Sedimentary processes and paleoenvironments in St. Jonsfjorden, western Spitsbergen

Elizabeth Bunin
GEO-3900 Master’s thesis in geology
May 2015
Abstract

Multibeam (swath) bathymetry, high-resolution two-dimensional seismic (chirp) profiles and four gravity cores have been analyzed to reconstruct the Late Weichselian and Holocene sedimentary environments and glacier dynamics in St Jonsfjorden, central western Spitsbergen (Svalbard). The sediment cores comprise stratified glacimarine sediments with varying amounts of ice rafted debris. Three cores terminate in dense diamict units; in the outer fjord (unit 956-1) this is interpreted as being of an ice rafted origin while in the inner fjord (units 953-1 and 954-1) this is interpreted as ice-contact till.

Fast-flowing, grounded ice draining the Svalbard-Barents Sea Ice Sheet during the Last Glacial is suggested to have formed flutes, drumlins and glacial lineations that are, in part, controlled by a complex underlying bedrock basement. Clusters of regularly spaced recessional moraines, interpreted to have been deposited annually, on the north shore of the outer fjord indicate that ice retreated stepwise during the last deglaciation. Thick packages of seismically stratified sediments in the outer fjord likely bury any indication of grounded ice in this area. However, the presence of glacial lineations in Forlandsundet is interpreted as evidence that ice flowed out of St Jonsfjorden and south through Forlandsundet during the Weichselian.

The mouth of the fjord was deglaciated by 12,625 cal BP; the ice margin had retreated inward to a bend in the fjord’s orientation (11.2 km from the fjord mouth) by 8,960 ± 70 cal BC and to 15 km from the fjord mouth – the location of a broad, shallow ridge – prior to 7,840 ± 90 (Forman 1989, re-calibrated herein), exposing open water in the inner fjord and providing a suitable habitat for marine microfauna. During this ice retreat, sediments may have been deposited on the floor of the inner fjord as five transverse zones of elevated topography, which were subsequently reworked during later Holocene advances of the Osbornebreen-Konowbreen glacier complex that occupied the fjord head.

During the Little Ice Age, and possibly also before, ice flowed to but not past a broad shallow ridge expected to be at least partially made of bedrock. Reworked foraminiferal tests from a gravity core (HH12-954) taken east of this ridge’s crest provide an AMS date of 2030 ± 113 cal BP, indicating that glacial advance reached the ridge crest after this point and is herein interpreted as a neoglacial advance which lasted until the Little Ice Age. Annual recessional moraines in the inner fjord indicate that the post-Little Ice Age deglaciation of St Jonsfjorden may have begun between 1840 and 1860, and that it occurred at a relatively constant rate since (average: 50 m/year); this rate is only slightly higher than the rate at which Osbornebreen has retreated on the period 1909-1986, known from historical photographs (average: 46 m/year). Lead decay-derived sedimentation rate estimates from a gravity core in the inner fjord, 2.2 km east of the ridge crest, indicate that the stratified glacimarine sediments of unit 953-3 are likely to represent sedimentation after the 1850s, supporting this estimate.
Forward

I am enormously grateful to both my supervisors, Matthias Forwick and Jan Sverre Laberg, for their suggestions and encouragement over the past fourteen months. Your patience is legendary. Thank you for sharing so much of your knowledge and experience with me; this thesis has benefitted tremendously from our discussions and your wisdom and suggestions.

Generous financial support from Norske Oljeselskap ASA paid for laboratory analyses and enabled me to participate in the 2015 NGF Vinterkonferansen, which was an incredible an valuable educational experience for me

Lukas Wacker, from the Laboratory of Ion Beam Physics at the ETH Zürich (Switzerland) performed AMS radiocarbon analyses presented in this thesis. Thank you!

Witold Szczuciński, from the Adam Mickiewicz University in Poznan, Poland determined the sediment accumulation rates presented herein through complementery $^{210}$Pb and $^{137}$Cs analyses. I’d also like to thank you and the rest of the AMU group for 17 fantastic hours coring in Hornsund (+ transit time) and wish you the best of luck videre!

Trine Dahl and Ingvild Hald helped me enormously in the lab, especially in my months spent on the particle size analyzer. Discussions with them, Juho Junttila and Simon Pind Jessen have helped me to understand and respect its capabilities and limitations; our conversations were always enjoyable and I appreciate all the time you have spent helping me.

Discussions with Patrycja Jernas helped me to choose foraminifera species for radiocarbon dating and Erna Ósk Arnardóttir and Björg Jónsdóttir helped me to identify specimens in the lab. Erna also helped in calibrating the radiocarbon dates herein. Thank you for all of your help.

The past two years have been lovely and I am so grateful to have been able to live here in Tromsø, the most beautiful and friendliest city in the world, and to have had the opportunity to work in this wonderful department, with the kindest and most helpful staff and students imaginable. I wish to thank all my friends in both brakker who have made my stay in Tromsø and at UiT so pleasant, especially Ingrid and all of the fjord-focused masterstudenter who have come before me, whose work inspires me, and my office neighbors Iselin, Elise and Solveig who are here with me at the very end!

Finally, my family: I love you very much and your support has meant everything to me. And Trym. I would never have managed this without you.

mvh, ejb.
OBJECTIVES

INTRODUCTION

HISTORY OF THE SVALBARD-BARENTS SEA ICE SHEET

HOLOCENE CLIMATE AND GLACIATION ON SVALBARD

BACKGROUND INFORMATION ABOUT FJORDS

STUDY AREA

PHYSIOGRAPHIC SETTING: ST JONSFJORDEN

EXPLORATION HISTORY AND PREVIOUS INVESTIGATIONS OF ST JONSFJORDEN

BATHYMETRY

OCEANOGRAPHY

GLACIOLOGY

BEDROCK GEOLOGY

MATERIALS AND METHODS

DATA ACQUISITION / MATERIALS

Geophysical (Acoustic) Data

Sediment Cores

DATA ANALYSIS / METHODS
ACOUSTIC RESULTS

BATHYMETRY

SEISMIC PROFILES

INTERPRETATION

LITHOLOGIC RESULTS

Gravity Core HH12-953
Gravity Core HH12-954
Gravity Core HH12-955
Gravity Core HH12-956

DISCUSSION

SOURCES OF ERROR AND UNCERTAINTY

SEISMOSTRATIGRAPHY

CORRELATION OF SEDIMENT CORES AND ACOUSTIC DATA

SEDIMENT SOURCES AND DEPOSITIONAL PROCESSES

Suspension Fallout
Ice Rafting
Subglacial Bedforms

GLACIAL HISTORY OF ST JONSFJORDEN

Deglaciation
The Early Holocene
Late Holocene (4 ka BP to present)

CONCLUSIONS
Objectives

This master’s thesis is an examination of sedimentary deposits (sediments, sedimentary structures and landscape elements) in St Jonsfjorden (Svalbard).

