Continental-scale transport of sediments by the Baltic Ice Stream elucidated by coupled grain size and Nd provenance analyses

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ABSTRACT

We introduce a methodology for determining the transport distance of subglacially comminuted and entrained sediments. We pilot this method on sediments from the terminal margin of the Baltic Ice Stream, the largest ice stream of the Fennoscandian Ice Sheet during the Last Glacial Maximum. A strong correlation ($R^2 = 0.83$) between the ϵ Nd and latitudes of circum-Baltic river sediments enables us to use ϵ Nd as a calibrated measure of distance. The proportion of subglacially transported sediments in a sample is estimated from grain size ratios in the silt fraction ($<63 \mu\text{m}$). Coupled ϵ Nd and grain size analyses reveal a common erosion source for the Baltic Ice Stream sediments located near the Åland sill, more than 850 km upstream from the terminal moraines. This result is in agreement with both numerical modeling and geomorphological investigations of Fennoscandinavian erosion, and is consistent with rapid ice flow into the Baltic basins prior to the Last Glacial Maximum. The methodology introduced here could be used to infer the distances of glacial sediment transport from Late Pleistocene and earlier glaciations.

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1. Introduction

Ice streams act as the arteries of ice sheets (Bennett, 2003), delivering large quantities of ice from interior regions of positive mass balance to the margins (i.e., Rignot et al., 2008). Ice streams flow via internal ice deformation, sliding of basal ice across the bed, and the deformation of subglacial sediments (Cuffey and Paterson, 2010). The transport of sediment provides an important constraint on the basal conditions and processes beneath former ice streams (e.g., Clark, 1987; Houmark-Nielsen, 1987; Kjær et al., 2003; Stokes et al., 2015).

The Baltic Ice Stream was the largest of the Fennoscandian Ice Sheet (FIS) ice streams (Hughes et al., 2016; Stroeven et al., 2016), flowing south into the topographic lows of the Bothnian and Baltic Sea basins (Fig. 1; Houmark-Nielsen, 1987; Kleman et al., 1997; Boulton et al., 2001; Patton et al., 2016). From the initiation of fast flow around 38–35 ka BP, the Baltic Ice Stream advanced and retreated over the Baltic Sea basin several times. The ice stream reached its southern terminus in Denmark, Germany, and Poland

during the Last Glacial Maximum ca. 23 to 21 ka BP (LGM; Hughes et al., 2016; Toucanne et al., 2015).

Numerical modeling of the FIS suggests an area of high subglacial erosion between the central Bothnian and central Baltic Seas (Fig. 1; Patton et al., 2016). This zone of maximal erosion is consistent with a wide region of glacial scouring and till removal near the Åland sill (Kleman et al., 2008), and 41–60% of LGM boulder and gravel indicators in the southern Baltic correspond to this erosion source (Overweel, 1977). The transport history of fine sediments by the Baltic Ice Stream, however, is less certain.

Isotopic provenance, in which transported sediments are fingerprinted and referenced to the signatures of other rocks in the basin, provides a robust interpretation of the source of Baltic Ice Stream sediments (i.e., Toucanne et al., 2015). Isotope methods are particularly valuable for fine sediments, as petrographic identification is difficult at this scale.

In this study, we present a new method for discerning the distance of fine sediment transport by ice sheets. We pair the grain size distributions and ϵ Nd of glacial silts to identify sources for far-traveled and distally-sourced sediments. For sediments collected near the LGM terminus of the FIS in Denmark, Germany, and Poland, we document continental scale transport of the coarse silt fraction. Our reinterpretation of the glacial sediment ϵ Nd record in the southern Baltic provides a convergence among geo-

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Fig. 1. (a) Modeled FIS flowlines, consistent with lineated flow sets (Kleman et al., 1997; Winsborrow et al., 2010), and a zone of high erosion potential (orange) beneath the Baltic Ice Stream (Patton et al., 2016). The main source of erosion for the sediments considered in this study (Sec. 4.3) is indicated by the yellow star (Fig. 4, projected here onto the 20°E line of longitude), with 95% uncertainty bounds (yellow highlight). (b) Archean and Proterozoic bedrock geology of Fennoscandia (Asch, 2005), with the sampling locations of the riverine (Table 1) and glacial (Table 2) sediments discussed in this manuscript. Note that the limits of the modeled FIS at 20.58 ka BP do not align exactly with the empirical LGM limits in the Northern European Lowlands (Patton et al., 2016). (For interpretation of the colors in the figure(s), the reader is referred to the web version of this article.)

chemical, geomorphological, and numerical model results of sediment sources and transport by the Baltic Ice Stream.

