Deglacial dynamics of the Vestfjorden - Trænadjupet paleo-ice stream, northern Norway

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Running title: Deglacial dynamics of a paleo-ice stream

ABSTRACT

Few well-dated records of the deglacial dynamics of the large paleo-ice streams of the major northern hemisphere ice sheets are presently available, a prerequisite for an improved understanding of the ice-sheet response to the climate warming of this period. Here we present a transect of gravity-core samples through Trænadjupet and Vestfjorden, northern Norway, the location of the Trænadjupet - Vestfjorden paleo-ice stream of the NW sector of the Fennoscandian Ice Sheet. Initial ice recession from the shelf break to the coastal area (~400 km) occurred at an average rate of about 195 m/yr, followed by two ice readvances, at 16.6 - 16.4 ka BP (the Røst readvance) and at 15.8 – 15.6 ka BP (the Værøy readvance), the former at an estimated ice advance rate of 216 m/yr. The Røst readvance has been interpreted to be part of a climatically-induced regional cold spell while the Værøy readvance was restricted to the Vestfjorden area and possibly formed as a consequence of internal ice-sheet dynamics. Younger increases in IRD content have been correlated to the Skarpnes (Bølling – Older Dryas) and Tromsø – Lyngen (Younger Dryas) Events. Overall, the decaying Vestfjorden paleo-ice stream responded to the climatic fluctuations of this period but ice response due to internal reorganization is also suggested. Separating the two is important when evaluating the climatic response of the ice stream. As demonstrated here, the latter may be identified using a regional approach involving the studies of several paleo-ice streams. The retreat rates reported here are of the same order of magnitude as rates reported for ice streams of the southern part of the Fennoscandian Ice Sheet implying no latitudinal differences in ice response and retreat rate for this ~1000 km sector of the Fennoscandian Ice Sheet (~60 – 68^{o}N) during the climate warming of this period.

Keywords: ice-rafted debris, paleo-ice stream, northern Norway, late Weichselian, deglaciation, recession rate

Introduction

There is significant interest in the reconstruction of the dynamics of the marine based part of the Fennoscandian Ice Sheet and its ice streams following the last (late Weichselian) glacial maximum (e.g. Vorren and Plassen, 2002; Nygård *et al*., 2004; Ottesen *et al*., 2005a, 2008; Eilertsen *et al*., 2005; Rydningen *et al*., 2013; Mangerud *et al*., 2013; Vorren *et al*., 2013, 2015; Stokes *et al*., 2014). However, few details are available on spatial and temporal fluctuations during the disintegration of the 45,000 km^2 of ice that covered the Norwegian continental shelf at its widest (up to 250 km), from 64 – 68^{o}N (e.g. Baumann *et al*., 1995; Dahlgren and Vorren, 2003; Mangerud, 2004; Mangerud *et al*., 2011). Such reconstructions are of importance for decoding the precise timing and time span over which the marine-based part of the NW Fennoscandian Ice Sheet disintegrated and thus the timing of the ice flux of this sector of the Fennoscandian Ice Sheet into the Norwegian – Greenland Sea, the establishment of the causal links to the forcing factors controlling the ice recession including halts/readvances, and from this, our ability to predict the response of modern ice sheets to future climate change (e.g. Conway *et al*., 1999; Alley *et al*., 2005).

The glacial impact on morphology and the deglaciation chronology of the onshore areas of the Lofoten – Vesterålen - Ofoten region of northern Norway has been examined extensively by earlier workers (e.g. Andersen, 1975; Andersen *et al*., 1981; Rasmussen, 1984; Olsen, 2002; Bargel, 2003; Knies *et al*., 2007; Vorren *et al*., 2013), but few datings and a general lack of Quaternary sediments have made it difficult to establish a detailed deglacial chronology in the area, including both the onshore and offshore areas (Vorren and Plassen, 2002; Bargel, 2003; Ottesen *et al*., 2005b; Vorren *et al*., 2015).