Reconstructing St Jonsfjorden’s history of glacial activity since the last glacial was the primary objective of this thesis, specifically the extent and dynamics of the Svalbard-Barents Sea Ice Sheet in St Jonsfjorden during and since the last glacial and the behavior of the Osbornebreen-Konowbreen glacier complex during and since the Little Ice Age. This was done by analyzing high-resolution two-dimensional seismic profiles, multibeam bathymetric data and the multi-proxy study of four sediment gravity cores to establish the following:

• The nature and distribution of the dominant depositional and postdepositional processes active in the fjord
• The timing of changes in the fjord paleoenvironment and sedimentation
• Rates of sediment accretion (net positive accumulation) in the fjord

Ultimately, this should contribute to a better understanding of the dynamics and breakup of the Svalbard-Barents Sea Ice Sheet on central western Spitsbergen and Holocene glacier dynamics of the study area.
Introduction

Fjords are glacially carved valleys occupied by estuaries (Syvitski et al 1987). As low-lying basins located between land ice and the oceans, fjords are in a unique position to accumulate sediments produced through glacial, glaciofluvial and glacimarine processes in addition to the tidal, wave, and mass wasting processes that affect sedimentation in estuarine basins unaffected by glacial and paraglacial processes. Sediments trapped in fjords can therefore be valuable records of changes in glacier activity, ocean circulation and terrestrial climate; they can further be used to understand complicated interactions between ice, the oceans and the atmosphere.

The Svalbard archipelago’s position (see Figure 1) at the known western margin of the former Svalbard-Barents Sea Ice Sheet (hereafter, SBSIS; e. g., Landvik et al 1998) and at the intersection of the North Atlantic, the Barents Sea, and the Arctic Ocean makes sedimentary records from this region valuable for paleoceanographic and paleoclimatic studies.

Figure 1: Location of Svalbard (red box) and SBSIS ice margin during the Last Glacial Maximum (white line). This figure is modified after Svendsen et al 2004.
History of the Svalbard-Barents Sea Ice Sheet

Much attention has been and continues to be paid to the glacial history of Svalbard from the late Weichselian to the present. Investigations of Svalbard’s Late Quaternary climatic history largely center on understanding the development and decay of the Svalbard-Barents Sea Ice Sheet, especially with respect to timing and extent (e.g., Gjermundsen et al 2013; Hald et al 2004; Hogan et al 2010; Jessen et al 2010). Our understanding of the SBSIS is continuously improved upon via the undertaking of targeted paleoclimatic, paleoceanographic, and paleoglaciologic studies and the reinterpretation of existing datasets in the light of new findings (e.g., Forman 1999; Mangerud et al 1998).

Although ice has most likely reached and retreated from the shelf edge west of Spitsbergen repeatedly over the past 1.6-1.3 million years (Sejrup et al 2005), the preservation potential of these older glacial deposits decreases with each successive glacial episode. The present-day understanding of the SBIS is that it was a dynamic ice sheet that formed and disintegrated at least four times over the course of the late Quaternary (last 150,000 years; Landvik et at 1998; Mangerud et al 1998;).

A review paper by Ingólfsson and Landvik (2013) details the development of our current understanding of the Svalbard-Barents Sea Ice Sheet from the Saalian to the present. These four most recent recognized Quaternary glaciations correspond to marine isotope stages 6 (Saalian), 5d (Early Weichselian), 4 (Middle Weichselian), and 2 (Late Weichselian; approximate extent given in Figure 1, white line) and are separated by interglacials or interstadials of milder climate (Mangerud et al 1998).

Some of the oldest terrestrial unconsolidated sedimentary records of early glaciation found on Svalbard are believed to belong to an extensive Saalian glaciation which likely destroyed and/or buried most of the previously deposited glacial sediments (Knies et al 2009; Olsen et al 2013). Sections of raised sediments representing two distinct Saalian glaciations are present at Poolepynten on Prins Karls Forland (Alexanderson et al 2011a and b; see Figure 2 for locations of places mentioned in this section).

Sediments from the Eemian interglacial – littoral and shallow marine deposits including dropstones such as those described at Kapp Ekholm (Mangerud et al 1998) and those from
Broggerhalvøya and Poolepynten (depicted in the composite stratigraphy published in Houmark-Nielsen & Funder 1999) – separate tills of Saalian age from overlying, younger tills. Elsewhere on Svalbard these younger tills can be of early and middle Weichselian age and represent episodes of local glaciation, such as those located in Kapp Ekkholm (Billefjorden; Mangerud & Svendsen 1992; Mangerud et al 1998) and Kongsøya (Ingólfsson et al 1995, reinterpreted in Mangerud et al 1998). Where divisible, early and middle Weichselian tills are separated by sediments deposited during the Phantomodden interstadial. However, cores from the southwestern Barents Sea show only Late Weichselian tills to be located stratigraphically above Eemain interglacial sediments, indicating that these glaciation events were likely local readvances of existing ice that did not reach the continental shelf (Mangerud et al 1998).

The Most Recent Glacial

The most modern iteration of the SBIS is believed to have begun to grow approximately 32,000 calibrated years before present (hereafter, cal BP), during the Middle Weichselian. This determination is based on the results of numerical models (Siegert and Dowdeswell 2002), the analysis of sediment cores from the western Svalbard continental slopes (IRD diamict facies; Andersen et al 1996), and the results of sediment core (magnetic susceptibility, lithostratigraphy) studies by Jessen et al (2010). The western margin of the SBIS likely reached the continental shelf at approximately 27,000 calibrated years before present (Jessen et al 2010 and references therein), shortly before advancing to its maximum position (see Figure 1) at the shelf edge at approximately 24,000 calibrated years before present, as evidenced by the presence of glacigenic debris flow deposits all along the continental slope (Jessen et al 2010 and references therein).

