The Kola Peninsula and Russian Lapland: a review of Late Weichselian glaciation

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Abstract

The Kola Peninsula and Russian Lapland (Murmansk Oblast, northwest Arctic Russia) represent a major sector of the Fennoscandian Ice Sheet (FIS) where empirical geomorphological, sedimentological, and chronological data are lacking and thus, where the pattern, style, and timing of glaciation is not well established. In this study, we present a critical review of published empirical data and interpretations of Late Weichselian (*c.* 40-10 ka) glaciation for the region. The review includes, for the first time, information published in Russian-language journal articles (n = 37), and is accompanied by a new Geographic Information System (GIS) numerical age database (spanning 472.3-6.2 ka) that collates known published numerical dates associated with the advance and retreat of the FIS in the study area.

Our review suggests that an ice mass existed over the Kola Peninsula and Russian Lapland during the Early-Middle Weichselian (*c.* 115-40 ka), and likely retreated during the Ålesund interstadial (*c.* 38-34 ka). During the Late Weichselian, it is likely that the FIS advanced eastwards across Russian Lapland

and the Kola Peninsula, establishing the White Sea Ice Stream before the local-Last Glacial Maximum (*c.* 19-15 ka). Through an evaluation of the existing Last Glacial-Interglacial Transition (*c.* 20-10 ka) glaciation models for the region, we propose that the Kola Peninsula and Russian Lapland was deglaciated by the FIS, rather than the Ponoy Ice Cap or the Kara Sea Ice Sheet. In collating, discussing, and critically evaluating empirical data and interpretations, this paper provides a valuable resource to inform FIS dynamics at both a regional- and ice sheet-scale, and offers a framework through which numerical ice sheet models can be constrained. Precise FIS dynamics on the Kola Peninsula and Russian Lapland, including the position of the Younger Dryas ice marginal zone, remain unclear due to low-resolution geomorphological data. In concluding, we recommend that further work is needed in the form of a revised glacial reconstruction using high-resolution, peninsula-wide geomorphological data.

Keywords: glacier reconstruction, geomorphology, Kola Peninsula, Russian Lapland, Late Weichselian, Fennoscandian Ice Sheet, Russia

1. Introduction

The response of the Antarctic and Greenland ice sheets to recent climate change has firmly positioned ice sheets as a priority on the global research agenda (e.g. Joughin et al., 2014; Oppenheimer et al., 2019; Miles et al., 2020; Slater et al., 2020; van Kampenhout et al., 2020). However, attempting to quantify the response of ice sheets to future climate change remains challenging because of inadequate understanding of contemporary ice sheet dynamics (Kleman and Applegate, 2014). To circumvent this, studies of palaeo-ice sheets, such as the Fennoscandian Ice Sheet (herein referred to as the FIS), can provide insights into ice discharge history and the scale of ice sheet responses to past climatic changes (Kleman and Applegate, 2014; Stokes et al., 2015; Stroeven et al., 2016). The reconstruction of palaeo-ice sheets can also provide the initial boundary conditions for climate models and can be used to test the accuracy of numerical ice sheet models, which are essential for predicting future ice sheet response to anthropogenic climate change (Stokes et al., 2015; Pearce et al., 2017).

The FIS has attracted considerable research interest since the mid-19th Century (Agassiz, 1840; De Geer, 1884, 1912; Sauramo, 1918, 1923; Mannerfelt, 1949; Lundqvist, 1972; Andersen, 1979; Andersen et al., 1995a; Punkari, 1995; Kleman et al., 1997; Svendsen et al., 2004; Hättestrand and Clark, 2006b; Hughes et al., 2016; Stroeven et al., 2016). The FIS nucleated in the Scandinavian Mountains of northern Sweden and southern Norway, terminating in lobate ice margins with strong topographic steering in pre-existing valleys (Patton et al., 2016; Stroeven et al., 2016). Geomorphological and chronometric data indicate that the FIS experienced asynchronous growth i.e. the timing of peak ice volume and maximum ice extent (referred to as the local-Last Glacial Maximum (local-LGM)) differed across the ice sheet (Svendsen et al., 2004; Hughes et al., 2016; Stroeven et al., 2016). For example, the FIS reached its maximum configuration *c.* 27-26 ka on the Norwegian continental shelf, but did not attain its maximum lateral extent until *c.* 23-22 ka in northern Europe (Hughes et al., 2016; Stroeven et al., 2016). However, it is generally regarded that the Last Glacial Maximum (LGM) of the FIS occurred between 26.5 and 20 ka (Fig. 1a; Clark et al., 2009).

Fig. 1: (a) Reconstructed maximum-achieved lateral extent of the Late Weichselian FIS (modified from Hughes et al., 2016). FIS growth was asynchronous; as such, the maximum lateral extent of ice was achieved at different times. The location of the Kola Peninsula and Russian Lapland is shown by the

black box. (b) FIS ice marginal zones during the Younger Dryas. Moraines attributed to the Younger Dryas have been mapped and dated almost continuously around Scandinavia (solid black lines), allowing near complete characterisation of the Younger Dryas ice sheet margin, except on the Kola Peninsula and Russian Lapland (dashed black lines) (modified from Hughes et al., 2016). (c) The Kola Peninsula and Russian Lapland within Murmansk Oblast – the federal subject of Russia (indicated by thick black border) in which the study area is located – showing the major geographical regions and mountainous areas discussed in the text. The administrative city of the region, Murmansk, is also shown. The division between the western and eastern Kola Peninsula is determined by the approximate topographic and Quaternary geological distinctions between the two regions, based on the work of Ramsey (1898), Lavrova (1960), Niemelä et al. (1993), and Hättestrand and Clark (2006a), among others. The division between Russian Lapland and the Kola Peninsula is also shown, although Russian Lapland is discussed alongside the western Kola Peninsula in this study because of similarities in topographic and sediment-landform assemblages in both regions. In this, and subsequent figures, topographic and bathymetric data are from the GEBCO Compilation Group (2020) GEBCO 2020 Grid. Topography is shaded dark-light grey with darker shades representing higher ground; bathymetry is shaded dark-light blue with darker shades representing greater depth. Country outlines are derived from the Database of Global Administrative Areas (GADM, 2018).

Increased summer insolation and climatic warming *c.* 14.6 ka, known as the Bølling oscillation (as seen in the Greenland Ice Core Chronology (GICC05)), induced the rapid retreat of the FIS (Lehman et al., 1991; Clark et al., 2009; Rasmussen et al., 2014; Stroeven et al., 2016; Patton et al., 2017). Retreat continued throughout the Last Glacial-Interglacial Transition (LGIT; *c.* 20-10 ka) until the final demise of the FIS at approximately 10 ka (Cuzzone et al., 2016; Stroeven et al., 2016; Regnéll et al., 2019). Like the growth of the ice sheet, the pattern of deglaciation was asynchronous (Boulton et al., 2001; Hughes et al., 2016), with subglacial conditions and topographic constraints influencing the rate of deglaciation (Kleman et al., 2008; Stroeven et al., 2016). In addition, the overall retreat of the FIS was interrupted by several standstills and readvances of the ice margin during which end moraines and glaciofluvial sediments were deposited in response to climatic cooling anomalies (Hughes et al., 2016; Stroeven et al., 2016; Patton et al., 2017). The last extensive zone of end moraines is attributed to the Younger Dryas stadial (*c*. 12.9-11.7 ka; Rasmussen et al., 2014), an ice marginal zone that can be traced almost continuously around Fennoscandia (Fig. 1b; Andersen et al., 1995a; Hughes et al., 2016; Stroeven et al., 2016). Particular emphasis has been given to this ice marginal zone (Fig. 1b) as it is often used to constrain numerical ice sheet models and understand how palaeo-ice sheets responded to abrupt climatic warming during the Younger Dryas-Holocene transition (*c*. 11.7 ka; Andersen et al., 1995a; Mangerud et al., 2011; Stroeven et al., 2016; Patton et al., 2017). As a result of this considerable body of research, the pattern, style, and timing of FIS glaciation during the Late Weichselian (*c.* 40-10 ka) is largely well-understood, allowing numerical ice sheet models to be tested across most of the FIS (Patton et al., 2016; Patton et al., 2017).

However, the details of glaciation remain elusive for one sector of the FIS; the Kola Peninsula and Russian Lapland (within Murmansk Oblast federal subject, northwest Arctic Russia; Fig. 1). The glacial landform and sediment assemblages of this region have been studied for over 100 years, either as small-scale mapping projects (e.g. Grigoryev, 1934; Yevzerov and Kolka, 1993; Superson, 1994; Pękala, 1998; Superson and Zgłobicki, 1998; Hättestrand et al., 2007; Hättestrand et al., 2008; Yevzerov, 2009; Yevzerov and Nikolaeva, 2010; Yevzerov, 2015; Lunkka et al., 2018; Yevzerov, 2018) or large-scale glacial reconstructions (e.g. Ramsey, 1898; Lavrova, 1960; Apukhtin and Ekman, 1967; Krasnov et al., 1971; Strelkov, 1976; Apukhtin et al., 1977; Niemelä et al., 1993; Svendsen et al., 2004; Hättestrand and Clark, 2006a, 2006b; Winsborrow et al., 2010; Petrov et al., 2014; Astakhov et al., 2016; Hughes et al., 2016; Stroeven et al., 2016). Early maps of the Quaternary geology of the Kola Peninsula and Russian Lapland, showing the spatial distribution of morainic and glaciofluvial deposits, and drumlins, permitted reconstructions of ice sheet extent (Lavrova, 1960; Niemelä et al., 1993). Most notably, Hättestrand and Clark (2006a) present a comprehensive glacial geomorphological map of the region that details the spatial distribution of subglacial bedforms (drumlins and ribbed moraine), morainic and glaciofluvial deposits, and meltwater channels (Fig. 2). While this research facilitates the reconstruction of ice flow configuration (Hättestrand and Clark, 2006b; Winsborrow et al., 2010), the resolution of the Landsat 7 ETM+ data (30 m ground resolution) used to compile the map precludes the identification of discrete, smaller-scale landforms.

The Kola Peninsula and Russian Lapland were uniquely situated at the confluence of three dynamically different ice masses during the Late Weichselian: (i) the FIS, a continental ice sheet with a largely cold-based central region that nucleated to the west of the peninsula (Kleman et al., 1997; Kleman and Hättestrand, 1999; Stroeven et al., 2016); (ii) the White Sea Ice Stream, a major ice stream located at the southern margin of the peninsula that drained the northeastern sector of the FIS; and (iii) the Barents Sea Ice Sheet, a marine ice sheet that coalesced with the FIS off the northern coastline of the peninsula (Boulton et al., 2001; Hättestrand and Clark, 2006a; Yevzerov, 2015; Hughes et al., 2016; Stroeven et al., 2016). Note that the White Sea Ice Stream and Fennoscandian Ice Sheet are in fact part of the same overall ice mass, despite being discussed here as two *dynamically different* ice masses. Some authors postulate that an ice mass on the Kola Peninsula and an ice lobe draining through the White Sea may have played an important role in ice sheet dynamics of the Barents Sea Ice Sheet and, later, influenced Arctic Ocean circulation during the LGIT (Hättestrand et al., 2007; Winsborrow et al., 2010; Hughes et al., 2016; Stroeven et al., 2016). However, a relatively sparse glacial chronology (48 age estimates included in existing glacial reconstructions e.g. Hughes et al., 2016; Stroeven et al., 2016) means that the timing of ice sheet extent and dynamics on the Kola Peninsula and Russian Lapland during the Late Weichselian is not well constrained (Hättestrand and Clark, 2006b; Hättestrand et al., 2007; Hughes et al., 2016; Stroeven et al., 2016). This stems, in part, from (i) the limited research attention given to the region, (ii) the publication of data mainly in Russian language journals where they remain largely unknown to non-Russian scientists, and (iii) the limited communication and transfer of knowledge between Russian and non-Russian scientists, particularly during Soviet times. As a result, the Kola Peninsula and Russian Lapland is one of several major sectors of the FIS where empirical geomorphological, sedimentological, and chronological data are lacking and thus, where the pattern, style, and timing of Late Weichselian glaciation is not well established (Hughes et al., 2016; Stroeven et al., 2016). Consequently, numerical ice sheet models for the FIS cannot be tested in this region (e.g. Patton et al., 2016; Patton et al., 2017), which precludes complete characterisation of an important sector of the FIS. Such information is crucial to understand how ice sheets respond to climatic changes over geological timescales.

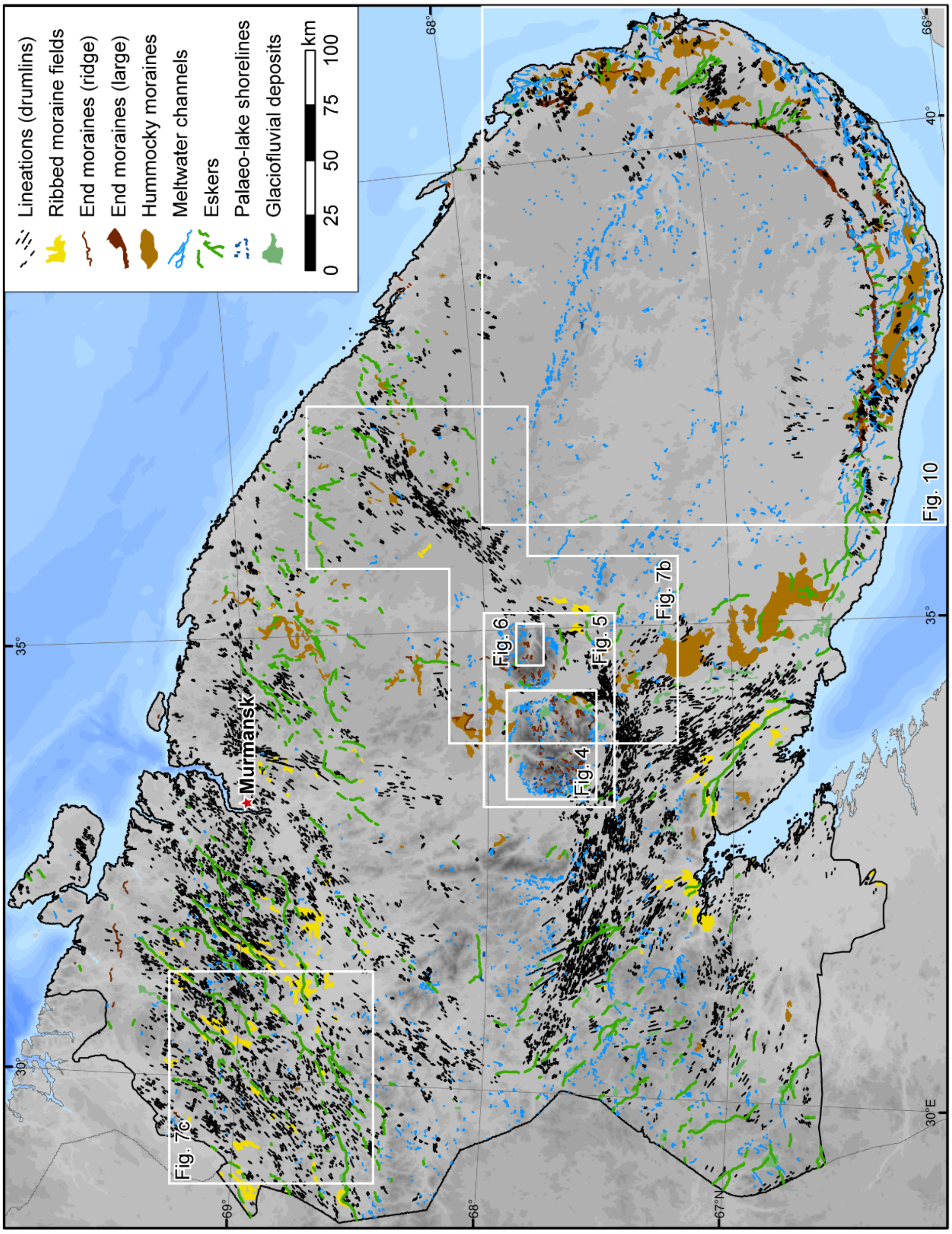

Fig. 2: The glacial geomorphology of the Kola Peninsula and Russian Lapland, modified from Hättestrand and Clark (2006a). This published map builds upon previous Quaternary maps (Niemelä

et al., 1993) and includes previously undocumented landforms such as meltwater channels and ribbed moraines. It was compiled using a combination of satellite imagery (Landsat 7 ETM+) and field observations (for details see Hättestrand and Clark, 2006a) and is the highest-resolution peninsula-wide geomorphological map currently available. Only glacial landforms within Murmansk Oblast (border indicated by black outline) are included; contemporary fluvial networks and lakes are not shown to avoid confusion with the mapped palaeo-meltwater channels.

In this contribution, we provide a critical review of all known (to the authors) previously published empirical data and interpretations of Late Weichselian glaciation on the Kola Peninsula and Russian Lapland. Our main aims are to discuss and critically evaluate (where possible) all available (i) glacial geomorphological mapping, sedimentary analyses, and numerical dating studies, including (for the first time) information published in Russian-language journals, (ii) conflicting interpretations of glacial dynamics, and (iii) LGIT glaciation scenarios and Younger Dryas ice margin positions. We stress *where possible*, as not all information presented in this paper can be critically evaluated, especially where data are sparse (or indeed the only data that exist) in a particular region, or where glacial interpretations have been presented with limited supporting data. In addition, we present a database of all known (to date) numerical ages from the Kola Peninsula and Russian Lapland. By collating, discussing, and critically evaluating empirical data and interpretations, this paper provides a resource for future researchers to inform FIS dynamics at both a regional- and ice sheet-scale. In addition, it acts as a critical first step in providing a framework through which numerical ice sheet models can be constrained. All such information is crucial to furthering our understanding of contemporary ice sheet dynamics in other Arctic, Antarctic, and Alpine regions. Given recent advances in palaeo-ice sheet reconstruction, we recommend that further work is needed in the form of a revised glacial reconstruction using high-resolution, regional-scale geomorphological data to determine, in greater detail, the pattern, style, and timing of Late Weichselian glaciation on the Kola Peninsula and Russian Lapland.

2. Methods

In this study, we critically evaluate all known published glacial geomorphological mapping, sedimentary analyses, and numerical dating studies and associated interpretations of glacial dynamics from the Kola Peninsula and Russian Lapland. Our approach is based on methods outlined by Hughes et al. (2016) and Stroeven et al. (2016) in which geomorphological data are examined in a Geographic Information System (GIS) setting using ESRI® ArcMap™ 10.7.1. Published shapefile data were imported and scanned maps were georectified, allowing comparisons to be made with published mapped data. To maintain the integrity of the original scanned maps, no cartographic changes were made to the landforms (Fig. 2 to 10). Generalised ice margin retreat patterns are reconstructed from the pattern of eskers and meltwater channels, ice-margin positions are identified from the spatial distribution of moraines (e.g. Hättestrand and Clark, 2006a; Petrov et al., 2014), and ice flow configurations are interpreted from generalised flow patterns of subglacial landforms (e.g. Kleman et al., 1997; Hättestrand and Clark, 2006b; Winsborrow et al., 2010). Ice retreat behaviour (i.e. still stands, readvances) are identified from sedimentary analyses (e.g. Lunkka et al., 2018; Vashkov and Nosova, 2018a). Glaciation timings are based on interpretations of numerical age estimations from previous glacial reconstructions (e.g. Stroeven et al., 2016; Lunkka et al., 2018). The resultant glacial models are not accurate 'timeslice reconstructions' (e.g. Hughes et al., 2016; Stroeven et al., 2016), rather they are visualisations and summaries of the proposed glaciation theories that are intended to inform numerical ice sheet models.

In addition, we have compiled a database of all known previously published numerical ages (to date) from within the Kola Peninsula and Russian Lapland relevant to the advance and retreat of ice during the Late Weichselian (Fig. 3). This includes dates derived from glacial and non-glacial studies (e.g. basal radiocarbon dates from Holocene climate reconstruction studies). This database guides our discussion of previously published glacial reconstructions on the Kola Peninsula and Russian Lapland, and, along with the critical review of published materials, is intended to be used to inform future glacial reconstructions of the region.