During the late Weichselian, fast flowing ice streams are believed to have existed in about 20 cross-shelf troughs along the western margin of the Fennoscandian – Barents Sea – Svalbard

Ice Sheet (Ottesen *et al*., 2005b), including three major ice streams in the sector overlying the shelf from 64 – 68^{o}N. Based on bathymetric and seismic data revealing large-scale submarine landforms, Ottesen *et al*. (2005a) showed that one of these ice streams (the northernmost) drained the north-western part of the Fennoscandian Ice Sheet, which was up to 2 km thick in the inland areas (Kleman *et al*., 1997). This ice stream was flowing into the Ofotfjorden – Vestfjorden - Trænadjupet troughs having a total length of ~400 km (Fig. 1). During its recession from the shelf break, two prominent push moraines were formed in outer Vestfjorden (Laberg *et al*., 2007). Here we use seismic, bathymetric, and core data from Vestfjorden, along with 33 radiocarbon dates to discuss: (1) the dynamics of the Vestfjorden - Trænadjupet paleo-ice stream from its maximum position at the shelf edge to its complete removal from the marine realm, (2) from a regional correlation of the established deglacial chronology and the paleoclimate of the area, discuss implications for the origin of the ice front oscillations, and 3) the ice flux of this sector of the Fennoscandian Ice Sheet into the Norwegian – Greenland Sea. Previous studies of the continental slope succession has suggested a highly dynamic paleo-ice stream in the Vestfjorden – Trænadjupet trough throughout the last glacial maximum (25.9 - 18.1 ka BP) through the studies of their IRD-record (Dahlgren and Vorren, 2003; Rørvik *et al*., 2010).

Physiographic and geological setting

The Vestfjorden trough (Figs. 1, 2) is located adjacent (south) of the Lofoten archipelago, and is an open, ~200 km long NE-SW trending embayment, joining the elongated, glacially scoured Trænadjupet trough (between 400 – 500 m deep) extending to the shelf break. Maximum depths and widths in Vestfjorden are 600 m and 15 km in the inner parts, shallowing and widening to 350-400 m and ~100 km in the outer part. It is the seaward continuation of Ofotfjorden, which has a maximum depth of more than 500 m, and is about 250 km long. Several tributary fjords enter Vestfjorden from southeast, the largest being Skjerstadfjorden, Folda, Sagfjorden, and Tysfjorden (Fig. 2).

Vestfjorden is influenced by two major ocean currents, the Norwegian Atlantic Current (NAC) and the Norwegian Coastal Current (NCC). The Lofoten archipelago and the large deep-silled Vestfjorden trough causes the NCC to bifurcate with one part continuing north to join the circulation of Vestfjorden and the other traveling westwards over the Trænadjupet Trough, then north along the outside of the Lofoten archipelago. The largest current velocities occur at the shelf break and uppermost slope (Sundby, 1983; Heathershaw *et al*., 1998).

The bedrock in the area consists of Mesozoic and Cenozoic sedimentary rocks in the outer parts (Rokoengen and Sættem, 1983; Løseth and Tveten, 1996), whereas crystalline rocks consisting of Precambrian granite and gneiss, and Caledonian schist, marble, and minor igneous rocks occur in the inner half of the Vestfjorden and onshore (Sigmond, 1992).

The sediments above acoustic basement (sedimentary rocks) in Vestfjorden can be divided into 3 main seismostratigraphic units (Laberg *et al*., 2009). The lowermost unit is characterized by pre-late Weichselian sediments, followed by late Weichselian glacial sediments (unit 2), overlain by latest Weichselian - Holocene glacimarine – marine sediments (unit 3). In the main part of Vestfjorden, the latest Weichselian-Holocene glacimarine – marine sediments are mainly found in the deepest part of the axial-parallel trough (Fig. 3). Little glaciomarine sediment was deposited elsewhere. Thus, ocean bottom currents have probably controlled their distribution, i.e., caused erosion and/or no deposition in the shallowest part of the basin and deposition in the deepest part (Laberg *et al*., 2009). The thick glacimarine succession in the innermost part of the fjord is inferred to mainly have been deposited during period(s) when the ice front was situated at or close to the mouth of Ofotfjorden and Tysfjorden (Laberg *et al*., 2009).