The Last Glacial Maximum lasted between three and five thousand years, during which time a large ice sheet existed over most of Svalbard and large parts of the Barents Sea. The records of postglacial emergence (sea level curves) and exposure (\(^{10}\)Be dates) suggest that this large ice sheet may have comprised more than one ice dome. It is believed that multiple ice streams drained the SBSIS through many of the island’s larger fjord systems (Landvik et al 2005; Ottesen et al 2005 & 2007). Areas between the ice streams probably housed slower-moving, dynamically less active ice which may have been thin and cold.
Introduction

based (Landvik et al 2005; Ottesen & Dowdeswell 2009). Maximum glacial loading is inferred – by Landvik et al (1998) from the postglacial emergence pattern – to have occurred over Kongsøya (eastern Svalbard). Ice streams fed by a local ice dome are suggested by Gjermundsen et al (2013) to have radially drained northwestern Spitsbergen, based on exposure dates, and additional ice domes may have existed east of Spitsbergen. The deglaciation of the Late Weichselian SBSIS is understood to have been stepwise and to have been interrupted by multiple re-advances (e.g., Baeten et al 2010; Elverhøi et al 1995; Forwick & Vorren 2010; Kempf et al 2013; Ottesen et al 2007; Polyak et al 1995). Retreat from the shelf edge is inferred from the presence of hemipelagic sediments above glacigenic debris flows on the continental slope and from elevated concentrations of IRD, presumably produced as calving increased. Dates from hemipelagic sediments overlying mass transport deposits in five western Svalbard slope cores were obtained by Jessen et al (2010) and indicate that ice must have begun to retreat from the shelf break by 20,000 calibrated years before present. Rasmussen et al (2007) also report an age of at least 20,000 calibrated years. The deglaciation of the Late Weichselian SBSIS is understood to have been stepwise and to have been interrupted by multiple re-advances (e.g., Baeten et al 2010; Elverhøi et al 1995; Forwick & Vorren 2010; Kempf et al 2013; Ottesen et al 2007; Polyak et al 1995). Retreat from the shelf edge is inferred from the presence of hemipelagic sediments above glacigenic debris flows on the continental slope and from elevated concentrations of IRD, presumably produced as calving increased. Dates from hemipelagic sediments overlying mass transport deposits in five western Svalbard slope cores were obtained by Jessen et al (2010) and indicate that ice must have begun to retreat from the shelf break by 20,000 calibrated years before present. Rasmussen et al (2007) also report an age of at least 20,000 calibrated years before present for the onset of deglaciation based on hemipelagic sediments in a core from Storfjordrenna (south of Spitsbergen, east of the shelf break. For location see Figure 3). The speed of retreat was likely not uniform, with the ice sheet retreating more quickly through the cross-shelf troughs than it did across the banks (Ottesen et al 2007).

Notable re-advances of the SBIS occurred during Heinrich Event 1, at 16,000 calibrated years before present (Knies et al 2007), and presumably during the Younger Dryas (11-12
Introduction

thousand radiocarbon years BP as per the chronology proposed in Mangerud et al 1974; Forwick & Vorren 2011). Sediments dated by Mangerud et al (1991) imply that ice had receded to the mouths of Isfjorden and Van Mijenfjorden by 11.6 ka BP (locations: Figure 2). This retreat continued well into the Holocene, and certain areas of western Spitsbergen may have become entirely ice free (e.g., Linnédalen, Svendsen & Mangerud 1997).

Figure 2 (opposite): Locations of fjords, sounds and bays of central western and northern Spitsbergen as well as terrestrial locations referred to in the History of the Svalbard-Barents Sea Ice Sheet section. Background image comes from TopoSvalbard (npolar.no)
Introduction

**Holocene Climate and Glaciation on Svalbard**

The Holocene climate on Svalbard can be described as an ameliorated climate punctuated by a series of cold snaps. These climatic fluctuations are often resolvable, at least locally, with very high temporal resolution due to the high preservation potential of the most recent sediments relative to older deposits, as discussed in Landvik et al (2014). Unfortunately, chronostratigraphic subdivision of the Holocene is complicated by the spatially transgressive nature of glacier and sedimentary responses to climate evolution, meaning that attempts to develop a fixed timescale for the Holocene are valid only on local-to-regional scales (Björk et al 1998).

The early Holocene experienced a climate warmer than the present (Holocene Climate Optimum), which likely reached its peak between 11,200 and 9,000 cal BP (Forwick & Vorren 2009). Despite this warm climate, Spitsbergen retained at least some ice; the presence of tidewater glaciers in Van Mijenfjorden and Isfjorden is confirmed by the continuous deposition of glacimarine muds with dropstones throughout the Holocene in both fjords (Baeten et al 2010; Forwick & Vorren 2009; Hald et al 2004; Kempf et al 2013).

After the end of the Holocene Climate Optimum, glaciers began to grow asynchronously (Baeten et al 2010; Forwick & Vorren 2009; Hald et al 2004; Hansen 2014; Velle 2012). By 5,000 cal BP the climate on Svalbard is believed to have cooled significantly, such that the middle Holocene climate was comparable to the present climate (Svendsen & Mangerud 1997).

A brief cooling event at 8.2 cal ka BP lasting 150 ±30 years (Daley et al 2009) is recognized in the isotopic signals of Greenland ice cores (Alley & Ágústdóttir 2005; Alley et al 1997) and numerous other proxy records throughout the northern hemisphere (see Walker et al 2012 and references therein). One hypothesis is that cooling event was caused by an increased flux of freshwater into the Atlantic, which slowed meridonal overturning circulation; this freshwater is believed to have come from the drainage of proglacial Lake Agassiz into the Hudson Bay (See Morrill et al 2013 and references therein). On Svalbard, the 8.2 event is recognized in cores from Van Mijenfjorden as an interval of depressed δ¹⁸O
in benthic foraminifera together with an increase in IRD and decrease in biologic productivity (Hald & Korsun 2008).

The timing of the Little Ice Age is accepted as being highly spatially variable and likely made up of more than one ice advance between 1300 and 1880 AD before ending with an abrupt rise in temperature at the beginning of the 20th century (Svendsen & Mangerud 1997; Werner 1993). Spitsbergen glaciers are believed to have reached their maximum Holocene extent during the Little Ice Age (D’Andrea et al 2012; Mangerud & Landvik 2007; Plassen et al 2004; Svendsen & Mangerud 1997). Ice-marginal features from the Little Ice Age and retreat thereafter are the most recent and best preserved but represent only the most recent section of a larger cool period on Svalbard, termed the Holocene Neoglacial (Werner 1993).

