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Asynchronous response of marine-terminating outlet glaciers during deglaciation of the Fennoscandian Ice Sheet --Manuscript Draft--

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Abstract:	<p>Recent studies have highlighted the dynamic behavior of marine-terminating outlet glaciers over decadal time-scales, linked to both atmospheric and oceanic warming. This helps explain episodes of near-synchronous flow acceleration, thinning and retreat, but non-climatic factors such as subglacial overdeepenings can also induce rapid recession. There is support for these topographic controls on glacier retreat, but there are few long-term records to assess their significance across a population of glaciers over millennial time-scales. Here, we present retreat chronologies alongside topographic data for eight major outlet glaciers that experienced similar climatic forcing during deglaciation of the Fennoscandian Ice Sheet (ca. 18-10 cal. kyr B.P.). Retreat rates averaged over several millennia (~30 m a⁻¹) are less than half those recently observed on modern-day outlet glaciers (>100 m a⁻¹), but deglaciation was punctuated by episodes of more rapid retreat (up to ~150 m a⁻¹) and re-advances. Significantly, phases of rapid retreat were not synchronous between glaciers and most occurred irrespective of any obvious atmospheric warming. We interpret this to reflect the complex interplay between external forcing and both topographic (e.g., bathymetry, width) and glaciological factors (e.g., ice catchments) that evolve through time, but conclude that basal over-deepening in wide fjords induce episodes of rapid retreat (>100 m a⁻¹), further exacerbated by their greater susceptibility to oceanic warming. This complicates attempts to predict the centennial-scale trajectory of outlet glaciers and suggests that modeling the interaction between neighboring catchments and the accurate description of subglacial topography beneath them is a priority for future work.</p>
Response to Reviewers:	See attached cover letter and response to reviewer comments

1 Asynchronous response of marine-terminating outlet
2 glaciers during deglaciation of the Fennoscandian Ice Sheet

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10 **ASBTRACT**

11 Recent studies have highlighted the dynamic behavior of marine-terminating
12 outlet glaciers over decadal time-scales, linked to both atmospheric and oceanic warming.
13 This helps explain episodes of near-synchronous flow acceleration, thinning and retreat,
14 but non-climatic factors such as subglacial overdeepenings can also induce rapid
15 recession. There is support for these topographic controls on glacier retreat, but there are
16 few long-term records to assess their significance across a population of glaciers over
17 millennial time-scales. Here, we present retreat chronologies alongside topographic data
18 for eight major outlet glaciers that experienced similar climatic forcing during
19 deglaciation of the Fennoscandian Ice Sheet (ca. 18–10 cal. kyr B.P.). Retreat rates
20 averaged over several millennia ($\sim 30 \text{ m a}^{-1}$) are less than half those recently observed on
21 modern-day outlet glaciers ($>100 \text{ m a}^{-1}$), but deglaciation was punctuated by episodes of
22 more rapid retreat (up to $\sim 150 \text{ m a}^{-1}$) and re-advances. Significantly, phases of rapid

23 retreat were not synchronous between glaciers and most occurred irrespective of any
24 obvious atmospheric warming. We interpret this to reflect the complex interplay between
25 external forcing and both topographic (e.g., bathymetry, width) and glaciological factors
26 (e.g., ice catchments) that evolve through time, but conclude that basal over-deepening
27 in wide fjords induce episodes of rapid retreat ($>100 \text{ m a}^{-1}$), further exacerbated by their
28 greater susceptibility to oceanic warming. This complicates attempts to predict the
29 centennial-scale trajectory of outlet glaciers and suggests that modeling the interaction
30 between neighboring catchments and the accurate description of subglacial topography
31 beneath them is a priority for future work.

32 INTRODUCTION

33 Ice sheets are organized into a pattern of tributaries feeding rapidly flowing ice
34 streams, separated by slow-flowing ice. In coastal regions, many ice streams are
35 influenced by topography and become confined within deep troughs as marine-
36 terminating outlet glaciers. Because of their disproportionate ice flux, they are a key
37 component of ice sheet mass balance and associated impacts on sea level (Thomas et al.,
38 2011; Nick et al., 2013). Indeed, there is an urgent need to understand the longer-term
39 significance of recent ‘dynamic’ changes that have been observed on outlet glaciers,
40 typically characterized by their accelerating flow, thinning and retreat (Howat et al.,
41 2007; Andresen et al., 2012; Nick et al., 2013). This, however, is difficult; partly because
42 of the complexity in identifying factors that drive such changes (e.g., air and ocean
43 temperatures, fjord geometry; see Carr et al., 2013), and partly because there are few
44 records of outlet glacier behavior over centennial to millennial time-scales. Moreover,

45 such records are mostly restricted to just one or two outlet glaciers (e.g., Briner et al.,
46 2009; Young et al., 2011; Hughes et al., 2012; Mangerud et al., 2013).

47 Theory suggests that the bathymetry beneath marine-terminating glaciers is an
48 important control on their advance and retreat (Weertman, 1974; Schoof, 2007).
49 However, few studies have examined its importance empirically, largely because
50 topography beneath modern-day glaciers is difficult to extract. This can be circumvented
51 by dating the retreat of palaeo-outlet glaciers, where formerly subglacial topography can
52 be measured, and some studies found that glaciers receding into deeper waters
53 experienced irreversible and rapid retreat (e.g. Briner et al., 2009), irrespective of any
54 climatic forcing. Others, however, note slow retreat across reverse bed slopes (e.g., Ó
55 Cofaigh et al., 2008; Jamieson et al., 2012), suggesting that factors such as fjord width
56 and the size of catchment area are also important (cf. Warren and Glasser, 1992;
57 Rydningen et al., 2013; Carr et al., 2014).

58 To investigate the controls on millennial-scale behavior of outlet glaciers under
59 similar external forcing, we reconstruct the retreat of eight neighboring outlet glaciers
60 that operated during deglaciation of the northern margin of the Fennoscandian Ice Sheet
61 (FIS) (Fig. 1). During the Last Glacial Maximum (LGM), the FIS was coalescent with the
62 marine-based Barents Sea Ice Sheet, and major fjords in northern Norway acted as
63 tributaries to ice streams that reached the continental shelf edge (Ottesen et al., 2008;
64 Winsborrow et al., 2010). Deglaciation from the shelf edge commenced after 19 cal. kyr
65 B.P. and the ice margin was close to or confined to fjords by 15 cal. kyr B.P. (Andersen,
66 1968; Sollid et al., 1973), with ice-free conditions in the south-west Barents Sea
67 (Winsborrow et al., 2010; Junttila et al., 2010).