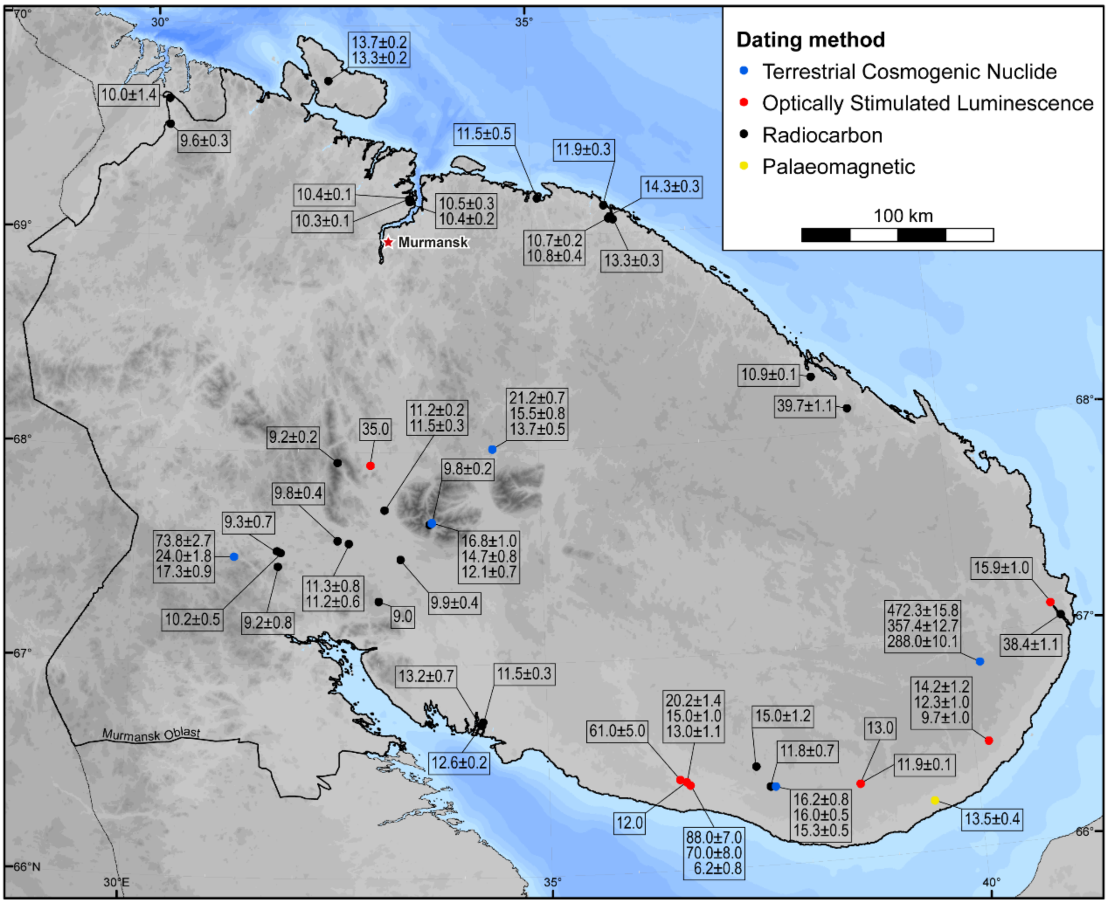

Fig. 3: Spatial distribution of numerical dates included within the database associated with the Late Weichselian glaciation of the Kola Peninsula and Russian Lapland (n=66). Multiple numerical dates from individual sites are grouped. Dating is predominantly from coastal areas and close to inhabited areas on the central Kola Peninsula (i.e. areas that are relatively easy to access in this largely remote region). All numerical dates included in the database can be viewed in Data S1 and S2 in the Supplementary Materials.

Each numerical age was entered into an Excel spreadsheet (see *Data S1* in the Supplementary Materials) and used to create an ArcGIS point shapefile (.shp; see *Data S2*) for use within a GIS. Each date is attributed to the source publication; where dates are obtained from secondary publications, citation is made to both the original source and secondary publication. The shapefile retains all metadata (Table 1) from the spreadsheet as fields within the associated attribute table. In some instances, geographical co-ordinates were missing (either from the original source or secondary publication). In such cases, where maps were provided alongside the published dates, co-ordinates were derived by identifying the site on Google Earth mapping software. Where neither co-ordinates nor map locations were given, the co-ordinate information was derived using a site name search in Google Earth mapping software; these sites are flagged within the metadata as "N/A". Coordinates are reported in decimal degrees and the WGS84 datum, following the recommendations of Hughes et al. (2016).

KOLA_ID	Unique database identification number
Location	Site name Latitude and Longitude: °N °E (WGS84) Comment on the precision of the location if not from original source e.g. taken from map, based on place name, etc.
Sample information	Laboratory ID number Site type: lake core, sediment exposure, surface Elevation (m above sea level) Depth (m) (if applicable) Terrestrial Cosmogenic Nuclide data: thickness (cm), density (g/cm^3), shielding, erosion rate (cm/yr), ^{10}Be concentration and uncertainty (atoms/g), ^{10}Be standard Dated material
Dating method	Radiocarbon (^{14}C) and Accelerator Mass Spectrometry Radiocarbon (^{14}C AMS); optically stimulated luminescence (OSL); Terrestrial Cosmogenic Nuclide (TCN^{10}Be); palaeomagnetic
Ages	Uncalibrated/uncorrected age and error (as reported) Calibrated/corrected age and error. Radiocarbon ages calibrated to INTCAL20 (Reimer et al., 2020). TCN^{10}Be ages recalculated using 'Arctic' production rate (Young et al., 2013). Any pertinent comment (e.g. reliability of date)
Citation information	Source reference (author, year, Digital Object Identifier (DOI)) Compilation reference (author, year) (where applicable)

Table 1: Metadata recorded for each date obtained from landforms and sediments across the Kola Peninsula and Russian Lapland; these metadata are included in both the database table and shapefile attributes (Data S1 and S2 in the Supplementary Materials).

Sixty-six known numerical ages from 31 locations across the Kola Peninsula and Russian Lapland (within Murmansk Oblast) are included within the database (see Fig. 3 and *Data S1*). To maintain consistency across the dataset, numerical ages were recalculated where appropriate. Radiocarbon ages were recalibrated using the OXCAL v4.4 radiocarbon calibration software with the INTCAL20 calibration curve (Reimer et al., 2020), and are reported here as cal. kyr BP (i.e. before 1950). Where radiocarbon dates cannot be recalibrated because insufficient data are provided within the original source (or secondary) publication, the reported date is recorded and flagged as “No details given” within the metadata. Terrestrial Cosmogenic Nuclide (TCN) age estimations are recalculated using the CRONUS-EARTH online calculator v3.0 (Balco et al., 2008) and the ‘Arctic’ production rate calibration dataset of Young et al. (2013); this was chosen on account of the study region’s position between 66° and 70°N within the Arctic circle. Ages derived from optically stimulated luminescence (OSL) and palaeomagnetic techniques are not recalculated since such dates are clearly reported with sufficient background data (e.g. dose rates) in order to confidently reassess their quality (Hughes et al., 2016). Where TCN, OSL, and palaeomagnetic dates are reported with insufficient data within the source (or secondary) publication, the reported date is recorded and flagged as “No details given” within the metadata. TCN, OSL, and palaeomagnetic ages are reported relative to the year of sampling/analysis (ka) following standard convention. A consistent datum is not imposed across dates derived from different methods since the maximum deviation between dates is negligible in terms of the time-scales involved within the reconstructions. Individual dates are discussed alongside glacial geomorphological and sedimentary evidence and glacial interpretations, as well as being evaluated for their reliability. It should be noted that, in many cases, numerical ages in the region are published alongside insufficient contextual, stratigraphical, or sample property information and thus do not provide reliable age estimations. Here, we highlight such instances and emphasise that dates without reliable contextual information should be used with caution.**3.**

Evidence for glaciation of the Kola Peninsula and Russian Lapland

The Kola Peninsula and Russian Lapland has an abundance of glacial, periglacial, and paraglacial landforms and sediments. This section critically evaluates the existing evidence for, and interpretations of, Late Weichselian glaciation in the region. To facilitate this, the study area is divided into three sectors guided by topography and ice sheet signatures (Fig. 1c): (i) the Khibiny and

Lovozero mountains; (ii) the western Kola Peninsula and Russian Lapland; and (iii) the eastern Kola Peninsula.

3.1. Khibiny and Lovozero mountains

The Khibiny and Lovozero mountains (Fig. 1c) are horseshoe-shaped massifs located in the central western Kola Peninsula, that are renowned as the world's largest and second largest alkaline plutons, respectively (Kalashnikov et al., 2016). The Khibiny Mountains attain heights of 1,201 m above sea level (asl) and typically have narrow plateau summits over 1,000 m asl, while the Lovozero massif is characterised by broad plateaux >800 m asl, attaining heights of 1,120 m asl.

The geomorphology of the Khibiny and Lovozero mountains, including cirques, parabolic valleys, and relict plateau surfaces, is widely documented (e.g. Armand, 1960; Lavrova, 1960; Strelkov, 1976; Niemelä et al., 1993; Hättestrand and Clark, 2006a; Korsakova and Kolka, 2007; Hättestrand et al., 2008; Yevzerov, 2009; Yevzerov and Nikolaeva, 2010). Sediment-landform assemblages, including moraines and meltwater channels, suggest that both massifs were inundated by the FIS during the Late Weichselian (Armand, 1960; Lavrova, 1960; Hättestrand and Clark, 2006a; Hättestrand et al., 2008; Yevzerov, 2009; Yevzerov and Nikolaeva, 2010). However, the well-developed morphology of the cirques and parabolic valleys suggest erosion over multiple phases of localised glaciation during the Quaternary (Hättestrand and Clark, 2006a; Hättestrand et al., 2008).

Early investigations of cirques by Armand (1960) and Strelkov (1976) identify a series of boulder-strewn hummocky deposits banked against valley headwalls and spread across cirque floors. These authors interpret the deposits – which have arcuate, convex margins up-valley parallel to cirque headwalls, with concave or irregular down-valley margins – as forming between cirque glaciers and FIS glacier snouts. Later investigations identify a total of 16 such 'cirque infills', estimated to be up to 50 m thick (Fig. 4), comprising locally-sourced coarse boulders (Hättestrand and Clark, 2006a; Hättestrand et al., 2008; Yevzerov and Nikolaeva, 2008). In addition, hummocky moraines are mapped in cirques where cirque infills are not identified (Fig. 4; Hättestrand and Clark, 2006a;

Hättestrand et al., 2008). Hättestrand et al. (2008) interpret these cirque infill and hummocky moraine deposits as forming between glacier snouts and cirque headwalls, where glacier thinning allowed thick accumulations of supraglacial and/or frost shattered debris. Hättestrand et al. (2008) and Sugden and Hall (2020) further propose that cirque infills and hummocky moraines also formed under so-called blue-ice conditions, where ablation is dominated by sublimation in a dry continental climate. Similar conditions are thought to occur during the formation of supraglacial moraines in East Antarctica (Hättestrand and Johansen, 2005; Sugden and Hall, 2020).

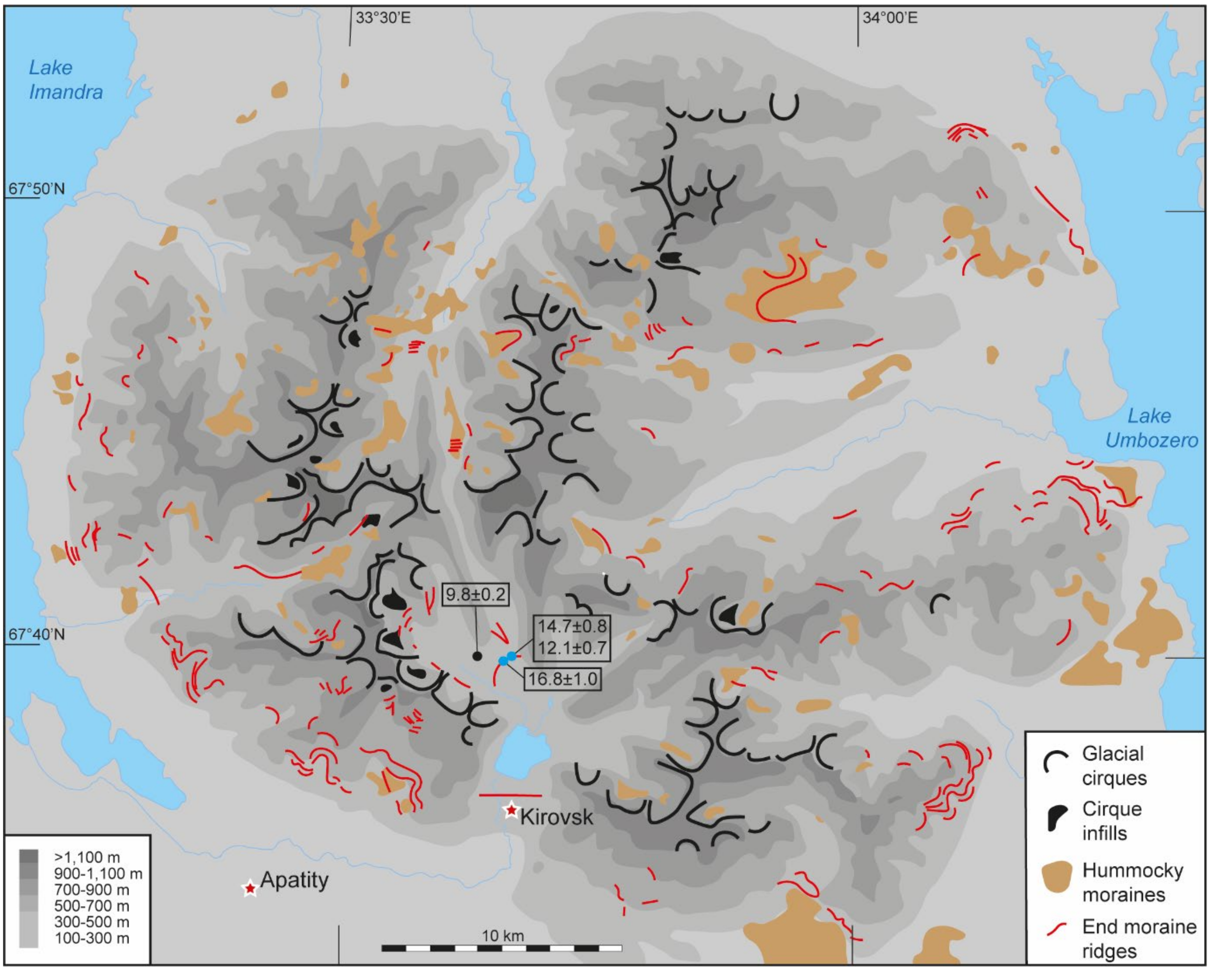

Fig. 4: Spatial distribution of mapped cirques, cirque infills, hummocky moraines, and end moraines in the Khibiny Mountains (modified from Hättestrand et al., 2008). The positions of all known numerical ages from the Khibiny Mountains are also shown (and included within the database – see Data S1). The location of Fig. 4 is shown in Fig. 2.

Studies also identify a series of arcuate end moraine ridges orientated up-valley (i.e. arcuate ridges with the vertex up-valley) in several valleys below the cirques at <600 m asl (Fig. 4; Armand, 1960; Strelkov, 1976; Hättestrand and Clark, 2006a, 2006b; Hättestrand et al., 2008; Yevzerov, 2009; Yevzerov and Nikolaeva, 2010). Such moraines are thought to be indicative of FIS thinning and ice lobe retreat, down-valley, during the LGIT (Hättestrand and Clark, 2006a, 2006b; Hättestrand et al., 2008; Yevzerov, 2009; Yevzerov and Nikolaeva, 2010). Additionally, palaeo-lake shorelines and glaciofluvial outwash terraces in some valleys in the Khibiny Mountains provide clear evidence that meltwater was impounded by glacier snouts (Armand, 1960; Hättestrand and Clark, 2006a). Palaeo-lake shorelines, glaciofluvial outwash, and lateral meltwater channels, first identified by Armand (1960) and later mapped in greater detail by Hättestrand and Clark (2006a), indicate the routing of meltwater across glacier surfaces, with meltwater accumulating in ice-dammed lakes (Armand, 1960; Hättestrand and Clark, 2006a; Yevzerov and Nikolaeva, 2008; Yevzerov, 2009). Yevzerov (2009) also suggests that an ice-dammed lake formed in the Seydozerskaya depression in the Lovozero Mountains, although Hättestrand and Clark (2006a) do not report palaeo-lake shorelines indicative of ice-dammed lakes within the massif.

De Geer moraines, which form at the grounding line of water-terminating ice margins, are also identified in the Khibiny Mountains in association with palaeo-lake shorelines and terraces, and are thought to indicate the lateral retreat (rather than thinning) of glacier snouts (Hättestrand et al., 2008). Pękala (1998) and Yevzerov and Nikolaeva (2008) consider predominantly locally-sourced clasts within the glaciofluvial outwash and De Geer moraines to indicate local glaciation (i.e. cirque and valley glaciers and ice cap outlet glaciers). However, the orientation of lateral meltwater channels (which form along the lateral margins of glaciers), sub-parallel to contours, suggest that the glaciofluvial outwash deposits and De Geer moraines were probably formed by glaciers retreating down-valley (Hättestrand and Clark, 2006b; Greenwood et al., 2007; Hättestrand et al., 2008).

Large cross-valley end moraine ridges, that predominantly contain shieldrock erratics from the surrounding lowlands, are documented at valley mouths around the perimeter of the Khibiny and Lovozero mountains (Fig. 4, 5, and 6; Hättestrand and Clark, 2006a; Hättestrand et al., 2008; Yevzerov and Nikolaeva, 2008; Yevzerov, 2009; Yevzerov and Nikolaeva, 2010). These end moraines are thought to have been deposited by glaciers as they retreated out of the massifs (Fig. 5; Hättestrand et al., 2008; Yevzerov and Nikolaeva, 2010). Hättestrand and Clark (2006a) also identify a flight of prominent, continuous lateral meltwater channels on the southern flank of the Lovozero massif that are thought to indicate ice flow to the east during ice sheet thinning (Hättestrand and Clark, 2006b; Yevzerov, 2009; Yevzerov and Nikolaeva, 2010).