Glacial history

Recently, Vorren *et al*. (2015) presented a revised model for the timing and extent of the Fennoscandian Ice Sheet on the continental shelf offshore from Lofoten based on the identification of glacial landforms and their correlation to the glacial landforms and stratigraphy onshore at Andøya, north Norway. The Fennoscandian Ice Sheet reached the shelf break twice. These expansions to the shelf break are locally termed Egga I and II advances, in accordance with Andersen (1968, 1975), separated by the Andøya interstadial (Vorren *et al*., 2015). Egga II was estimated to have occurred prior to 22.2 ka BP offshore of Andøya (Vorren *et al*., 2015). Results from the continental margin to the south of Lofoten showed that the sedimentary environment was significantly influenced by fluctuations of the Fennoscandian Ice Sheet, repeatedly reaching the outermost shelf between ~25.9 – 18.1 ka BP (Dahlgren and Vorren, 2003; Rørvik *et al*., 2010).

The Fennoscandian Ice Sheet recession from the continental shelf south of Lofoten occurred from ca. 18.1 ka (Dahlgren and Vorren, 2003; Rørvik *et al*., 2010, see also Olsen *et al*., 2001) and was characterized by several halts and/or readvances once close to the coastal areas, including the events of the Older Dryas Stadial (ca. 14.3–14 ka), the Younger Dryas Stadial (ca. 12.8–11.7 ka), and the Preboreal Events (ca. 11.7–10.2 ka) (e.g., Andersen, 1968, 1975; Møller and Sollid, 1972; Andersen *et al*., 1981; Rasmussen, 1984; Lyså and Vorren, 1997; Olsen, 2002; Bergstrøm *et al*., 2005; Knies *et al*., 2007; Fløistad *et al*., 2009). The Røst and Værøy Morainal Banks in outer Vestfjorden (Figs. 2, 4) have been interpreted as push moraines formed by ice-readvance (Laberg *et al*., 2007, 2009). Vorren *et al*. (2015) correlated the Røst Moraine to the Skogvoll Event on Andøya/Andfjorden.

Materials and methods

In 2001, the University of Tromsø collected one gravity core in Vestfjorden (JM01-604), while in 2003, nine more cores were collected using a 6 m long steel pipe and inner PVC liners of 110 mm diameter (Fig. 1, Tab. 1). Two of the cores contain only postglacial sediments, while the rest also contain deglacial sediments. Various investigations have been carried out on the sediment cores, including shear strength by the fall cone test (Hansbo, 1957), grain-size analysis using wet sieving of the >0.063 mm fractions, and by Micromeritics SediGraph 5100 analysis of the < 0.063 mm fractions. Prior to opening, the physical properties of the cores were measured using a GEOTEK Multi-Sensor Core Logger (MSCL). This included bulk density, estimated from gamma ray attenuation, magnetic susceptibility using a loop sensor as well as fractional porosity, and p-wave velocity. Splitted cores were x-radiographed and the clast content was counted according to the method described by Grobe (1987). Lithological description of the cores was performed to document lithologies, bedding characteristics and bedding surfaces of the sediments. The foraminiferal fossil fauna was examined at 5 levels in core JM01-604.

33 samples for AMS (Accelerator Mass Spectrometry) radiocarbon dating were prepared by the Radiological Dating Laboratory in Trondheim (6 also presented and discussed by Laberg *et al*. (2007)), and the measurements carried out by the University of Uppsala (Tab. 2). The dates have been corrected for a marine reservoir effect of 440 years (Mangerud and Gulliksen, 1975), although the reservoir age has probably varied through time (Bondevik *et al*. 1999, 2001, 2006; Haflidason *et al*. 2000). The ^{14}C-dates were (re)calibrated using Calib Rev 7.0.4 based on Stuiver and Reimer (1993) and Stuiver *et al*. (1998), the Marine13 calibration curve (Reimer *et al*., 2013) and a local reservoir age (ΔR) of 67 ±37 years based on the nearest existing data (Mangerud and Gulliksen, 1975). The calibration was constrained

to a 1 σ range. We will use the calibrated dates only in the following (sensu Bartlein *et al*, 1995). Chronozone nomenclature follows that of Mangerud *et al*. (1974).

Age-depth models for the cores were constructed based on linear interpretation of the calibrated ages (Fig. 5). The model is considered robust as it includes 33 dated samples, most of them (26) from the deglacial succession. None of the dated samples represents inverted ages (Tab. 2). Linear interpolation between the dates was used to make a chronology through the cores, however the sedimentation rates were probably not linear, especially around the transition between glaciomarine and marine deposition, i.e. when the glacier front had redrawn from the marine realm. Therefore, the model should be considered the simplest fit from the available data.