1.1. Provenance characterization of ϵNd in the Baltic Sea region

The origin of glacial sediments can be inferred from the Sm–Nd isotope system and corresponding ϵNd values. ϵNd is the parts per ten thousand deviation of $^{143}\text{Nd}/^{144}\text{Nd}$ in a sample from the chondritic uniform reservoir, or CHUR, where ^{143}Nd is the alpha decay product of ^{147}Sm . Sm is less incompatible during mantle melting than Nd, and thus Nd is preferentially enriched in the melts that separate from the mantle. Because the Sm/Nd ratio of continental crust is relatively low and mostly uniform (Taylor and McLennan, 1985; Goldstein and Jacobsen, 1988), the ϵNd of continental crust and its detritus becomes progressively more negative (non-radiogenic) with age (e.g., Taylor and McLennan, 1985; McLennan and Hemming, 1992; Bayon et al., 2015). The ϵNd of crustal material is typically unaffected by weathering, subsequent thermal events, or other forms of alteration (Taylor and McLennan, 1985). Processes occurring during sediment transport do not change the ϵNd of rock fragments, so the sources of glacial sediments deposited by the Baltic Ice Stream can be ascertained from their ϵNd .

Because rivers integrate the sediments within their catchments, the ϵNd values of detrital grains reflect the relative formation ages of constituent crust. The interpretation of these ϵNd values (Soulet et al., 2013; Freslon et al., 2014; Toucanne et al., 2015) is consistent with the bedrock geology of the Fennoscandian re-

gion (Fig. 1; Gaál and Gorbatschev, 1987; Gorbatschev and Bogdanova, 1993). Archean bedrock (>2.5 Ga) lies to the north and east of the Gulf of Bothnia in northern Finland and northwest Russia. Much of the remaining terrane in circum-Baltic Fennoscandia (i.e., Sweden, Finland, and Denmark) is Paleoproterozoic in age (2.0–1.65 Ga), with the youngest bedrock concentrated in Denmark and southwest Sweden. Fennoscandia was altered by both the Sveconorwegian–Grenville orogeny (1.2–0.9 Ga) and Caledonian orogeny (0.5–0.4 Ga), but these produced relatively little new continental crust in the region (e.g., Mykkeltveit et al., 1980; Gaál and Gorbatschev, 1987; Gorbatschev and Bogdanova, 1993). The relationship between river sediment ϵNd and regional outcrop geology is obscured in the southern and southeastern Baltic, however, by the extensive coverage of Phanerozoic sedimentary rocks (i.e., Winterhalter et al., 1981). Because the depositional ages of these sedimentary units do not necessarily reflect the formation ages of their constituent fragments, we rely on the ϵNd of the rivers that drain these terranes to rigorously fingerprint their crustal signatures.

1.2. Comminution

Subglacial sediments are crushed and abraded against the bedrock and other clasts by the flow of ice. This reduction of clasts to smaller fragments is termed comminution. The integrated effect of comminution leads to characteristic grain size distributions (e.g., Haldorsen, 1981; Hooke and Iverson, 1995). Dreimanis and Vagners (1971, 1972) observed that there is a comminution

Table 1

Information for the circum-Baltic river sediments considered in this project, compiled from Toucanne et al. (2015), Freslon et al. (2014), and Soulet et al. (2013). The locations of the sediments are displayed in Fig. 1. The estimated uncertainty of the measurements is $\pm 0.3 \text{ } \epsilon\text{-units}$ (2σ) based on the external reproducibility of replicate analyses of the JNdi-1 standard solution (Toucanne et al., 2015).

River	Country	Latitude (°N)	Longitude (°E)	$\epsilon\text{Nd} \pm 2\sigma$
Elbe	Germany	53.703	9.449	-11.3 ± 0.3
Oder	Poland	53.841	14.121	-12.4 ± 0.3
Vistula	Poland	54.651	19.287	-14.3 ± 0.3
Neman	Lithuania	55.362	21.257	-14.3 ± 0.3
Daugava	Latvia	57.060	24.039	-15.5 ± 0.3
Gauja	Latvia	57.133	24.684	-14.1 ± 0.3
Narva	Estonia	59.489	28.040	-16.4 ± 0.3
Neva	Russia	60.070	29.279	-17.7 ± 0.3
Kymijoki	Finland	60.260	26.496	-18.7 ± 0.3
Kiiminkijoki	Finland	65.133	25.731	-23.3 ± 0.3
Luleälv	Sweden	65.682	21.820	-19.1 ± 0.3
Umeälven	Sweden	63.718	20.267	-17.9 ± 0.3

threshold, or terminal grade, at which sediment grains are not reduced beyond a particular size, even with continued subglacial transport. The terminal grade is likely controlled by a minimum fracture length that is influenced by the presence of microcracks, the sizes of crystals, and mineral hardnesses (e.g., Barenblatt, 1962; Dreimanis and Vagners, 1971). When a clast is reduced below the terminal grade, fracturing does not continue and comminution ceases. Each mineral has its own terminal grade (Dreimanis and Vagners, 1971, 1972).

