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

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Abstract:	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-deepenings 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.
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Asynchronous response of marine-terminating outlet glaciers during deglaciation of the Fennoscandian Ice Sheet

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ASBTRACT

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 andmost 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-deepenings 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.

INTRODUCTION

Ice sheets are organized into a pattern of tributaries feeding rapidly flowing ice streams, separated by slow-flowing ice. In coastal regions, many ice streams are influenced by topography and become confined within deep troughs as marine-terminating outlet glaciers. Because of their disproportionate ice flux, they are a key component of ice sheet mass balance and associated impacts on sea level (Thomas et al., 2011; Nick et al., 2013). Indeed, there is an urgent need to understand the longer-term significance of recent ‘dynamic’ changes that have been observed on outlet glaciers, typically characterized by their accelerating flow, thinning and retreat (Howat et al., 2007; Andresen et al., 2012; Nick et al., 2013). This, however, is difficult; partly because of the complexity in identifying factors that drive such changes (e.g., air and ocean temperatures, fjord geometry; see Carr et al., 2013), and partly because there are few records of outlet glacier behavior over centennial to millennial time-scales. Moreover, such records are mostly restricted to just one or two outlet glaciers (e.g., Briner et al., 2009; Young et al., 2011; Hughes et al., 2012; Mangerud et al., 2013).

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

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

METHODS

Ice sheet retreat in northern Norway following the LGM has been known in general outline for several decades, and eight major sub-stages have been identified and dated in Finnmark (Sollid et al., 1973) and Troms (Andersen, 1968). These are based on extensive mapping of moraines, traced over considerable distances, together with raised shorelines cut into end moraines or extending beyond ice-contact deltas; further augmented by radiocarbon dates from marine sediments pre- or post-dating moraines (e.g., Andersen, 1968; Sollid et al., 1973). We identified known ice front positions from previous work that have been assigned to the established regional sub-stages (see Table DR1 in GSA Data Repository[1]). We then reviewed and, in a small number of cases, revised positions and ages based on new bathymetric data, new mapping of glacial geomorphology (e.g., Ottesen et al., 2008; Winsborrow et al., 2010; Rydningen et al., 2013), and more recent radiocarbon dates (Vorren and Plassen, 2002; Eilertsen et al., 2005).

Transects were then drawn to extract bathymetric data from each fjord using the MAREANO multibeam dataset collected by the Norwegian Hydrographic Service (www.mareano.no), and we estimated the width of the outlet glacier at the calving front at 50 m increments during retreat. Transects extend from the outermost part of the fjord to the marine limit at the head of the fjord at the time it was deglaciated (Fig. 1). They therefore cover the entire palaeofjord, projecting landward into what are now fjord-valleys, and are depth-adjusted for former sea level and postglacial infill, where known. Retreat rates between each sub-stage were calculated in each fjord and we assign errors based on: (1) the reported radiocarbon age uncertainty and an appraisal of the number of dates in each region, (2) stratigraphic evidencerelating to the dates and glacial events, and (3) the strength of regional correlations of marginal moraines and raised shorelines related to particular events and dates. These errors (Table DR1; Figures 2 and 3) capture the maximum range of 'known' uncertainty and do not influence the broad patterns of ice margin recession in each fjord, which is our focus (i.e. relative changes in retreat rate between the dated sub-stages).