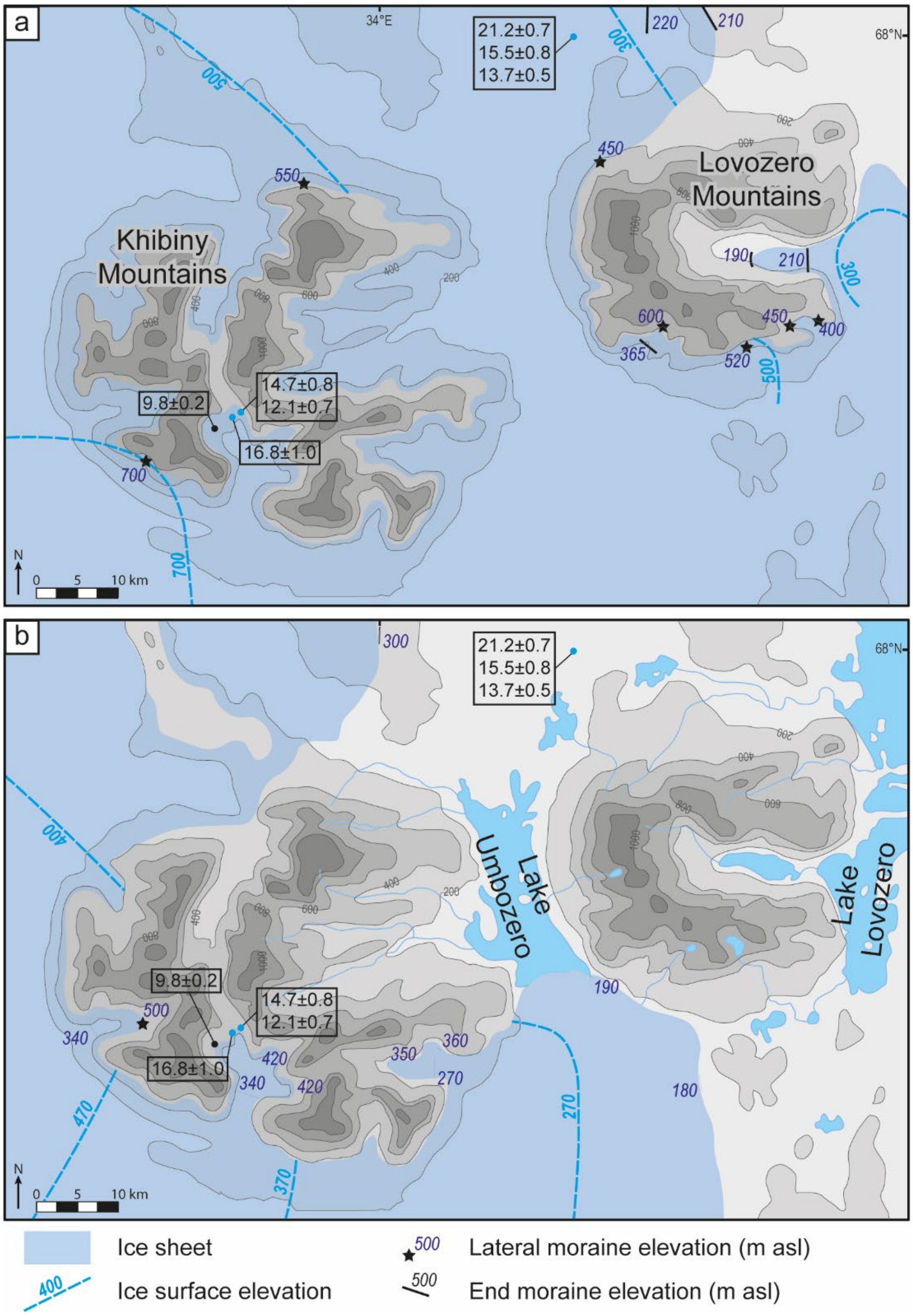

*Fig. 5: Fennoscandian Ice Sheet retreat pattern around the Lovozero and Khibiny Mountains during the (a) Older Dryas stadial (*c. *14.1-13.9 ka) and (b) the Younger Dryas stadial (*c. *12.9-11.7 ka) (modified from Yevzerov and Nikolaeva, 2010). The ice margin retreat pattern and ice surface elevations are reconstructed using the positions and elevations of moraines. Numerical ages included in the database are also shown. The location of Fig. 5 is shown in Fig. 2.*

Lavrova (1960) presents an early local-LGM reconstruction of FIS glaciation on the Kola Peninsula, proposing that relict plateaux summits of the Khibiny and Lovozero mountains were exposed as nunataks. However, lateral meltwater channels identified on some of the summit surfaces suggest that cold-based ice covered and preserved the relict plateaux summits (Armand, 1960; Hättestrand and Clark, 2006a; Hättestrand et al., 2008), indicating that initial FIS deglaciation is likely to have been characterised by ice sheet thinning (Fig. 5). The lack of meltwater landforms and dominance of cirque infill and hummocky moraine deposits between 600 and 1,000 m asl suggest ablation dominated by cool, dry conditions (Hättestrand and Clark, 2006b; Hättestrand et al., 2008; Yevzerov and Nikolaeva, 2010). Hättestrand et al. (2008) propose that these conditions existed during the Younger Dryas stadial, although as yet there are no numerical ages to verify this interpretation. Subsequent deglaciation likely occurred during climatic warming (although, to date, there are no published LGIT temperature estimates from this region) when meltwater was released during positive temperatures; this is indicated by the presence of ice-dammed lake features below 600 m asl (Armand, 1960; Hättestrand and Clark, 2006a; Hättestrand et al., 2008). The FIS ice lobes, flowing into the Khibiny and Lovozero mountains, are thought to have retreated down-valley with a number of standstills and/or readvances evidenced by large, cross-valley end moraines (Fig. 5) (Ekman and Iljin, 1991; Rainio et al., 1995; Hättestrand and Clark, 2006a; Yevzerov and Nikolaeva, 2010). As the FIS is thought to have retreated east to west, it is likely that the Lovozero Mountains were free of FIS ice before the Khibiny Mountains (Fig. 5), possibly prior to the Younger Dryas, although there are no numerical ages to verify this (Hättestrand and Clark, 2006b; Yevzerov, 2009; Yevzerov and Nikolaeva, 2010).

Yevzerov and Nikolaeva (2008) suggest that a prominent arcuate ridge in the central Khibiny Mountains – which predominantly contains shieldrock erratics from the surrounding lowlands, and which they further interpret as a push moraine – was formed during the Younger Dryas. However, TCN ages (16.8±1.0, 14.7±0.8, 12.1±0.7 ka; Stroeven et al., 2016) from the crest of this moraine (Fig. 3, 4, 5) suggest that this particular moraine ridge was formed prior to the Younger Dryas. In our view, the timing of formation is inconclusive due to the spread of the TCN dates. In addition, a single AMS radiocarbon date (9.8±0.2 cal. kyr BP) is available from plant macrofossils taken from a lake core retrieved from the distal side of the moraine ridge (Fig. 4). However, we do not regard this as deglacial in age since the date is retrieved from gyttja ~12 cm above the base of the core, rather than the lower ~10 cm silty gyttja basal lake sediments that Ilyashuk et al. (2013) consider to be characteristic of a recently deglaciated lake basin in the region. This age therefore represents a minimum deglacial date.

Following FIS retreat and the onset of the Younger Dryas cold stadial, Yevzerov and Nikolaeva (2008) propose that the Khibiny Mountains were glaciated by local ice masses (i.e. cirque glaciers, valley glaciers, or small ice fields). They suggest that this is indicated by glaciofluvial outwash terraces and moraines, composed of locally-sourced debris, which are likely to have formed at the margins of local topographically confined glaciers. However, the orientation of moraines and lateral meltwater channels, which all indicate FIS ice margins retreating down-valley (Hättestrand and Clark, 2006a; Hättestrand et al., 2008), does not support this conclusion. Hättestrand et al. (2008) also suggest that some cirque infill deposits in the Khibiny Mountains display clear evidence of overriding by glacier ice in the form of flutings, possibly by cirque glaciers. However, these authors state neither which, nor how many, cirque infills display flutings, and instead they use an example from the Lovozero Mountains to illustrate these features.

In contrast, hummocky moraines and arcuate end moraines aligned down-valley in the Lovozero Mountains (Fig. 6) clearly indicate cirque and valley glaciers, as well as a piedmont lobe glacier, that are thought to have existed during the Younger Dryas stadial (Hättestrand and Clark, 2006a; Hättestrand et al., 2008; Yevzerov, 2009). Additionally, the fluted surface of cirque infill deposits in the Lovozero Mountains suggests glacier overriding, possibly by a cirque glacier (Hättestrand et al., 2008). Overall, while the extent of localised glacial events in the Khibiny Mountains remains uncertain, there is clearer geomorphological evidence of localised glaciation in the Lovozero massif. However, numerical dating is required to verify the timings of localised glaciation events.

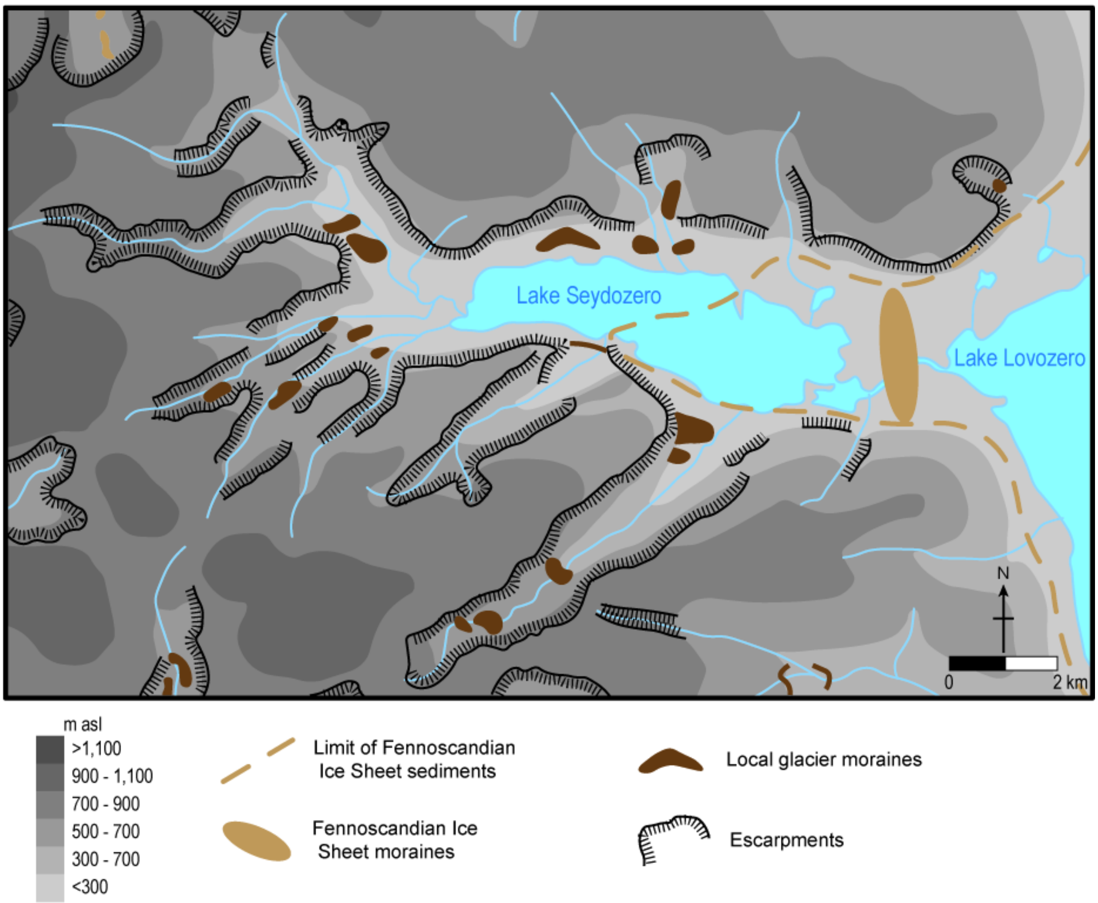

Fig. 6: Spatial distribution of moraines associated with the FIS and inferred local glaciers in the Seydozerskaya depression in the Lovozero Mountains (modified from Yevzerov, 2009). Smaller, topographically controlled ice-masses are thought to indicate local glaciation (i.e. cirque glaciers, valley glaciers, or small ice fields) after FIS deglaciation, possibly during the Younger Dryas stadial. The location of Fig. 6 is shown in Fig. 2.

3.2. Western Kola Peninsula and Russian Lapland

The western Kola Peninsula and Russian Lapland are discussed alongside each other on account of their similarities in terms of topography and glacial signatures. This western sector (Fig. 1c) is characterised by undulating topography, with mountainous terrain along the west-east axis of the region. The highest point (excluding the Khibiny and Lovozero mountains) is the Chuna Tundra Mountains (1,074 m asl) west of the Khibiny massif, although the majority of the terrain only attains a maximum altitude of 500 m asl.

3.2.1. Weichselian till deposits

The distribution and stratigraphic position of Quaternary sediments on the western Kola Peninsula and Russian Lapland are documented by Russian scientists in a series of studies from the mid- to late-20th Century (e.g. Lavrova, 1960; Grave et al., 1964; Nikonov, 1964, 1966; Apukhtin and Ekman, 1967; Armand et al., 1969; Apukhtin et al., 1977; Yevzerov and Koshechkin, 1991). However, historically this information has not been easily accessible to non-Russian scientists due to limited transfer of knowledge. Niemelä et al. (1993) compiles the available data into the first comprehensive Quaternary geological map, which details the spatial distribution of glacial sediments across the entire Kola Peninsula and Russian Lapland, facilitating ice sheet-scale glacial reconstructions. This map is converted to a digital format, with minor updates, by Petrov et al. (2014).

Several sedimentological investigations identify Eemian (*c.* 130-115 ka) interglacial deposits overlain by two till units (Grave et al., 1964; Nikonov, 1966; Apukhtin and Ekman, 1967; Armand et al., 1969; Yevzerov and Koshechkin, 1991; Petrov et al., 2014). The two tills are separated by non-glacial sediments, with several previous authors suggesting the two tills indicate two separate glaciations on the Kola Peninsula and Russian Lapland during the Weichselian (Grave et al., 1964; Nikonov, 1966; Apukhtin and Ekman, 1967; Armand et al., 1969; Yevzerov and Koshechkin, 1991; Ikonen and Ekman, 2001; Mangerud et al., 2004; Svendsen et al., 2004; Petrov et al., 2014; Hughes et al., 2016). However, the spatially discontinuous lower till unit has been identified only in the interior of the Kola Peninsula (Grave et al., 1964; Yevzerov and Koshechkin, 1991; Svendsen et al., 2004), which may indicate localised glaciation rather than an ice sheet. Svendsen et al. (2004) present OSL ages (KOLA_ID 61, Data S1) of glaciofluvial deltaic sediments from the "southern part of the peninsula" (p.1245) that suggest the lower till was deposited during the Early-Middle Weichselian (*c.* 115-40 ka), possibly by an ice mass flowing south. In addition, a single OSL date of ~35 ka (Fig. 3) from glaciofluvial sediments, which are overlain by the upper till, is thought to suggest that a significant part of the region was ice-free during the Ålesund interstadial (*c.* 38-34 ka) (Svendsen et al., 2004). However, these ages are not supported by contextual, stratigraphical, and age calculation information. As such, while we include the ages in our database they cannot be verified and therefore are listed as “N/A” in Data S1. Furthermore, the ages reported by Svendsen et al. (2004) may also be included in Mangerud et al. (2004) and Lunkka et al. (2018). However, since this is unclear, we include them here in Data S1.

3.2.2. Ice flow patterns

Previous authors argue that the extensive distribution and stratigraphic position of the upper till unit on the western Kola Peninsula and Russian Lapland demonstrates that the entire region was glaciated by the FIS during the LGM (Apukhtin and Ekman, 1967; Niemelä et al., 1993; Svendsen et al., 2004; Petrov et al., 2014). Kleman et al. (1997) attempt to reconstruct the FIS flow configuration in the region using glacial lineations (drumlins and mega-scale glacial lineations) identified by Niemelä et al. (1993). However, the resolution of the map precludes a detailed reconstruction. The region was subsequently investigated by Hättestrand and Clark (2006a) who identify glacial lineations (up to 10 km long and 1 km wide) and ribbed moraine. The orientations of these subglacial bedforms indicate a general west to east ice flow direction, suggesting that the FIS advanced east across Russian Lapland and the Kola Peninsula from the Scandinavian Mountains during the Late Weichselian (Hättestrand and Clark, 2006a; Winsborrow et al., 2010). However, previous authors suggest lineations and bedrock landforms, such as roche moutonnées, along the northern coast of the Kola Peninsula and Russian Lapland, indicate that ice flowed towards the northeast (Superson, 1994; Kleman et al., 1997; Hättestrand and Clark, 2006a; Yevzerov et al., 2007; Winsborrow et al., 2010). Authors such as Ekman and Iljin (1991) and Hättestrand and Clark (2006b) propose that the local-LGM ice divide was centred along the elevation axis (i.e. mountainous areas) of the western Kola Peninsula and Russian Lapland, with ice flowing towards the coast. However, since Kleman et al. (1997) and Winsborrow et al. (2010) consider many of the glacial lineations on the Kola Peninsula and Russian Lapland to be deglacial landforms, and because large areas of the interior of the FIS were probably cold-based (which will not permit the formation of subglacial bedforms) during the LGM (Kleman et al., 1997; Kleman and Hättestrand, 1999; Hättestrand and Clark, 2006b; Kleman et al., 2006; Winsborrow et al., 2010; Stroeven et al., 2016), we cannot say with certainty where the ice divide existed in this region, if at all.

Punkari (1993, 1995, 1997), and later Kleman et al. (1997), interpret abundant lineations and eskers in fan-shaped flow patterns as the locations of three deglacial palaeo-ice streams in the western Kola Peninsula and Russian Lapland (Fig. 7): (i) the White Sea Ice Stream, flowing west to east through Kandalaksha Gulf and along the White Sea; (ii) the Kola Ice Stream, flowing eastwards south of the Khibiny Mountains, after which it flows northeast; and (iii) the Tuloma Ice Stream, flowing south to north along the Russian-Norwegian border. However, the development of robust identification criteria of ice streaming – including highly convergent flow patterns, highly attenuated bedforms, and abrupt lateral margins (Stokes and Clark, 1999, 2001; Stokes et al., 2016b) – had not been established when these initial interpretations were made. Furthermore, recent advances in palaeo-ice stream identification suggest (i) the likely occurrence of ribbed moraines along ice stream pathways, which are thought to develop concomitantly with glacial lineations (Stokes et al., 2016b; Vérité et al., 2021), and (ii) esker development is not necessarily contemporaneous with ice streaming (Stokes and Clark, 1999, 2001; Stokes et al., 2009; Hughes et al., 2014). These recent advances in understanding, and identifying, palaeo-ice streaming cast doubt on the early ice streaming interpretations of Punkari (1993, 1995, 1997) and Kleman et al. (1997). In addition, despite these advances, subsequent studies have rarely reconsidered ice stream pathways and dynamics on the Kola Peninsula or Russian Lapland (*cf*. Boulton et al., 2001; Hättestrand and Clark, 2006a; Hättestrand et al., 2007; Winsborrow et al., 2010), nor the impact of ice streaming in this particular region on dynamics of the wider FIS (*cf*. Punkari, 1997; Boulton et al., 2001; Winsborrow et al., 2010; Stroeven et al., 2016).