68 **METHODS**

69 Ice sheet retreat in northern Norway following the LGM has been known in
70 general outline for several decades, and eight major sub-stages have been identified and
71 dated in Finnmark (Sollid et al., 1973) and Troms (Andersen, 1968). These are based on
72 extensive mapping of moraines, traced over considerable distances, together with raised
73 shorelines cut into end moraines or extending beyond ice-contact deltas; further
74 augmented by radiocarbon dates from marine sediments pre- or post-dating moraines
75 (e.g., Andersen, 1968; Sollid et al., 1973). We identified known ice front positions from
76 previous work that have been assigned to the established regional sub-stages (see Table
77 DR1 in GSA Data Repository¹). We then reviewed and, in a small number of cases,
78 revised positions and ages based on new bathymetric data, new mapping of glacial
79 geomorphology (e.g., Ottesen et al., 2008; Winsborrow et al., 2010; Rydningen et al.,
80 2013), and more recent radiocarbon dates (Vorren and Plassen, 2002; Eilertsen et al.,
81 2005).

82 Transects were then drawn to extract bathymetric data from each fjord using the
83 MAREANO multibeam dataset collected by the Norwegian Hydrographic Service
84 (www.mareano.no), and we estimated the width of the outlet glacier at the calving front
85 at 50 m increments during retreat. Transects extend from the outermost part of the fjord
86 to the marine limit at the head of the fjord at the time it was deglaciated (Fig. 1). They
87 therefore cover the entire palaeofjord, projecting landward into what are now fjord-
88 valleys, and are depth-adjusted for former sea level and postglacial infill, where known.
89 Retreat rates between each sub-stage were calculated in each fjord and we assign errors
90 based on: (1) the reported radiocarbon age uncertainty and an appraisal of the number of

91 dates in each region, (2) stratigraphic evidence relating to the dates and glacial events,
92 and (3) the strength of regional correlations of marginal moraines and raised shorelines
93 related to particular events and dates. These errors (Table DR1; Figures 2 and 3) capture
94 the maximum range of 'known' uncertainty and do not influence the broad patterns of ice
95 margin recession in each fjord, which is our focus (i.e. relative changes in retreat rate
96 between the dated sub-stages).

97 **RESULTS**

98 As in previous work (e.g., Briner et al., 2009; Mangerud et al., 2013), we present
99 time-distance diagrams for each fjord system (Figs. 2 and 3). Retreat rates averaged over
100 several millennia ($\sim 30 \text{ m a}^{-1}$) were less than half those observed on modern-day outlet
101 glaciers over decadal time-scales ($> 100 \text{ m a}^{-1}$; Howat and Eddy, 2011). However,
102 maximum rates typically exceeded 100 m a^{-1} . Due to the inherent uncertainties of the
103 dated ice margin positions, we focus on the broad patterns of retreat within each fjord,
104 with particular attention as to when and where the major sub-stages are recorded, and
105 when retreat rates increased. In this regard, five out of the eight glaciers experienced their
106 most rapid retreat during early deglaciation (before 15 cal. kyr B.P.), when air
107 temperatures were relatively cold (Fig. 1c). This typically occurred across major
108 overdeepenings on the continental shelf (e.g., Andfjorden, Malangen) or through the mid-
109 to outer-fjord areas (e.g., Altafjorden, Varangen). In some cases, rapid retreat occurred
110 over a reverse bed slope into progressively deeper water (e.g., Andfjorden, Malangen), or
111 simply where water depths were greatest (e.g., Porsangen, Tanafjorden). Whilst the
112 expectation is that glaciers will tend to retreat more rapidly through deeper water (cf.
113 Schoof, 2007), the correlation between water depths and retreat rates is perhaps not as

114 strong as might be expected ($R^2 = 0.17$; Fig. DR1b in the Data Repository). There
 115 are
 116 cases where rapid retreat shows no obvious correlation with bathymetric changes
 117 (e.g., Laksefjorden) and where retreat was relatively slow ($\sim 40 \text{ m a}^{-1}$) through deep
 118 water, often coinciding with narrow troughs or localized constrictions (e.g., Lyngen,
 119 Malangen). Indeed, fjord width shows a stronger correlation with retreat rate ($R^2 = 0.21$;
 120 but this is not straightforward, in that some ~~Figure DR1c~~ ^{Figure DR1c} troughs (2–3 km wide) were
 121 evacuated relatively quickly (e.g., Altafjorden).

121 **DISCUSSION**

122 A key outcome of our millennial-scale reconstruction is that the retreat rates were
 123 asynchronous, despite a similar regional climate forcing. Retreat through some fjords was
 124 comparatively slow and steady (e.g., Lyngen) while others were evacuated rapidly (e.g.,
 125 Varangerfjorden). One might have expected glaciers to have undergone phases of rapid
 126 retreat during or after periods of warming (cf. Young et al., 2011; Hughes et al., 2012),
 127 i.e., during the transition into the Bølling-Allerød (ca. 14.7 cal. kyr B.P.) or the early
 128 Holocene (ca. 11.7 cal. kyr B.P.), but this is not obviously evident (Fig. 2; Fig. DR1a).

129 An explanation for the asynchronous pattern of retreat is the variable topography
 130 within each fjord (Fig. 1). There are clear cases where maximum retreat rates coincide
 131 with reverse bed slopes (e.g., Andfjorden, Malangen) and/or deep (200–300 m) water
 132 (e.g., Porsangen, Tanafjorden). These cases support the importance of water depth in
 133 inducing episodes of rapid retreat (e.g., Schoof, 2007; Briner et al., 2009). It also explains
 134 why the highest retreat rates in most fjords occurred during early deglaciation, because
 135 this is when outlet glaciers were more likely to encounter basal overdeepenings (Fig. 1).
 136 Thus, we find a clear indication that deep and wide fjords, characterized by subglacial

137 overdeepenings, always induce episodes of rapid retreat (e.g., $>100 \text{ m a}^{-1}$ in Andfjorden,
138 Malangen, Altafjorden, and Varangerfjorden). Thus, although atmospheric warming will
139 inevitably lead to deglaciation by inducing a negative ice sheet mass balance, there is not
140 always an obvious correlation between climate forcing and the rate of retreat of outlet
141 glaciers over centennial to millennial time-scales. Measurements of glacier retreat
142 following the Bølling-Allerød warming at ca. 14.5 cal. kyr B.P., for example, would
143 reveal terminus positions ranging from mid to inner-fjord areas and retreat rates
144 ranging from $<30 \text{ m a}^{-1}$ (Laksefjorden) to $>140 \text{ m a}^{-1}$ (Varangerfjorden).