RESULTS

As in previous work (e.g., Briner et al., 2009; Mangerud et al., 2013), we present time-distance diagrams for each fjord system (Figs. 2 and 3). Retreat rates averaged over several millennia (~30 m a^{-1}) were less than half those observed on modern-day outlet glaciers over decadal time-scales (>100 m a^{-1}; Howat and Eddy, 2011). However, maximum rates typically exceeded 100 m a^{-1}. Due to the inherent uncertainties of the dated ice margin positions, we focus on the broad patterns of retreat within each fjord, with particular attention as to when and where the major sub-stages are recorded, and when retreat rates increased. In this regard, five out of the eight glaciers experienced their most rapid retreat during early deglaciation (before 15 cal. kyr B.P.), when air temperatures were relatively cold (Fig. 1c). This typically occurred across major overdeepenings on the continental shelf (e.g., Andfjorden, Malangen) or through the mid- to outer-fjord areas (e.g., Altafjorden, Varangen). In some cases, rapid retreat occurred over a reverse bed slope into progressively deeper water (e.g., Andfjorden, Malangen), or simply where water depths were greatest (e.g., Porsangen, Tanafjorden). Whilst the expectation is that glaciers will tend to retreat more rapidly through deeper water (cf. Schoof, 2007), the correlation between water depths and retreat rates is perhaps not as strong as might be expected ($R^2 = 0.17$; Fig. DR1b in the Data Repository). There are cases where rapid retreat shows no obvious correlation with bathymetric changes (e.g., Laksefjorden) and where retreat was relatively slow (~40 m a^{-1}) through deep water,often coinciding with narrow troughs or localized constrictions (e.g., Lyngen, Malangen). Indeed, fjord width shows a stronger correlation with retreat rate ($R^2 = 0.21$; Fig. DR1c), but this is not straightforward, in that some narrow troughs (2–3 km wide) were evacuated relatively quickly (e.g., Altafjorden).

DISCUSSION

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

An explanation for the asynchronous pattern of retreat is the variable topography within each fjord (Fig. 1). There are clear cases where maximum retreat rates coincide with reverse bed slopes (e.g., Andfjorden, Malangen) and/or deep (200–300 m) water (e.g., Porsangen, Tanafjorden). These cases support the importance of water depth in inducing episodes of rapid retreat (e.g., Schoof, 2007; Briner et al., 2009). It also explains why the highest retreat rates in most fjords occurred during early deglaciation, because this is when outlet glaciers were more likely to encounter basal overdeepenings (Fig. 1). Thus, we find a clear indication that deep and wide fjords, characterized by subglacial overdeepenings, always induce episodes of rapid retreat (e.g., >100 m a^{-1} in Andfjorden, Malangen, Altafjorden, and Varangerfjorden). Thus, although atmospheric warming will inevitably lead to deglaciation by inducing a negative ice sheet mass balance, there is not always an obvious correlation between climate forcing and the rate of retreat of outlet glaciers over centennial to millennial time-scales. Measurements of glacier retreat following the Bølling-Allerød warming at ca. 14.5 cal. kyr B.P., for example, would reveal terminus positions ranging from mid to inner-fjord areas and retreat rates ranging from <30 m a^{-1} (Laksefjorden) to >140 m a^{-1} (Varangerfjorden).

Although topographic factors can clearly influence glacier retreat, the fact that these relationships are not stronger (Figs. DR1b and DR1c) indicates a complex interplay between them. Glaciers may retreat slowly in deep fjords if they are narrow, or in wide fjords if they are shallow. It is also important to note that we only measure fjord depth and width at the inferred glacier terminus, and not the longitudinal gradients of glacier depth and width. Retreat rates will be affected by thinning at the glacier terminus, which is further affected by the longitudinal flux gradient. As such, glaciological factors further modulate outlet glacier behavior, and the size and slope of the catchment are likely to be important. Those with larger, higher catchment areas are more likely to be able to sustain ice fluxes and maintain a stable grounding line position in deep water or across reverse bed slopes (Schoof, 2007; Jamieson et al., 2012). If glaciers are unable to balance calving by draw-down of ice, it is likely to lead to thinning and retreat. The anomalous period of rapid retreat in Tanafjorden between ca. 15.5 and 15 cal. kyr B.P. might be a reflection of a small catchment area that was rapidly diminished by drawdown caused by retreat in Varangenfjorden. We suggest, therefore, that interactions between adjacent ice stream catchments (ice piracy and capturing) are likely to be an important control on outlet glacier dynamics over centuries to millennia (cf. Payne and Dongelmans, 1997). This complicates attempts to numerically model the behavior of individual outlet glaciers over these time-scales, which are often targeted at specific glaciers and necessarily omit interactions with neighboring catchments (e.g., Jamieson et al., 2012; Nick et al., 2013).