The terminal grades of feldspars, amphiboles, pyroxenes, and heavy minerals are between ~ 30 and $63 \text{ } \mu\text{m}$ (Dreimanis and Vagners, 1972). The terminal grain sizes of quartz and carbonate minerals contain “major” modes in this coarse silt fraction, though their terminal grades extend to $4 \text{ } \mu\text{m}$ (Dreimanis and Vagners, 1972). Feldspars are the predominant mineral of Earth’s continental crust, and heavy minerals dominate the rare earth element concentrations and thus the ϵNd of silts and sand-sized sediments (e.g., Garçon et al., 2014). Accordingly, the coarse silt fraction ($30\text{--}63 \text{ } \mu\text{m}$) of glacial sediments increases with transport distance and contains the mineral assemblage that is mixed through comminution and transport along the ice stream path. Heavy minerals would be comminuted to the terminal grade after approximately 80 to 180 km of transport at the glacial bed, with less transport being necessary for comminution to the terminal grade among the less resistant minerals (Dreimanis and Vagners, 1971). Therefore, heterogeneities in the ϵNd of fine glacial sediments are minimized by the long transport distances required for terminal comminution of heavy minerals.

2. Materials and methods

2.1. Glacial and riverine sediments

Modern riverine sediments were collected from the estuaries and lagoons of various rivers draining into the Baltic and Bothnian Sea basins (Table 1; Fig. 1; Toucanne et al., 2015). Rivers of the western and northern Baltic basin, including the Umeälven, Luleälv, Kiiminkijoki, and Kymijoki, drain the widely-exposed crystalline bedrocks of the Fennoscandian Shield while the rivers of the southern and southeastern Baltic, including the Elbe, Oder, Vistula, Neman, Daugava, Gauja, and Narva, integrate the diverse Phanerozoic sedimentary rocks of the East European Platform (Table 1; Winterhalter et al., 1981; Andersson et al., 1992). The catchment of the Neva river, with its terminus in the easternmost Gulf of Finland, includes both Precambrian and Phanerozoic basement rocks.

Glacial sediments from the Late Weichselian (Last Glacial) were collected from moraines, ice-marginal valleys (IMVs), and

proglacial lakes along the southern margin of the Baltic Ice Stream terminus in Denmark, Germany, and Poland (Table 2; Toucanne et al., 2015). Outwash sediments from the Brandenburg–Leszno advance phase (23.4 ± 0.3 to $20.3 \pm 0.2 \text{ ka BP}$; Toucanne et al., 2015) were acquired from the Głogów–Baruth IMV in Beelitz, eastern Germany (Lühtens et al., 2010) and Karchowo and Hetmanice, western Poland (Krzyszowski et al., 1999). Tills and glacial diamicts from the Frankfurt–Poznań phase (18.7 ± 0.3 to $18.2 \pm 0.2 \text{ ka BP}$) were gathered in Kozłowo, Glaznoty, Chrostkowo, and Obórki villages in eastern Poland (Wysota et al., 2009; Tylman et al., 2013; Lesemann et al., 2010; Narloch et al., 2012). Outwash plain and glaciolacustrine sediments of the Pomeranian phase (16.7 ± 0.2 to $15.7 \pm 0.3 \text{ ka BP}$) were recovered from Althüttendorf and Macherslust, respectively, in the Toruń–Eberswalde IMV (Lühtens et al., 2011). Additional glaciolacustrine sediments were sourced from the Brodtener Ufer cliff exposure on the Baltic shore in Travemünde (Kabel, 1983).

In Denmark, samples were collected from cliff exposures and gravel mining pits on the islands of Sjælland and Møn. Samples from Møn were identified in consultation with the lithostratigraphy of Houmark-Nielsen (1999). From a beach cliff in Tøvelde, we procured till samples from the pre-LGM Baltic advance (Klintholm, ~ 35 to $30.7 \pm 0.7 \text{ ka BP}$; Toucanne et al., 2015), the main LGM advance (Mid-Danish), and post-LGM readvance (Young Baltic). The upper half-meter of the Klintholm till contained pronouncedly larger gravel clasts than the lower section; we sampled both. At a nearby beach cliff in Hvide, we sampled a diamict from the Kobbegård beds that post-dates the main LGM advance and has been correlated to the Allarp till in Skåne, Sweden (Houmark-Nielsen, 1999). At a mining pit in Ledreborg, we acquired outwash sediments from the Mid-Danish advance and till from the overlying Young Baltic advance. Mid-Danish outwash sediments were also gathered at Knarbo Klint, south of Ordrup Næs. The Mid-Danish advance is equivalent to the Brandenburg–Leszno advance in Germany and Poland, while the Young Baltic re-advances are correlative to the Frankfurt–Poznań and Pomeranian phases (Houmark-Nielsen and Kjær, 2003; Toucanne et al., 2015).