145 Although topographic factors can clearly influence glacier retreat, the fact that
146 these relationships are not stronger (Figs. DR1b and DR1c) indicates a complex interplay
147 between them. Glaciers may retreat slowly in deep fjords if they are narrow, or in wide
148 fjords if they are shallow. It is also important to note that we only measure fjord depth
149 and width at the inferred glacier terminus, and not the longitudinal gradients of glacier
150 depth and width. Retreat rates will be affected by thinning at the glacier terminus, which
151 is further affected by the longitudinal flux gradient. As such, glaciological factors further
152 modulate outlet glacier behavior, and the size and slope of the catchment are likely to be
153 important. Those with larger, higher catchment areas are more likely to be able to sustain
154 ice fluxes and maintain a stable grounding line position in deep water or across reverse
155 bed slopes (Schoof, 2007; Jamieson et al., 2012). If glaciers are unable to balance calving
156 by draw-down of ice, it is likely to lead to thinning and retreat. The anomalous period of
157 rapid retreat in Tanafjorden between ca. 15.5 and 15 cal. kyr B.P. might be a reflection of
158 a small catchment area that was rapidly diminished by drawdown caused by retreat in
159 Varangenfjorden. We suggest, therefore, that interactions between adjacent ice stream

160 catchments (ice piracy and capturing) are likely to be an important control on outlet
161 glacier dynamics over centuries to millennia (cf. Payne and Dongelmans, 1997). This
162 complicates attempts to numerically model the behavior of individual outlet glaciers over
163 these time-scales, which are often targeted at specific glaciers and necessarily omit
164 interactions with neighboring catchments (e.g., Jamieson et al., 2012; Nick et al., 2013).

165 A further complication is that the longitudinal flux gradient can be affected by
166 back-pressure from an ice shelf. There are few proxies available to reconstruct the
167 presence of ice shelves, but evidence of numerous well-developed shorelines and raised
168 beaches correlating with end moraines suggests that open water conditions prevailed as
169 ice retreated within the fjords (e.g., Sollid et al., 1973; see the Data Repository).
170 Exceptions might include the cold reversals, where the development of ice shelves may
171 have provided a stabilizing influence during re-advances. Indeed, Junttila et al. (2010)
172 note the possibility of extensive, seasonal or semi-perennial sea-ice cover during the
173 Skarpnes readvance. Any ice shelves are more likely to have formed in narrow fjords,
174 where lateral resistance and the effect of pinning points is proportionally higher; and are
175 likely to have been maintained in settings that prevented incursion of warm sub-surface
176 Atlantic Water, such as shallower fjords or those with sills. In contrast, wide fjords with
177 major overdeepenings are less likely to support ice shelves, and would have been more
178 susceptible to the incursion of Atlantic water, which is thought to have occurred between
179 16 and 15 cal. kyr B.P. (Junttila et al., 2010; Rørvik et al., 2013). This oceanic forcing
180 might further contribute to the high retreat rates we reconstruct across major
181 overdeepenings early in deglaciation, when atmospheric temperatures were relatively
182 cool (see Fig. 1C).

183 Finally, our data provide a context to gauge the magnitude and significance of
184 recent changes in modern-day ice sheets. In our study, maximum retreat rates averaged
185 over a few hundred years typically exceed 100 m a^{-1} , which is higher than those reported
186 as ‘rapid’ during early Holocene retreat of the Laurentide ($>58 \text{ m a}^{-1}$ by Briner et al.
187 [2009]) and Greenland Ice Sheets ($>80 \text{ m a}^{-1}$ by Hughes et al. [2012]; $\sim 100 \text{ m a}^{-1}$ by
188 Young et al. [2011]) over similar time-scales. Mangerud et al. (2013) reported higher
189 rates of retreat in two fjord systems in the south-western FIS ($240\text{--}340 \text{ m a}^{-1}$), but even
190 these are an order of magnitude lower than those observed on major outlet glaciers in
191 modern-day ice sheets, albeit over much shorter time-scales, e.g., Thwaites (1000 m a^{-1}
192 from 1996 to 2009; Tinto and Bell, 2011) and Pine Island Glacier in West Antarctica
193 (1000 m a^{-1} from 2004 to 2009; Thomas et al., 2011), and Helheim Glacier in southeast
194 Greenland ($2,500 \text{ m a}^{-1}$ from 2000 to 2005; Howat et al., 2007). Calculation of palaeo-
195 retreat rates are necessarily averaged over long time-scales, and are likely to mask any
196 episodes of extreme retreat, but they clearly demonstrate that current retreat rates in
197 excess of 1000 m a^{-1} are an order of magnitude higher than the average rates which led to
198 the disappearance of the last mid-latitude ice sheets. It is, perhaps, unlikely that these
199 high rates can be sustained, but our data suggest that it will largely depend on the
200 topography that these glaciers encounter and their interactions with neighboring
201 catchment areas.

202 **CONCLUSIONS**

203 Recent work suggests a rapid and near-synchronous response of outlet glaciers to
204 large-scale oceanic and atmospheric conditions over decadal time-scales (Andresen et al.,
205 2012). This paper reconstructs the retreat of eight major outlet glaciers during

206 deglaciation of the FIS (18–10 cal. kyr B.P.) and shows that, despite experiencing a
207 similar climate forcing, their response was asynchronous over millennial time-scales. We
208 interpret this to reflect the complex interplay between climate forcing and both
209 topographic (e.g., bathymetry, width) and glaciological factors (e.g., the evolution of
210 catchment areas) that evolve through time, but there is clear evidence that basal
211 over-deepenings in wide fjords induce episodes of rapid retreat ($>100 \text{ m a}^{-1}$), further
212 exacerbated by their susceptibility to oceanic warming. Thus, high resolution data of
213 subglacial topography beneath the catchments of modern-day outlet glaciers is likely to
214 be a crucial requirement for modeling and assessment of future ice sheet dynamics
215 (Durand et al., 2011). Such modeling will offer further opportunities to assess the
216 sensitivity of outlet glaciers to a range of forcing factors and, in this regard, this paper
217 offers a valuable data set much longer than the current observational record.

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314 **FIGURE CAPTIONS**

315 Figure 1. **A:** The study area (red box) at the Last Glacial Maximum (LGM) (BR—
316 Bjørnøyrenna). **B:** Topography of northern Fennoscandia and ice margin chronology for
317 eight major fjord systems in this study (labeled). Note the location of transects in each

318 fjord (black lines); estimated lateral boundaries between catchment areas at LGM (dashed
319 white line); and dates associated with known or interpolated ice-front positions based on
320 our synthesis of previous work (see Table DR1 [see footnote 1]). C: Stable oxygen
321 isotopes ($\delta^{18}\text{O}$) from the North Greenland Ice Core Project (NGRIP members, 2004)
322 together with marine stable oxygen isotopes ($\delta^{18}\text{O}$) from sediment core MD99 2294,
323 Lofoten (Rørvik et al., 2013). OD—Oldest Dryas; B—Bølling; A: Allerød; YD—
324 Younger Dryas; PB—Preboreal.

325

326 Figure 2. Time-distance diagrams for glacier terminus position and width within each
327 fjord system in Troms, Norway, alongside bathymetric and geological data. Retreat rates
328 are calculated between each of the identified sub-stages (linked from fjord to fjord by
329 vertical dashed lines) and values in brackets capture the maximum possible range of
330 values based on the age uncertainties (see the Data Repository [see footnote 1]). Marine
331 limit (ML) shows approximate relative sea level during glacier retreat.

332

333 Figure 3. Time-distance diagrams for glacier terminus position and width within each
334 fjord system in Finnmark, Norway, alongside bathymetric and geological data as in Fig.
335 2.