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

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

CONCLUSIONS

Recent work suggests a rapid and near-synchronous response of outlet glaciers to large-scale oceanic and atmospheric conditions over decadal time-scales (Andresen et al., 2012). This paper reconstructs the retreat of eight major outlet glaciers during deglaciation of the FIS (18–10 cal. kyr B.P.) and shows that, despite experiencing a similar climate forcing, their response was asynchronous over millennial time-scales. We interpret this to reflect the complex interplay between climate forcing and both topographic (e.g., bathymetry, width) and glaciological factors (e.g., the evolution of catchment areas) that evolve through time, but there is clear evidence that basal over-deepenings in wide fjords induce episodes of rapid retreat (>100 m a^{-1}), further exacerbated by their susceptibility to oceanic warming. Thus, high resolution data of subglacial topography beneath the catchments of modern-day outlet glaciers is likely to be a crucial requirement for modeling and assessment of future ice sheet dynamics (Durand et al., 2011). Such modeling will offer further opportunities to assess the sensitivity of outlet glaciers to a range of forcing factors and, in this regard, this paper offers a valuable data set much longer than the current observational record.

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FIGURE CAPTIONS

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

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

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

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

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

Figure 1

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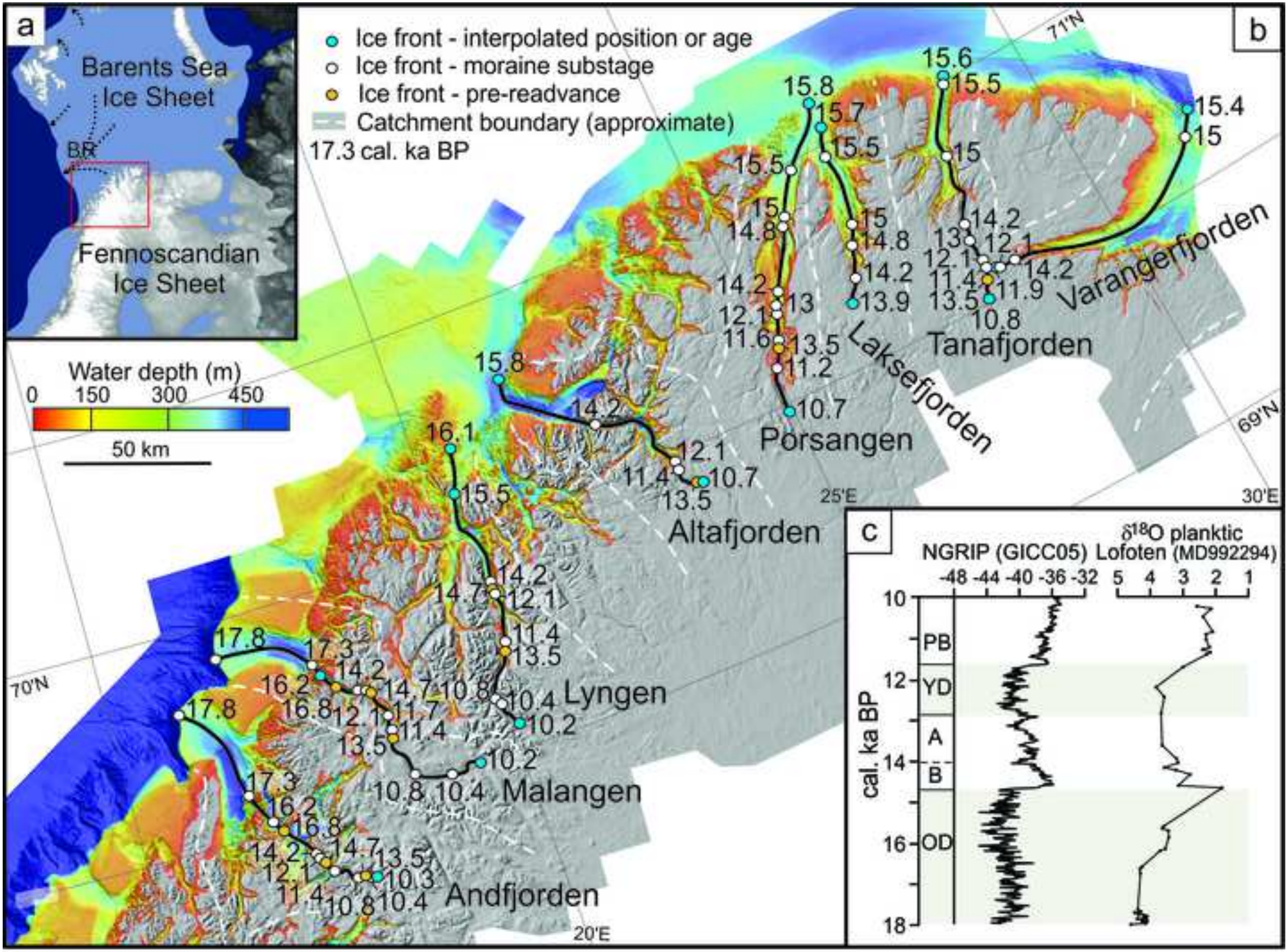