2.2. Grain size and Nd isotope analyses

The grain size distributions of the bulk glacial sediments were measured by laser diffractometry on an LS200 Coulter at IFREMER in Brest, France ($n = 23$, Table S1). Samples were wet-sieved to remove gravels and then dried. Representative aliquots of the dried fractions were sonicated immediately prior to analysis to prevent clay flocculation. Volume percentages of grain size in each class (Table S1) were calculated from the mean of at least 3 replicates for each sample. We conservatively estimate the internal reproducibility of each measurement at the 1% level.

Toucanne et al. (2015) reported Nd isotope analyses performed on the $<63 \text{ } \mu\text{m}$ fractions of circum-Baltic river sediments and glacial sediments from Germany and Poland (Tables 1, 2). Here, we report Nd isotope analyses from the $<63 \text{ } \mu\text{m}$ fraction of glacial sediments from Denmark using the same procedure (Table 2). Prior to digestion by alkaline fusion (Bayon et al., 2009), organic material, carbonates, and Fe–Mn oxides were removed in the manner described by Bayon et al. (2002). Nd was then isolated by ion-exchange chromatography. All Nd isotope measurements (Toucanne et al., 2015; this study) were performed at the Pôle Spectrométrie Océan in Plouzané, France using a Thermo Scientific Neptune multi-collector ICP-MS. Nd isotopic compositions of the samples were calculated by sample-standard bracketing, with the JNdi-1 standard solution analyzed every two samples. Exponential mass-bias corrections were applied to the Nd isotope ratios using $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, and

Table 2

Information for the glacial sediments considered in this project. Glaciolacustrine and select outwash sediments (indicated by an asterisk) were excluded from the provenance analysis (Sec. 4.2) based on evidence that the grain size distributions of these samples are corrupted by post-depositional sorting (Sec. 4.1). EM 1 corresponds to the concentration of fine-grained sediments by glaciofluvial and lacustrine sorting. EM 2 is the comminution process. EM 3 represents till genesis (Sec. 4.1). The ϵNd of sediments from Denmark are first reported in this study; those from Germany and Poland were reported by Toucanne et al. (2015). The estimated uncertainty of the measurements is ± 0.3 ϵ -units (2σ) based on the external reproducibility of replicate analyses of the JNdi-1 standard solution (i.e., Sec. 2.2, this study; Toucanne et al., 2015). The internal precision of the Danish sediment ϵNd analyses ranges from 0.08 to 0.12 ϵ -units (2 SE).

Sample	Lat. (°N)	Lon. (°E)	Type	Regional (ϵNd)	Sample (ϵNd)	$\Delta\epsilon\text{Nd}$	30–63/ <63 μm	30–125/ <125 μm	EM 1 Score	EM 2 Score	EM 3 Score
Althüttendorf-a	52.963	13.872	Outwash	Oder (–12.4)	–16.5	4.1	0.64	0.83	0	23	77
Althüttendorf-b*	52.963	13.872	Outwash	Oder (–12.4)	–15.7	3.3	0.11	0.11	62	38	0
Travermünde S1*	53.971	10.883	Glaciolac.	Elbe (–11.3)	–12.4	1.1	0.21	0.28	46	54	0
Travermünde S5*	53.971	10.883	Glaciolac.	Elbe (–11.3)	–14.2	2.9	0.24	0.29	33	67	0
Machterslust-a*	52.848	13.838	Glaciolac.	Oder (–12.4)	–14.8	2.4	0.12	0.13	62	38	0
Machterslust-b*	52.848	13.838	Glaciolac.	Oder (–12.4)	–15.6	3.2	0.17	0.24	82	2	16
Beelitz-a*	52.288	12.937	Outwash	Elbe (–11.3)	–13.9	2.6	0.10	0.18	89	0	11
Beelitz-b	52.288	12.937	Outwash	Elbe (–11.3)	–13.8	2.5	0.53	0.57	0	100	0
Oborki	53.152	19.381	Till	Vistula (–14.3)	–16.1	1.8	0.20	0.40	37	17	47
Hetmanice ST/17	51.858	16.265	Outwash	Oder (–12.4)	–14.9	2.5	0.56	0.60	0	100	0
Karchowo ST/14	51.889	16.834	Outwash	Oder (–12.4)	–15.4	3.0	0.54	0.58	0	100	0
Kozłowo	53.341	18.341	Till	Vistula (–14.3)	–15.8	1.5	0.29	0.53	17	20	63
Glaznoty	53.535	19.904	Till	Vistula (–14.3)	–15.8	1.5	0.36	0.54	8	42	50
Chrostkowo 1	52.943	19.253	Diamict	Vistula (–14.3)	–14.5	0.2	0.27	0.49	25	19	56
Chrostkowo 2	52.943	19.253	Diamict	Vistula (–14.3)	–15.2	0.9	0.25	0.46	30	16	54
Ledreborg MD	55.626	12.046	Outwash	Denmark (–14.3)	–16.4	2.1	0.41	0.68	3	25	73
Ledreborg YB	55.626	12.046	Till	Denmark (–14.3)	–15.0	0.7	0.19	0.27	63	29	8
Ordруп MD	55.817	11.377	Outwash	Denmark (–14.3)	–14.9	0.5	0.21	0.40	37	21	42
Tøvelde KL	54.946	12.301	Till	Denmark (–14.3)	–15.4	1.0	0.23	0.29	52	48	0
Tøvelde KU	54.946	12.301	Till	Denmark (–14.3)	–15.5	1.2	0.25	0.36	42	47	11
Tøvelde MD	54.946	12.301	Till	Denmark (–14.3)	–15.5	1.2	0.14	0.22	74	21	5
Tøvelde YB	54.949	12.307	Till	Denmark (–14.3)	–15.3	0.9	0.18	0.27	59	34	7
Hvide 3	54.935	12.278	Diamict	Denmark (–14.3)	–15.5	1.2	0.26	0.45	31	24	45