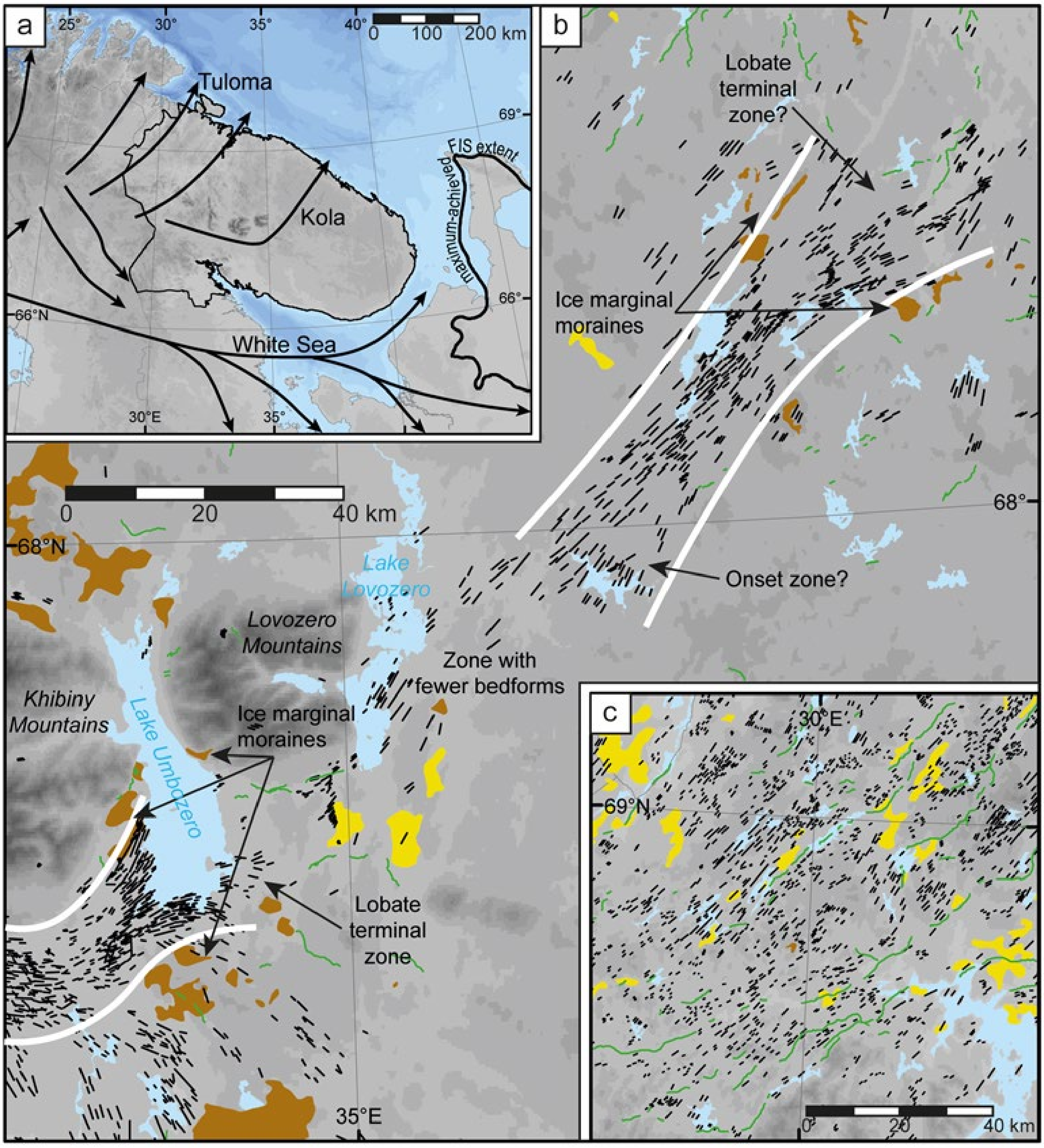

Fig. 7: (a) Locations of FIS ice stream pathways, modified from Punkari (1997). Three ice streams are thought to have glaciated the Kola Peninsula and Russian Lapland: (i) the White Sea Ice Stream, (ii) the Kola Ice Stream, and (iii) the Tuloma Ice Stream. (b) Subglacial bedforms (lineations and ribbed moraine), moraines, and eskers mapped by Hättestrand and Clark (2006a) at the eastern section of the Kola Ice Stream. Here, Punkari (1993, 1995, 1997) suggests the ice stream pathway flowed around the Khibiny and Lovozero mountains. However, subglacial bedforms and moraines mapped by Hättestrand and Clark (2006a) indicate there may have been two separate ice stream pathways (indicated by the white lines). (c) Subglacial bedforms (lineations and ribbed moraine), moraines, and eskers mapped by Hättestrand and Clark (2006a) in the Tuloma river basin. Punkari (1993, 1995,

1997) interpret abundant lineations and eskers as evidence of ice streaming. The locations of panels b and c is shown in Fig. 2; the symbology of lineations, ribbed moraines, hummocky moraines, and eskers matches that in Fig. 2.

Densely spaced, highly attenuated lineations at the head of the Kandalaksha Gulf (Fig. 1c) are widely interpreted as evidence of ice streaming in the White Sea basin (Kleman et al., 1997; Boulton et al., 2001; Hättestrand and Clark, 2006a; Hättestrand et al., 2007; Winsborrow et al., 2010). These reconstructions suggest the ice stream obliquely crossed the elongated White Sea depression (Fig. 7a; Punkari, 1993, 1995; Kleman et al., 1997; Punkari, 1997; Boulton et al., 2001), but detailed geomorphological data of the White Sea floor are lacking, and the ice stream pathway cannot be verified. Flowsets (groupings of bedform patterns over large areas) later constructed by Winsborrow et al. (2010) depict shifting flow patterns over time. Northwest-southeast orientated lineations at the head of the Kandalaksha Gulf and south of the Khibiny Mountains are thought to indicate ice flowing into the White Sea during the eastward advance of the FIS across Russian Lapland and the Kola Peninsula (Hättestrand and Clark, 2006a; Winsborrow et al., 2010). Winsborrow et al. (2010) therefore suggest the west-east orientated lineations, in a fan-shaped distribution, are associated with the retreat stages of the White Sea ice stream. Regardless, these data imply that the White Sea Ice Stream pathway shifted over time.

Highly attenuated lineations along the Kola Ice Stream pathway display highly convergent flow patterns (Fig. 2, 7b; Hättestrand and Clark, 2006a; Winsborrow et al., 2010), which appear to undergo a flow direction change from eastwards to northeast (Fig. 7a; Punkari, 1993, 1995, 1997; Winsborrow et al., 2010). This is the only palaeo-ice stream in the FIS to have such a profound flow direction change (Punkari, 1993, 1995; Kleman et al., 1997; Punkari, 1997; Boulton et al., 2001; Winsborrow et al., 2010). However, flowset reconstructions by Kleman et al. (1997) and Winsborrow et al. (2010) separate lineations along the Kola Ice Stream pathway into two individual flowsets, one terminating in Lake Umbozero south of the Khibiny Mountains and the other beginning ~30 km northeast at the eastern margins of the Lovozero Mountains (Fig. 7b). In addition, lineations and hummocky moraines mapped by Hättestrand and Clark (2006a) display fan-shaped flow patterns at the head of each flowset, typical of the terminal zone of ice streams (Fig. 7b) (Stokes and Clark, 1999, 2001). These observations suggest the Kola Ice Stream is in fact two palaeo-ice streams rather than one. It is likely that the two flowsets on the Kola Peninsula were active at different times during the LGIT, but additional data on bedform superimposition are required to determine relative ages.

In northern Russian Lapland, ice streaming is identified from abundant flow parallel, highly attenuated lineations and eskers orientated approximately south-north (Punkari, 1993, 1995, 1997; Boulton et al., 2001). Fields of ribbed moraine (Fig. 2, 7c), identified by Hättestrand and Clark (2006a), support ice streaming interpretations, and may indicate the lateral margins of individual ice stream trunks within the Tuloma Ice Stream (Stokes et al., 2016b; Vérité et al., 2021). However, there are several instances of cross-cutting bedforms and apparent divergent lineations orientated west-east (Hättestrand and Clark, 2006a). Punkari (1993, 1995, 1997) and Kleman et al. (1997) suggest cross-cutting bedforms are produced time-transgressively during ice margin retreat. In contrast, we suggest cross-cutting bedforms and divergent lineation orientations may represent multiple flow events (not necessarily ice streaming) in the geomorphological record that require further investigation. In addition, esker development is not considered contemporaneous with ice streaming (Stokes and Clark, 1999, 2001; Greenwood and Clark, 2009a; Stokes et al., 2009; Hughes et al., 2014). Instead, eskers in the region are more likely to indicate warm-based ice margin retreat following ice streaming.

3.2.3. Ice marginal zones

Areas of hummocky moraine and moraine ridges in the western Kola Peninsula and Russian Lapland are widely considered to be ice marginal zones indicative of retreating ice margin positions (Fig. 2) (Ramsey, 1898; Lavrova, 1960; Ekman and Iljin, 1991; Niemelä et al., 1993; Yevzerov and Kolka, 1993; Rainio et al., 1995; Yevzerov and Nikolaeva, 1997, 2000; Hättestrand and Clark, 2006a; Yevzerov, 2015). Sedimentary studies on moraines in this area identify glaciotectonised sediments overlain by undeformed glacial sediments (Yevzerov and Kolka, 1993; Yevzerov, 2015, 2017; Vashkov and Nosova, 2018b; Yevzerov, 2018; Vashkov and Nosova, 2019; Vashkov, 2020). For example, the lower units of a prominent ridge near the southern coast (Fig. 8) display folded glacial diamicton within which there are sheared lenses of clayey sediments. Vashkov and Nosova (2018a) interpret these lower units as deformation beneath an advancing ice margin. Overlying the lower deformed sediments, Vashkov and Nosova (2018a) interpret laminated diamicton with clasts displaying preferred southwest-northeast orientations as lodgement till deposited beneath the ice margin. The upper units of the morainic ridge are characterised by undeformed laminated sands and gravels, and small boulders within a clayey-sandy matrix, which Vashkov and Nosova (2018a) interpret as glaciofluvial sediments and ablation tills. Consequently, Vashkov and Nosova (2018a) propose that this prominent ridge formed at an oscillating ice margin that underwent a period of sustained readvance during retreat (Fig. 8). Similar internal structures are identified in morainic deposits more widely across the western Kola Peninsula (e.g. Yevzerov and Kolka, 1993; Yevzerov, 2015, 2017; Vashkov and Nosova, 2018b; Yevzerov, 2018; Vashkov and Nosova, 2019; Vashkov, 2020), which we suggest indicate oscillatory retreat and periods of sustained readvance of the FIS during the LGIT.

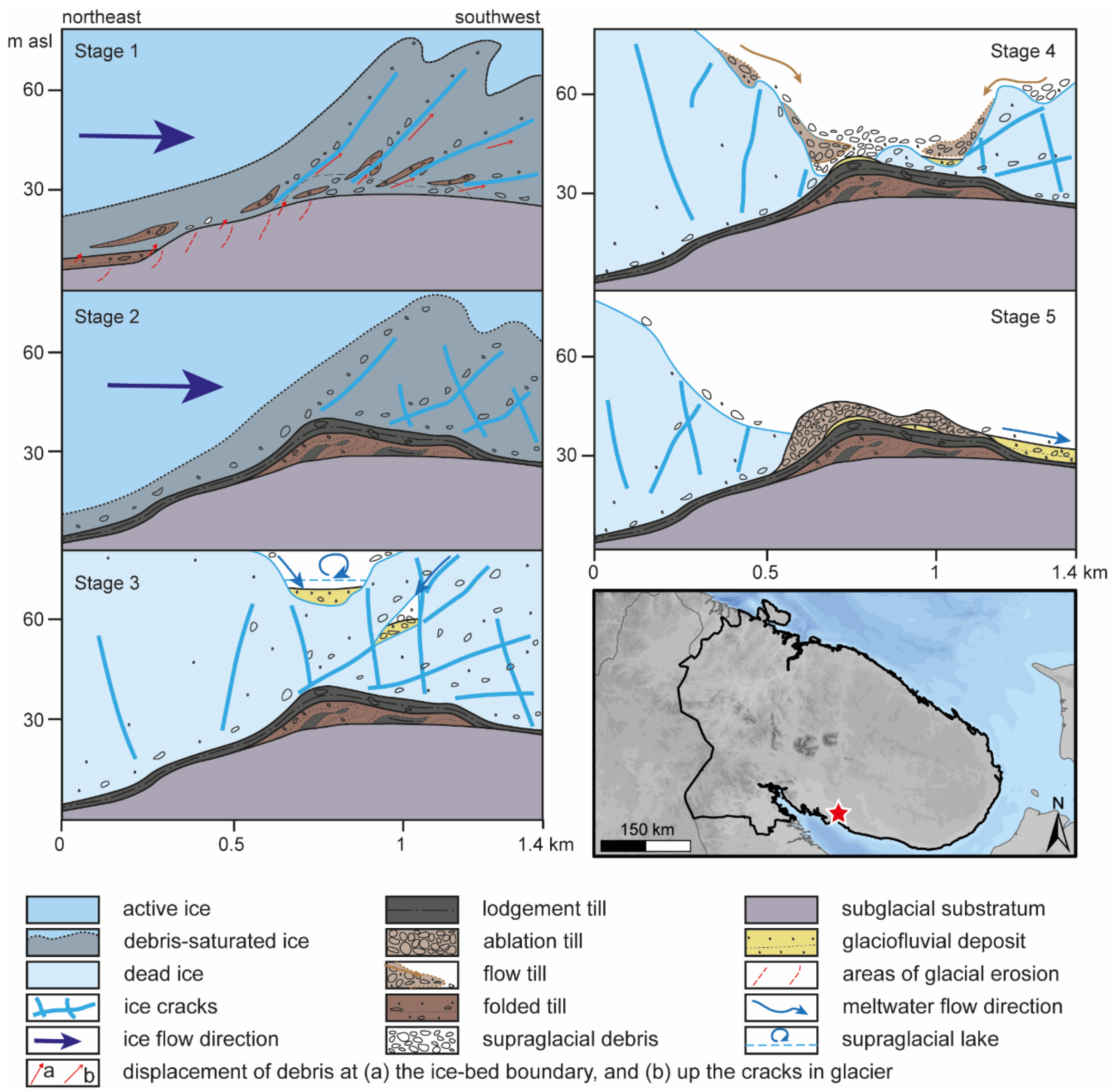

Fig. 8: Reconstruction of the formation of a moraine ridge on the southern coast of the western Kola Peninsula (modified from Vashkov and Nosova, 2018a). The location of the sediment exposure is indicated by the red star on the inset map (bottom right).

Ekman and Iljin (1991) present the first synthesis of ice margin positions based on morainic landforms and attempt to correlate them with known glacial stages, using morphostratigraphy (Fig. 9). On the western Kola Peninsula and Russian Lapland, Ekman and Iljin (1991) and Yevzerov and Nikolaeva (2000) identify two ice marginal zones (zones I and II) dominated by ring and ridge hummocky moraines and end moraine ridges (Fig. 9). However, hummocky moraines and end moraines are not necessarily synchronous over large distances, so inferring age from correlation is fraught with difficulties. This can be particularly problematic where ice marginal landform evidence is lacking. For example, many of the ice marginal zones inferred by Ekman and Iljin (1991) and Yevzerov and Nikolaeva (2000) are not identified by Hättestrand and Clark (2006a) (Fig. 9). Where moraines are lacking, reworked, or spatially discontinuous, Hättestrand and Clark (2006b) use the orientation and spatial distribution of lateral meltwater channels and eskers to reconstruct the ice margin retreat pattern. Their reconstruction suggests the ice margin retreated inland from the coasts on the western Kola Peninsula and Russian Lapland, following the ice flow configuration indicated by the orientation of glacial lineations (Hättestrand and Clark, 2006b; Winsborrow et al., 2010; Stroeven et al., 2016). In addition, this reconstruction suggests increasingly topographically controlled ice margins due to thinning of the FIS as it retreated across Russian Lapland (Hättestrand and Clark, 2006b). However, while Hättestrand and Clark (2006b) note that the ice marginal zones inferred by Ekman and Iljin (1991) and Yevzerov and Nikolaeva (2000) corresponds with their ice margin reconstruction from the meltwater landform record in coastal areas, the inferred ice margin retreat patterns are almost perpendicular across much of the western Kola Peninsula and Russian Lapland. As such, a new geomorphological map of this region based on high-resolution remotely-sensed imagery and/or field mapping is required to resolve these conflicts.

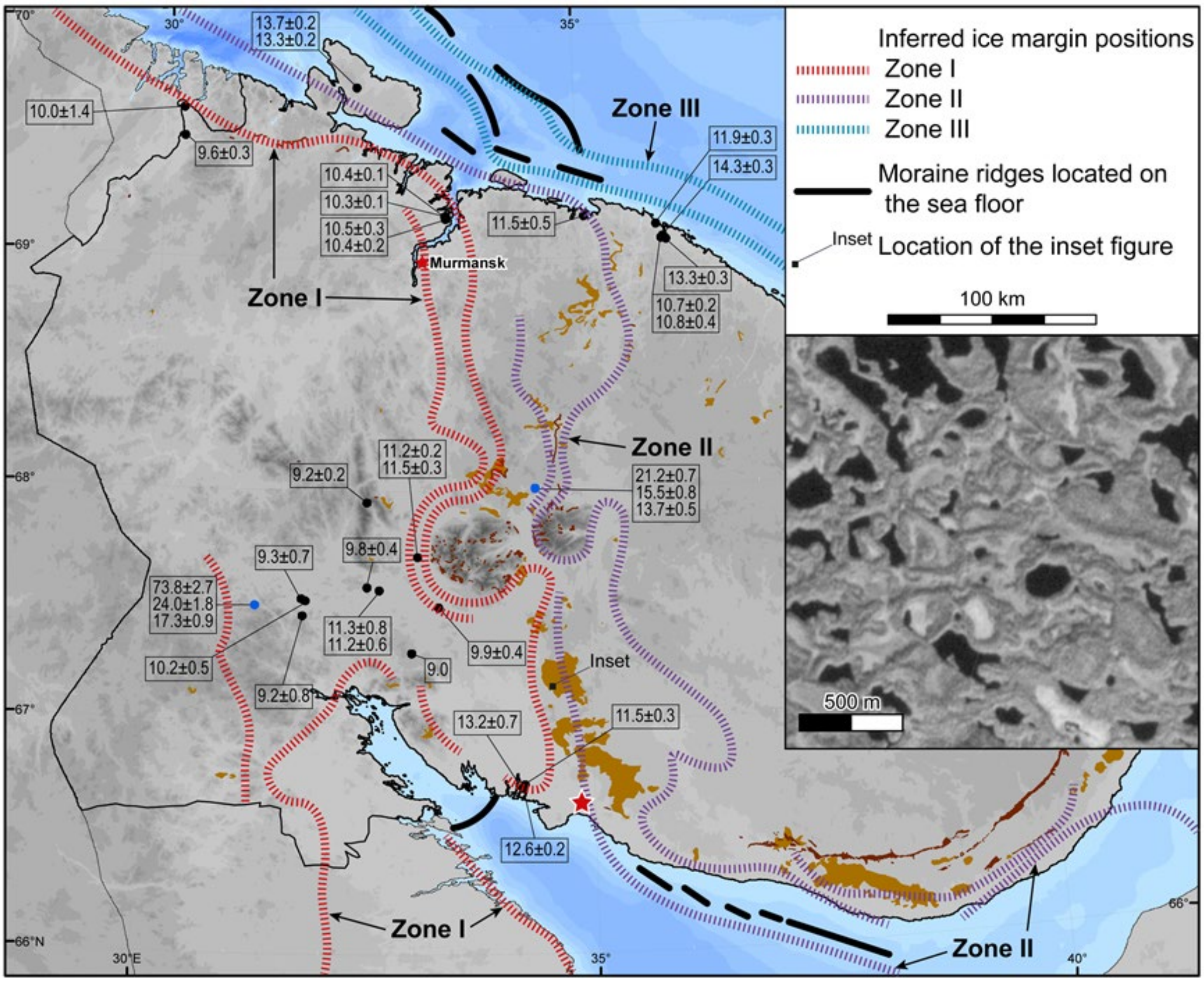

Fig. 9: Ice marginal zones of the Kola Peninsula and Russian Lapland, inferred from the spatial distribution of moraines by Ekman and Iljin (1991) (modified from Yevzerov and Nikolaeva, 2000). Moraines mapped by Hättestrand and Clark (2006a) (using the same symbology as Fig. 2) and moraine positions on the sea floor identified by Yevzerov and Nikolaeva (2000) are included for comparison. The positions of numerical dates discussed in Section 3.2 are also shown. Yevzerov and Nikolaeva (2000) correlate each ice marginal zone to ice marginal landforms identified elsewhere in Fennoscandia, based on their morphostratigraphic relationships: zone I – Younger Dryas (c. 12.9-11.7 ka); zone II – Older Dryas (c. 14.1-13.9 ka); zone III – Oldest Dryas (c. 14.7-16.1 ka). The location of the sedimentary reconstruction from Fig. 8 is indicated by the red star. Inset: examples of ring and ridge moraines, the most prolific hummocky moraine type on the western Kola Peninsula.