Figure 2
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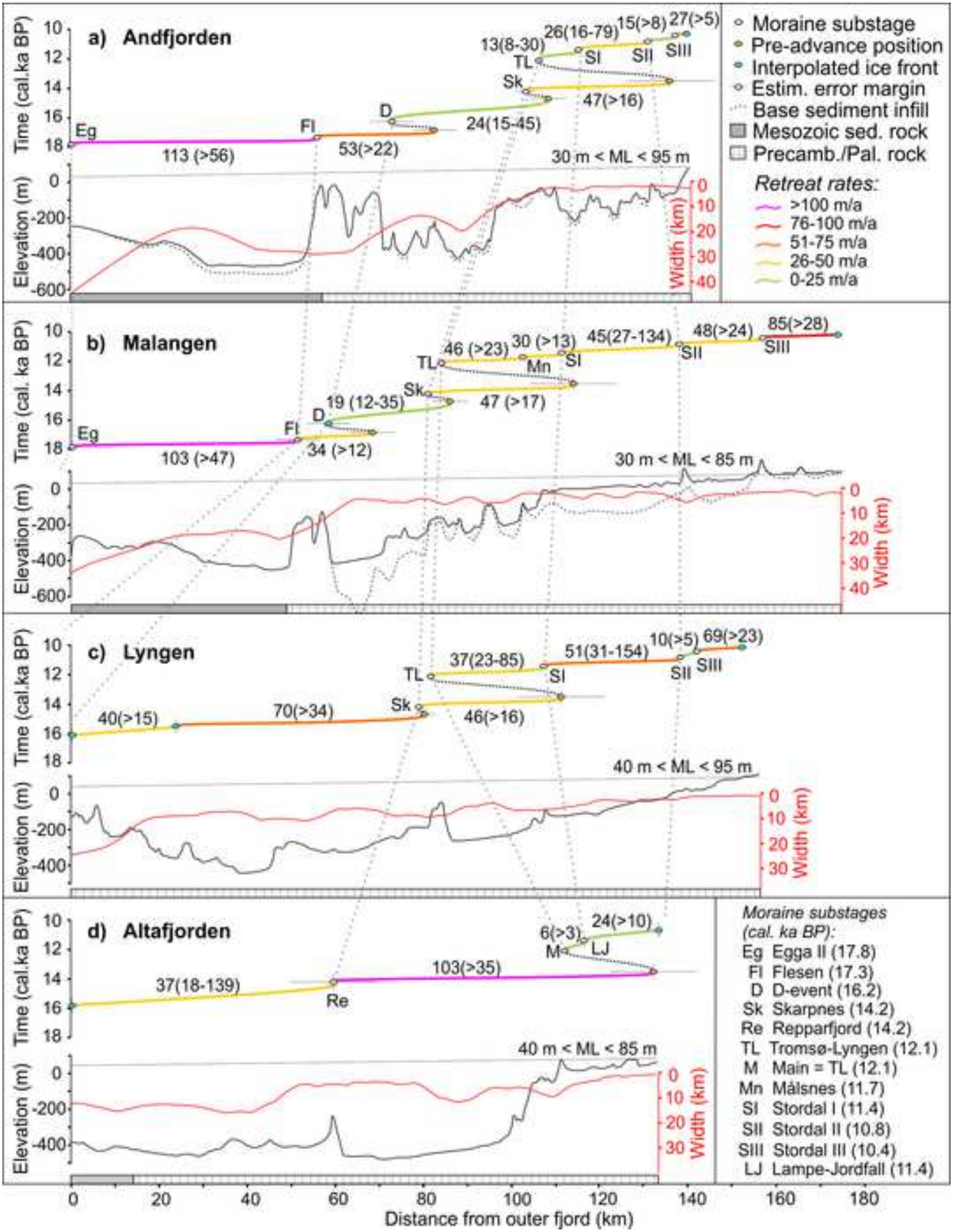