the mass-bias corrected $^{143}\text{Nd}/^{144}\text{Nd}$ values were normalized to a JNdi-1 value of $^{143}\text{Nd}/^{144}\text{Nd} = 0.512115$ (Tanaka et al., 2000). The estimated uncertainty of our measurements is ± 0.3 ϵ -units (2σ) based on the external reproducibility of replicate analyses of the JNdi-1 standard solution ($^{143}\text{Nd}/^{144}\text{Nd} = 0.512115 \pm 0.000009$, 2σ , $n = 31$). Procedural blanks contained 7 and 12 pg of Nd while the samples contained between 1.4×10^5 and 9.0×10^5 pg of Nd. We report $^{143}\text{Nd}/^{144}\text{Nd}$ ratios in ϵNd notation, $[(^{143}\text{Nd}/^{144}\text{Nd})_{\text{sample}}/(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} - 1] \times 10^4$, using the $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}$ value of 0.512638 (Jacobsen and Wasserburg, 1980).

2.3. Endmember modeling of sedimentary processes

Silts compose a volumetrically significant but underutilized fraction of subglacially transported sediments since, due to their small size, petrographic identification of provenance is ineffective. As such, we seek to infer the relative transport distances of glacial sediments from their grain size distributions, specifically the relative concentrations of coarse silts. This strategy is based on empirical observations that the proportion of coarse silt in a till increases with transport distance (e.g., Dreimanis and Vagners, 1971, 1972). Here, we quantitatively verify the grain size signature of comminution in our samples by endmember modeling analysis (EMMA).

EMMA is an eigenspace method of solving the bilinear mixing problem in which the number of endmembers (EMs) and their relative proportions are not known in advance (Weltje, 1997). EMMA explains the structure of observations in a dataset as a smaller number of non-negative EMs (Dietze et al., 2012). The grain size distributions of sediments are dominated by a discrete number of processes that can act to comminute and transport them. EMMA allows us to discriminate individual processes operating on the sediment distributions.

To evaluate the EMs, we perform singular value decomposition on the matrix of our grain size data and obtain the characteristic

eigenvalues and eigenvectors following the method of Dietze et al. (2012). Each eigenvalue is paired to an eigenvector, and the normalized eigenvalues correspond to the percentage of variance in the data explained by the associated eigenvector. In practice, only a few eigenpairs are relevant because they explain most of the variance in the dataset. The selected eigenvectors are rotated orthogonally following the method of Kaiser (1958) and normalized. The EM scores, corresponding to the relative proportions of each EM in the grain size distributions, are estimated by least squares with non-negativity constraints (as introduced by Lawson and Hanson, 1974) from the grain size data matrix and the rotated and normalized eigenvectors. Lastly, the EMs are defined by rescaling the rotated and normalized eigenvectors to non-negative percentages such that each sums to 100% (Dietze et al., 2012).

To test whether the number of studied grain size distributions is enough to have certainty in our EMs, we resample the data with replacement, known as the bootstrap method (i.e., Efron and Tibshirani, 1994). For 1000 bootstrap samples, the EMMA was performed on 23 randomly selected grain size distributions with replacement allowing for the same distribution to be chosen multiple times per bootstrap.

3. Results

3.1. Grain sizes

The grain size distribution data obtained for this study are reported in Table S1. The proportion of subglacially transported sediment in each glacial sample is estimated from the relative abundances of coarse silt within the silt and clay fraction (30–63 $\mu\text{m}/<63 \mu\text{m}$). We term this ratio the comminution index. The comminution indices of the glaciolacustrine samples cluster between 0.12 and 0.24 (Table 2). The till and diamict samples range from 0.14 to 0.36. There is greater variability in the comminution indices of the glacial outwash deposits, from 0.10 to 0.64.

