, with a dominant coarse silt peak between 16 μm to 32 μm and a fine-grained mode whose origin depends on the local setting. This finer component may represent background long-term suspension or, in some cases, contributions from pedogenic or chemical processes (Schwamborn et al., 2012). Higher proportions of the fine fraction are associated with periods dominated by strong upper-level winds (Sun et al., 2002).

Finally, natural sediments rarely reflect a single sedimentary system or transport mechanism. Mixed deposits are common, for example when aeolian loess is reworked by fluvial processes. Such sediments may form through direct deposition of loess into hydraulic environments or by fluvial reworking of previously deposited aeolian material. Reworked loess typically exhibits three to four distinct grain-size components, each with its own modal peak (Sun et al., 2002). As discussed in Chapter 5, the stratigraphic framework of Yedoma deposits on Kurungnakh Island and Holocene environmental conditions strongly favour the formation of such mixed, polygonal sediments.

9.2 Additional data and data visualisations

9.2.1 Source region: core D

Core D has a total length of 175 cm. As no laboratory analyses have been conducted for this core to date, CT scans form the sole basis for cryostructure classification and the subdivision into active layer, sediment, and wedge-ice sections (Figure 26). Based on the CT-scan data, I separated the core into an active layer extending from 0 cm to 32 cm, an overlying sediment section between 32 cm to 63 cm, and underlying wedge ice down to the base of the core at 175 cm. The ice-wedge section accounted for approximately 64 % of the total core length. Within the sediment section, reticulate and organic-matrix cryostructures dominated the cryostratigraphy. Due to the lack of complementary laboratory data, I did not include core D in the main chapters of this thesis, as the available information does not allow for robust sedimentological, geochemical, or chronological interpretations.

Future investigations of this TEV on Kurungnakh Island would benefit from targeted laboratory analyses of core D. Given its position in the source region, I expect core D to record older stratigraphic units with limited Holocene sedimentation and possibly erosional gaps.

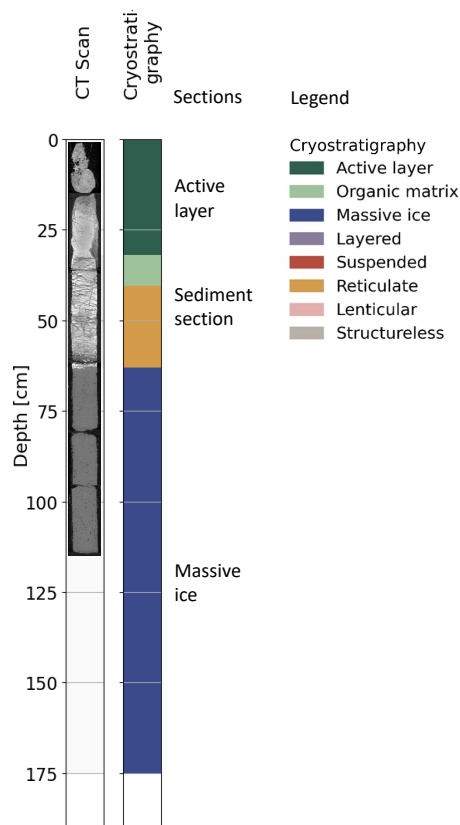


Figure 26: Data from the source region core D, plotted by depth with a CT scan and cryostratigraphy classification.