Temperatures have likely been increasing on Svalbard since the end of the Little Ice Age. In the past century annual average temperatures on Svalbard have increased approximately 2.6°C (Nordli et al 2014); widespread thinning and retreat of glaciers on central Spitsbergen has been documented over this same time period (e.g., Lønne & Lyså 2005; Rachelwicz et al 2007). Presently, Svalbard is 60% glaciated and the average temperature was -3.6°C in Longyearbyen in 2014 (yr.no). Also in Longyearbyen, the Norwegian Meteorological Institute reports a 2.8% increase in precipitation each decade over the 20th century (Hanssen-Bauer 2002).

Western Svalbard’s warm climate is linked to the transport of warm and saline Atlantic Water (AW) northward by the West Spitsbergen Current (WSC), the northernmost branch of the Norwegian Atlantic Current (see Figure 3; Aagaard et al 1987; Swift & Aagaard 1981). Traditionally, increases in the transport of Atlantic water northward and into the fjords have been associated with warmer terrestrial and marine climate on Svalbard (e.g., Hald et al 2004; Rasmussen et al 2012; Rasmussen et al 2014 and references therein). In addition to this warm, northward moving water, Arctic Water from the Barents Sea east of Svalbard is transported around the southern tip of Spitsbergen and travels north both atop and west of the WSC as a branch of the East Spitsbergen Current (Rasmussen et al 2014).
As they pass north of Svalbard, waters of the West Spitsbergen Current sink and spread and a portion of the current is recirculated southwardly along the East Greenland coast (Aagaard et al 1987).

Figure 3: Currents in the area surrounding the Svalbard archipelago. This figure is modified from Rasmussen et al 2014.
Introduction

Background Information about Fjords

Morphology
Classically U-shaped and elongate, fjords are the glacially carved, submerged, often overdeepened channels exposed as ice melts and retreats into the fjord valley, (Syvitski et al 1987). This ice mass loss and retreat is often associated with isostatic rebound and sea level changes, causing shorelines and beach sediments to form terraces along the fjord walls (Syvitski et al 1987). Fjords, and associated fjord valleys, often form dendritic networks that may follow the paths of older river plains that have repeatedly been uplifted and incised (Syvitski et al 1987).

Spitsbergen fjords are commonly divided into one or many basins by sills, a non-genetic term used to describe transverse ridges and shoals which may be formed of resistant bedrock or morainal bank sediments. If they are continuous across the width of the fjord, they can isolate deeper water, inhibiting fjord-shelf exchange and causing the accumulation of dense bottom water.

Oceanography
Circulation in fjords is largely controlled by the density differences in ambient and input water masses, the seasonal modification of water density via extreme cooling and/or brine formation and tidally driven mixing (Cottier et al 2010 in Howe et al 2010).

In fjords without sills, or with less prominent sills, both the WSC-derived Atlantic Water and ESC-derived Arctic Water have been observed to infiltrate Spitsbergen fjords, where density-driven fjord-shelf exchange leads to a seasonally variable distribution of water masses in the fjords (Cottier et al 2005; Nilsen et al 2008). Rapid fjord-shelf exchange, where warm waters are continuously transported to tidewater glacier termini, has been suggested as a possible mechanism for recent mass loss and acceleration at Helheim glacier, East Greenland (Straneo et al 2009). In fjords with sills, shelf-fjord exchange can be limited, causing the densest water to become trapped in the fjord basin and residence times of isolated basin water are variable depending on the ability of the overlying layers to mix and the rate at which deeper water can be formed. Notably, beneath polynyas, a complete vertical mixing of the water column can occur over a single winter via brine formation (Cottier et al 2010 in Howe et al 2010).
Sedimentation

The majority of sediments entering fjords with active, fast-moving glaciers are expected to be delivered either as ice rafted debris or by way of meltwater streams, where sediment-laden glacial meltwater enters the fjord either directly at the tidewater cliff or after first flowing over land as a proglacial river or stream. At the margins of the tidewater glaciers ice-contact deposition is also expected. Failures along the fjord walls (e.g., avalanches, rockfalls) can also contribute sediments to the fjord basin, as can aeolian and biogenic sedimentation (Hambrey 1994)

Once delivered to the fjord, currents, tides, wind and the Coriolis force all contribute to the distribution of sediments and the energy of the water column controls the amount of time needed for sediments of varying sizes to fall settle through the water column (Nemec 1995; Syvitski et al 1987; Syvitski 1989). After deposition, these sediments can be further redistributed and redeposited via mass wasting, current sorting and re-entrainment by glacial ice. Figure 55 outlines the processes by which sediments enter and are redistributed in fjords.
Study Area: St Jonsfjorden

Physiographic Setting: St Jonsfjorden

St. Jonsfjorden is located on the west coast of Spitsbergen and opens westward into Forlandsundet, a sound which separates the islands of Spitsbergen and Prins Karls Forland (see Figure 2). The Fram Strait lies beyond Prins Karls Forland. The effect of the warm West Spitsbergen Current on the westward-opening Spitsbergen fjords is well documented (e.g., Cottier et al 2007; Cottier et al 2010; Saloranta & Svendsen 2001). The degree to which Prins Karls Forland restricts hydrographic communication between the fjord and the Norwegian Sea remains, however, unknown. St Jonsfjorden is located between two known Late Weichselian paleoicestreams: Kongsfjorden to the north and Isfjorden to the south (see Figure 4 e.g., Landvik et al 2005 and Ottesen et al 2007).

The fjord has a distinct elbow shape, where the fjord axis in the outermost 12 kilometers is oriented northwest-southeast while the innermost 10 kilometers is oriented northeast-southwest. This transition is marked by the protuberance of Gjertsenodden into the fjord from the north shore, opposite the bay in front of Holmesletbreane on the south shore, as shown in Figure 5. Of the inner ten kilometers, the five nearest to the fjord head are oriented northnortheast-southsouthwest.
Figure 5: Glaciers and mountains in the St Jonsfjorden catchment. Inset shows the location of the study area on central western Spitsbergen. Catchment is outlined in black. Two unnamed glaciers are shown, without numbers, one between Charlesbreen and Vegardbreen and one between Ankerbreen and Gaffelbreen. Osbornebreen is shown divided into two polygons based on the position of an ice convergence zone and medial moraine – it is the western (left) portion of this glacier that was involved in the 1986-1989 surge (Dowdeswell et al 1991). This figure was created in ArcGIS (ArcMap10.1) using data from npolar.no (König et al 2013; labels were added in CorelDraw)
Exploration History and Previous Investigations of St Jonsfjorden

Until the 20th century, St. Jonsfjorden was known as Osborne Bay or Osborne Inlet, after an early English whaler who was active along the western Spitsbergen coast (http://stadnamn.npolar.no). Whalers and sealers were present in this area intermittently since at least the early seventeenth century and the earliest accounts of oceanographic conditions and climate in this area come from their observations (Hoel 1929).