Based on morphostratigraphy, Yevzerov and Nikolaeva (2000) attribute ice marginal zone II, comprising north-south trending moraine assemblages (Fig. 9), to the Older Dryas (*c.* 14.1-13.9 ka) stadial. Although Yevzerov and Nikolaeva (2000) do not support this with numerical age estimates, a number of numerical ages included within Data S1 (Fig. 3) can be associated with this ice marginal zone. For example, Snyder et al. (1997) and Kremenetski et al. (2004) investigated lake sediments on the northern coastline of the Kola Peninsula, ~30 km east of hummocky moraines mapped by Hättestrand and Clark (2006a), using AMS radiocarbon dating. Although these dates provide some constraint on the timing of deglaciation in the area, only the 13.3±0.3 cal. kyr BP date (Snyder et al., 1997) can be interpreted as representing a minimum deglaciation age. This is because the macrofossils for this date were sampled directly above basal diamict (Snyder et al., 1997). We regard the older age estimation (14.3±0.3 cal. kyr BP) presented by Snyder et al. (1997), which is sampled ~15 cm above basal diamict in non-laminated, sandy marine sediments, to be an overestimation due to an increased marine reservoir effect and/or reworking of sediments. Snyder et al. (1997) also present an AMS radiocarbon date (11.9±0.3 cal. kyr BP) of algal macrofossils sampled in laminated lake sediments ~3 cm above non-laminated, sandy marine sediments; however, with a sampling location ~86 cm above the basal sediments, the date cannot represent the initiation of ice-free conditions. We interpret dates presented by Kremenetski et al. (2004) (10.7±0.2, 10.8±0.4 cal. kyr BP) as representing ice-free conditions during the Younger Dryas, but not the timing of deglaciation. This is because the pollen stratigraphy of the lake core suggests ice-free conditions prior to the Younger Dryas, and records the cooling climate during the Younger Dryas through the rise of *Salix* and corresponding increase of *Artemisia* pollen (Kremenetski et al., 2004). Yevzerov (2012) investigated lake sediments on the Rybachy Peninsula, west of the sites sampled by Snyder et al. (1997) and Kremenetski et al. (2004), near the Norwegian border. We interpret bulk radiocarbon dates (13.3±0.2, 13.7±0.2 cal. kyr BP; Yevzerov, 2012) as representing a minimum deglaciation age, because the dated plant macrofossils were sampled directly above a basal diamict. Stroeven et al. (2016) use TCN methods to date a number of boulders to the south of the sites investigated by Snyder et al. (1997) and Kremenetski et al. (2004), ~10 km north of the Lovozero Mountains. While two of the dates (13.7±0.5, 15.5±0.8 ka) presented by Stroeven et al. (2016) may indicate the timing of deglaciation in ice marginal zone II, the TCN dates cannot be verified due to a lack of contextual information such as whether the sampled boulders are associated with moraines identified at the margins of the Lovozero Mountains. Additionally, a third TCN age (21.2±07; Stroeven et al., 2016) from the same location is possibly an inherited age. We suggest this because the age predates the local-LGM (Yevzerov, 2015; Stroeven et al., 2016), but again, a lack of contextual information precludes interpretation of this date.

Ice marginal zone I, comprising north-south trending moraine assemblages on the western Kola Peninsula and east-west trending moraine assemblages on northern Russian Lapland (Fig. 9), is considered by many to be the last significant zone of ice marginal landforms in northwest Arctic Russia (Lavrova, 1960; Strelkov, 1976; Ekman and Iljin, 1991; Niemelä et al., 1993; Yevzerov and Kolka, 1993; Andersen et al., 1995a; Rainio et al., 1995; Yevzerov and Nikolaeva, 2000; Svendsen et al., 2004; Hättestrand and Clark, 2006b, 2006a; Hughes et al., 2016; Stroeven et al., 2016). As such, this ice marginal zone has been correlated with the Rugozero and Kalevala moraines in the Republic of Karelia, northwest Russia (Ekman and Iljin, 1991; Rainio et al., 1995; Yevzerov and Nikolaeva, 2000; Kolka et al., 2013; Yevzerov, 2015), and the Tromsø-Lyngen moraines in Norway (Ekman and Iljin, 1991; Rainio et al., 1995; Yevzerov, 2015), and is attributed to the Younger Dryas stadial.

Unfortunately, chronological dating of the zone I moraine assemblages on the Kola Peninsula and Russian Lapland is lacking. Several studies have focussed on dating lake sediments within the vicinity of ice marginal zone I, which can be used in an attempt to chronologically constrain the moraines to the Younger Dryas stadial, and provide insights into Younger Dryas ice dynamics on the western Kola Peninsula and Russian Lapland (Fig. 3, 9). Kolka et al. (2013) present “bulk organic” (p.75) radiocarbon dates (11.5±0.3, 12.6±0.2, 13.2±0.7 cal. kyr BP) of gyttja sampled above glacial varves identified in lake cores ~10 km southeast of the Younger Dryas ice marginal zone on the White Sea coast. While Kolka et al. (2013) interpret these numerical ages to indicate ice-free conditions at this location during the Younger Dryas, they do not discuss the possibility that the glacial varves may indicate sedimentation during the Younger Dryas stadial. ~100 km northwest, Kolka et al. (2020) investigate lake sediments ~15 km southwest of the Khibiny Mountains using bulk radiocarbon dating. Kolka et al. (2020) interpret the lacustrine basal sandy sediments to indicate sedimentation during the early Holocene when the lake was connected to a much larger lake system. Thus, the 9.9±0.4 cal. kyr BP date of gyttja sampled directly above the abrupt transition from sandy to organic sediments in the lake core reflects the isolation of the present-day lake (Kolka et al., 2020). As such, the radiocarbon date presented by Kolka et al. (2020) can only indicate ice-free conditions after deglaciation. On the western flanks of the Khibiny Mountains, in Lake Imandra (Fig. 1c, 4), Lenz et al. (2021) present AMS radiocarbon dates from a sedimentary sequence of silty and very coarse gravel basal sediments, glacial varves, and overlying silty gyttja. Lenz et al. (2021) suggest that the radiocarbon dates from the basal sediments and glacial varves are biased from ancient carbon introduced into the lake basin, and therefore interpret the 11.2±0.2 and 11.5±0.3 cal. kyr BP dates, which are taken from the silty gyttja layer, to represent ice free conditions after the Younger Dryas stadial. Lenz et al. (2021) further suggest deglaciation took place during the Bølling-Allerød interstadial (*c.* 14.7-12.9 ka). However, the landform assemblage around Lake Imandra and in the Khibiny Mountains indicates that the location where Lenz et al. (2021) retrieved their dates is on the proximal side of ice marginal zone I, which is widely interpreted as the Younger Dryas ice marginal zone (Lavrova, 1960; Strelkov, 1976; Ekman and Iljin, 1991; Niemelä et al., 1993; Yevzerov and Kolka, 1993; Andersen et al., 1995a; Rainio et al., 1995; Yevzerov and Nikolaeva, 2000; Svendsen et al., 2004; Hättestrand and Clark, 2006b, 2006a; Hughes et al., 2016; Stroeven et al., 2016). This suggests Lake Imandra would have been occupied by the FIS during the Younger Dryas stadial. Moreover, the assumption that the varved sediments are proglacial (Lenz et al., 2021) may not necessarily be true, as they could be associated with a subglacial lake (Livingstone et al., 2015; Shackleton et al., 2018). We, therefore, suggest sediment preservation beneath the FIS – in, for example, subglacial lakes – should be considered when interpreting sediments from large lakes like Lake Imandra. Other lake sediments within ice marginal zone I were sampled along the northern Russian Lapland coastline. For example, Corner et al. (1999) present two bulk radiocarbon dates (9.6±0.3, 10.0±1.4 cal. kyr BP) from lake cores near the Norwegian border, ~15 km to the south of the Younger Dryas ice marginal zone. We regard these dates as representing ice-free conditions after the Younger Dryas stadial. This is because the radiocarbon dated samples are from the centre of the lower, weakly laminated gyttja unit, ~205 and ~12 cm respectively from the base of the core (Corner et al., 1999). Similarly, we regard four AMS radiocarbon dates (10.3±0.1, 10.4±0.1, 10.4 ±0.2, 10.5±0.3 cal. kyr BP; Corner et al., 2001) from lake sediments ~30 km north of Murmansk to indicate ice-free conditions after the Younger Dryas stadial. This is because the relative position of the investigated lake is on the proximal side of the Younger Dryas ice marginal zone (Rainio et al., 1995; Yevzerov and Nikolaeva, 2000; Yevzerov, 2015). This is in contrast to Corner et al. (2001), who argue that pollen stratigraphic evidence from this lake core suggests that the area around Murmansk was deglaciated before the onset of the Younger Dryas.

Several numerical ages are published for southern Russian Lapland, on the proximal side of ice marginal zone I, which can be used to provide insights into the retreat of the FIS after the Younger Dryas stadial (Fig. 3, 9). Tolstobrova et al. (2016) investigate lake sediments south of the Chuna Tundra Mountains (Fig. 1c) using bulk gyttja radiocarbon dating. The diatom assemblages in the basal sediments suggest deposition in a proglacial environment (Tolstobrova et al., 2016). We therefore suggest the 11.2±0.6 and 11.3±0.8 cal. kyr BP dates represent a minimum deglaciation age. In addition, other radiocarbon ages from southern Russian Lapland can only be used to confirm ice-free conditions after the Younger Dryas: (i) bulk radiocarbon ages (9.2±0.8, 9.3±0.7, 9.8±0.4, 10.2±0.5 cal. kyr BP) of gyttja above possible proglacial basal lake sediments are associated with Boreal sedimentation (Nikolaeva et al., 2015; Nikolaeva et al., 2016); and (ii) an AMS radiocarbon date (9.2±0.2 cal. kyr BP; Solovieva and Jones, 2002) of lake sediments in the Chuna Tundra Mountains, which we consider to be representative of ice-free conditions because it is unclear whether the sampled gyttja is from basal lake sediments. A final bulk radiocarbon date (9.0 kyr BP), presented by Davydova and Servant-Vildary (1996), from a lake ~60 km southeast of the Chuna Tundra Mountains is not supported by chronological or stratigraphical information. As such, we cannot use it to verify the timing of ice-free conditions. Stroeven et al. (2016) sample a number of boulders and cobbles to the southwest of the Chuna Tundra Mountains for TCN dating (17.3±0.9, 24.0±1.8, 73.8±2.7 ka; Fig. 3, 9). Although it is possible that these dates reflect inherited ages since they pre-date the local-LGM, a lack of contextual information precludes interpretations of ice dynamics at this location.

3.3. Eastern Kola Peninsula

The eastern Kola Peninsula is characterised by low-relief topography with weathered bedrock surfaces and tors, exhibiting an east-west aligned elevation axis rising to a maximum elevation of 410 m asl. The spatial distribution and stratigraphic context of Quaternary sediments on the eastern Kola Peninsula is reasonably well documented (e.g. Ramsey, 1898; Lavrova, 1960; Nikonov, 1966; Apukhtin and Ekman, 1967; Apukhtin and Yakovleva, 1967; Armand et al., 1969; Krasnov et al., 1971; Apukhtin et al., 1977; Niemelä et al., 1993; Petrov et al., 2014).

3.3.1. Keiva Ice-marginal Zone

The largest glacial landform system on the eastern Kola Peninsula is the ~300 km long Keiva Ice-marginal Zone (KIZ), which is aligned approximately parallel to the Tersky coastline (Fig. 10; Svendsen et al., 2004; Hättestrand and Clark, 2006a; Hättestrand et al., 2007). In early geomorphological investigations, Grigoryev (1934), Vvdenski (1934), and Richter (1936) identify two morainic ridges on the Tersky coastline; a smaller fragmentary morainic ridge and a larger morainic ridge named ‘Keiva I’ and ‘Keiva II’, respectively (Fig. 10; Lavrova, 1960). However, subsequent geomorphological investigations (e.g. Armand, 1960; Apukhtin and Ekman, 1967; Apukhtin and Yakovleva, 1967; Krasnov et al., 1971; Strelkov, 1976; Punkari, 1985, 1993, 1995; Grosswald, 1998; Boulton et al., 2001; Grosswald and Hughes, 2002; Demidov et al., 2004; Svendsen et al., 2004; Hättestrand and Clark, 2006a; Hättestrand et al., 2007) identify three prominent, near-continuous morainic ridges (named 'Keiva I', 'Keiva IIa', and 'Keiva IIb' by Hättestrand et al., 2007), as well as hummocky moraines and glaciofluvial deposits. These are overridden by glacial lineations and interspersed with meltwater channels and eskers, the latter of which grade into both sides of the

Keiva II deposits (Fig. 10). There are contrasting sediment-landform assemblages on either side of the KIZ. On the Tersky coastline side of the KIZ there is an abundance of glacial landforms and sediments, whereas west and north of the KIZ, on the central-eastern Kola Peninsula, there are no mapped subglacial bedforms and sediments (Apukhtin and Ekman, 1967; Svendsen et al., 2004; Hättestrand et al., 2007; Petrov et al., 2014).

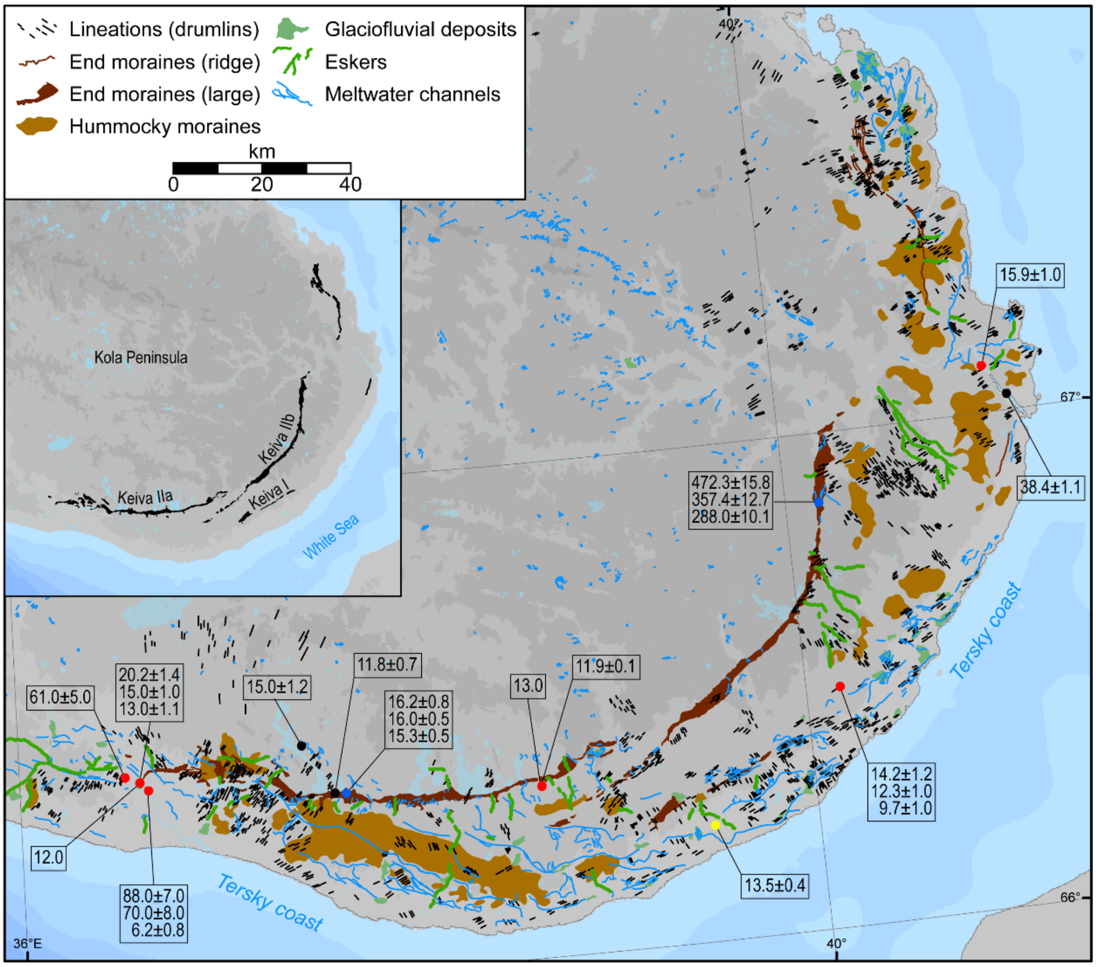

Fig. 10: The glacial geomorphology of the Keiva Ice-marginal Zone on the eastern Kola Peninsula (modified from Hättestrand and Clark, 2006a; Hättestrand et al., 2007). The positions of numerical dates included in Data S1 in the Supplementary Materials are also shown.

Despite the impressive geomorphology in this region, sedimentary descriptions representative of the entire KIZ are limited due to the inaccessibility of the area. Bakhmutov et al. (1992), Korsakova (2009), and Lunkka et al. (2018) identify cross-bedded medium- to coarse-grained sand deposits (which they interpret as glacial outwash sediments) at several sedimentary exposures along the KIZ (Fig. 10). In addition, at the western end of the KIZ, Lunkka et al. (2018) document two matrix-supported, silty-sandy diamicts with occasional cobble- and pebble-sized clasts, interpreting the sandy diamicts as tills separated by sand and silt deposits of possible fluvial or glaciolacustrine origins. We suggest that these two stratigraphically distinct till units, which are positioned above an Eemian interstadial marker horizon (Korsakova, 2009; Lunkka et al., 2018), along with numerous lateral meltwater channels across the eastern Kola Peninsula (Hättestrand and Clark, 2006a), indicate that the region must have been glaciated during the Weichselian. This is in direct contrast to Ramsey (1898), Apukhtin and Ekman (1967), and Krasnov et al. (1971), among others, who suggest that the region largely was ice-free during the Weichselian. Furthermore, it is widely suggested that the two distinct till units indicates two different major glacial advances in northwest Arctic Russia during the Weichselian (Kjær et al., 2003; Mangerud et al., 2004; Larsen et al., 2006; Korsakova, 2009; Lunkka et al., 2018). This interpretation is supported by several numerical ages. For example, Lunkka et al. (2018) present OSL dates (70.0±8.0, 88.0±7.0 ka) of fluvial sediments retrieved from between the two till in the western end of the KIZ (Fig. 3, 10). Although these dates provide some constraint on the timing of glaciation at the KIZ, they are more useful in determining the timing of ice-free conditions between the deposition of the two till units. Lunkka et al. (2018) suggest a third OSL date (61.0±5.0 ka) from sands deposited in a possible glaciolacustrine environment is inaccurate because sediments are poorly bleached due to deposition in a deep-water environment.

Punning et al. (1971) provide additional radiocarbon dates from shell fragments from the northeastern coastline of the Kola Peninsula, ~200 km north and northeast of the OSL dates taken by Lunkka et al. (2018). We infer these radiocarbon dates to represent ice-free conditions *c.* 38 ka. However, the first date (38.4±1.1 cal. kyr BP) is not provided with stratigraphical information (Punning et al., 1971), and so cannot be correlated with any glacial deposits in the region. The second date (39.7±1.1 cal. kyr BP) is slightly more reliable, since it is provided with some, ableit limited, stratigraphic information, which in turn suggests the dated macrofossil sample is from a sedimentary unit “overlying reddish-brown loamy moraine” (p.82, Punning et al., 1971). We suggest, therefore, this radiocarbon age could indicate ice-free conditions after the deposition of the lower Early-Middle Weichselian till.