Rates of ice recession and readvance for the Røst and Værøy readvances were estimated from the derived time – distance diagram based on the established age model (see below) assuming that: 1) the peak readvance corresponded to increased IRD input as recorded in cores recovered from outside the ice front, 2) the rates of ice recession and readvance were linear, and 3) the ice was present at its maximum position for a very short time period only. If the peak glacial periods were longer, this would imply higher rates of recession and readvance, therefore the rates presented below should be considered minimum estimates.

Lithostratigraphy, faunal characteristics, radiocarbon dates and correlation to the established seismic stratigraphy

Four units are distinguished in the cores based on sedimentological and faunal characteristics (A-D; Fig. 5). Below, each unit is described, the depositional environment interpreted and the chronology discussed based on the radiocarbon dates.

Unit A (basal till)

Unit A is found at the bottom of four of the cores (Figs. 5, 6), with a thickness ranging from ~70 to ~150 cm. It consists of a dark grey, unsorted, homogeneous and clast-rich diamicton, with maximum clast size up to 10 cm. No macrofossils were found, only a few foraminifera with worn surfaces. The boundary to the overlying unit is distinct (Fig. 5). Both magnetic susceptibility and bulk density values are relatively high, and make a good base for correlation between the cores. This unit is also recognized in the seismic data (seismic unit 2), with thicknesses up to 200 msec (two-way travel time (twt)), and is widely distributed throughout the study area (Laberg *et al.*, 2009).

Unit A is interpreted as a basal till based on the diamicton, clast content, and high bulk density values. It was probably deposited at the base of an ice stream draining the central part of the Fennoscandian Ice Sheet into Vestfjorden. The seismic data also suggest that the flutings and moraines originate in this unit (Laberg *et al.*, 2007; 2009), thus supporting the interpretation. The foraminifera were probably picked up and reworked during an ice advance (see below).

Unit B (glaciomarine deposits)

Unit B is found in 2 of the cores, directly overlying unit A (Figs. 5, 6). The boundary between unit A and B is distinct but with no indication of being an unconformity, and the upper boundary is transitional. The unit is between 8 and 45 cm thick, and consists of laminated, clayey silt to silty clay sediments at the base, which in core JM03-499 grades into massive clay to silty clay at the top. In core JM03-501 two distinct, 1-2 cm thick normally-graded sand beds are present within the laminated unit. Also, the massive layer grades into laminated sediments at the top. The unit is further characterized by low magnetic susceptibility values, and very little clasts (Fig. 5). No fossils were found in this unit.

Unit B is interpreted as deposited by suspension plumes in front of an ice margin with rapid deposition. The lack of IRD in such a setting might have several explanations, as discussed by Vorren and Plassen (2002), including low detritus content of the ice stream, lack of icebergs reaching the core site, rapid calving and evacuation of icebergs, low temperatures preventing melting of icebergs, and the presence of a floating ice tongue. Low detritus content of the ice stream seems unlikely as the diamicton below (unit A) and the unit above (unit C) contains abundant clasts and IRD. Icebergs are known to get stuck in 'sikussaks' (Syvitski *et al*., 1996), however 'sikussaks' (or "ice-melanges") are related to confined settings, like fjords or embayments. Such a bathymetric confinement was not present in Vestfjorden. Rapid calving and evacuation of the icebergs could prevent the icebergs from melting and subsequent release IRD. Also, the temperature was probably very low, and may have prohibited melting of the icebergs (Vorren and Plassen, 2002) along with seasonal sea ice cover. We favor rapid calving and retreat of the ice-stream front as the most plausible explanation (see discussion below). The two distinct sand layers are interpreted as deposited from turbidity currents.

Unit C (glaciomarine deposits)

Unit C is found in all cores except JM03-504, JM03-497 and JM03-528 (Figs. 5, 6). It is from 120 to > 535 cm thick, separated by underlying units A and B, and overlying unit D by transitional boundaries. Overall, the unit appears to thicken eastwards, as confirmed by the seismic data (seismic unit 3). It consists of moderately bioturbated massive clay to clayey silt. Occasional normal graded layers are present. Clast content varies from relatively low to very high (Fig. 5). Typically, magnetic susceptibility and density values are low and decreasing upwards, although slightly higher than that measured for the underlying unit B. 26 radiocarbon dates from paired shells, shell fragments and foraminifera span from 16.4 to 11.0

ka BP. The fossil foraminiferal fauna in this unit recorded in core JM01-604 is diverse with 8 species found, but dominated by *Elphidium excavatum* and *Cassidulina reniforme* (Tab. 3).