Fig. 2. The latitudinal distribution and ϵNd of circum-Baltic river sediments (Fig. 1; Table 1). The linear trendline indicates that the Nd isotopes of circum-Baltic sediments, a measure of continental crust formation age, are increasingly non-radiogenic (older) towards the north. This reflects the distribution of Proterozoic and Archean bedrock in Fennoscandia (Fig. 1; Gaál and Gorbatshev, 1987). For reference, the global average ϵNd of suspended river loads, approximating the ϵNd of the upper continental crust, is -10.4 (Goldstein and Jacobsen, 1988). The trendline and its 95% prediction interval (dashed lines) were calculated by bootstrapping.

3.2. Neodymium isotopes

The ϵNd of the glacial sediments from Denmark are reported in Table 2 ($n = 8$). The ϵNd of these samples range from -16.4 to -14.9 . For comparison, Toucanne et al. (2015) reported ϵNd of glacial sediments from Germany and Poland between -16.5 and -12.4 ($n = 15$, Table 2). From the ϵNd of terrigenous river sediments that drain into the Baltic and Bothnian Seas (Table 1), we document an approximately north–south trend ($R^2 = 0.83$, $p < 10^{-5}$) in the formation age of exposed Fennoscandian crust (Fig. 2). Prediction intervals (95%) for the trendline in Fig. 2 are calculated by bootstrapping of the observations with replacement ($n = 1000$).

3.3. Model endmembers

We perform EMMA on the grain size distributions of the glacial sediments ($n = 23$). We restrict our domain of consideration to particles between 2 and 125 μm because we are focusing on the processes that affect fine sediment size distributions. The EMMA identifies 3 EMs that can be mixed to explain 97% of variance in the data, satisfying a common criterion for EM selection ($>95\%$, e.g., Reymont and Joreskog, 1996). The first EM is predominantly composed of fine silt grains ($79\% \leq 16 \mu\text{m}$). The second EM mainly consists of medium and coarse silts ($68\% 16\text{--}63 \mu\text{m}$). The third EM is mostly fine sand ($51\% 63\text{--}125 \mu\text{m}$) and fine silt ($31\% \leq 16 \mu\text{m}$), but lacks medium and coarse silt ($18\% 16\text{--}63 \mu\text{m}$). The robustness of the model is verified by bootstrapping the 95% confidence intervals for each EM (Fig. 3) and by ensuring that the form of the EMs is insensitive to the specific choice of grain size domain boundaries.

4. Discussion

4.1. Transport and sorting processes

Each of the three EMs appears to represent a different set of processes. We interpret the first EM to represent the concentration of fine silts that have been sorted hydraulically through glaciofluvial and lacustrine processes. The first EM scores highly for all of the glaciolacustrine and two of the outwash sediments. The tills

Fig. 3. The three EMs determined by endmember modeling analysis. The EMs represent independent processes controlling the grain size distributions of the glacial sediments. EM 1 illustrates the concentration of fine-grained sediments by glaciofluvial and lacustrine sorting, EM 2 is the comminution process of Dreimanis and Vagners (1971, 1972), and EM 3 represents a poorly sorted, washed-out till distribution. Confidence intervals (95%) about the EMs were determined by bootstrapping.

from eastern Denmark also have elevated scores for the first EM, consistent with their incorporation of chalk, limestone, and sandy marl substrata (i.e., Houmark-Nielsen and Kjær, 2003). The second EM has a pronounced mode in the terminal grain sizes of heavy minerals and feldspar, exhibiting characteristics of the comminution process (Dreimanis and Vagners, 1972). The third EM appears to characterize a poorly-sorted till that lacks medium and coarse silts that would have been concentrated by comminution (EM 2). Our interpretation is consistent with the third EM scoring highly for the German and Polish till samples and at low values for each of the glaciolacustrine sediments (Table 2).

We suspect that some of the glaciolacustrine and outwash sediments (Table 2) experienced significant post-transport sorting by water. Such sorting would alter the grain size proxy and possibly change the amount and concentration of the heavy minerals that control the ϵNd . We compare the proportions of coarse silt and fine sand ($30\text{--}125 \mu\text{m}/<125 \mu\text{m}$) in the glaciolacustrine and outwash sediments to the average of our German and Polish tills ($0.48 \pm 0.06 \mu\text{m}$, 1σ) to quantify the relative sorting of these samples. All four glaciolacustrine sediments (from 0.13 to 0.29 μm) and two outwash sediments (0.11 and 0.18 μm) show depletion of coarse silt and fine sand relative to the German and Polish till deposits (Table 2). These six samples are likewise depleted in the third EM and enriched in the first EM. In contrast to the other outwash samples, the two outwash samples with low $30\text{--}125 \mu\text{m}/<125 \mu\text{m}$ ratios (Althüttendorf-b and Beelitz-a) also have low scores for both EMs 2 (comminuted sediments) and 3 (washed-out till). Accordingly, we do not consider these six samples in the determination of sediment transport distance by coupled ϵNd and grain size analyses (Sec. 4.2).