336

337 ¹GSA Data Repository item 2014xxx, xxxxxxxx, is available online at
338 www.geosociety.org/pubs/ft2014.htm, or on request from editing@geosociety.org or
339 Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

Figure 1
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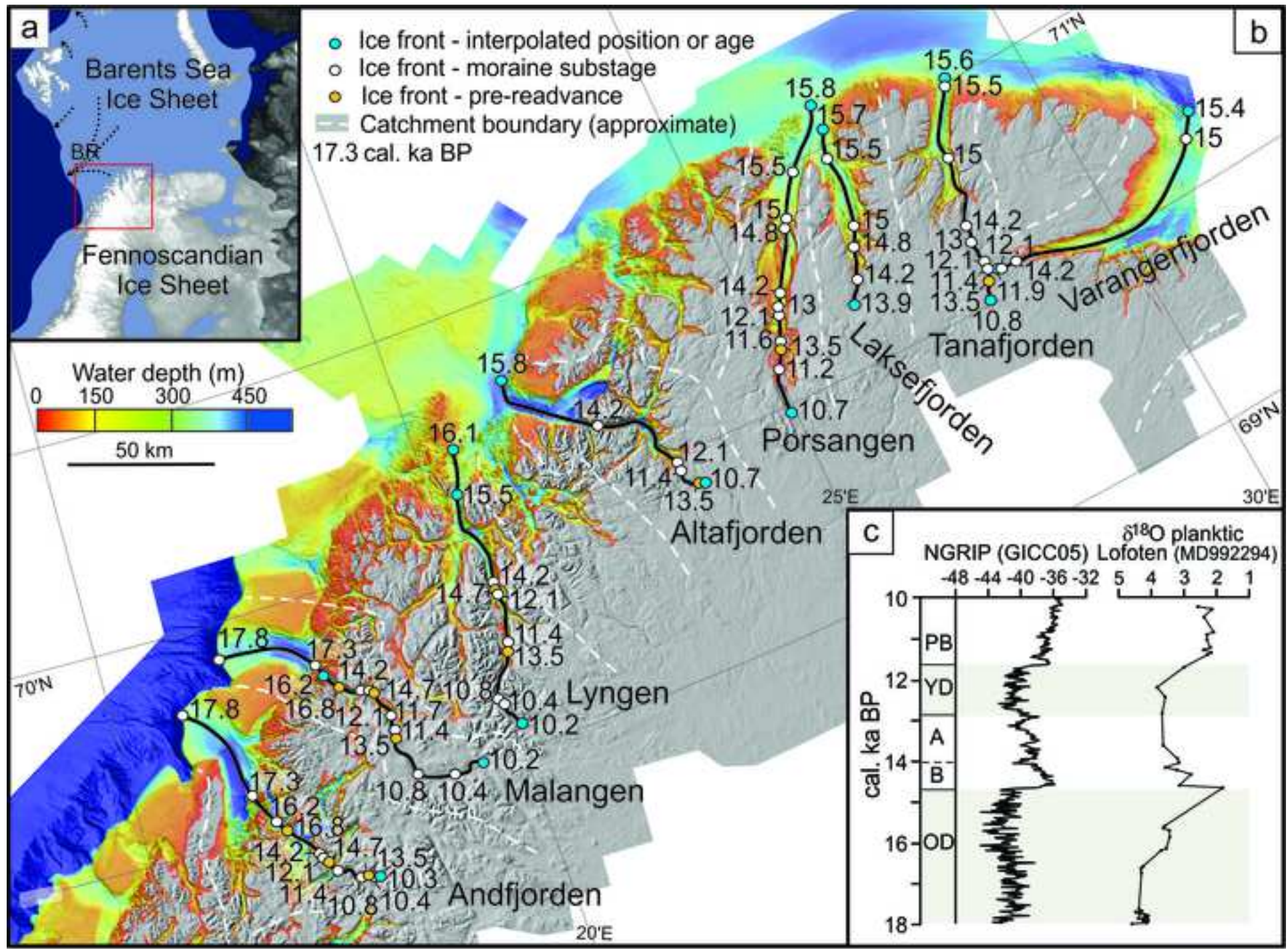