Unit C is interpreted as glaciomarine sediments based on the sediment structures, clast content, low magnetic susceptibility and density values. The radiocarbon dates and the cold-water fossil fauna (e.g. Vorren *et al*., 1984) are supporting the interpretation that unit C was deposited in a glaciomarine environment (see below).

Unit D (postglacial marine deposits)

Unit D is present at the top of all cores (Figs. 5, 6), and consists of 12 to 170 cm thick, moderate to heavily bioturbated, massive clay and silt. The unit appears to thicken towards east. It is generally finer grained than unit C, with lower magnetic susceptibility and density values. The boundary between unit C and D is marked by a clear decrease in clast content, although sporadic clasts do occur within the unit. Occasional normally-graded layers are present. 7 radiocarbon dates from paired shells, shell fragments and foraminifera span from 6.910 to 940 years BP. The fossil foraminiferal fauna within this unit found in core JM01-604 is dominated by *Trifarina angulosa* and *Cassidulina terretis*. An increase in *Cibicides lobatilus* and *Cassidulina terretis* is recorded towards the top of the unit (Tab 3).

Unit D is interpreted as postglacial, marine sediments, deposited as the ice-stream front had withdrawn from the marine realm. The radiocarbon dates and a relative warm-water fossil fauna, as indicated by the species *Hyalina balthica*, *Uvigerina mediterranea* and *Brizalina skagerakensis* (Husum and Hald, 2004a, b), support a Holocene age of the sediments. The increase in *Cibicides lobatilus* and *Cassidulina terretis* towards the core top suggests increased current activity (e.g. Hald and Steinsund, 1992) at the site of core JM01-604, and could thus explain the occurrence of clasts within the unit as being reworked from the underlying glaciomarine sediments.

Discussion

From the above results we can now do a detailed reconstruction of the recession of one of the major paleo-ice streams of the NW Fennoscandian Ice Sheet from its maximum position at the shelf edge to its complete removal from the marine realm including its timing, rate and origin. This we do by addressing: 1) the origin of the IRD-events in Vestfjorden, 2) the ice-front oscillations and the resulting ice retreat and re-advance rates of this part of the Fennoscandian Ice Sheet, 3) the correlation to regional events and the implications for the origin of the ice front oscillations, and 4) the ice flux of this sector of the Fennoscandian Ice Sheet into the Norwegian – Greenland Sea.

IRD-events in Vestfjorden

Two cores were collected distal of the Røst Morainal Bank (JM03-505 and JM01-604; Fig. 1). The former has a very condensed section of deglacial and Holocene sediments while two distinct IRD peaks occur early in core JM01-604, one between 16.5 and 16.3 ka BP and one between 15.8 and 15.0 ka BP (Fig. 7). The former was interpreted to correspond to the ice front advance terminating at the Røst Morainal Bank, and the latter corresponds to the advance and deposition of the Værøy Morainal Bank (see below) representing more precise age estimates of these events as compared to previous studies (Knies *et al*., 2007; Laberg *et al*., 2007). The remaining cores are positioned proximal to the moraines. Here, an overall increase in IRD between 14.6 and 14 (up to 15 ka BP in core JM03-524), can be detected. This increase is attributed to an ice front advance, probably the Skarpnes Event (Fig. 7) identified in coastal areas both north and south of the study area (Andersen *et al*., 1981; Lyså and Vorren, 1997). The remaining periods with increased IRD input are scattered during the Allerød and Younger Dryas chronozones (Fig. 7).

IRD-peaks corresponding to glacial advances were also found by Vorren *et al*. (2015) studying the offshore record. However, they also envisage a correspondence between the release of icebergs at glacial retreats and IRD peaks. This is not seen in Vestfjorden for the later stage of the ice recession, possibly because when the ice was located inside the tributary fjords during Younger Dryas, icebergs were mostly trapped inside the pronounced thresholds at the mouth of Ofotfjorden and Tysfjorden (e.g. Fløistad *et al*., 2009).