4.2. Sediment transport from coupled ϵNd and grain size analyses

To calculate distances of subglacial sediment transport, we determine proxy measures for both the mean transport distance of a glacial sample and the proportion of sediments in the sample that were transported subglacially. The proportion of subglacially transported sediment in the sample is estimated from the relative abundances of coarse silts in the silt and clay fraction (the comminution index). Because ϵNd provides a robust geographic fingerprint for the origin of sediments (Sec. 1.1), the mean distance of transport, for all sediments in the sample, is estimated from the

Fig. 4. $\Delta\epsilon\text{Nd}$ (the difference between the ϵNd of the glacial sediment and ϵNd signature of the catchment in which the sample was collected) vs. the comminution index (the proportion of glacial modification) of the glacial samples. The fully comminuted endmember ($\text{CI} = 1$) is represented by the yellow star (projected onto a common line of longitude bisecting the Baltic Sea in Fig. 1). This endmember reveals that the main geographical origin of the subglacially-transported sediments have an Nd isotope signature that is 5.4 ϵ -units less radiogenic (older, more northward in this case) than the locale they were recovered from. The trendline and its 95% prediction interval (dashed lines) were calculated by bootstrapping. Transport distance is inferred from $\Delta\epsilon\text{Nd}$ per the relationship in Fig. 2. The locations of the tills (blue) and outwash sediments (green) are displayed in Fig. 1.

difference between the ϵNd of a glacial sample and the ϵNd of non-glacial sediments in the region where the glacial sample was collected. We quantify this difference as $\Delta\epsilon\text{Nd}$. Calculating $\Delta\epsilon\text{Nd}$ in this manner corrects for the longitudinal variation in ϵNd across the Northern European Lowlands (i.e., Toucanne et al., 2015), allowing us to project our samples onto a common transport path (Figs. 1 and 4). The $\Delta\epsilon\text{Nd}$ is translated into a transport distance via the relationship between the ϵNd and latitude of terrigenous sediments from rivers that drain into the Baltic and Bothnian Seas ($R^2 = 0.83$, Fig. 2). From this relationship, we discern that each ϵ -unit in the $\Delta\epsilon\text{Nd}$ corresponds to ~ 160 km of transport in the north–south sense. For samples collected in Germany and Poland, the regional ϵNd signature is measured from the river sediments of major catchments (Table 1). No comparable river system exists in Denmark, however. Therefore, for the samples collected in Denmark, we estimate the regional ϵNd signature (-14.3) from the y-intercept of the linear regression of Danish sample ϵNd values and comminution indices (i.e., Fig. 4).

To examine the relationship between the sediment transport distance and grain size, we plot the $\Delta\epsilon\text{Nd}$ of each sample versus its comminution index (Fig. 4). A linear trend is fit to the data that we infer to define a mixing line between two EMs. The two EMs are the comminuted and non-comminuted fractions of the sediments, whose comminution indices are one and zero, respectively. The comminuted EM ($\text{CI} = 1$) represents far-transported sediments whereas the non-comminuted EM ($\text{CI} = 0$) represents sediments that were entrained near the terminal margin and therefore not comminuted to the terminal grade before deposition. By excluding the six samples described as hydraulically sorted in Sec. 4.1, we find a strong correlation ($R^2 = 0.78$, $p < 10^{-5}$) between the $\Delta\epsilon\text{Nd}$ of our glacial sediments and their comminution indices (Fig. 4; $n = 17$). The linear correspondence between $\Delta\epsilon\text{Nd}$ and the comminution index is consistent with mixing between two well-defined EMs having similar Nd concentrations. The distal EM ($\text{CI} = 0$) is composed of fine silts and clays with a $\Delta\epsilon\text{Nd}$ value of -0.2 ± 0.7 (95% prediction interval). The far-transported EM ($\text{CI} = 1$) is composed of coarse silts and has a $\Delta\epsilon\text{Nd}$ value of 5.4 (asymmetric 95% prediction interval from 3.9 to 6.5). The prediction intervals

(95%) are calculated by bootstrapping with replacement (Fig. 4, $n = 1000$).

By assuming that the Baltic Ice Stream flowed nearly along lines of longitude, a reasonable approximation to its flow path (e.g., Kleman et al., 1997; Boulton et al., 2001; Patton et al., 2016; Fig. 1), the $\Delta\epsilon\text{Nd}$ indicates the source region of our samples in a simple north–south sense (Fig. 2). From the $\Delta\epsilon\text{Nd}$ of the comminuted EM ($\text{CI} = 1$), we estimate that the comminuted fractions of our sediments were deposited 850 ± 250 km (95% confidence) from their source region (Fig. 1). The uncertainty estimate for this transport distance accounts for the analytical uncertainty of the ϵNd analyses for both the glacial and river sediments, the prediction uncertainty in $\Delta\epsilon\text{Nd}$ for the comminuted EM ($\text{CI} = 1$, Fig. 4), and the prediction uncertainties about the trendline between river ϵNd and latitude (Fig. 2). Because the $\Delta\epsilon\text{Nd}$ of our samples are normalized to a common region of deposition at the ice sheet margin, we calculate the inferred source region of the comminuted sediments from a projected site of deposition at 53.7°N , 20°E (Fig. 1).