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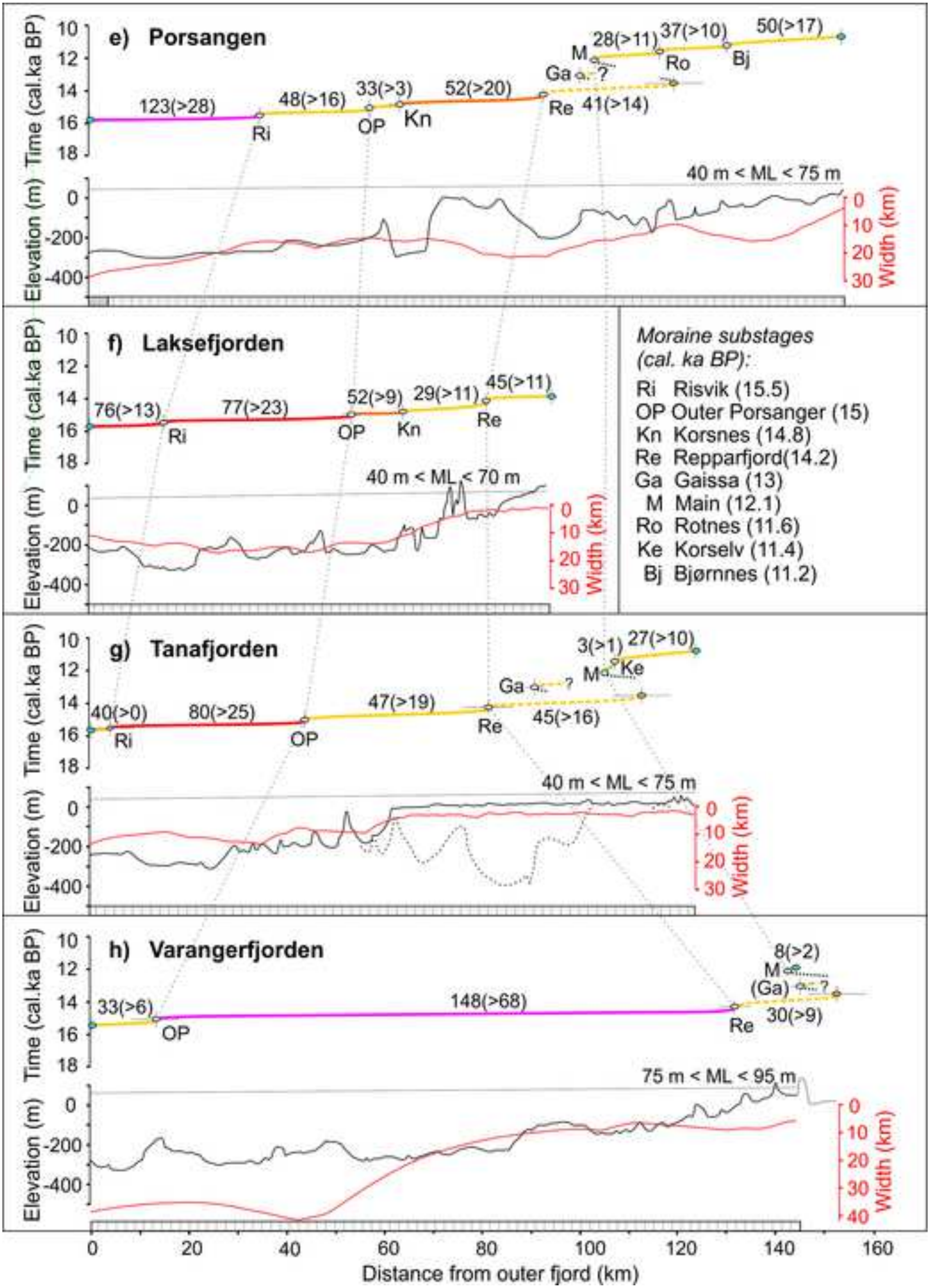