Figure 2
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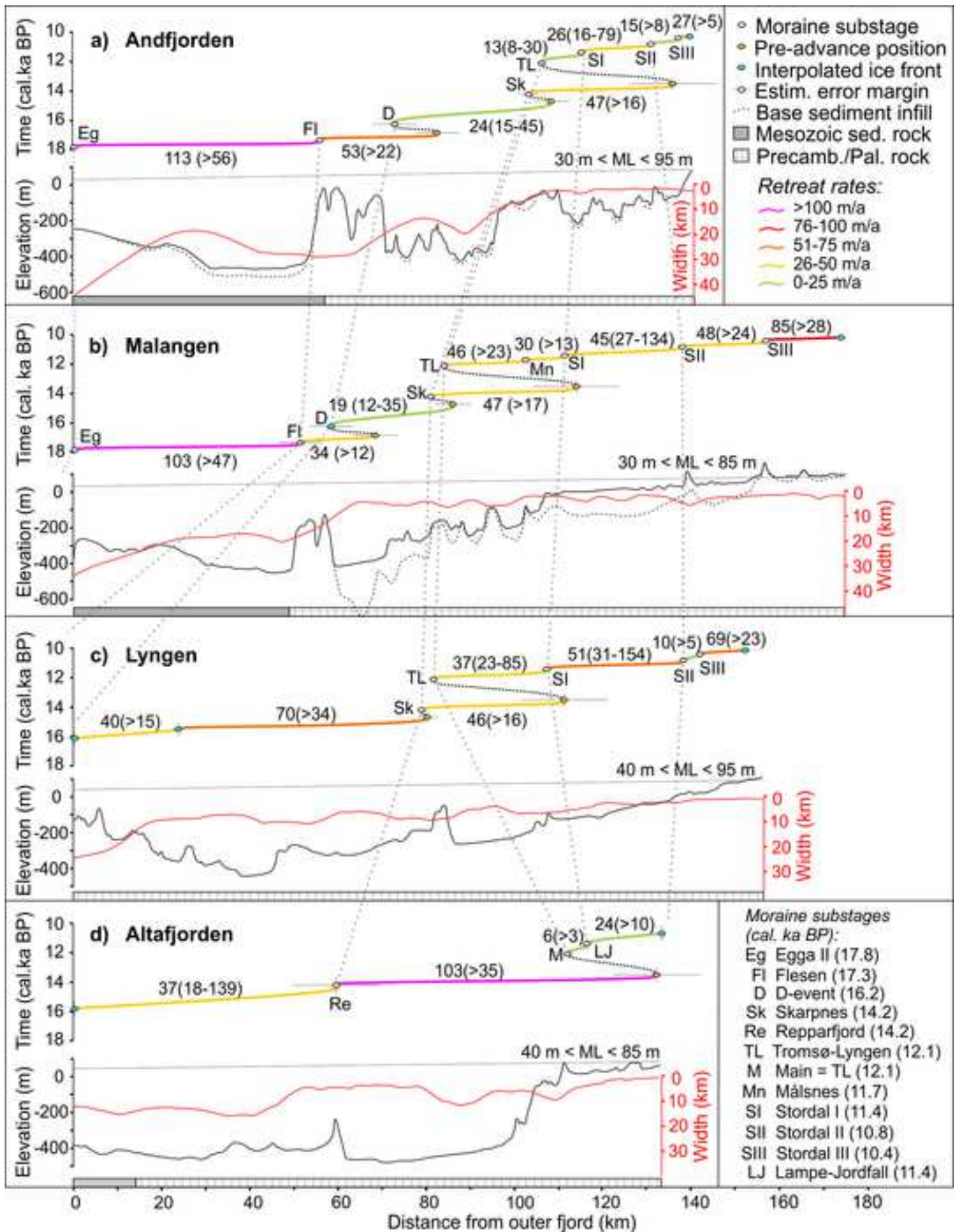
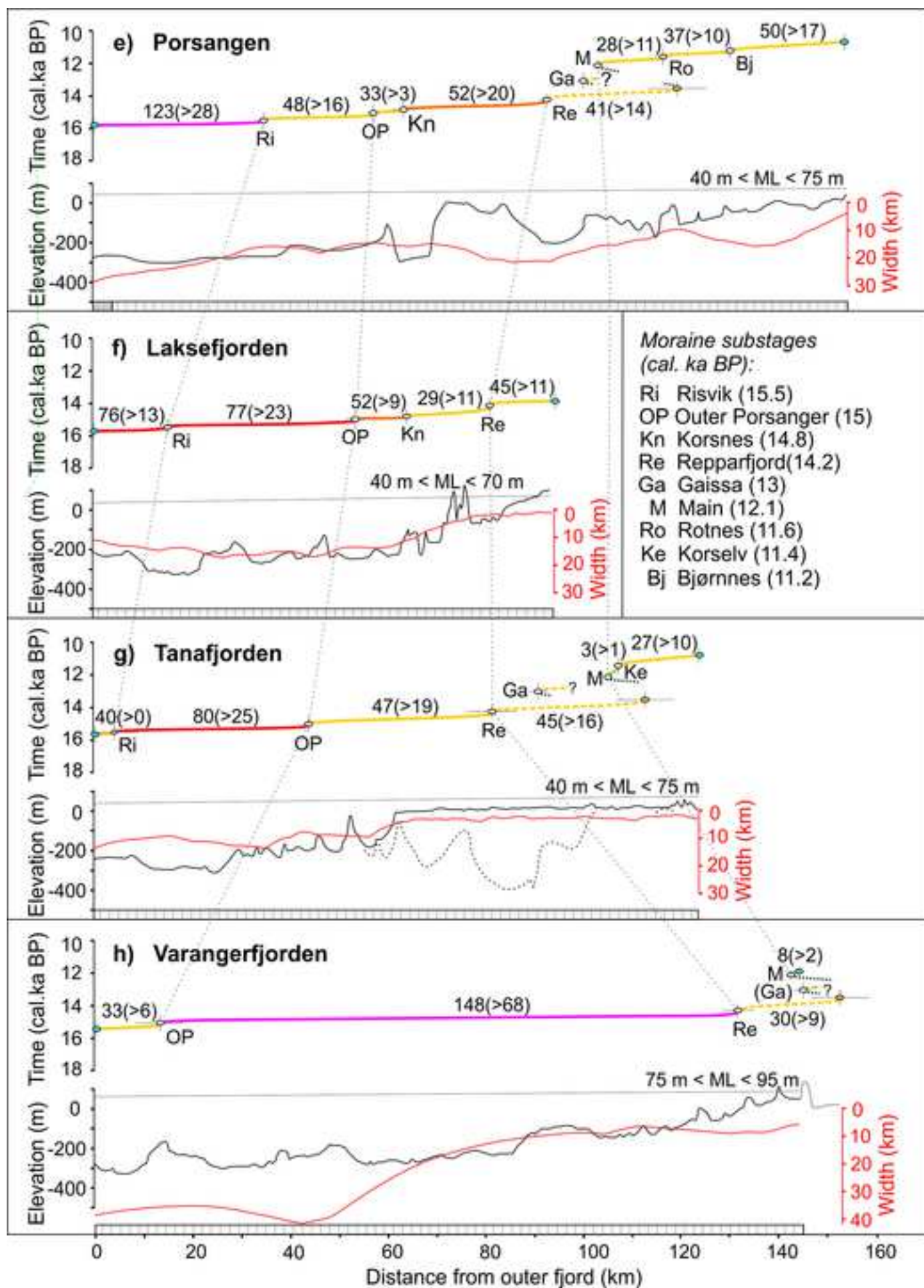


Figure 3

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1 **Supplementary Material for Data Repository (DR):**

2
3 **Manuscript Title:** Asynchronous response of marine-terminating outlet glaciers during
4 deglaciation of the Fennoscandian Ice Sheet

5 **Authors:** Stokes *et al.*
6

7 **1. Supplementary Methods: Dating and Correlating Ice Margin Retreat**

8 The pattern and chronology of glacier retreat in northern Norway following the Last
9 Glacial Maximum (LGM) has been known, in general outline, for several decades and up to
10 eight major sub-stages have been identified in Finnmark (Sollid *et al.*, 1973) and Troms
11 (Andersen, 1968). These sub-stages are based on extensive and detailed mapping of marginal
12 moraines that can be traced over considerable distances, together with raised shorelines cut
13 into end moraines or extending beyond ice-contact deltas; further augmented by radiocarbon
14 dates from marine sediments pre- or post-dating moraines (e.g. Marthinussen, 1962;
15 Andersen, 1968; Sollid *et al.*, 1973; Vorren and Elsborg, 1979; Corner, 1980). Much of this
16 work laid the foundation for a regional ice margin chronology, compiled by Andersen and
17 Karlsen (1986).

18 Andersen and Karlsen's (1986) map represents the only detailed attempt to both correlate
19 and date ice-front positions across the entire region. It clearly demonstrates the asynchrony in
20 ice margin retreat between different fjords, which is the focus of our investigation, but it
21 requires updating. This is because several studies have carried out new mapping and collected
22 new dates (Anderson *et al.*, 1995; Fimreite *et al.*, 2001; Olsen *et al.*, 2001a, b; Forwick and
23 Vorren, 2002; Vorren and Plassen, 2002; Eilertsen *et al.*, 2005, 2006; Vorren and Mangerud,
24 2008; Winsborrow *et al.*, 2010; Romundset *et al.*, 2011; R  ther *et al.*, 2011). Our objective,
25 therefore, is to synthesise data from these studies into an updated ice margin chronology for
26 each of our studied fjords (summarised in Table DR1). As in previous work, we do this by
27 correlating major ice marginal positions across the region, constrained by moraines of known
28 age (e.g. from radiocarbon dates) and based on previous work within individual fjords. Here,
29 we briefly summarise this work, which underpins the chronology in Table DR1 (below) that
30 is used in the manuscript (e.g. Fig's 1-3).