Ice-front oscillations in Vestfjorden

Assuming that the ice started to retreat from the shelf edge around 18.1 ka BP, and receded past the position of the Værøy Morainal Bank, the ice front retreated at a rate of about 195 m/yr (Table 4). Low input of glaciomarine sediments in the deeper part of outermost Vestfjorden (and most of Trænadjupet) has been interpreted to indicate that ice withdrawal was due to more rapid calving rather than melting because melting would have resulted in a higher glacimarine sediment flux (Laberg *et al*., 2009). The retreat rate is comparable with the values given by Vorren and Plassen (2002) for the final retreat from the shelf edge to the Flesen Event (310 m/yr) in Andfjorden (Fig. 1) during the same time period, although the retreat distance was four times longer in Vestfjorden. This also compares to the retreat rate reported from Jakobshavn Isbrae on Greenland of 280 m/yr between 1850 and 1960 (Knight, 1999). However, while Vorren and Plassen (2002) reported a further retreat rate of between 31 and 67 m/yr for the remaining recession periods during the deglaciation, the situation appears to have been somewhat different in Vestfjorden. Based on the inside termination of the Røst Morainal Bank beneath the more proximal Værøy Morainal Bank (Laberg *et al*., 2007), the ice front must have advanced close to 65 km in ~300 years before depositing the

Røst Morainal Bank around 16.4 ka BP due to the lack of buttressing from the grounded, marine-based part of the ice sheet. This gives an advance rate of 216 m/yr. Following the deposition of the Røst Morainal Bank, the ice front receded eastward at least to the position of core JM03-499 based on the presence of moraine ridges, at a rate of ~274 m/yr, before advancing and depositing the Værøy Morainal Bank. This gives an advance rate of ~51.5 km in ~200 years, or ~258 m/yr. The retreat rate from the Værøy Morainal Bank (pre-Skarpnes) was rapid as indicated by the oldest date in core JM03-524 (15115 years BP) (Tab. 2), suggesting a retreat rate of ~175 km in 620 years, or ~283 m/yr. At present, we do not know how far the ice front receded before the Skarpnes Event nor the exact location of the ice front during the culmination of the Skarpnes readvance. Most likely, it was deposited at roughly the same position as the Tromsø-Lyngen Moraine (Figs. 1, 2), as noted for nearby fjords as well (Lyså and Vorren, 1997), and this will be further discussed below. Also, the deep basin distal to the threshold may have prohibited an ice-front advance during this time.

The retreat rates reported here are of the same order of magnitude as rates reported by Mangerud *et al*. (2013) for ice streams of the southern part of the Fennoscandian Ice Sheet implying no latitudinal differences in retreat rate for this ~1000 km sector of the Fennoscandian Ice Sheet (~60 – 68^{o}N). From modern outlet glaciers larger rates by an order of magnitude have been found although over a much shorter period (e.g. Howat *et al*., 2007). Thus the results reported here support the findings of Stokes *et al*. (2014) that modern retreat rates are up to an order of magnitude higher than the rates associated with the recession of the large ice sheets of the last glacial.

The ice re-advance rate of up to 216 m/yr is for instance nearly twice the rate estimated for the advance of the Puget Lobe of the Cordilleran Ice Sheet where a rate of 135 m/yr was reported by Porter and Swanson (1998). Thus the NW Fennoscandian Ice Sheet responded rapidly to climatic events of the deglaciation as further discussed below.

Correlation to regional events – implications for the origin of the ice front oscillations

The Røst and Værøy readvances occurred within an initial phase of overall northern Hemisphere temperature increase (Shakun *et al*., 2012) but coinciding with a cold spell observed on the adjacent Norwegian mainland (Alm, 1993; Vorren and Alm, 1999), an ice advance at the Mid-Norwegian shelf (Nygård *et al*., 2004), and IRD peaks in the northern Barents Sea (Kleiber *et al*., 2000; Knies *et al*., 2001). Knies *et al*. (2007) suggested a connected response of the ice sheets and attributed the re-advance to meltwater pulses into the North Atlantic and Nordic Seas associated with Heinrich Event 1 (>15.3-17.7 ka BP; see also Sarnthein *et al*., 2001), causing disruptions to the thermohaline circulation in the North Atlantic (Broecker, 2003; Ganopolski and Rahmstorf, 2001).