Supplementary Material for Data Repository (DR):

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

Authors: Stokes *et al.*

1. Supplementary Methods: Dating and Correlating Ice Margin Retreat

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

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

1.1.Troms (Andfjorden, Malangen, Lyngen)

The chronology of ice margin recession is best known in the western part of the study area, in Troms, where most radiocarbon dates have been obtained (cf. Andersen 1968, Vorren and Elvsborg 1979; Corner 1980; Forwick and Vorren, 2002; Vorren & Plassen 2002; Eilertsen *et al*., 2005). Indeed, previous work has already produced time-distance diagrams for outlet glacier positions in Andfjorden-Vågsfjorden (Vorren and Plassen 2002) and the inner part of Malangen-Målselv (Eilertsen *et al.*, 2005), which we utilise. These depict phases of both glacier retreat and re-advance, interspersed with more stable ice margin positions of variable duration. Moraines have been correlated across the region based on their relationship to raised shorelines. Radiocarbon dates give maximum or minimum ages for these moraines or for deglaciation events (see Marthinussen, 1962; Andersen, 1968; Vorren and Elvsborg, 1979; Corner, 1980; Fimreite et al., 2001; Olsen et al., 2001a; Vorren and Plassen, 2002; Forwick and Vorren, 2002; Eilertsen et al., 2005; Romundset et al., 2011).

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

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

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

This chronology has been applied to all three fjord systems in Troms included in this study. It is estimated (conservatively) to be reliable to within ±200-500 yrs. The uncertainty increases with increasing age and, for pre-Skarpnes ice front positions, distance from the reference area in Andfjorden-Vågsfjorden. Ice-front positions corresponding to the Skarpnes and younger events have been positively identified in all three fjord systems. Pre-Skarpnes glacier positions in Malangen and Lyngen have not been identified and their position is inferred from comparison with Andfjorden-Vågsfjorden. Malangen is situated fairly close to Andfjorden and has a similar setting with regard to topography and distance to the shelf edge. Its early deglaciation history, therefore, is assumed to be similar to that of Andfjorden (cf. Rydningen et al., 2013). Lyngen, however, is located much farther from Andfjorden and much farther from the shelf edge, and probably has more in common with Finnmark than Troms regarding its early deglaciation history (see below).

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

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

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

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

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

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

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

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

1. Risvik	15.5 (± 0.5)
2. Outer Porsanger	15 (± 0.5)
3. Korsnes	14.8 (± 0.5)
4. Repparfjord	14.2 (± 0.5)
5. Allerød IS	13.5 (± 0.5)
6. Gaissa	13 (± 0.5)
7. Main	12.1 (± 0.3)
8. Rotnes	11.6 (± 0.5)
9. Korselv/Lampe-Jordfall	11.4 (± 0.5)

10.	Bjønnes	11.2 (± 0.5)

1.3. Age Uncertainties

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

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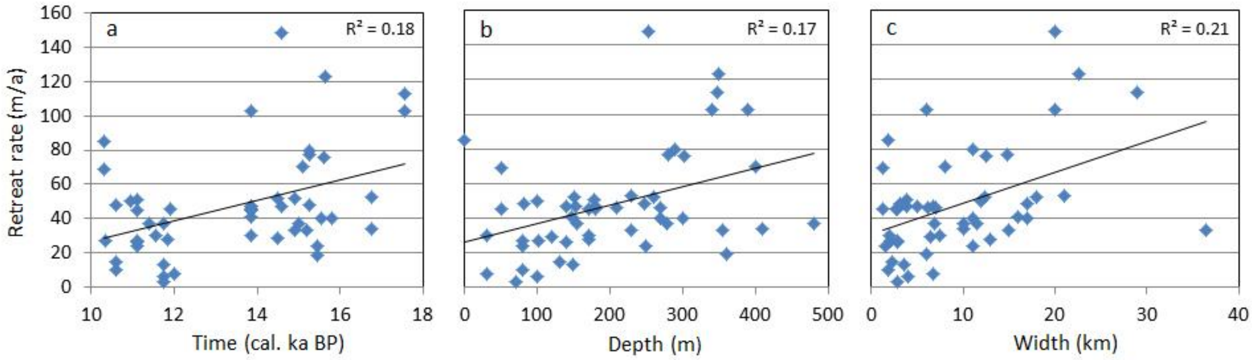