The first scientific expedition to visit St. Jonsfjorden was the Isachsen Spitsbergen Expedition of 1909-1910 – a broad survey of Spitsbergen topography, geology, and hydrography carried out over the course of two summers. In St. Jonsfjorden, the expedition participants mapped bedrock on the southern shore of the fjord in 1909 and returned in the spring of 1910, at which time they traversed the ice cap Løvenskioldfonna on ski while mapping the topography of the mountains north of the fjord (Hoel et al 1915-1917).

Observations of raised beach positions and their faunal assemblages compose some of the earliest studies of environmental change on Svalbard. Peach (1916) was the first to describe raised beaches in the Forlandsundet area (at 15 and 21 meters above sea level on Prins Karls Forland). In 1950, R. W. Feyling-Hanssen and F. A. Jørstad suggested correlating west Spitsbergen raised beaches by their included faunas (Feyling-Hanssen & Jørstad 1950 in Dinely et al 1953b). St. Jonsfjorden has been the subject of several studies concerning the nature and positions of raised shorelines beginning in 1951 with the second Birmingham University Spitsbergen Expedition and the work of D. L. Dinely, who divided the terraces into three biostratigraphic zones, an upper zone 21-30 masl, a middle zone 10-21 masl and a lower zone up to 12 masl.

Dinely’s biostratigraphic zones are likely to represent only the most recent paleoshorelines. Forman (1989) investigated two levels of raised beaches (60 and 35-45 masl) in river valleys and low-lying areas as well as raised subglacial deposits. They suggest St Jonsfjorden was occupied by a Late Weichselian glacier at least 55 m thick (lateral moraine crest at 55 masl) and terminating 5 km from the fjord mouth based on the presence of lodgement till at 6 masl). Distal to this, a 10-meter rise in the marine limit supports this interpretation. Paired *Mya truncata* shells found at 13.5 masl in the same
section – where the marine limit begins to rise – are radiocarbon dated to 9,110 ± 90 cal BP. Furthermore, they note that raised beaches are infrequent and discontinuous in the fjord and that Pre-Late Weichselian beach deposits are exposed only discontinuously along the southern side of the fjord (as did Dinely 1953b), though are more continuous at the fjord mouth. Forman et al (1989) interpreted this as evidence of ice advancing through the fjord before reaching a late Weichselian glacial maximum.

In 1996, the Glasgow University Spitsbergen Expedition (see Evans and Rice 2005) noted and surveyed in more detail the same two sets of shorelines described in Forman et al 1989 and described the fjord’s Holocene landforms (moraine ridges, kames). New radiocarbon dates from *Mya truncata* shells in terrace sediments 3.5-20 masl were used to reconstruct the shoreline from 10 ka to 9.1 ka BP and suggest rapid uplift leading up to the 8.2 event (Evans & Rea 2005).

![Figure 6: Locations of radiocarbon dates and geomorphologic data collected by Evans & Rea (2005) and including the results of Forman (1SJ and 2SJ; 1989). Marine limits by Forman are circled; the remainders come from Evans & Rea. This figure was originally published in Evans & Rea 2005.](image)

A number of studies carried out in the 1970s and 1980s describe undisturbed pre-late Weichselian raised beach terraces occurring higher than the postglacial marine limit on both shores of Forlandsundet (e.g., Andersson 2000; Forman & Miller 1984; Forman 1989;
Miller 1982). These have often been interpreted as not having been overridden during the late Weichselian. However, Landvik et al (1998) suggest that preservation of these landforms is not enough to prove non-glaciation on the basis that landforms overrun by glaciers on other parts of Spitsbergen have been preserved, citing examples from Kongsfjordsletta (Lehman & Forman 1992) and Linnédalen (Mangerud et al 1987, 1992 and 1998). Evans & Rea (2005) also mention cold-based ice as a possible explanation for the preservation of older shorelines, in addition to the possibility that the area was ice-free. In St Jonsfjorden the Osbornebreen surge of 1986-1989 was observed by Arnt Iver Kverndal to slide over an older, ice-cored moraine without any glaciotectonic disturbance (Kverndal 1991).

Further geological mapping was done in the 1970s and 1980s by Les Kanat, and Alan Morris (Kanat & Morris 1988; Morris 1988; Kanat 1984). Studies on the geochemistry and deformation history of units south of St Jonsfjorden have been undertaken by Yoshihide Ohta (1985) and J. J. Auge and Alan Morris (1985). The stratigraphy of central Oscar II Land was the subject of Les Kanat’s dissertation and aims to reconcile observations made in St Jonsfjorden with earlier studies done on the stratigraphy of western Spitsbergen (for proposed stratigraphy, see Kanat and Morris 1988).

Until recently, very limited information on the bathymetry of St Jonsfjorden has been available; navigational maps of this area show two shoals at the mouth of St Jonsfjorden and shallow zone in the inner fjord has been known since at least the 1980s (referenced in Forman 1989). Single-beam echosounder data was collected by the Norwegian Hydrographic Service between 1965 and 1994. In St Jonsfjorden this dataset has been interpreted twice; Król et al (2010) published this data and is reprinted below as Figure 7. From the singlebeam survey data, Ottesen et al (2005) interpreted the relief of the fjord as megascale glacial lineations continuous with additional megascale glaciations in the south of Forlandsundet. In their reconstruction of the Svalbard Barents Sea Ice Sheet and its ice streams, Ottesen et al (2005) depict St Jonsfjorden as a tributary to the Isfjorden ice stream. Król et al, however, places the Late Weichselian ice limit some kilometers from the fjord mouth and mention no indicators of fast-flowing ice, in agreement with Forman et al 1989.
Bathymetry

The bathymetry of St Jonsfjorden is broadly known from the NHS dataset published in Król et al (2010), reproduced as Figure 7. The outer fjord is divided into independent, smooth-floored basins by a series of low ridges while the seafloor in the inner fjord is irregular and shallow. A 3-km wide shallow zone crosscuts the fjord east of Gaffelbreen and Løvliebreen, between Konowfjellet and Piriepynten (see Figure 5 for locations). Król et al (2010) identified this as a broad, smooth-sided ridge and linked it to the terrestrial morainal sediments and erratics described on the ridge’s landward extensions in the dissertation of Zdzisław Preisner (Preisner 1988 in Król et al 2010). They interpret this ridge as a pre-existing feature that likely prohibited ice from advancing past this point during the Little Ice Age. Beyond this ridge, Król et al have interpreted glacial lineations and flutes with initiating boulders oriented along the axis of the fjord.