9.2.2 EMMAgeo outputs

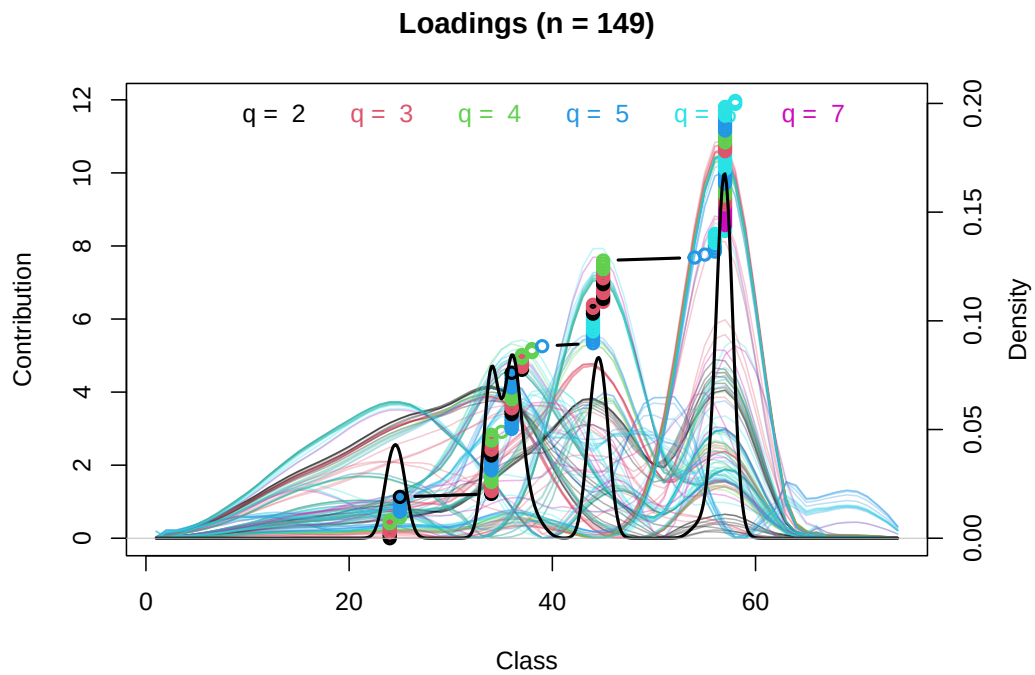


Figure 27: EMMAgeo loadings for possible amounts of robust endmembers (2-7), suggested by the weight transformation values. The figure is a EMMAgeo output of the decision making process for the number of rEMs.

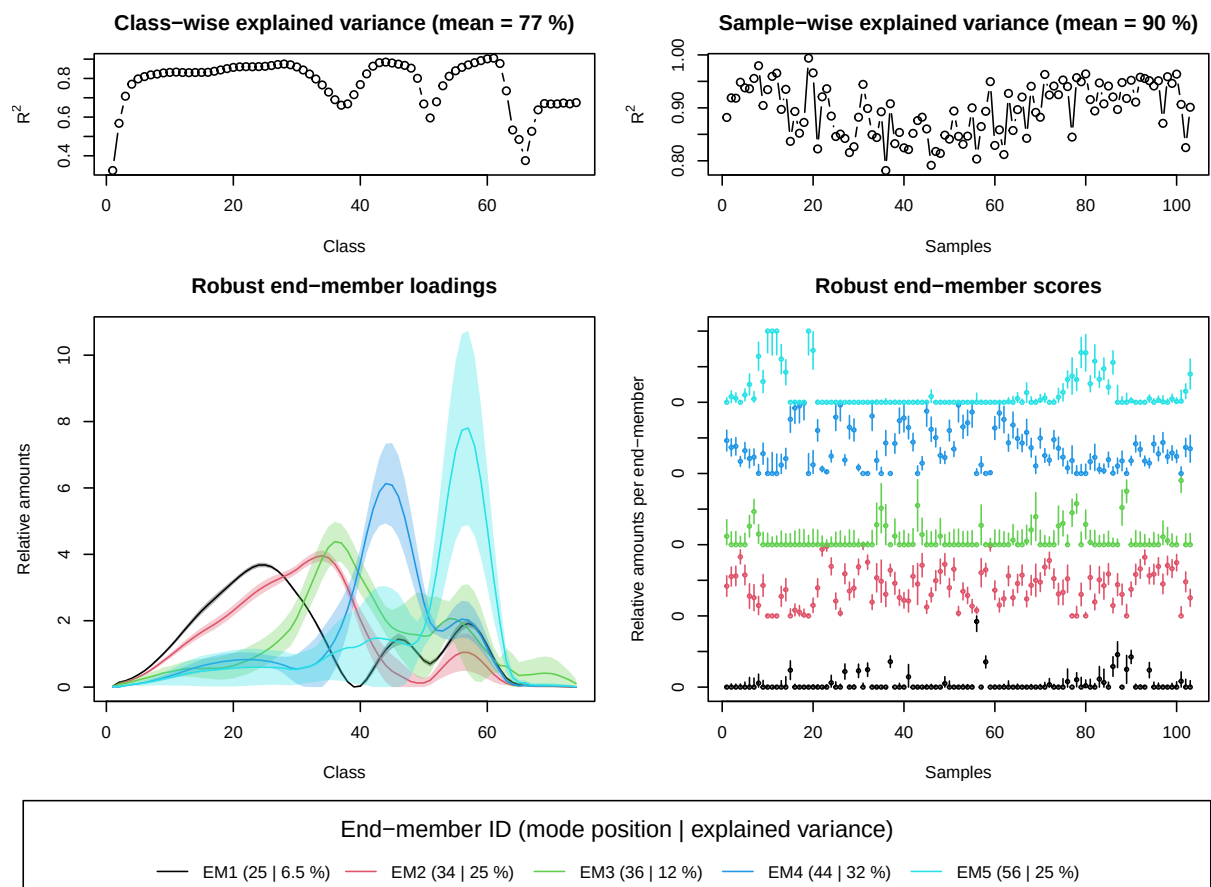


Figure 28: Final EMMAgeo output including rEM loadings and the rEM scores by sample. Additionally, the explained variance is plotted respectively.