Figure DR1: Retreat rate versus time (**a**), depth (**b**) and width (**c**). Depth takes account of isostatic rebound and post-retreat sediment fill, where known.

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]						
Location:	Andfjorden[1,7]		Malangen[1,7,8]		Lyngen[9]	
Marine limit:	30-95 m		30-85 m		40-95 m	
Glacial event:	*Age cal.ka BP*	*Error ± ka/km*	*Age cal.ka BP*	*Error ± ka/km*	*Age cal.ka BP*	*Error ± ka/km*
Egga II	17.8	0.3/0	17.8	0.3/0		
Flesen	17.3	0.2/0	17.3	0.2/5		
F-D IS	16.8	0.3/5	16.8	0.3/5		
D-event	16.2	0.3/5	(16.2)	0.3/5	(16.1)	0.5/0
					(15.5)	0.5/0
Bølling IS	14.7	0.3/5	14.7	0,3/5	14.7	0.3/5
Skarpnes	14.2	0.3/2	14.2	0.3/0	14.2	0.3/0
Allerød IS	13.5	0.4/10	13.5	0.4/10	13.5	0.4/10
Tromsø-Lyngen[10]	12.1	0.2/0	12.1	0.2/0	12.1	0.2/0
Målsnes			11.7	0.2/0		
Stordal I	11.4	0.2/0	11.4	0.2/0	11.4	0.2/0
Stordal II	10.8	0.2/0	10.8	0.2/0	10.8	0.2/0
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

FINNMARK[2,11] and Barents Sea[3,4,5]										
Location:	Altafjorden[2]		Porsangen[2]		Laksefjorden[2]		Tanafjorden[2]		Varangerfjorden[2]	
Marine limit:	40-85 m		40-75 m		40-70 m		45-75 m		75-95 m	
Glacial event:	*Age cal.ka BP*	*Error ± ka/km*	*Age cal.ka BP*	*Error ± ka/km*	*Age cal.ka BP*	*Error ± ka/km*	*Age cal.ka BP*	*Error ± ka/km*	*Age cal.ka BP*	*Error ± ka/km*
Stage 1	(Barents Sea, shelf edge: c. 19)[3,5]									
Stage 2	(Barents Sea, Outer Bjørnøyrenna sediment wedge: 17.1 - 16.6)[3,4,5]									
Ende Stage 2	(Barents Sea, retreat from OBSW, c. 16.5)[3,4]									
Stage 3	(Barents Sea mostly ice-free, c. 16)[3,5]									
Coast deglaciation[5),6)]	(15.8)	0.5/0	(15.8)	0.5/0	(15.7)	0.5/0	(15.6)	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	15.0	0.5/5
Korsnes			(14.8)	0.5/2.5	(14.8)	0.5/1				
	(14.5)	(no plot)								
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	0.5/10	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*
Main	12.1	0.3/0	12.1	0.3/1			12.1	0.3/0	12.1	0.3/0
Rotnes			11.6	0.5/1					(11.9)	0.3/0
LJ/Korselv/Bjørnnes	11.4 (LJ)	0.5/0	11.2 (B)	0.5/1			11.4 (K)	0.5/0		
	(10.7)	0.5/0	(10.7)	0.5/0			(10.8)	0.5/0		

* Gaissa substage margin in Varangerfjorden located just inside (overridden by) the Main substage moraine.

Key data sources for correlating and dating glacial events:

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

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

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

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

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

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

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

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

[9]Corner (1980): Moraine and shoreline correlation; **3 radiocarbon dates** from raised marine sediments (shells).

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

[11]Olsen et al. (2001): **Radiocarbon dates** from glacigenic sediments containing low carbon content.