In the outer fjord the two shoals, Farmgrunnen and St. Jonsgrunnen, are known from navigational charts from this area. In the NHD bathymetric survey, this area appears to be one continuous ridge oriented roughly 55 degrees to the fjord’s opening. The ridge terminates in a teardrop-shaped plateau in the northwest. Because the dataset extends to the fjord walls, a deep (over 160 m) and narrow basin can be seen to separate the shoal from the fjord wall in the northwest, while the ridge can clearly be seen to extend all the way to the fjord wall just east of Bulltinden in the southeast (for locations see Figure 5). Furthermore, a sinuous ridge extends from the southwest of the plateau’s western endpoint; this ridge is poorly visible in the multibeam dataset but can also be seen to extend to the fjord wall, at Müllerneset. At the fjord mouth, the submarine portion of the northern fjord wall is significantly steeper than that of the southern fjord wall.
Figure 7: Singlebeam bathymetric survey performed by the Norwegian Hydrographic Service and published in Król et al (2010)
Oceanography
The stratification and circulation of St Jonsfjorden is largely unknown, as is the degree to which it is in communication with the water masses of the shelf, the West Spitsbergen Current and the East Spitsbergen Current. The results of one CTD (conductivity, temperature, depth) cast are available and presented below, as Figure 8. This profile shows fresh but fairly warm surface water becoming rapidly salty and warm with depth in the uppermost ten meters of the water column. This may be caused by the mixing of a plume of glacial meltwater at the surface with warmer, ambient (possibly Atlantic) water entering the fjord from Forlandsundet.

Figure 8: CTD profile taken from outer St Jonsfjorden [78° 31.797’ N, 12° 32.361’ E] on 31 July 2012
Study Area
Study Area

Glaciology

Nineteen named glaciers exist within St Jonsfjorden’s 575 km$^2$ catchment, occupying a total area of 335 km$^2$ – nearly 69% of the terrestrial portion of the catchment as of 2007 (calculated using data from Korona et al 2009 distributed by Norsk Polarinstittut). The extent of the catchment and the locations of the glaciers are shown in Figure 5.

The largest of the fjord’s glaciers, Osbornebreen, is a tidewater glacier fed by multiple tributary glaciers, all but two of which are unnamed. Including its tributary glaciers, Osbornebreen accounts for 47% of the ice cover in this catchment and is the only glacier draining to St Jonsfjorden known to have surged.

The only observed surge of Osbornebreen occurred from 1986-1989. This surge was the subject of a study by Rolstad et (1997), who used Landsat images to determine the glacier’s velocity through the surge and aerial photographs to develop a digital elevation model based on the glacier’s geometry in 1990. Based on the evolution of the crevasse pattern and deformation of individual crevasses, Rolstad et al (1997) suggest that the surge may have begun as a zone of extension which traveled up the ablation zone, rather than as a zone of compression traveling downstream. This is not a view shared by Sund et al (2009), who believe that this surge was already well underway when the upward-traveling extensional zone was identified, giving the illusion that the surge initiated at the terminus. The duration and extent of this surge’s active phase is also discussed in Dowdeswell et al (1991), who suggest that the surge may have begun in the fall of 1986 and note that the glacier is divided into two major flow units only one of which (the western portion, visible in Figure 5) was affected by the surge. The glacier front reached its maximum extension in July 1988 and held this position until 1990 (Rolstad et al 1997).

Presently, two additional tidewater glaciers flow into the fjord from the north: Gaffelbreen and Konowbreen. Konowbreen drains the Løvenskioldfonna ice cap to the north and, prior to the 1960s, shared one tidewater face with Osbornebreen. Like many fjords on western Spitsbergen, glaciers in the St Jonsfjorden catchment have undergone rapid retreat over the past hundred years. In addition to the three extant tidewater glaciers, additional glaciers are present within the fjord’s drainage basin that terminated in the fjord as recently as 100
years ago but have since retreated to terminate on land, including Bullbreen, Ankerbreen and Charlesbreen.

Table 1: Surface areas, presented in square kilometers, of the glaciers surrounding St Jonsfjord. ID numbers and 1982 areas are from Hagen et al (1993) while areas as of 2007 were calculated using ArcGIS Spatial Analyst from König et al (2003) published on npolar.no

<table>
<thead>
<tr>
<th>ID</th>
<th>Name</th>
<th>Area, 1982</th>
<th>Area, 2007</th>
</tr>
</thead>
<tbody>
<tr>
<td>15301</td>
<td>Gislebreen</td>
<td>1.9</td>
<td>1.1</td>
</tr>
<tr>
<td>15302</td>
<td>Bullbreen</td>
<td>14.2</td>
<td>14.5</td>
</tr>
<tr>
<td>15303</td>
<td>Holmsletbreen W</td>
<td>1.05</td>
<td>0.9</td>
</tr>
<tr>
<td>15304</td>
<td>Holmsletbreen E</td>
<td>1</td>
<td>0.7</td>
</tr>
<tr>
<td>15305</td>
<td>Løvliebreen</td>
<td>12.7</td>
<td>11</td>
</tr>
<tr>
<td>15306</td>
<td>Gunnarbreen</td>
<td>4</td>
<td>3.2</td>
</tr>
<tr>
<td>15307</td>
<td>Anna Sofiebreen</td>
<td>3</td>
<td>3.2</td>
</tr>
<tr>
<td>15308</td>
<td>Charlesbreen</td>
<td>27.8</td>
<td>25.6</td>
</tr>
<tr>
<td>15309</td>
<td></td>
<td>2.15</td>
<td>2</td>
</tr>
<tr>
<td>15310</td>
<td>Vegardbreen</td>
<td>11.4</td>
<td>10.1</td>
</tr>
<tr>
<td>15311</td>
<td>Paulbreen</td>
<td>3.3</td>
<td>2</td>
</tr>
<tr>
<td>15312</td>
<td>Vintervegen</td>
<td>31.6</td>
<td>29.3</td>
</tr>
<tr>
<td>15313</td>
<td>Osbornebreen</td>
<td>152</td>
<td>156.7</td>
</tr>
<tr>
<td>15314</td>
<td>Konowbreen</td>
<td>49.6</td>
<td>47</td>
</tr>
<tr>
<td>15315</td>
<td>Smalgangen</td>
<td>1.8</td>
<td>1.6</td>
</tr>
<tr>
<td>15316</td>
<td>Gaffelbreen</td>
<td>23.4</td>
<td>21.2</td>
</tr>
<tr>
<td>15317</td>
<td></td>
<td>1.4</td>
<td>1</td>
</tr>
<tr>
<td>15318</td>
<td>Ankerbreen</td>
<td>4.9</td>
<td>4.2</td>
</tr>
</tbody>
</table>