Early investigations concluded that the disparity in the density and preservation of glacial landforms on the eastern Kola Peninsula, in comparison to the west, is a consequence of an independent ice mass (the Ponoy Ice Cap), thought to have been centred on the higher ground at ~400 m asl throughout the Late Weichselian (Armand, 1960; Strelkov, 1976). More recent published interpretations of the landforms on the eastern Kola Peninsula now suggest the FIS glaciated the entire peninsula (e.g. Lavrova, 1960; Ekman and Iljin, 1991; Kleman et al., 1997; Demidov et al., 2006; Hättestrand and Clark, 2006a; Hättestrand et al., 2007; Hughes et al., 2016; Stroeven et al., 2016). In the absence of subglacial landforms and sediments (Fig. 2; Hättestrand and Clark, 2006a; Petrov et al., 2014), many authors interpret an abundance of lateral meltwater channels (aligning west to east) as indicating cold-based ice sheet conditions during the Late Weichselian (Kleman and Hättestrand, 1999; Hättestrand and Clark, 2006b; Kleman et al., 2008; Stroeven et al., 2016). In contrast, an abundance of subglacial landforms, including glacial lineations and eskers, on the Tersky coastline suggest warm-based ice sheet conditions in the White Sea basin (Fig. 2, 10; Hättestrand and Clark, 2006a, 2006b; Hättestrand et al., 2007; Petrov et al., 2014; Stroeven et al., 2016). We suggest the KIZ is therefore likely to be an interlobate formation that developed between an ice stream in the White Sea and a predominantly stagnant FIS on the eastern Kola Peninsula; this is supported by Punkari (1985, 1993, 1995), Boulton et al. (2001), and Hättestrand et al. (2007). Routing of meltwater and trapping of glacial sediments between the eastwardly extending, cold-based FIS on the Kola Peninsula and the advancing White Sea ice lobe likely formed the Keiva II moraines (Punkari, 1993, 1997; Boulton et al., 2001; Hättestrand et al., 2007). Hättestrand et al. (2007) propose that an apparent break and overlap between the Keiva IIa and IIb moraines is a consequence of oscillatory advances of the White Sea ice lobe, induced by climatic fluctuations during the Late Weichselian. However, because there is a lack of subglacial bedforms and sedimentary evidence from the KIZ, we are not able to verify this. Subsequent local-LGM ice dynamics are also uncertain on the eastern Kola Peninsula owing to the absence of extensive sediment-landforms associations. Ekman and Iljin (1991), Svendsen et al. (2004), and Lunkka et al. (2018), among others, propose that an ice dome established itself on the central eastern Kola Peninsula. However, geomorphological and sedimentary evidence is not available to substantiate this interpretation. In contrast, others suggest that ice streaming in the White Sea dominated FIS dynamics on the eastern Kola Peninsula, resulting in the drumlinisation and deformation of the KIZ (Punkari, 1985, 1993; Boulton et al., 2001; Hättestrand and Clark, 2006a; Hättestrand et al., 2007; Hughes et al., 2016). Some numerical ages along the KIZ could be associated with the early formational stages of the KIZ. For example, Lunkka et al. (2018) OSL date sands in a deltaic deposit at the western end of Keiva IIa moraine (20.2±1.4 ka). Although this could be interpreted as deposition prior to the local-LGM, Lunkka et al. (2018) interpret this age as an overestimation due to the potential for incomplete bleaching of sediments in an inferred deep water, glaciolacustrine environment. Stroeven et al. (2016) also present TCN dates from boulders (northeast from the OSL date presented by Lunkka et al. (2018)) on the Keiva IIb moraine. The three anomalously older TCN dates (288.0±10.1, 357.4±12.7, 472.3±15.8 ka; Stroeven et al., 2016) probably indicate preservation of relict rock surfaces, thus resulting in inherited ages (Fig. 3, 10). However, Stroeven et al. (2016) do not provide contextual information with these ages, and so they cannot be verified.

3.3.2. Deglaciation theories on the eastern Kola Peninsula

During the LGIT, two styles of glaciation have been proposed on the eastern Kola Peninsula: (i) the separation of an ice dome that centred on the central eastern Kola Peninsula, which established itself as the dynamically independent Ponoy Ice Cap (see section 4.3.3 and Strelkov, 1976; Rainio et al., 1995; Svendsen et al., 2004; Astakhov et al., 2016; Lunkka et al., 2018); and (ii) FIS and White Sea ice lobe retreat (Figure 8a, b; e.g. Lavrova, 1960; Kleman et al., 1997; Boulton et al., 2001; Hättestrand and Clark, 2006a; Yevzerov, 2015; Hughes et al., 2016; Stroeven et al., 2016).

Proponents of the Ponoy Ice Cap theory suggest that the disparity of the sediment-landform assemblages between the eastern and western Kola Peninsula indicates the FIS was unlikely to have glaciated the eastern peninsula during the LGIT and that a dynamically different ice mass must have existed (Armand, 1960; Strelkov, 1976; Svendsen et al., 2004; Astakhov et al., 2016). In support of this theory, the lack of subglacial landforms and sediments on the central eastern Kola Peninsula is suggested to indicate that the Ponoy Ice Cap was mostly stagnant throughout the LGIT, only undergoing advances in response to Younger Dryas climatic cooling (Rainio et al., 1995; Svendsen et al., 2004; Lunkka et al., 2018). Proponents of the Ponoy Ice Cap glaciation theory regard the LGIT as the main formation stage of the KIZ (Rainio et al., 1995; Svendsen et al., 2004; Astakhov et al., 2016; Lunkka et al., 2018). A number of numerical ages support this. For example, Lunkka et al. (2018) investigate the age of the KIZ using OSL dating on deltaic glaciofluvial sediments deposited in shallow glaciolacustrine environments across the Keiva II moraines. Although Lunkka et al. (2018) consider several OSL dates to be indicative of Late Weichselian glaciation by the Ponoy Ice Cap (13.0±1.1, 15.0±1.0, 15.9±1.0 ka; Fig. 3, 10), they only interpret the 11.9±0.1 ka from sand and gravel deltaic foresets, possibly deposited in a glaciolacustrine basin, as indicative of a Younger Dryas ice margin of the Ponoy Ice Cap along the Keiva II moraines. Svendsen et al. (2004) present additional OSL dates for the Keiva II moraines. Although the 12 and 13 ka dates from “glaciofluvial deltas associated with the Keiva II moraines” (p.1246, Svendsen et al., 2004) suggest a Younger Dryas ice margin position, a lack of stratigraphic and chronological information provided with these dates means they cannot be used robustly. Furthermore, since it is unclear whether the ages reported by Svendsen et al. (2004) are included in Mangerud et al. (2004) and Lunkka et al. (2018), we include them in Data S1. Lunkka et al. (2018) also investigate glacial sediments in the Keiva I moraine and identify glaciotectonised glaciofluvial sediments surrounding a glacial till, which suggests an advancing White Sea ice lobe. Lunkka et al. (2018) interpret OSL dates (9.7±1.0, 12.3±1.0, 14.2±1.2 ka) of these glaciofluvial sediments as the likely deposition of till during the Younger Dryas.

In contrast, proponents of FIS deglaciation on the eastern Kola Peninsula argue that the west to east orientation of lateral meltwater channels – which are overlooked by supporters of the Ponoy Ice Cap theory – indicate a westward retreating ice margin (Hättestrand and Clark, 2006b; Stroeven et al., 2016). Hättestrand and Clark (2006b) argue this invalidates the concept of the Ponoy Ice Cap, which would have produced lateral meltwater channels orientated east to west. Consequently, several previous authors suggest that the FIS ice margin retreated westwards inland from the coast of the eastern Kola Peninsula during the LGIT (Kleman et al., 1997; Boulton et al., 2001; Hättestrand and Clark, 2006b; Hättestrand et al., 2007; Hughes et al., 2016; Stroeven et al., 2016). Several authors suggest that possible marine instability of the White Sea Ice Stream induced rapid deglaciation of the White Sea basin soon after the local-LGM (Punkari, 1993, 1995; Boulton et al., 2001; Hättestrand and Clark, 2006b; Hättestrand et al., 2007; Stroeven et al., 2016), leaving the eastern Kola Peninsula and White Sea basin ice free before the Younger Dryas stadial (Hättestrand and Clark, 2006b; Stroeven et al., 2016). Alternatively, Hughes et al. (2016) suggest a more gradual retreat of the White Sea ice lobe, which may have extended beyond the Younger Dryas FIS margin. However, the geomorphological and sedimentary evidence that is necessary to reconstruct glaciation in the White Sea basin is lacking (*cf*. Petrov et al., 2014) and is largely restricted to the Tersky coastline (Fig. 10), so we cannot say with certainty which theory is correct.

Several numerical dates (using multiple techniques) have been published for the KIZ (Fig. 3, 10) that can be used to constrain the timing of deglaciation on the eastern Kola Peninsula. For example, Stroeven et al. (2016) TCN date boulders at the western end of the KIZ. These dates (15.3±0.5, 16.0±0.5, 16.2±0.8 ka; Stroeven et al., 2016) suggest deglaciation *c.* 16 ka, but since they are not presented with contextual information they cannot be confidently employed. However, in support of the TCN dates, is a bulk organic radiocarbon date presented by Kremenetski and Patyk-Kara (1997), which is taken from a lake on the northern margins of the Keiva IIa moraine, ~1 km from the TCN dates. We regard this 15.0±1.2 cal. kyr BP date of sand-rich basal sediments as a minimum deglaciation age. Pollen analyses of overlying loam sediments indicate arboreal vegetation after deglaciation, which radiocarbon dating (11.8±0.7 cal. kyr BP; Kremenetski and Patyk-Kara, 1997) suggests occurred during the Younger Dryas stadial. Kolka (1996) also identifies 606 glacial varves within a deltaic deposit north of the Keiva I moraine (Bakhmutov et al., 1992, 1993, 1994; Kolka, 1996). We regard a 13.5±0.4 ka palaeomagnetic date from these glacial varves (Bakhmutov et al., 1992, 1993, 1994; Kolka, 1996) to be a minimum deglaciation age. Finally, Grönlund and Kauppila (2002) bulk radiocarbon dated lake sediments on the northern coastline of the Kola Peninsula (Fig. 3, 10). Although a radiocarbon date of gyttja sampled ~5 cm above the bottom of the core may represent a minimum Younger Dryas deglaciation age, we regard the 10.9±0.1 cal. kyr BP date as representing ice-free conditions during the Younger Dryas. This is because Grönlund and Kauppila (2002) suggest that the basal gravel layer identified in the sediment core, which they found to be impenetrable, may not actually be basal lake sediments, but rather marine sediments possibly deposited after deglaciation.

4. Glacial reconstructions for the Kola Peninsula and Russian Lapland

The preceding review demonstrates that the Kola Peninsula and Russian Lapland are rich in glacial landforms and sediments. Despite this, the pattern, style, and timing of glaciation during the Late Weichselian across the region remains under debate (Hughes et al., 2016; Stroeven et al., 2016). In this section we draw together the current understanding of glaciation on the Kola Peninsula and Russian Lapland.

4.1. Ice sheet build-up

While sedimentary and dating evidence of two separate till deposits above the Eemian marker horizon across the Kola Peninsula could suggest an oscillating ice margin, the two till units are more widely considered to indicate there were two separate major glacial episodes during the Weichselian (Grave et al., 1964; Nikonov, 1966; Apukhtin and Ekman, 1967; Grave and Koshechkin, 1969; Apukhtin et al., 1977; Svendsen et al., 2004; Demidov et al., 2006; Hughes et al., 2016; Lunkka et al., 2018; Korsakova et al., 2019). Several authors suggest the lower till unit represents an ice mass on the Kola Peninsula during the Early and Middle Weichselian (Fig. 11a; Nikonov, 1966; Grave and Koshechkin, 1969; Apukhtin et al., 1977; Svendsen et al., 2004; Lunkka et al., 2018; Korsakova et al., 2019), although the maximum extent and configuration of this ice mass is uncertain. Glaciofluvial sediments above the lower till suggest that the Early-Middle Weichselian ice mass retreated prior to subsequent FIS glaciation during the Late Weichselian, possibly in response to climatic warming during the Ålesund interstadial (Fig. 11b; Grave and Koshechkin, 1969; Svendsen et al., 2004; Hughes et al., 2016; Lunkka et al., 2018; Korsakova et al., 2019). The timing of ice-free conditions between the Early-Middle and Late Weichselian glaciations is also supported by several scattered numerical ages (Punning et al., 1971; Svendsen et al., 2004; Lunkka et al., 2018), although further stratigraphical and numerical dating evidence is required to fully constrain the ice-free period.

Subsequently, the FIS likely advanced eastwards across Russian Lapland and the Kola Peninsula from the Scandinavian Mountains (Fig. 11c), inundating any active cirque and valley glaciers in mountainous areas and later establishing the White Sea Ice Stream (Hättestrand et al., 2007; Hughes et al., 2016; Stroeven et al., 2016). We regard the local-LGM ice margin reconstruction presented by Larsen et al. (2014) as the best estimate for the maximum lateral extent of the FIS in northwest Arctic Russia (Fig. 11d). However, the maximum extent and the timing of the local-LGM in northwest Arctic Russia is uncertain, with variable (between 19 and 15 ka) supporting numerical age estimations from the Kanin Peninsula and elsewhere in northwest Russia (Fig. 11d; Svendsen et al., 2004; Linge et al., 2006; Hughes et al., 2016; Stroeven et al., 2016). Despite this, several authors believe the FIS reached its maximum extent on the Kola Peninsula later than elsewhere in Fennoscandia owing to its location in a precipitation shadow: a consequence of its distance from both the dominant moisture source of the Atlantic Ocean and the main ice nucleation centres over the Scandinavian Mountains (Hughes et al., 2016; Patton et al., 2016; Stroeven et al., 2016).

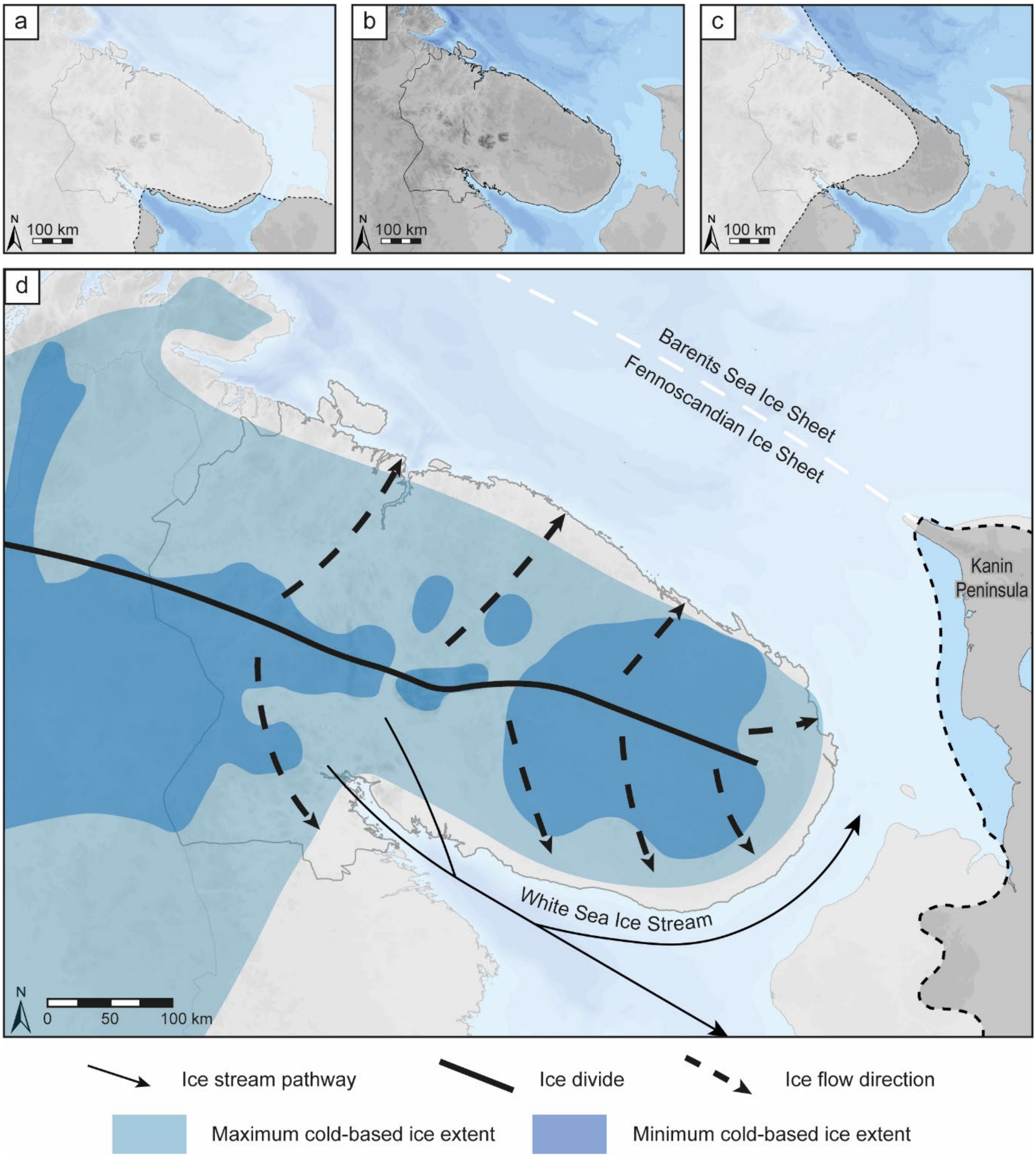

Fig. 11: Model of ice sheet build-up at various points during the Weichselian; (a) possible Early Weichselian (MIS 3) ice mass extent; (b) no independent ice mass on the Kola Peninsula during the Ålesund interstadial (c. 38-34 ka); (c) the FIS extending eastwards across the Kola Peninsula after c. 34 ka; and (d) local-LGM (c. 19-15 ka) FIS extent, showing ice flow directions, the White Sea Ice Stream pathway, and the minimum and maximum extent of cold-based ice (Ekman and Iljin, 1991; Punkari, 1995, 1997; Svendsen et al., 2004; Hättestrand and Clark, 2006b; Winsborrow et al., 2010;

Hughes et al., 2016; Stroeven et al., 2016; Batchelor et al., 2019). The position of the local-LGM ice sheet margin is based on the ice margin reconstruction presented by Larsen et al. (2014).

4.2. Ice dynamics during the local-LGM

During the local-LGM, the FIS ice divide on the Kola Peninsula and Russian Lapland was probably centred along the west-east elevation axis (Fig. 11d; Ekman and Iljin, 1991; Hättestrand and Clark, 2006b; Hättestrand et al., 2007). Ice flowed out radially to the coasts, as indicated by the orientation of glacial lineations (Fig. 11d; Ekman and Iljin, 1991; Svendsen et al., 2004; Hättestrand and Clark, 2006b; Stroeven et al., 2016). Areas with fewer subglacial landforms or glacial sediments indicate that large areas of ice on the Kola Peninsula and Russian Lapland were cold-based (Niemelä et al., 1993; Kleman et al., 1997; Kleman and Hättestrand, 1999; Hättestrand and Clark, 2006b; Kleman et al., 2008; Stroeven et al., 2016; Lunkka et al., 2018), although the maximum extent of cold-based ice is uncertain. Fig. 11d shows two possible scenarios. Niemelä et al. (1993) and Hättestrand and Clark (2006b) identify glacial lineations and ribbed moraines on the western Kola Peninsula and Russian Lapland, which they attribute to represent ice flow during the local-LGM. Thus, under the inferred minimum cold-based extent scenario, persistent cold-based ice was likely restricted to the central eastern Kola Peninsula as well as mountainous areas of the western Kola Peninsula and Russian Lapland (Niemelä et al., 1993; Hättestrand and Clark, 2006b). In contrast, Kleman et al. (1997) identify eskers alongside glacial lineations and ribbed moraines, which they suggest formed during the LGIT. As such, under the inferred maximum cold-based extent scenario, the majority of the Kola Peninsula and Russian Lapland was cold-based during the local-LGM (Kleman et al., 1997; Kleman and Hättestrand, 1999; Kleman et al., 2008; Stroeven et al., 2016).

However, densely spaced lineations at the head of the Kandalaksha Gulf and the KIZ are widely considered to represent persistent warm-based conditions in the White Sea basin during the local-LGM (Fig. 11d: Punkari, 1995; Kleman et al., 1997; Punkari, 1997; Hättestrand and Clark, 2006a; Hättestrand et al., 2007; Winsborrow et al., 2010). Similar to Hättestrand et al. (2007) and

Winsborrow et al. (2010), we suggest the White Sea Ice Stream likely flowed southeast through the Kandalaksha Gulf, and along the Kola Peninsula coastline towards the Barents Sea, where the ice stream probably dominated the ice dynamics of the peninsula-based ice mass. In addition, we suggest previous White Sea Ice Stream reconstructions by Punkari (1993, 1995, 1997) and Kleman et al. (1997), among others, which show ice flowing eastwards across Karelia (Fig. 7a), reflect ice stream configuration during the LGIT.