31 32 *1.1. Troms (Andfjorden, Malangen, Lyngen)*

33 The chronology of ice margin recession is best known in the western part of the study
34 area, in Troms, where most radiocarbon dates have been obtained (cf. Andersen 1968, Vorren
35 and Elvsborg 1979; Corner 1980; Forwick and Vorren, 2002; Vorren & Plassen 2002;
36 Eilertsen *et al.*, 2005). Indeed, previous work has already produced time-distance diagrams
37 for outlet glacier positions in Andfjorden-V  gsfjorden (Vorren and Plassen 2002) and the
38 inner part of Malangen-M  lselv (Eilertsen *et al.*, 2005), which we utilise. These depict phases
39 of both glacier retreat and re-advance, interspersed with more stable ice margin positions of
40 variable duration. Moraines have been correlated across the region based on their relationship
41 to raised shorelines. Radiocarbon dates give maximum or minimum ages for these moraines

42 or for deglaciation events (see Marthinussen, 1962; Andersen, 1968; Vorren and Elvsborg,
43 1979; Corner, 1980; Fimreite et al., 2001; Olsen et al., 2001a; Vorren and Plassen, 2002;
44 Forwick and Vorren, 2002; Eilertsen et al., 2005; Romundset et al., 2011).

45 Glacier re-advances have been documented in several cases, in the form of overridden or
46 over-consolidated sediments observed in sections (Andersen 1968, Corner 1980, Vorren &
47 Elvsborg 1979) and on seismic profiles (Lyså and Vorren, 1997; Vorren and Plassen 2002;
48 Eilertsen et al., 2005), and their extent reconstructed in the form of time-distance diagrams.
49 The largest re-advance occurred during the late Allerød (Tromsø-Lyngen re-advance), for
50 which there is evidence of overridden moraines at least 25 km proximal to these moraines
51 (Vorren & Plassen 2002, Eilertsen et al., 2005). Vorren and Plassen (2002) assumed a re-
52 advance of at least 40 km at this time, comparable to conditions in western Norway (cf.
53 Andersen et al 1995). To capture this uncertainty in our time-distance diagrams (Fig's 2 and
54 3), we have assumed a distance of 30 km (± 10 km) in the fjords in Troms.

55 As noted above, eight moraine sub-stages have been recognised between the shelf break
56 and the innermost fjords in Troms (Andersen 1968; Vorren and Plassen 2002). They are
57 (from oldest to youngest): Egga II, Flesen, D-event, Skarpnes (late Bølling-Allerød), Tromsø-
58 Lyngen (Younger Dryas), and Stordal I, II and III (Preboreal). The most complete chronology
59 has been obtained from Andfjorden-Vågsfjorden, where Vorren & Plassen (2002) combined
60 evidence of early deglacial sub-stages (Egga II, Flesen, D-event) recognised in marine
61 sediments, with evidence of younger glacial sub-stages (Skarpnes, Tromsø-Lyngen and
62 Stordal I, II and III) identified on the basis of regional terrestrial evidence (Andersen, 1968).
63 Their resulting chronology spans the time of glacier retreat from the shelf edge to the
64 innermost fjords. We have adjusted this chronology for the Preboreal (Stordal I, II and III),
65 based on more detailed radiocarbon and shoreline correlation dating of Preboreal moraines in
66 Lyngen-Storfjord (Corner, 1980) and Malangen-Målselv (Eilertsen et al., 2005), where two
67 major, and one minor, climatically controlled moraine sub-stages and several topographically
68 controlled ice-front accumulations have been recognised. Based on this body of work, the
69 resulting chronology (in cal. ka BP) for moraine sub-stages in Troms is as follows (cf. Table
70 DR 1):

71

72	1. Egga II	17.8 (± 0.3)
73	2. Flesen	17.3 (± 0.2)
74	3. D-Event	16.2 (± 0.3)
75	4. Skarpnes	14.2 (± 0.3)
76	5. Tromsø-Lyngen	12.1 (± 0.2)
77	6. Stordal I (= Ørnes in Lyngen, Kjerresnes in Målselv)	11.4 (± 0.2)
78	7. Stordal II (= Skibotn in Lyngen, Bardu-Storskog in Målselv)	10.8 (± 0.2)
79	8. Stordal III (= Nyli in Lyngen-Storfjord, Alapmoen in Målselv)	10.4 (± 0.2)

80

81 This chronology has been applied to all three fjord systems in Troms included in this
82 study. It is estimated (conservatively) to be reliable to within ± 200 -500 yrs. The uncertainty
83 increases with increasing age and, for pre-Skarpnes ice front positions, distance from the
84 reference area in Andfjorden-Vågsfjorden. Ice-front positions corresponding to the Skarpnes

85 and younger events have been positively identified in all three fjord systems. Pre-Skarpnes
86 glacier positions in Malangen and Lyngen have not been identified and their position is
87 inferred from comparison with Andfjorden-Vågsfjorden. Malangen is situated fairly close to
88 Andfjorden and has a similar setting with regard to topography and distance to the shelf edge.
89 Its early deglaciation history, therefore, is assumed to be similar to that of Andfjorden (cf.
90 Rydningen et al., 2013). Lyngen, however, is located much farther from Andfjorden and
91 much farther from the shelf edge, and probably has more in common with Finnmark than
92 Troms regarding its early deglaciation history (see below).

93

94 1.2. Finnmark (*Altafjorden, Porsangen; Laksefjorden; Tanafjorden and Varangerfjorden*)

95 In Finnmark, the dynamics and timing of glacier retreat and re-advance are less well
96 known than in Troms. Fewer radiocarbon dates are available, and the chronology of glacier
97 retreat is based largely on morpho-stratigraphic correlation of marginal moraines and raised
98 shorelines (Sollid et al., 1973). The following regional moraine succession (from oldest to
99 youngest) has been established: Risvik, Outer Porsangen, Korsnes, Repparfjord, Gaissa,
100 Main, and up to two successive Preboreal moraines (Lampe-Jordall in Altafjorden (Follestad,
101 1979), Rotnes and Bjørnnes in Porsangen, and Korselv in Tanafjorden). At least one glacier
102 re-advance is indicated by the way in which the Younger Dryas ('Main') sub-stage moraine
103 overrides the Gaissa moraine in some areas (Sollid et al. 1973). This indicates a glacier re-
104 advance of at least several kilometres, possibly corresponding to the late-Allerød re-advance
105 in Troms (see below).

106 The 'Main' sub-stage can be reliably correlated with the Tromsø-Lyngen (Younger
107 Dryas) sub-stage in Troms on the basis of raised shoreline correlation and moraine continuity.
108 An age of 12.1 ± 0.3 cal. ka BP is assumed for this sub-stage. Younger moraine sub-stages
109 (Lampe-Jordall in Alta, Rotnes and Bjørnnes in Porsangen, and Korselv in Tanafjorden) are
110 assigned a Preboreal age, approximately equivalent to the Stordal I, or possibly Stordal II in
111 Troms, on the basis of shoreline correlation (cf. Sollid et al., 1973), i.e. ca. 11.4 ± 0.5 cal. ka
112 BP. The Repparfjord sub-stage has been correlated with the Skarpnes sub-stage in Troms
113 based on raised shoreline evidence (cf. Marthinussen 1962) and is consequently assigned an
114 age of 14.2 ± 0.3 cal. ka BP, although a younger age has also been suggested (discussed
115 below).