Bedrock Geology

The bedrock geology of St Jonsfjorden is highly diverse and, in places, poorly understood (Kanat and Morris 1988). The complicated geologic history of St Jonsfjorden has caused a diverse range of lithologies to be exposed therein. In the outer fjord valley, Svalbard's Proterozoic basement (Hecla Hoek) is exposed at lower elevations and locally overlain by Paleozoic (Ordovician) basement at higher elevations (See bedrock map, Figure 9). The basement rocks are highly diverse and include igneous and sedimentary rocks which have undergone various degrees of metamorphism (Harland et al 1974). These units are often
Study Area

referred to as Svalbard’s Pre-Old Red Basement. Sedimentary rocks deposited in the later parts of the Paleozoic (Devonian and thereafter) and Mesozoic dominate the mountains of the inner fjord.

The most recent stratigraphy published by Kanat & Morris (1998) divides the bedrock geology into six groups. The four most common being the Strandflat Strip deformed sedimentary rocks, Bullbreen Group low-grade metasedimentary rocks, marbles and slates of the Sarsøyra Formation, Comfortlessbreen Group greenschist-facies (low temperature, moderate pressure) metamorphosed glaciomarine sedimentary rocks (including tillites and quartzites and phyllites), and St Jonsfjorden Group metamorphosed sedimentary rocks which represent a fining-downward succession with quartzites at the top and banded marbles at the base, all of which overlies a basic, Pre-Cambrian igneous unit (Ohta 1985). Shales and schists with intercalated carbonates of the Müllerneset Formation are exposed locally in southwest St. Jonsfjorden. The Vestgötabreen Structural Complex high pressure-low temperature metamorphic rocks overlie the Bullbreen group in the southern section of the catchment and clasts from the Vestgötabreen are found in conglomerates of the Bullbreen (Kanat and Morris 1988).

Mineralogy of overbank sediments

The geochemistry of overbank sediments larger than 0.06 mm was measured at 20 points around St Jonsfjorden in 1986-87 by NGU and NVE. These results are presented in the Geochemical Atlas of Norway (Ottesen et al 2010) and show that sediments sampled at different points around the fjord’s shoreline are enriched in different elements. According to this study, the highest concentrations of Gold (Au), Calcium (Ca) and Strontium (Sr) are found on the northern shore of the outer fjord, especially in front of Ankerbreen and Lowzowfjella, an area with the lowest concentrations of Iron (Fe) and Aluminum (Al). Overbank sediments recovered from the fjord head appear modestly enriched in Rubidium (Rb), Potassium (K), Manganese (Mn) and Zirconium (Zr).

Figure 9 (opposite): Bedrock map of St Jonsfjorden (catchment outlined in red) and surrounding areas. Data is from Berggrunnskart over Svalbard [Svalbard Bedrock Map] 1:750 000 published by the Norwegian Polar Institute. This figure was made using ArcMap 10.1
Study Area
Materials and Methods

Data Acquisition / Materials

The multibeam bathymetric survey, nine high resolution, two-dimensional acoustic profiles and four sediment gravity cores presented in this thesis were collected from St. Jonsfjorden during the Geology course GEO-3144/8144 (Marine Geology and Geophysics) class cruise on August 1, 2012 (R/V Helmer Hanssen). Details pertaining to these datasets and their acquisition are presented below.

Geophysical (Acoustic) Data

The bathymetry and sub-sea sediments have been imaged using the University of Tromsø’s hull-mounted Kongsberg-Simrad EM300 multibeam echosounder and EdgeTech 3300 sub-bottom (chirp) profiling system, respectively. Both instruments use digitally produced sound waves and the principles of reflection seismology to image landforms, as well as surficial and buried sediments, and bedrock features.

The transducers (‘projectors’) that produce these sound waves direct them downward, where they reflect off interfaces (reflectors) between intervals of contrasting acoustic impedance (the product of density and velocity of the compressional wave through the medium). The penetration of the signal is dependent on its frequency, with lower frequencies penetrating furthest. Reflections – the manifestations of reflectors on seismograms – appear strongest where the difference in acoustic impedance is greatest.

Reflectors exist at the sediment-water interface and the interface between the sediment package and underlying bedrock as well as within the sediment package. There they denote changes in, for example, structure and stratification, grain size and lithology, sorting, and compaction. Furthermore, reflectors often exist at unconformities and faults, where the properties of the units on either side of the interface are markedly different.

The returned acoustic signal is received by a second set of transducers (‘hydrophones’). The time between the acoustic signal’s generation and return is used, along with the speed
of sound through the water column, to determine the distance from the ship to the reflector. The speed of sound is dependent on temperature and density; these values are measured at the start of the survey via CTD casts, in which instruments which measure and record conductivity, temperature and depth are lowered through the water column while the boat remains stationary. Because the hydrophones produce multiple signals simultaneously it is necessary that they be distinguishable from one another upon return. In sonar units where multiple projectors are used to ensonify a large area this can be done by assigning each projector a unique frequency (see Multibeam description below). In sonar units where a larger amount of energy must be used to penetrate and image shallow sub-sea units the use of a time-variable frequency enables the use of long signals (see the description of chirp profilers below).

Kongsberg-Simrad EM300 Multibeam

In the EM300, the projecting and receiving transducers are arranged in a Mills cross configuration, wherein the projectors are mounted along the ship's centerline with the hydrophones aligned perpendicular. The projector array comprises several sections (transducer clusters) which are independently steered to best control for motion of the vessel (pitch, roll and yaw) as sensed by an assemblage of motion sensors, gyrocompasses and positioning units in continuous communication with the transceiver unit. These clusters each emit a signal of unique frequency 30-34 kHz, a technique which improves the resolvability of features in the returned signal (Kongsberg 2006). This signal, composed of 135 individual beams, is in the shape of a fan, allowing a wide (150°) swath of the seafloor to be ensonified simultaneously. The width of this fan is depth-dependent, with greater coverage in deeper areas. One CTD cast was performed prior to the start of data acquisition in order to calibrate the echosounder.