Based on a detailed regional ice-front correlation, Vorren *et al*. (2015) found the Røst readvance to correlate to the Skogvoll event on Andøya/Andfjorden (the other area in North Norway where the early stages of the deglaciation chronology of the NW Fennoscandian Ice Sheet is known in detail) and thus to represent a regional halt and/or readvance of the NW sector of the Fennoscandian Ice Sheet due to a climatic cold spell. This implies that there are no events in Vestfjorden coinciding with the Flesen Event (at ~17. ka BP) of Andøya/ Andfjorden (Vorren *et al*., 2015). The Værøy readvance on the other hand seems to be an ice readvance restricted to the Vestfjorden area, no event of this age has been found in the Andfjorden area that experienced a temperature increase following the Skogvoll cold spell (Vorren *et al*., 2015) (Fig. 7). Seismic data reveal no obvious topographic control, neither on the Røst nor on Værøy Morainal Bank (Laberg *et al*., 2009). Thus, we favour a climatic origin of the regional Røst event while the Værøy readvance possibly was controlled by internal ice dynamics (“ice downdraw”). In terms of morphology the Røst and Værøy Morainal Banks are similar, both has been assigned to an ice readvance forming a push moraine. Thus, no relationship to their forcing factor can be deduced from their morphology.

An overall increase in IRD between 14.6 and 14 ka BP is attributed to the Skarpnes ice front readvance (Fig. 7). However, as discussed above no morainal bank have been identified in Vestfjorden that can be associated with this event. This implies that the ice front was located at the bedrock threshold at the Vestfjorden – Ofoten transition or within Ofotfjorden. The latter is considered less likely as the threshold (max water depth of ~ 330 m in a narrow bathymetric depression, see Fløistad *et al*. (2009), their Fig. 3) would block calving icebergs from entering Vestfjorden and thus significantly reducing the ice and IRD fluxes into Vestfjorden. It is considered more likely that the ice front was located at the Vestfjorden – Ofoten threshold allowing for input of iceberg directly into Vestfjorden. Icebergs may also have been derived from ice lobes re-entering the marine realm from the Lofoten Islands, however very few details are presently available on deglaciation of this area.

The favoured position of the ice front during the Skarpnes event coincides with the position of the ice front during the Younger Dryas as proposed by Fløistad *et al*. (2009), slightly revising the position proposed by Bergstrøm *et al*. (2005). Thus although these events were climatically induced and initiated by a cold spell, the ice front positions were topographically controlled.

Timing of ice discharge

Our reconstruction shows that Trænadjupet and most of Vestfjorden was deglaciated in the period from 18.1 – 16.6 ka BP. As this represented the northern part of the 45,000 km^2 of grounded ice that covered the Norwegian continental shelf at its widest (e.g. Baumann *et al*., 1995; Dahlgren and Vorren, 2003; Mangerud, 2004, Mangerud *et al*., 2011), we find that most of the grounded ice must have disintegrated during the period from 18.1 – 16.6 ka BP.

Assuming an average ice thickness of 600 m, this would then imply an average ice discharge of up to 18 km^3/year into the Norwegian – Greenland Sea. These yearly rates are comparable to the modern rates following the collapse of the Larsen B ice shelf, Antarctic Peninsula in 2002 (27 km^3/year) (Rignot *et al*., 2004) but occurring over a much longer period.

Conclusions

- The Vestfjorden paleo-ice stream of the NW sector of the Fennoscandian Ice Sheet had an initial ice recession from the shelf break to the coastal area at a rate of about 250 m/yr, followed by two ice readvances, at 16.6 - 16.4 ka BP (the Røst readvance) and at 15.8 – 15.6 ka BP (the Værøy readvance), the former at an estimated ice advance rate of 216 m/yr.
- The Røst readvance has been interpreted to be part of a climatically induced regional event while the Værøy readvance was restricted to the Vestfjorden area and possibly formed as a consequence of internal ice-sheet dynamics. Neither of these seems to be influenced by local topography.
- Younger increases in IRD content has been correlated to the Skarpnes (Bølling – Older Dryas) and Tromsø – Lyngen (Younger Dryas) Events when the ice front was located more up-fjord, in the Vestfjorden – Ofotfjorden transitional area where a prominent bedrock threshold is located. Thus although these events were likely climatically induced, the ice front position was topographically controlled.
- Overall, the Vestfjorden paleo-ice stream responded to the climatic fluctuations of this period (Røst, Skarpnes and Tromsø – Lyngen readvance) but ice response due to internal reorganization is also suggested (Værøy readvance). As demonstrated here, the latter may be identified from a regional approach involving the studies of several paleo-ice streams.