116 Among the older sub-stages, there are various ways to approach correlation. One
117 approach, using a direct comparison with Troms, would be to correlate the prominent Outer
118 Porsanger sub-stage with either the Flesen moraine (17.3 cal. ka BP) or the D-event (16.2 cal.
119 ka BP) in Andfjorden. Accordingly, the Risvik event, represented by marginal moraines on
120 the outermost coast at Porsangen, would be even older, suggesting coastal deglaciation
121 around 17 cal. ka BP. Ages close to these dates or younger were suggested by Olsen et al.
122 (1996; 2001a; b).

123 A second approach would be to correlate the Flesen and D-events in Andfjorden with
124 deglacial stages recognised in the Barents Sea. Such a correlation seems more likely given
125 that the coast of Finnmark is much farther from the shelf edge than the coast of western
126 Troms. Thus, Winsborrow *et al.* (2010) tentatively correlated their Stages 2 and 3 in the

127 Barents Sea with the Flesen and D-events, respectively. They suggested dates of 16 cal. ka
128 BP for late stage (Stage 3) ice-stream activity off the coast of Finnmark, and 15 cal. ka BP for
129 retreat of the ice margin to an onshore position. This is consistent with recent dating evidence
130 from the southern Barents Sea (Juntilla et al., 2010; R  ther et al. 2011), e.g. an age of 16.6
131 cal. ka BP for the Stage 2 Outer Bj  rn  yrenna sediment wedge.

132 A third approach was taken by Romundset *et al.* (2011). They used maximum ages of
133 14.1 and 14.3 cal. ka BP obtained from marine fossils in basal sediments in coastal lake
134 basins to infer deglaciation of the outer coast of Finnmark (Rolf  ya) at around that time.
135 Comparing these dates and glacial sub-stages in the fjords with cold events indicated by the
136 NGRIP ice core (NGRIP Members, 2004), they correlated the Outer Porsanger sub-stage with
137 the Older Dryas (B  lling-Aller  d) cold period (ca. 14 cal. ka BP), and the Repparfjord and
138 Skarpnes sub-stages with an inter-Aller  d cold event (ca. 13 cal. ka BP). Their correlation
139 makes these moraines more than 1,000 years younger than previously assumed. However, it
140 should be noted that their dated lake sediments represent only minimum ages for the date of
141 deglaciation of the outer coast, and other dating evidence suggesting an age of around 14 cal.
142 ka BP for the Skarpnes event in Troms (see above) cannot easily be dismissed. Thus,
143 assuming that the Skarpnes and Repparfjord sub-stages have the same age, which seems
144 reasonable on the basis of raised shoreline evidence, we assign the Repparfjord sub-stage an
145 age closer to 14 cal. ka BP than 13 cal. ka BP.

146 Because of the uncertainty regarding the age of sub-stages in Finnmark, the error
147 estimates are larger than those for Troms. The Korsnes sub-stage is assigned an intermediate
148 age closer to the age of the Outer Porsanger than Repparfjord, based on its position. Gaissa
149 sub-stage moraines, which in places are overrun by moraines of the Main sub-stage, are
150 tentatively assigned a late Aller  d age based on a correlation with an early phase of the
151 Troms  -Lyngen event in Troms, for which there is some evidence (Andersen 1968; Vorren
152 and Plassen, 2002). For inter-stadial ice front positions in Finnmark, we have assumed a
153 similar age for the Aller  d inter-stadial as in Troms, and a position estimated at 20 km (rather
154 than 30 km) behind the ‘Main’ sub-stage moraines, based on the assumption that glacier
155 fluctuations in the more continental setting of Finnmark were more subdued than in maritime
156 Troms. Thus, our chronology (in cal. ka BP; cf. Table DR 1) is based on: (i) age constraints
157 for coastal deglaciation and outer moraine sub-stages provided by recent data from the
158 Barents Sea; (ii) correlation with Troms for Repparfjord and later sub-stages, and (iii), an
159 inferred Aller  d inter-stadial position:

160

161	1. Risvik	15.5 (± 0.5)
162	2. Outer Porsanger	15 (± 0.5)
163	3. Korsnes	14.8 (± 0.5)
164	4. Repparfjord	14.2 (± 0.5)
165	5. Aller��d IS	13.5 (± 0.5)
166	6. Gaissa	13 (± 0.5)
167	7. Main	12.1 (± 0.3)
168	8. Rotnes	11.6 (± 0.5)
169	9. Korselv/Lampe-Jordfall	11.4 (± 0.5)

170 10. Bjønnes 11.2 (± 0.5)

171

172 1.3. Age Uncertainties

173 It is important to acknowledge the inherent uncertainty in any reconstruction of palaeo-
174 glacier behaviour and we acknowledge that subsequent dating may lead to revisions of the ice
175 margin chronology depicted in this paper. However, we note that our work builds on a rich
176 legacy of previous work, giving this region one of the most detailed ice margin chronologies
177 available for any palaeo-ice sheet. Indeed, because of the extensive morpho-stratigraphic
178 correlations, it is unlikely that new dates will lead to a radical revision of the broad pattern of
179 retreat within each fjord. The time-distance diagrams from each fjord (e.g. Fig's 2 & 3) show
180 error bars to clarify areas of uncertainty. Age uncertainties are based on an appraisal of: (i)
181 the radiocarbon dating error, (ii) the number of dates, (iii) the correlation uncertainty and age
182 relationships to older and younger events. They give a relative measure of uncertainty and are
183 believed to be maximum estimates that span the complete range of 'known' uncertainty.
184 Distance error bars are used in cases where the approximate, rather than precise, position of
185 the ice front (substage) is known. Importantly, we note that these uncertainties are largely
186 insignificant to our aim of reconstructing the broad patterns of ice margin recession in each
187 fjord (e.g. relative, rather than absolute, changes in retreat rate, as shown by the vertical
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189

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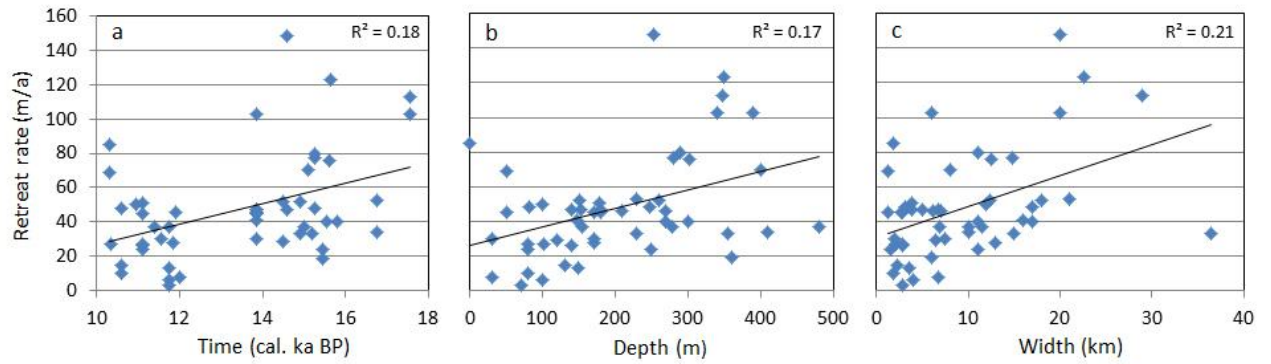
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Figure DR1: Retreat rate versus time (a), depth (b) and width (c). Depth takes account of isostatic rebound and post-retreat sediment fill, where known.