9.2.3 Time lapse camera



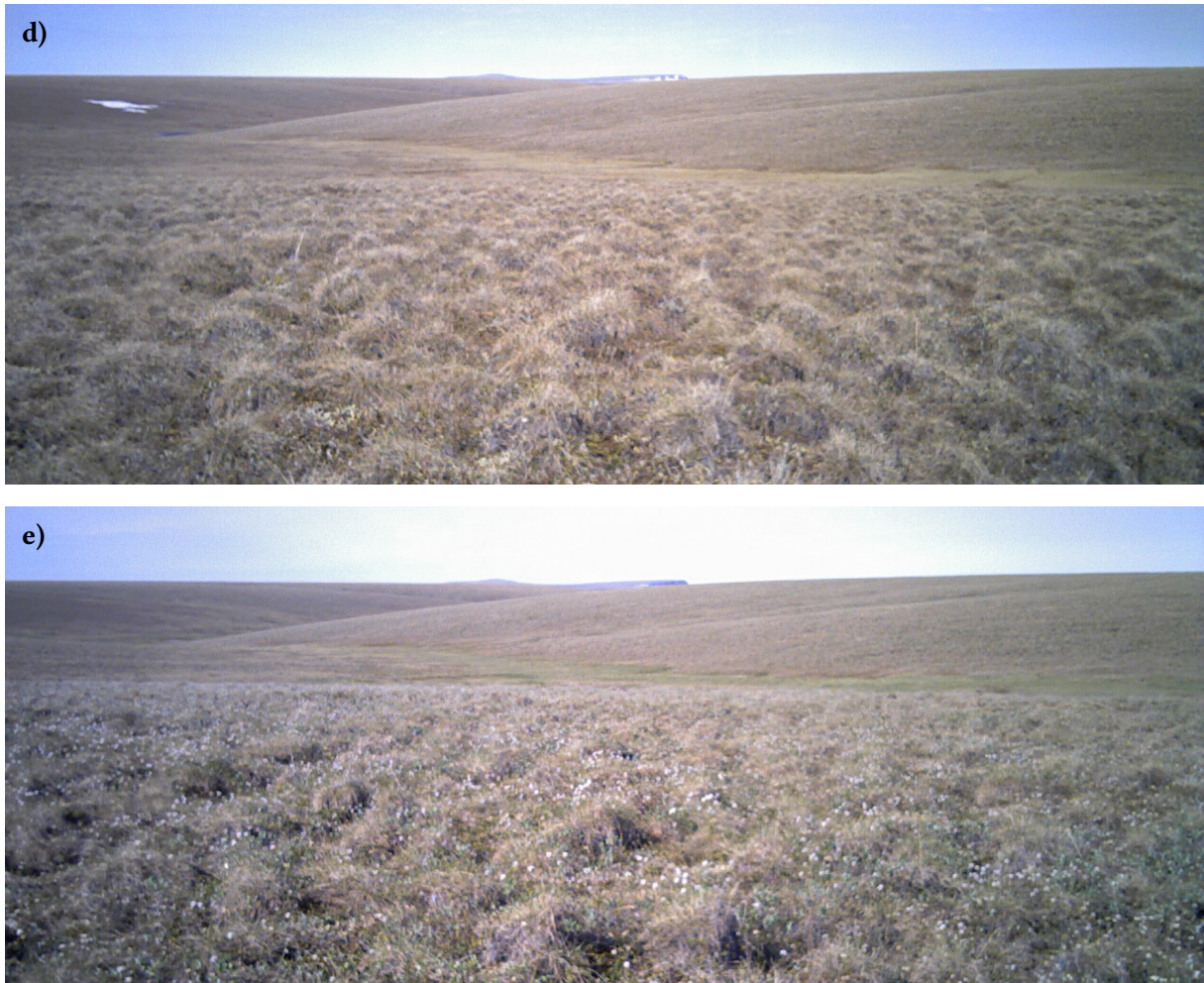


Figure 29: Time lapse camera images from the analysed thermo-erosional valley, taken on: (a) 29.04.2018; (b) 25.05.2018; (c) 06.06.2018; (d) 13.06.2018; (e) 22.06.2018 (Morgenstern et al., 2021a).

10 Personal declaration

I hereby declare that the submitted thesis is the result of my own, independent work. All external sources are explicitly acknowledged in the thesis.

A handwritten signature in black ink that reads "Jael Mencke". The script is cursive and fluid, with the first letters of the first and last names being capitalized and prominent.

Jael Mencke

30.1.2026