4.3. LGIT glaciation on the Kola Peninsula

Currently, there is no consensus surrounding the nature of the LGIT FIS glaciation in this region, and in particular the position of the Younger Dryas ice marginal zone (e.g. Ekman and Iljin, 1991; Andersen et al., 1995a; Rainio et al., 1995; Svendsen et al., 2004; Hättestrand and Clark, 2006a, 2006b; Hughes et al., 2016; Stroeven et al., 2016; Lunkka et al., 2018). As a consequence, there are several conflicting LGIT deglaciation theories (Fig. 12); these can be summarised into the following four models (which are evaluated in section 4.4).

4.3.1. Model 1: Initial deglaciation on the eastern Kola Peninsula

In this model, the central eastern Kola Peninsula was the first area to deglaciate, with ice lobes (believed to be former ice streams during the local-LGM; Punkari, 1995, 1997) occupying the surrounding lowlands and straits (Fig. 12a; Apukhtin and Ekman, 1967; Niemelä et al., 1993; Yevzerov and Nikolaeva, 2000; Yevzerov, 2001; Demidov et al., 2006). Niemelä et al. (1993) and Yevzerov (2001) suggest that the lack of glacial landforms and sediments on the central eastern Kola Peninsula indicates that this area must have been deglaciated before other parts of the peninsula, particularly where glacial lineations indicate the presence of ice lobes, such as in the White Sea. The style of deglaciation on the central eastern Kola Peninsula is proposed as either a complete loss of ice (indicated by the lack of glacial landforms and sediments) in the early Bølling interstadial (*c.* 14.7-14.1 ka; Apukhtin and Ekman, 1967; Niemelä et al., 1993), or large-scale stagnation and down-wasting, forming hummocky moraines at the margins of areas of dead ice (Demidov et al., 2006;

Yevzerov and Nikolaeva, 2007). This is similar to the style of deglaciation thought to have occurred on the FIS margin elsewhere in northwest Russia, where Demidov et al. (2006) identify large areas of hummocky moraine interspersed with ice-dammed lakes.

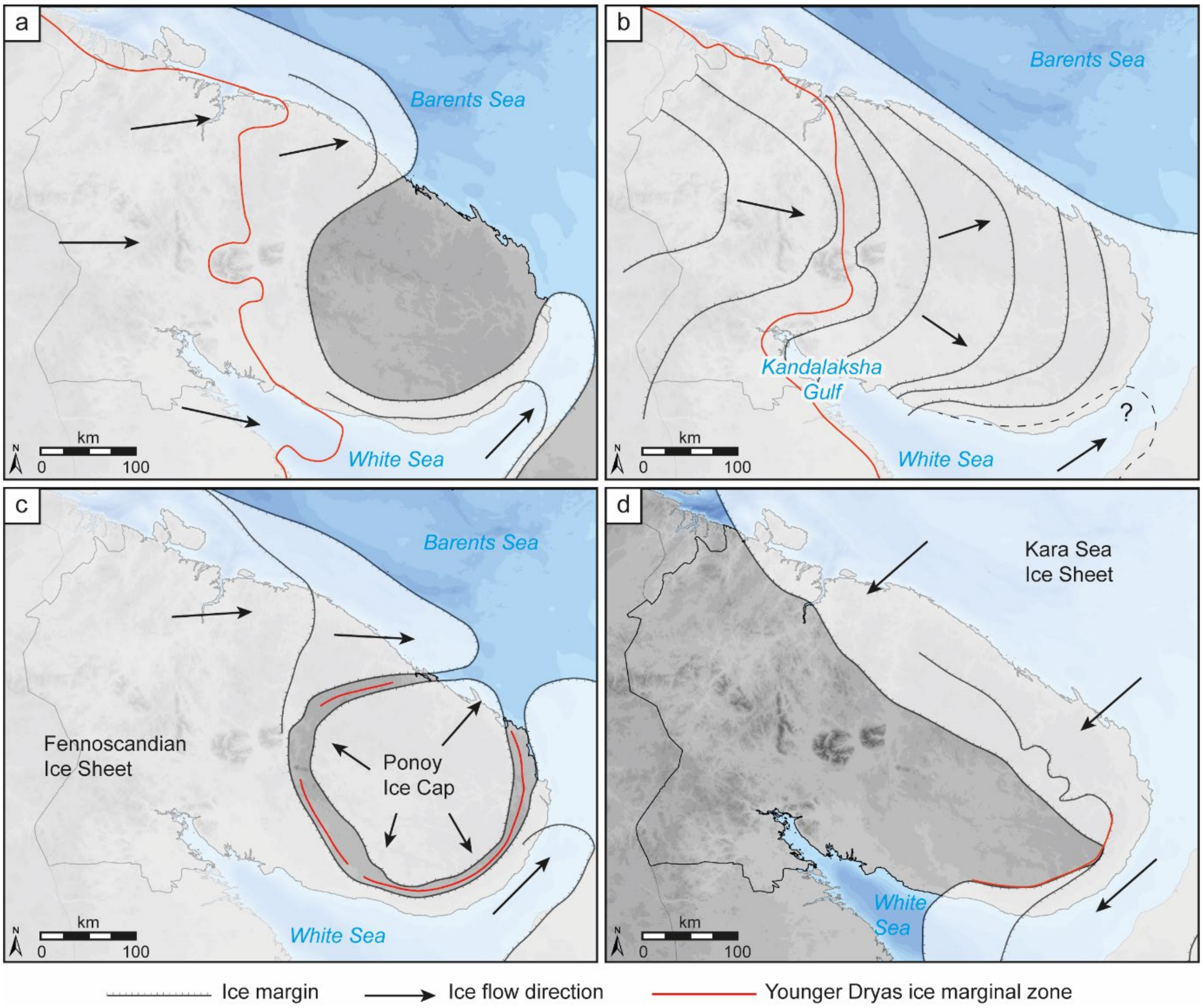

Fig. 12: Existing deglaciation models and associated possible Younger Dryas (c. 12.9-11.7 ka) ice marginal zones for the Kola Peninsula and Russian Lapland: (a) Model 1: Deglaciation of the Fennoscandian Ice Sheet (FIS) occurs on the eastern Kola Peninsula, leaving ice lobes occupying former ice stream pathways (Niemelä et al., 1993). Dead ice (areas of ice detached from the main ice sheet) may have occupied depressions (i.e. contemporary lakes) during deglaciation (Demidov et al., 2006); (b) Model 2: Coherent ice marginal retreat of the FIS, shown here with the development of an embayment in the Kandalaksha Bay (Hättestrand and Clark, 2006b). Alternatively, an ice lobe may have flowed into the White Sea (dashed line and question mark) as the FIS underwent coherent retreat (Hughes et al., 2016); (c) Model 3: The dynamically independent Ponoy Ice Cap, which deglaciated independently from the FIS (Rainio et al., 1995); and (d) Model 4: Retreat of the FIS

before the Younger Dryas stadial, followed by growth and advance of the Kara Sea Ice Sheet (KSIS) onto the northern Kola Peninsula and along the White Sea. Deglaciation of the KSIS occurred after the Younger Dryas (Grosswald and Hughes, 2002).

Following initial deglaciation of the central eastern Kola Peninsula, standstills and/or readvances of the FIS margin are documented on the western Kola Peninsula. A prominent belt of hummocky moraines aligned approximately north-south on the Kola Peninsula and moraine ridges east-west along the northern coast of Russian Lapland have been consistently attributed to the Younger Dryas FIS marginal zone (Fig. 12a; Ekman and Iljin, 1991; Niemelä et al., 1993; Yevzerov and Nikolaeva, 2000; Yevzerov, 2001). The hummocky moraine belt can be correlated with the Rugozero and Kalevala Younger Dryas moraines in Russia (Niemelä et al., 1993; Andersen et al., 1995a; Rainio et al., 1995), and the moraine ridges on the northern coastline are thought to be a continuation of the Tromsø-Lyngen Younger Dryas moraines in Norway (Andersen et al., 1995a; Andersen et al., 1995b; Olsen et al., 2013). Large, cross-valley end moraines in the Khibiny Mountains, also linked to the Younger Dryas, suggest the ice margin flanked the majority of the Khibiny massif (Yevzerov and Nikolaeva, 2010). The timing of the Younger Dryas ice margin is supported by numerous radiocarbon age estimations (10.3±0.1, 10.4±0.1, 10.0±1.4 10.4 ±0.2, 10.5±0.3 cal. kyr BP; Corner et al., 1999; Corner et al., 2001) from Russian Lapland and the Kola Peninsula.

4.3.2. Model 2: Coherent FIS retreat

This model depicts coherent east-west ice marginal retreat inland from the coastline of the Kola Peninsula (Fig. 12b; Punkari, 1993; Kleman et al., 1997; Boulton et al., 2001; Hättestrand and Clark, 2006a, 2006b; Hughes et al., 2016; Stroeven et al., 2016). In this model, the cold-based FIS on the Kola Peninsula was bordered by the White Sea Ice Stream along the southern and eastern coastlines during the local-LGM. The White Sea Ice Stream is thought to have collapsed and rapidly retreated into the Kandalaksha Gulf forming an embayment in the ice margin *c.* 14 ka (Hättestrand and Clark, 2006b; Stroeven et al., 2016) – this is supported by radiocarbon dates (11.5±0.3, 12.6±0.2, 13.2±0.7 cal. kyr BP; Kolka et al., 2013) that suggest the White Sea coastline was ice-free during the Younger

Dryas. This retreat pattern is based on the generally west-east orientation of meltwater landforms, which represent ice margins across the peninsula (Hättestrand and Clark, 2006b; Stroeven et al., 2016). Hughes et al. (2016) propose a variation to this model, suggesting that a White Sea ice lobe extended beyond the FIS margin on the Kola Peninsula (Fig. 12b).

Similar to Model 1, in Model 2 moraine ridges and hummocky moraines trending north-south are associated with a Younger Dryas ice marginal zone (Fig. 12b; Ekman and Iljin, 1991; Niemelä et al., 1993; Yevzerov and Nikolaeva, 2000; Yevzerov, 2001), which is also correlated with Younger Dryas moraines elsewhere in Russia and Norway (Fig. 12b; Yevzerov and Kolka, 1993; Andersen et al., 1995a; Hättestrand and Clark, 2006a, 2006b; Hughes et al., 2016; Stroeven et al., 2016). It is unclear, however, whether the distribution of hummocky moraines indicates an ice lobe flowing into the White Sea during the Younger Dryas or an embayment in the Kandalaksha Gulf (Hättestrand and Clark, 2006a, 2006b; Hughes et al., 2016). Finally, in this model, cirque infill deposits in the Khibiny Mountains are Younger Dryas deposits that formed at the terminus of ice lobes that flowed into the mountains (Hättestrand et al., 2008).

4.3.3. Model 3: Ponoy Ice Cap glaciation

In a third model, the ice dispersal centre on the central eastern Kola Peninsula became detached from the FIS soon after the local-LGM, thus forming the dynamically-independent Ponoy Ice Cap (Fig. 12c; Strelkov, 1976; Ekman and Iljin, 1991; Lundqvist and Saarnisto, 1995; Rainio et al., 1995; Svendsen et al., 2004; Astakhov et al., 2016; Lunkka et al., 2018). The clearest argument in support of the Ponoy Ice Cap is the lack of glacial landforms and sediments on the eastern Kola compared to the western Kola (Strelkov, 1976; Ekman and Iljin, 1991; Lundqvist and Saarnisto, 1995; Rainio et al., 1995; Svendsen et al., 2004; Astakhov et al., 2016; Lunkka et al., 2018). The minimal glacial landforms also suggests that the Ponoy Ice Cap was largely inactive during the LGIT, undergoing advances in response to cold interstadials, with final deglaciation occurring after the Younger Dryas-Holocene transition *c.* 11.7 ka (Svendsen et al., 2004; Lunkka et al., 2018). The KIZ is thought to be the Younger Dryas margin of the Ponoy Ice Cap and is correlated with the Salpausselkä Younger Dryas moraines in Finland (Fig. 12c; Rainio et al., 1995; Svendsen et al., 2004; Astakhov et al., 2016), which is supported by numerical age estimations (11.9±0.1, 12, 13, 13.0±1.1, 15.0±1.0, 15.9±1.0 ka; Svendsen et al., 2004; Lunkka et al., 2018).

Svendsen et al. (2004) suggest retreat of FIS ice lobes (occupying ice stream pathways and enveloping the Ponoy Ice Cap after the local-LGM; Fig. 12c) began *c.* 16 ka and continued throughout the LGIT. Subsequently, several authors argue that the Younger Dryas margins of the FIS in Model 3 are in a similar position to those in Model 1 (e.g. Ekman and Iljin, 1991; Rainio et al., 1995; Astakhov et al., 2016), with an ice lobe extending into the White Sea and along the KIZ. This is supported by OSL dating of the Keiva I moraine (9.7±1.0, 12.3±1.0, 14.2±1.2 ka; Lunkka et al., 2018).

4.3.4. Model 4: Kara Sea Ice Sheet glaciation

In Model 4, the FIS retreated from the Kola Peninsula and Russian Lapland, prior to the Younger Dryas, and an ice margin – thought to be attributed to the southward advance of the Kara Sea Ice Sheet (KSIS) – advanced onto the northern Kola Peninsula during the Younger Dryas, accompanied by an ice lobe surge into the White Sea *c.* 10 ka (Fig. 12d; Grosswald, 1980; Grosswald and Hughes, 2002). Grosswald and Hughes (2002) argue that this pattern of glaciation can be identified in the geomorphic record of the Kola Peninsula, with glacial lineations indicating an ice flow direction from the northeast, and the Younger Dryas margin of the ice sheet marked by the KIZ and lobate moraine ridges across the northern peninsula.

4.4. LGIT Models 1-4: the most likely scenario

Four models for the glaciation of the Kola Peninsula and Russian Lapland during the LGIT were presented in section 4.3., which we discuss and critically evaluate in this section. We begin by discussing Models 4 and 3, which we suggest are unlikely scenarios based on the evidence presented in Section 3.

4.4.1. Evidence against LGIT Model 4 – glaciation by the Kara Sea Ice Sheet

Model 4 (Fig. 12d) – which suggests the FIS retreated from the Kola Peninsula and Russian Lapland soon after the local-LGM, after which the KSIS advanced onto the northern Kola Peninsula and into the White Sea during the Younger Dryas (Grosswald, 1980; Grosswald and Hughes, 2002) – cannot be correct for two reasons. First, all available ice flow directional evidence (glacial lineations, lateral meltwater channels, and eskers; Hättestrand and Clark, 2006a), without exception, indicate easterly ice flow on the Kola Peninsula. This is in direct conflict with the proposal of ice flow from the north. Second, available numerical ages (e.g. 15.1±0.4, 15.5±0.3, 15.7±0.4, 15.8±0.3 cal. kyr BP; recalibrated from Polyak et al., 1995) suggest that the Barents Sea region to the north of the Kola Peninsula was ice-free by *c.* 15 ka, with no subsequent readvance of an ice sheet (Winsborrow et al., 2010; Hughes et al., 2016). This implies that the KSIS did not glaciate the region during the Younger Dryas.

4.4.2. Evidence against LGIT Model 3 – glaciation by the Ponoy Ice Cap

It is unlikely that the Ponoy Ice Cap glaciated the central eastern Kola Peninsula during the LGIT (Model 3; Fig. 12c) for two reasons. First, the abundant west to east orientated lateral meltwater channels on the eastern Kola Peninsula (Hättestrand and Clark, 2006a), which are often overlooked by proponents of the Ponoy Ice Cap theory, indicate westward retreat of an ice margin. This is in direct conflict with the presence of the possible Ponoy Ice Cap, which would have had an eastward retreating ice margin resulting in east to west orientated lateral meltwater channels. Second, evidence of subglacial bedforms on the western Kola Peninsula, and some glacial lineations on the eastern Kola Peninsula, suggests an eastward ice flow direction during deglaciation (Kleman et al., 1997; Hättestrand and Clark, 2006b, 2006a; Winsborrow et al., 2010).

One of our main criticisms of the Ponoy Ice Cap theory is that it appears to be based on the reduced presence of subglacial landforms and/or sediments on the eastern Kola Peninsula in comparison to the west (Armand, 1960; Strelkov, 1976; Ekman and Iljin, 1991; Svendsen et al., 2004; Lunkka et al., 2018). Ekman and Iljin (1991), Svendsen et al. (2004), and Lunkka et al. (2018), among others, suggest the difference in abundance of landforms between the eastern and western Kola Peninsula and Russian Lapland indicates two separate and dynamically different ice masses during the LGIT. However, polythermal ice sheets such as the FIS can have zones of cold-based conditions surrounded by corridors of warm-based ice (Kleman et al., 1997; Kleman et al., 2008). As such, FIS ice occupying the eastern Kola Peninsula could have remained entirely cold-based during the Late Weichselian. Given that the orientation of lateral meltwater channels indicates only a westward retreating ice margin (Hättestrand and Clark, 2006b), we suggest that the eastern Kola Peninsula was a large zone of cold-based FIS ice throughout the Late Weichselian.

The KIZ, which is often associated with the Ponoy Ice Cap (Strelkov, 1976; Ekman and Iljin, 1991; Rainio et al., 1995; Svendsen et al., 2004; Astakhov et al., 2016; Lunkka et al., 2018), has a lobate morphology that is compared with the Salpausselkä moraines in Finland (Lavrova, 1960; Ekman and Iljin, 1991; Rainio et al., 1995; Svendsen et al., 2004; Astakhov et al., 2016; Lunkka et al., 2018). This comparison has been used to suggest that the Keiva moraines are of approximately Younger Dryas age (Lavrova, 1960; Ekman and Iljin, 1991; Svendsen et al., 2004; Lunkka et al., 2018). For example, Lunkka et al. (2018) use this analogy to interpret sedimentary exposures and OSL ages of the Keiva IIa moraine in support of the possible Younger Dryas ice margin position along the KIZ. However, most of the numerical ages included within our database indicate pre-Younger Dryas deglaciation. In addition, Lunkka et al. (2018) interpret glaciofluvial sediments at the western end of the Keiva IIa moraine as deltaic deposits from Ponoy Ice Cap meltwaters. However, palaeo-meltwater flow direction measurements and glaciotectonised glaciofluvial sediments directly underlying till deposits indicate palaeo-ice flow directions from the west (Lunkka et al., 2018). This is in direct conflict with the Ponoy Ice Cap theory, since the ice cap is likely to have flowed south and also routed meltwater discharge to the south (Hättestrand and Clark, 2006b). Consequently, it is possible that the numerical ages (13.0±1.1, 15.0±1.0 ka; Lunkka et al., 2018) from these sediments reflect FIS deglaciation prior to the Younger Dryas in this area.

Elsewhere on the Keiva IIa moraine, Lunkka et al. (2018) interpret deltaic glaciofluvial sediments as reflecting a Younger Dryas ice margin position of the Ponoy Ice Cap, which is supported by a single OSL date of 11.9±0.1 ka. However, Hättestrand et al. (2007) suggest relict shorelines surrounding present-day lakes on the northwestern side of Keiva IIa moraine indicate moraine-dammed lakes formed subsequent to deglaciation. The spillways for these lakes are likely occupied by present-day outlet drainage routes that cut through the Keiva II moraine (Hättestrand et al., 2007), and are the location of the sedimentary exposure described and dated by Lunkka et al. (2018). It is therefore possible that the deltaic deposit and associated single OSL numerical age taken by Lunkka et al. (2018) indicates a drainage event of the moraine-dammed lakes. Additional OSL dates from Svendsen et al. (2004) are reported with insufficient chronological or stratigraphical information, thus they cannot convincingly support the Ponoy Ice Cap theory.