4.3. Sediment sources

The inferred transport distance of 850 ± 250 km suggests that the main source of sediment erosion was most likely located near the Åland sill between the Bothnian and Baltic Seas (Fig. 1). The extensive glacial scouring and till removal in this region indicates that till cover and weathered mantle overburden were eroded in a single or a few successive glacial cycles, providing a supply of sediments for subglacial transport (Kleman et al., 2008). Consistent with this interpretation, previous researchers have documented the transport of large clasts from the northern Baltic by the LGM Baltic Ice Stream (e.g., Overweel, 1977; Houmark-Nielsen, 1987; Smed, 1993). Furthermore, our determination of a main erosion source near the Åland sill is in agreement with numerical results of Patton et al. (2016) who show a potential zone of maximum erosion here. Although the mean ϵNd values of the glacial silts are dominated by non-comminuted sediments that reflect a southern Baltic source (Toucanne et al., 2015), we demonstrate that the signature of subglacial erosion and transport can be extracted by coupling grain size and ϵNd analyses.

The Baltic Ice Stream advanced over the zone of maximum erosion at least twice before reaching its final LGM terminus position (Hughes et al., 2016), providing subglacially eroded sediments that could be entrained englacially (Hambrey et al., 1997; Alley et al., 1997) and transported to the ice margin. We consider that rapid ice flow into the Baltic basins is responsible for the long transport distances of the comminuted sediments (i.e., Clark, 1987), a scenario that is supported by empirical and numerical demonstrations of rapid Fennoscandian Ice Sheet growth in the lead up to the LGM (Hughes et al., 2016; Patton et al., 2016).

In terms of its future application, the methodology introduced here could be used to infer the distances of sediment transport by other LGM ice streams, particularly those of the Laurentide Ice Sheet given the geometric similarities of its continental shield bedrock to the Baltic Ice Stream (i.e., Stokes and Clark, 2003; Stokes et al., 2016). We speculate that the most fruitful application of our method, however, would be to older glaciations, as geomorphological evidence of subglacial erosion is increasingly obscured and boundary conditions for numerical models are increasingly uncertain prior to the Last Glacial Period. As evidenced by the distribution of tills and outwash sediments in Fig. 4, future studies should take care to collect glacial sediments with diverse concentrations of coarse silt to ensure that mixing lines, should they exist, can be fit robustly. Furthermore, we note that our method is unlikely to be appropriate for understanding glacial processes and sediment transport at scales of tens of kilometers or less. The robustness of the relationship between the $\Delta\epsilon\text{Nd}$ of our glacial

sediments and their comminution indices (Fig. 4) evinces a regional homogenization of ϵNd in the silt and clay fractions due to the long distances of transport required for the comminution of heavy minerals to their terminal grade. To infer the transport distances of glacial sediments at shorter scales, researchers would be well-advised to track spatial trends in the concentration of distinctive rock fragments or metals (i.e., Hooke et al., 2013). The applicability of such dispersal train methods is limited at great lengths, however, by exponential decreases in the concentration of target rocks and metals with distance (i.e., Clark, 1987). An exponential relationship between $\Delta\epsilon\text{Nd}$ and distance is not discerned in the present study, likely because our method measures the integrated entrainment of sediments along lines of ice flow instead of erosion from a point source.

5. Conclusion

In this manuscript, we present a methodology for estimating the transport distance of fine subglacial sediments. We determine proxy measures for the mean transport distance of fine sediments in a sample as well as the proportion of the sediments that were transported subglacially. The proportion of subglacially transported sediment in a sample is estimated from the relative abundances of coarse silt in the silt and clay fraction. The conceptual basis for using grain size ratios to quantify transport is provided by empirical studies of subglacial comminution and the resulting grain size signatures. The mean distance of transport for the silts is estimated from the difference between the ϵNd value of the glacial sediment and the regional ϵNd signature that corresponds to the locale in which the former was collected ($\Delta\epsilon\text{Nd}$). A relationship between the ϵNd and latitude of circum-Baltic river sediments provides a calibrated framework for translating the $\Delta\epsilon\text{Nd}$ into transport distances along lines of ice flow. Endmember modeling analysis reveals the signature of comminution in the studied sediments, validating the methodology.

We determine that the comminuted silt fractions of glacial sediments from the LGM terminal margin in Denmark, Germany, and Poland originated at least 850 ± 250 km to the north. This result is consistent with a main erosion source near the Åland sill and rapid ice flow into the Baltic and Bothnian Sea basins in the lead up to the LGM.

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Appendix A. Supplementary material

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