- The retreat rates reported here are of the same order of magnitude as rates reported for ice streams of the southern part of the Fennoscandian Ice Sheet implying no latitudinal differences in ice response and retreat rate for this ~1000 km sector of the Fennoscandian Ice Sheet (~60 – 68^{o}N) during the climate warming of this period.

Acknowledgements

This work was financed by the University of Tromsø and the Research Council of Norway as a part of the project *Sedimentary processes and Palaeoenvironment on Northern Continental Margins* (SPONCOM) led by the late prof. Tore O. Vorren. We offer our sincere thanks to science engineer S. Iversen for all his technical support during the cruises, the captain and crew of RV *Helmer Hanssen* and RV *Johan Ruud* for their assistance collecting the data for this study, K. Husum for valuable discussions and J.P. Holm and T. Grytå who did part of the drafting work.

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Table captions

Table 1. Geographical position, length and the water depth of the gravity core samples of this study.

Table 2. Radiocarbon dates from the gravity cores retrieved in the study area. The dates have been corrected for a reservoir age of 440 years (Mangerud and Gulliksen, 1975). The ^{14}C-dates were calibrated using Calib Rev 7.0.4 based on Stuiver and Reimer (1993) and Stuiver *et al*. (1998), the Marine13 calibration curve (Reimer *et al*., 2013) and a local reservoir age (ΔR) of 67 ±37 years based on the nearest existing data (Mangerud and Gulliksen, 1975). The calibration was constrained to a 1 σ range.

Table 3. Foraminiferal list summarizing the taxa identified in lithological units C (samples at 270.5, 171 and 111 cm core depth) and D (30 and 5 cm depth) of core JM01-604. The number of foraminifera pr. gram is also included. See Fig 5 for more core details.

Table 4. Distances, estimated durations and rates of recession and advances of the Fennoscandian Ice Sheet during the Late Weichselian and early Holocene chronozones in Trænadjupet – Vestfjorden - Ofotfjorden, northern Norway.

Figure captions

Figure 1. Location map showing the study area including the gravity core samples. The location of the Vestfjorden – Trænadjupet paleo-ice stream and the position of the ice front during the last glacial maximum (LGM) is also indicated. Contours every 100 m for the fjords and shelf (east of the LGM ice front position), every 200 m for the continental slope and the deep sea. R= Røst Morainal Bank, V = Værøy Morainal Bank, M = Malangsdjupet Trough. 1 = core JM03-505, 2 = core JM03-504. The bathymetry was adapted from Jakobsson *et al*. (2012).

Figure 2. Shaded relief map of Vestfjorden showing the location of the Røst, Værøy, and Skarpnes/Younger Dryas Moraines. The location of the studied cores (located in Vestfjorden) and Figs. 3 - 4 is also indicated. R = Røst Morainal Bank, V = Værøy Morainal Bank.

Figure 3. Part of 3.5 kHz profile across the axial-parallel of Vestfjorden forming the deepest part of the fjord basin in this area. The latest Weichselian-Holocene glacimarine – marine sediments are mainly found here. Little glaciomarine sediment was deposited elsewhere. The figure is slightly modified from Laberg *et al*. (2009). See Fig. 2 for location of the profile.

Figure 4. Axial-parallel geoseismic profile showing the sediment distribution along the fjord axis including the Røst and Værøy Morainal Banks (slightly modified from Laberg *et al*., 2009). For location of the profile, see Fig. 2.

Figure 5. Lithostratigraphy, physical properties, ice-rafted debris (clasts > 2 mm), and age – depth model of the studied cores. See Figs. 1, 2 for core location.

Figure 6. Overview of the main lithological facies identified and their interpretation.

Figure 7. Time-distance diagram with ages in calibrated years showing the ice-front position during the last glacial maximum and the deglaciation of the Trænadjupet – Vestfjorden - Ofotfjorden area. The timing of the last glacial maximum was adapted from Dahlgren and Vorren (2003). The results from the Anfjorden area based on Vorren and Plassen (2002) is also included. Chronozones are after Mangerud *et al*. (1974).