A primary issue surrounding the validity of the Ponoy Ice Cap theory is the interpretation of the geomorphological signature, which should always form the precursor to any glacial reconstruction (Lowe and Walker, 2014; Chandler et al., 2018). We acknowledge that sedimentary investigations and chronological reconstructions are somewhat restricted due to inaccessibility of many areas of the Kola Peninsula and Russian Lapland. However, some reconstructions of the eastern Kola Peninsula have overlooked the regional geomorphology, which has resulted in confusion rather than progress. Consequently, we recommend that future glacial reconstructions of the eastern Kola Peninsula consider the geomorphological signature of the KIZ as part of the peninsula-wide landsystem.

4.4.3. Support for LGIT Models 1 and 2 – glaciation solely by the FIS

We propose that the FIS solely glaciated the Kola Peninsula and Russian Lapland during the LGIT. Subglacial bedforms, predominantly in the western Kola Peninsula and the White Sea basin, indicate a general eastward ice flow direction of the FIS across the Kola Peninsula (Kleman et al., 1997; Hättestrand and Clark, 2006a; Winsborrow et al., 2010). In addition, we regard the sediment-landform assemblages of the KIZ and the central eastern Kola Peninsula (Punkari, 1993, 1995; Kleman et al., 1997; Punkari, 1997; Hättestrand and Clark, 2006a; Hättestrand et al., 2007) to be indicative of cold-based ice on the Kola Peninsula and warm-based ice in the White Sea both flowing eastwards across the region throughout the Late Weichselian. Lateral meltwater channels and eskers across the entire Kola Peninsula indicate a westward (inland from the coasts) ice margin retreat pattern (Hättestrand and Clark, 2006b; Stroeven et al., 2016). In addition, up to three discontinuous belts of hummocky moraines and moraine ridges on the western Kola Peninsula and Russian Lapland indicate periodic standstills or readvances of the ice margin during overall ice retreat (Ekman and Iljin, 1991; Yevzerov and Nikolaeva, 2000; Hättestrand and Clark, 2006a; Yevzerov, 2015). Furthermore, sedimentary evidence, including glaciotectonised till and glaciofluvial outwash sediments (Yevzerov and Kolka, 1993; Yevzerov, 2015, 2017; Vashkov and Nosova, 2018b, 2018a; Yevzerov, 2018; Vashkov and Nosova, 2019; Vashkov, 2020), from across the peninsula suggest the FIS may have undergone periods of sustained readvance during oscillatory retreat. However, low-resolution geomorphological data and limited numerical age estimations preclude which, if either, LGIT Models 1 or 2 (Fig. 12a, b) is correct.

It is also unclear whether ice margins retreated synchronously or asynchronously. Lateral meltwater channel orientations depict straight or slightly curved ice margins on the shield terrain of the Kola Peninsula during FIS retreat (Model 2; Hättestrand and Clark, 2006b; Stroeven et al., 2016). However, the spatial distribution of hummocky moraines and moraine ridges indicates a highly lobate ice margin (Model 1; Ekman and Iljin, 1991; Yevzerov, 2015). The absence of glacial landforms and sediments on the central eastern Kola Peninsula is indicative of cold-based conditions that inhibit landform development (Hättestrand and Clark, 2006b; Stroeven et al., 2016). However, it is equally possible that a lack of landforms is a product of rapid deglaciation (Model 1; Niemelä et al., 1993; Yevzerov, 2001; Demidov et al., 2006).

Discontinuous belts of hummocky moraines and moraine ridges across the western Kola Peninsula and Russian Lapland may also demarcate the Younger Dryas ice margin position (Ekman and Iljin, 1991; Yevzerov and Nikolaeva, 2000; Hättestrand and Clark, 2006a; Yevzerov, 2015; Hughes et al., 2016; Stroeven et al., 2016). We propose that the Khibiny Mountains were ice-free prior to the Younger Dryas stadial, similar to Model 1. This is supported by TCN ages of a moraine ridge in the central Khibiny Mountains (16.8±1.0, 14.7±0.8, 12.1±0.7 ka; Stroeven et al., 2016). It remains unclear, however, whether an ice lobe extended into the White Sea during the Younger Dryas. Numerical dating of glacial varves suggests not (12.6±0.2, 13.2±0.7 cal. kyr BP; Kolka et al., 2013), but existing geomorphological mapping cannot accurately reconstruct this ice marginal zone. Consequently, we propose that FIS glaciation on the Kola Peninsula and Russian Lapland during the LGIT was a combination of LGIT Models 1 and 2, with lobate, westward ice margin retreat, inland from the coast on the Kola Peninsula. However, it is not currently possible to reconstruct accurately the pattern, style, and timing of glaciation from previously published, low-resolution maps.

4.4.4. Timing of glaciation

The precise timing of FIS glaciation is also poorly constrained for the Kola Peninsula and Russian Lapland since the majority of dates included in our numerical age estimation database (Fig. 3) were not collected with the purpose of informing glacial dynamics. Additionally, many of the numerical ages are not necessarily targeted at specific landforms and thus, are sampled at seemingly sporadic locations when compared with the glacial geomorphological record. This is most likely due to accessibility constraints, but also a focus on local, rather than regional, glacial reconstructions. As such, numerical age constraints cannot be applied confidently to possible ice margin positions in the region. Moreover, although some studies target ice marginal landforms for dating, they do not always comprehensively consider ice sheet-scale sediment-landform signatures, thus allowing development of alternative glacial interpretations (i.e. Ponoy Ice Cap or KSIS glaciation). Although such chronological information is limited in value for a glacial reconstruction, the dates presented in our review serve to fill a significant data gap for the FIS.

5. Discussion

This review has (i) consolidated and critically evaluated (where possible) the suite of glacial landforms and sediments, as well as previous glacial interpretations, for the Kola Peninsula and Russian Lapland for the first time, and (ii) added new detail to existing FIS reconstructions (e.g. Hughes et al., 2016; Stroeven et al., 2016). It has done so by drawing together previously unutilised and/or inaccessible Russian publications and compiling a comprehensive database of all known (to date) numerical age estimations (Fig. 3). We have shown that the Kola Peninsula and Russian Lapland was likely solely glaciated by the FIS throughout the Late Weichselian, although there is still some uncertainty on specific FIS dynamics. We now discuss the palaeoglaciological implications of our preferred glaciation model, and explore where future glacial reconstructions should focus to better understand FIS dynamics in this region.

5.1. Subglacial conditions on the Kola Peninsula and Russian Lapland

Several authors suggest that the maximum extent and timing of the FIS in northwest Russia occurred up to 5 kyr later than other sectors of the ice sheet (Svendsen et al., 2004; Linge et al., 2006; Hughes et al., 2016; Stroeven et al., 2016). This disparity in timing can, in part, be explained by the eastward migration of the ice divide from the Scandinavian Mountains over the Gulf of Bothnia, which facilitated the lateral extension of the ice margin into northwest Russia (Kleman et al., 1997; Winsborrow et al., 2010; Hughes et al., 2016). However, a migrating ice divide cannot explain the overstretching of ice lobes in northwest Russia, including the White Sea ice lobe.

Basal sliding over water-saturated sediments as well as soft-sediment subglacial deformation is thought to have enabled a fast-flowing low-gradient ice lobe to expand into the White Sea basin (Hättestrand et al., 2007; Larsen et al., 2014; Stroeven et al., 2016). This is supported by glacial lineations on the Tersky coastline of the KIZ (Fig. 10). We also suggest that fast ice flow in the White Sea Ice Stream, which was facilitated by low-shear soft subglacial sediments, resulted in deformation and drumlinisation of the KIZ, thus supporting the interpretations of Hättestrand et al. (2007).

In contrast, a lack of low-shear basal sediments on the Kola Peninsula, especially in the east, likely prevented a switch from cold-based, slow-advancing ice to warm-based, rapidly advancing ice on the peninsula (Kleman et al., 1997; Hättestrand and Clark, 2006b; Hättestrand et al., 2007; Kleman et al., 2008; Winsborrow et al., 2010; Petrov et al., 2014; Stroeven et al., 2016). As such, we suggest the FIS slowly advanced and thickened on the Kola Peninsula and Russian Lapland, and remained persistently cold-based throughout the Late Weichselian. This is supported by the occurrence of lateral meltwater channels in isolation of other glacial landforms across large areas of the peninsula (Fig. 2; Hättestrand and Clark, 2006a). Finally, we suggest the advance of cold-based ice across the region is likely dependent on prolonged cool climatic conditions since the region’s location in a precipitation shadow would have restricted the formation of a local ice dispersal centre (Mangerud et al., 2008a; Mangerud et al., 2008b; Patton et al., 2017).

5.2. Climatic controls on FIS dynamics

We also propose that climate significantly influenced the pattern and style of deglaciation on the Kola Peninsula and Russian Lapland during the LGIT. From the existing ice marginal geomorphological and sedimentary record, we infer that the ice sheet margin periodically readvanced or halted in its retreat. Moraines associated with glacial surges, which display lobate morphologies and are associated with fan-shaped distributions of subglacial bedforms and eskers (Stokes and Clark, 1999, 2001; Stroeven et al., 2016), are identified in parts of the Salpausselkä moraine complex in Finland (Rainio et al., 1995; Stroeven et al., 2016). However, many of the ice marginal formations on the Kola Peninsula and Russian Lapland are often not identified in association with subglacial bedforms. Thus, an ice margin advance on the peninsula must have been climatically influenced, rather than glacier surge induced.

Ring and ridge hummocky moraines (Fig. 9) dominate the ice marginal landform record on the western Kola Peninsula. Such moraines are thought to form from water-saturated subglacial sediments that are squeezed into basal cavities under glacial ice as well as supra-glacial debris accumulating in crevasses at the ice surface (Gravenor and Kupsch, 1959; Lagerbäck, 1988; Johnson et al., 1995; Mollard, 2000; Boone and Eyles, 2001; Knudsen et al., 2006; Evans, 2009; Yevzerov, 2015; Vashkov and Nosova, 2019). Furthermore, large spreads of ring and ridge moraines are thought to indicate an advancing ice margin, which then stagnates and gradually down-wastes (Gravenor and Kupsch, 1959; Johnson et al., 1995; Evans, 2009). However, the ring and ridge moraines on the western Kola Peninsula are not associated with subglacial bedforms indicative of warm-based flow events. We therefore suggest that these moraines are indicative of a cold-based ice margin, which would have advanced in response to deteriorating climatic conditions, with the preservation of stagnant ice margin landforms suggesting large areas of ice detached from the ice margin, possibly in response to a warming climate. Moreover, the most significant belt of ring and ridge moraines are widely attributed to the Younger Dryas cold period (Ekman and Iljin, 1991; Rainio et al., 1995; Yevzerov and Nikolaeva, 2000; Hättestrand and Clark, 2006a; Yevzerov, 2015). Thus, they are broadly linked to a major climatic cooling event that was proceeded by rapid climatic warming, and traditionally have been seen as correlatives to the Younger Dryas ice marginal zone in Finland (the Salpausselkä II Moraine) and elsewhere in northwest Russia (the Rugozero Moraine) (Rainio et al., 1995). Such continuity – over 1,000 km of ice margin distance – combined with little evidence for fast ice flow dynamics (glacial lineations) alongside the moraine belt on the Kola Peninsula, supports climatic control for the formation of the ring and ridge moraines.

The situation with regard to moraines in northern Russian Lapland is, however, more complex as it is unclear whether abundant glacial bedforms and eskers in northern Russian Lapland (which are traditionally interpreted as evidence of the Tuloma Ice Stream) represent ice streaming or warm-based ice margin retreat during deglaciation (see section 3.2.2.). Moraine ridges in northern Russian Lapland display similar morphologies to moraine ridges in northern Norway (Andersen et al., 1995b;

Hättestrand and Clark, 2006a; Stroeven et al., 2016; Romundset et al., 2017), which some interpret as examples of climatically controlled standstills in ice margin retreat (Andersen et al., 1995b; Stroeven et al., 2016). However, the more pronounced topographic relief of northern Norway (characterised by deep fjords and mountains) will inhibit glacier surging (Stroeven et al., 2016). This contrasts with the subdued topography of Russian Lapland, which may be more conducive to both glacier surging and ice streaming. Detailed geomorphological mapping and sedimentary analyses are therefore crucial to verify ice dynamics and climatic associations in northern Russian Lapland.

5.3. Ice streams on the Kola Peninsula and Russian Lapland

We regard two corridors of highly attenuated, densely spaced lineations on the Kola Peninsula (previously known as the Kola Ice Stream) as convincing evidence of two ice stream pathways that were active during the LGIT (see Section 3.2.2). We propose that pronounced warming and surface melting during the Bølling-Allerød interstadial (*c.* 14.7-12.9 ka) triggered the onset of the Lovozero Ice Stream (flowing northeast from the Lovozero Mountains). However, the timing of this ice stream event is unclear as there is a lack of numerical dating. Furthermore, little evidence of time-transgressive bedform patterns and no association with an ice marginal zone (Kleman et al., 1997; Hättestrand and Clark, 2006a; Winsborrow et al., 2010) suggests this ice stream “switched off” and retreated without reworking the landform assemblage. As such, the deactivation of the Lovozero Ice Stream may have been a response to climatic changes and alterations in ice sheet configuration (Conway et al., 2002; Siegert et al., 2003; Stokes and Clark, 2003; Rippin et al., 2006; Stokes et al., 2016a).

The Imandra Ice Stream (flowing eastwards, through Lake Imandra, south of the Khibiny Mountains) is located within the inferred Younger Dryas FIS extent. We therefore propose that ice streaming at this location was possibly triggered by abrupt climatic warming during the Younger Dryas-Holocene transition. This would have contributed to the drawdown and thinning of ice in Russian Lapland, and facilitated the rapid deglaciation of Russian Lapland. The White Sea Ice Stream was also likely active throughout the LGIT and would have significantly contributed to the drawdown of ice in Russian Lapland. Time-transgressive bedform patterns of the Imandra and White Sea Ice Streams are not apparent in existing glacial reconstructions (Kleman et al., 1997; Winsborrow et al., 2010). However, possible bedform superimposition identified by Hättestrand and Clark (2006a) may suggest several episodes of bedform generation. Eskers superimposed upon glacial lineations suggests the Imandra Ice Stream may have deactivated before final deglaciation, possibly in response to changes in ice sheet configuration. Furthermore, ice marginal landforms (predominantly glaciofluvial deposits; Yevzerov and Nikolaeva, 2000; Hättestrand and Clark, 2006a) in southern Russian Lapland may indicate a temporary halt in retreat during the Preboreal oscillation (*c.* 11.4 ka) – a short-lived cooling event after the Younger Dryas stadial (Rasmussen et al., 2014). Similar ice stream responses have been recorded in Finland and northern Karelia (Ekman and Iljin, 1991; Rainio et al., 1995; Stroeven et al., 2016); however, detailed geomorphological mapping and targeted numerical dating are required to further explore this.

5.4. Future work

Future investigations on the Kola Peninsula and Russian Lapland should adopt a comprehensive, high-resolution geomorphology-based glacial reconstruction approach that permits reconstruction of small-scale characteristics of the FIS while simultaneously considering the scale of the ice sheet system (Clark and Meehan, 2001; Greenwood and Clark, 2009a, 2009b). The recent development of high-resolution, remotely-sensed digital elevation models (DEM) and satellite imagery such as the ArcticDEM (Porter et al., 2018) will now permit such a reconstruction of this remote and largely inaccessible region. In turn, this will provide a framework into which sedimentology-based and chronological studies can be included and compared (Clark and Meehan, 2001; Greenwood and Clark, 2009a, 2009b; Clark et al., 2018).

To better understand ice sheet dynamics on the Kola Peninsula, Russian Lapland, and in the wider FIS system, we recommend a particular focus on the following landform assemblages:

1. The KIZ – one of the most prominent landform systems on the Kola Peninsula, which is a key element for reconstructing the history and dynamics of the FIS.
2. Subglacial bedforms i.e. glacial lineations and ribbed moraine across the Kola Peninsula and Russian Lapland – to determine ice flow configuration throughout the Late Weichselian.
3. Hummocky moraine and moraine ridge belts on the western Kola Peninsula and in Russian Lapland – to identify LGIT ice margin positions, including a possible Younger Dryas ice marginal zone.
4. Meltwater landforms i.e. lateral meltwater channels and eskers across the region – to determine ice margin retreat patterns where ice marginal landforms (i.e. moraines) are not present and ice sheet dynamics during the LGIT.

Future glacial reconstructions should also identify a relative chronology with which our numerical age estimation database – which includes all known numerical age estimations from the Kola Peninsula and Russian Lapland to date (Fig. 3; Data S1) – can be compared. Developing a high-resolution geomorphological-based glacial reconstruction as described above will also provide the means for identifying and selecting ideal locations for targeted numerical dating. Developing a relative chronology on the Kola Peninsula and Russian Lapland is crucial in order to better understand FIS dynamics more widely during the Late Weichselian, and would permit numerical ice sheet models to be tested in the region.

6. Conclusion

Understanding ice sheet processes and rates of ice advance and retreat is important given current and projected responses of the Antarctic and Greenland ice sheets to climate change. However, previous glacial reconstructions have thus far failed to determine the pattern, style, and timing of FIS glaciation on the Kola Peninsula and Russian Lapland. This is due to an incomplete understanding of glacial geomorphology and sedimentology in the region, combined with a limited glacial chronology.

In this study, we have compiled and critically evaluated (where possible) published geomorphological, sedimentary, and chronological data, as well as glacial interpretations of Late Weichselian glaciation on the Kola Peninsula and Russian Lapland. This includes often overlooked data published in Russian-language journals.

Sedimentary till evidence indicates that the Kola Peninsula and Russian Lapland was subject to two glacial advances during the Weichselian: (i) an Early-Middle Weichselian (*c.* 115-40 ka) ice mass advance, which likely retreated during the Ålesund interstadial (*c.* 38-34 ka), and (ii) FIS glaciation during the Late Weichselian (*c.* 40-10 ka). During the Late Weichselian, the FIS extended eastwards across Russian Lapland and the Kola Peninsula from the Scandinavian Mountains, reaching its maximum extent in northwest Arctic Russia *c.* 19-15 ka. During the local-LGM, the FIS flowed radially out to the coasts, and an ice stream was established in the White Sea basin. Three other ice streams – the Tuloma Ice Stream and two separate ice streams along the Kola Ice Stream pathway – likely existed in the region during the LGIT.

Differing glacial interpretations have resulted in four scenarios for the LGIT (see Fig. 8):

> Model 1 – Initial rapid deglaciation of the FIS on the eastern Kola Peninsula, leaving ice lobes occupying the surrounding lowlands and straits, which retreated westward during the LGIT.
>
> Model 2 – Coherent westward retreat of the FIS, inland from the coasts, throughout the LGIT.
>
> Model 3 – Ponoy Ice Cap glaciation, which established itself on the eastern Kola Peninsula after the local-LGM.
>
> Model 4 – Rapid westward deglaciation of the FIS, and subsequent glaciation by the Kara Sea Ice Sheet (KSIS) during the LGIT.

Based on the geomorphological signature of the region, we suggest that the Kola Peninsula was neither glaciated by the Ponoy Ice Cap (Model 3) nor the KSIS (Model 4) during the LGIT. Subglacial and meltwater landform evidence suggests it is more likely that the FIS solely glaciated the Kola

Peninsula during the LGIT (Models 1 and 2). However, how the FIS retreated across the Kola Peninsula and Russian Lapland is unclear, and the position of the Younger Dryas ice margin remains inconclusive. This uncertainty stems from previously employed low-resolution glacial geomorphological mapping in the region. Future investigations on the Kola Peninsula and Russian Lapland should address this knowledge gap by establishing a high-resolution geomorphology-based glacial reconstruction, which will provide a framework into which sedimentology-based and chronological studies can be included and compared.

Author contributions

BMB, LDL, DP, VVK, and DJN conceived the original study. Overall manuscript development was led by BMB. All authors contributed to the preparation of the final manuscript.

Competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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