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Table DR 1: Correlated and dated moraine sub-stages (green) and inter-stadials (IS: red) from a synthesis of the literature used for calculating ice margin retreat rates in Troms and Finnmark. Estimated error limits for their age (ka) and position (km) in each fjord are given and used to assign uncertainty to the retreat rates in Figure's 2 and 3. Interpolated dates for ice margin positions at the mouth and end of each fjord (yellow: cf. Fig's 1, 2 and 3), or where correlation is more tentative, are shown in brackets. LJ: Lampe-Jordfall.

TROMS ¹						FINNMARK ^{2,11} and Barents Sea ^{3,4,5}													
Location:		Andfjorden ^{1,7}		Malangen ^{1,7,8}		Lyngen ⁹		Location:		Altafjorden ²		Porsanger ²		Laksefjorden ²		Tanafjorden ²		Varangerfjorden ²	
Marine limit:		30-95 m		30-85 m		40-95 m		Marine limit:		40-85 m		40-75 m		40-70 m		45-75 m		75-95 m	
Glacial event:		Age	Error	Age	Error	Age	Error	Glacial event:		Age	Error	Age	Error	Age	Error	Age	Error	Age	Error
		cal.ka BP	± ka/km	cal.ka BP	± ka/km	cal.ka BP	± ka/km			cal.ka BP	± ka/km	cal.ka BP	± ka/km	cal.ka BP	± ka/km	cal.ka BP	± ka/km	cal.ka BP	± ka/km
Egga II		17.8	0.3/0	17.8	0.3/0			Stage 1		(Barents Sea, shelf edge: c. 19) ^{3,5}									
Flesen		17.3	0.2/0	17.3	0.2/5			Stage 2		(Barents Sea, Outer Bjørnøyrenna sediment wedge: 17.1 - 16.6) ^{3,4,5}									
F-D IS		16.8	0.3/5	16.8	0.3/5			Ende Stage 2		(Barents Sea, retreat from OBSW, c. 16.5) ^{3,4}									
D-event		16.2	0.3/5	(16.2)	0.3/5	(16.1)	0.5/0	Stage 3		(Barents Sea mostly ice-free, c. 16) ^{3,5}									
								Coast deglaciation ^{7,6}		(15.8)	0.5/0	(15.8)	0.5/0	(15.7)	0.5/0	(15.6)	0.5/0		
						(15.5)	0.5/0	Risvik				15.5	0.5/1	15.5	0.5/2.5	15.5	0.5/5	(15.4)	0.5/0
								Outer Porsanger				15.0	0.5/0	15.0	0.5/2	15.0	0.5/1		
								Korsnes				(14.8)	0.5/2.5	(14.8)	0.5/1				
Bølling IS		14.7	0.3/5	14.7	0.3/5	14.7	0.3/5			(14.5)	(no plot)								
Skarpnes		14.2	0.3/2	14.2	0.3/0	14.2	0.3/0	Repparfjord		(14.2)	0.6/10	14.2	0.4/1	14.2	0.4/1	14.2	0.4/5	14.2	0.4/3
Allerød IS		13.5	0.4/10	13.5	0.4/10	13.5	0.4/10	Allerød IS		13.5	0.5/10	13.5	0.5/6	(13.9)	0.5/0	13.5	0.5/6	13.5	0.5/6
								Gaissa				13	0.5/1.5			13	0.5/2	13	0.5/2*
Tromsø-Lyngen ¹⁰		12.1	0.2/0	12.1	0.2/0	12.1	0.2/0	Main		12.1	0.3/0	12.1	0.3/1			12.1	0.3/0	12.1	0.3/0
Målsnes				11.7	0.2/0			Rotnes				11.6	0.5/1					(11.9)	0.3/0
Stordal I		11.4	0.2/0	11.4	0.2/0	11.4	0.2/0	LJ/Korselv/Bjørnnes		11.4 (LJ)	0.5/0	11.2 (B)	0.5/1			11.4 (K)	0.5/0		
Stordal II		10.8	0.2/0	10.8	0.2/0	10.8	0.2/0			(10.7)	0.5/0	(10.7)	0.5/0			(10.8)	0.5/0		
Stordal III		10.4	0.2/0	10.4	0.2/0	10.4	0.2/0												
Final deglac.		(10.3)	0.2/0	(10.2)	0.2/0	(10.2)	0.2/0			* Gaissa substage margin in Varangerfjorden located just inside (overridden by) the Main substage moraine.									

Key data sources for correlating and dating glacial events:

¹Andersen (1968): Moraine, shoreline and equilibrium line altitude correlations; biostratigraphy and 17 radiocarbon dates (mostly shells from raised marine sediments)

²Sollid et al. (1973): Regional moraine mapping and shoreline correlation; shorelines used as 'dating lines'

³Winsborrow et al. (2010): Reconstructed flow patterns and dynamics (e.g. location of ice streams) during deglaciation; compilation of 14 radiocarbon dates from marine samples (foraminifera and macrofossils).

⁴Rüther et al. (2011) Seismic- and lithostratigraphy-based reconstruction of deglaciation; 6 radiocarbon dates from marine samples (foraminifera and macrofossils)

⁵Junttila et al. (2010): Dating marine sediments and deglaciation; 6 radiocarbon dates from foraminifera and macrofossils.

⁶Romundset et al. (2011): Raised coastal lake stratigraphy; 2 radiocarbon dates (shell and algae) relating to deglaciation.

⁷Vorren & Plassen (2002): Marine stratigraphy and compilation of 44 radiocarbon dates relating to deglaciation of Andfjorden.

⁸Eilertsen et al. (2005): Litho- and seismic stratigraphy and mapping; 25 radiocarbon dates (mostly shells from raised marine sediments).

⁹Comer (1980): Moraine and shoreline correlation; 3 radiocarbon dates from raised marine sediments (shells).

¹⁰Fimreite et al. (2001): Pollen stratigraphy of lake sediments; 5 radiocarbon dates from samples of gyttja.

¹¹Olsen et al. (2001): Radiocarbon dates from glacial sediments containing low carbon content.

270