The bathymetric survey covers 62.6323 km² (plan view) of the fjord's central area at a resolution of 5 m. Bad soundings were removed using the processing software Neptune and the survey has been interpreted in Interactive Visualization Systems Inc's Fledermaus – The Art of Visualization software, version 7.3.0c. The width of the seafloor swath mappable in any given pass of the research vessel is proportional to the water depth in the survey area; because of this, mapping shallow areas requires multiple passes. In the inner
fjord, the multibeam dataset reaches 2km from the present tidewater cliff of Osbornebreen and 3.5 km from the calving front of Konowbreen. From the fjord mouth, the dataset extends south as a limited amount of data was collected while the ship was in transit to St. Jonsfjorden.

**EdgeTech 3300 Chirp**

The EdgeTech 3300, in contrast, ensonifies only a limited area beneath the transducers. Instead of using different frequencies to ensonify a large lateral area, the EdgeTech 3300 uses a single acoustic pulse containing several frequencies (500 Hz to 20,000 Hz); this significantly increases the vertical resolution of the survey when compared to fixed-frequency source surveys (EdgeTech 2015).

The seismic survey consists of nine chirp lines: six inlines and three crosslines. Five inlines (numbers 1, 2, 4, 6 and 8) cover the entire length of the fjord while one, number 9, spans only the innermost three kilometers. Two of the three crosslines, numbers 3 and 7, cross the fjord at the mouth, while number 5 crosses the fjord at its head.

The nine chirp lines taken in St Jonsfjorden were interpreted using Seismic Micro-Technology Inc’s The KINGDOM software, version 8.6. Horizons have been picked manually.
Figure 10: Locations and extents of the multibeam and seismic surveys
Sediment Cores

The four sediment cores were recovered using a gravity corer, in which a steel barrel lined with a stiff plastic tube falls into the sediment package under the weight of an approx. 1600 kg weight. Sediments pass through a flexible metal cone (‘core catcher’) before entering the tube. This cone prevents them from falling out as the core is lifted. Once on the deck of the Helmer Hanssen, the liner was removed from the barrel and cores were split into approximately one-meter sections. After being capped and sealed, the sections were stored at +4C. Lengths and locations of the four recovered cores are presented as Figure 11 and Table 2, below.

Figure 11: Locations of the four gravity cores presented in this thesis

<table>
<thead>
<tr>
<th>Core</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Water Depth</th>
<th>Recovered Sediments</th>
<th>Sections</th>
</tr>
</thead>
<tbody>
<tr>
<td>HH12-953-GC-MF</td>
<td>78° 32.202’ N</td>
<td>13° 05.544’ E</td>
<td>82 m</td>
<td>346 cm</td>
<td>4</td>
</tr>
<tr>
<td>HH12-954-GC-MF</td>
<td>78° 31.732’ N</td>
<td>12° 59.024’ E</td>
<td>37 m</td>
<td>360 cm</td>
<td>4</td>
</tr>
<tr>
<td>HH12-955-GC-MF</td>
<td>78° 30.691’ N</td>
<td>12° 45.429’ E</td>
<td>88 m</td>
<td>255 cm</td>
<td>3</td>
</tr>
<tr>
<td>HH12-956-GC-MF</td>
<td>78° 31.792’ N</td>
<td>12° 27.085’ E</td>
<td>111 m</td>
<td>351 cm</td>
<td>4</td>
</tr>
</tbody>
</table>
Data Analysis / Methods

Laboratory work was carried out at the University of Tromsø – The Arctic University of Norway (UiT) from April 2014 to March 2015.

Geophysical Data: Inversion Models and the Process-Response Relationship

Processes that operate within depositional environments generate characteristic responses that can be recorded in the deposited sediments (Dalrymple 2010). This principle is known as the Process-Response Relationship, which further asservates that the relationships between sedimentary processes and products are both direct and predictable. The examination of sediments, sedimentary successions and landscape elements, therefore, can, with some effort, be used to understand the conditions that led to their formation (Dalrymple 2010).

Inverse problems are those where observations of systems and objects are collected and analyzed to understand the processes that formed them. Deducing model parameters from data is a method commonly used in physical science, where physical processes themselves are not always easy or possible to observe (Snieder & Trampert 1999).

When applied to the glacial landform record, inverse modeling can be used to reconstruct the dynamics and configurations of older glaciers from the sediments, erosional features and sedimentary structures they leave behind. The Glacial Inversion Method is one process by which the histories of glaciers and ice sheets can be reconstructed via such an analysis of landforms and landform assemblages exposed on former glacier beds. In this method, the complete landform record is utilized along with the relative ages of the landforms to provide information on the internal dynamics of the ice body and the positions of the ice margins. This method was first outlined by Johan Kleman and Ingmar Borgström for the purpose of reconstructing the regional dynamics of terrestrial paleo-ice sheets on millennial timescales (Kleman and Borgström 1996). In this thesis, landforms are identified on the fjord floor in the bathymetric data and described in terms of their lengths, shapes, elongation ratios, heights, and curvatures. Where possible, these landforms are further grouped into clusters (‘swarms’) believed to be of like depositional age.
Sediment Gravity Cores

Prior to opening, measurements on the whole core sections were performed using the University of Tromsø’s GEOTEK Multi-Sensor Core Logger (MSCL; see Figure 13) and X-Radiographs were taken using a GEOTEK x-ray machine. The gravity cores were opened in May 2014 using twin-mounted vibrating, circular saws (shown in Figure 12) to split the liners and an osmotic knife to split the contained sediments. Color photographs were taken of the split core surfaces and the qualitative elemental composition (bulk geochemistry) of the sediments was measured using the Avaatech XRF core scanner (Figure 14) and attached camera. Finally, all four cores were sampled for grain size analysis and dateable materials. These methods are described in detail below.

Multi-Sensor Core Logger

The GEOTEK Multi-Sensor Core Logger (MSCL) is capable of measuring multiple physical properties of sediments concurrently as a motor pushes the core section down a set of tracks, through individual sensors mounted on the instrument. As the diameter and temperature of the core can be slightly variable but are both important for the acquisition of the physical properties data obtained by the logger, both parameters were computed and recorded at each measuring point: the diameter using a set of calipers and the temperature using a thermometer. Furthermore, the cores were left at room temperature for at least one day before measuring. Explanations of the instruments installed on the multi-sensor core logger are given below; all information is taken from the manufacturer’s manual (Geotek 2014).
Materials and Methods

Figure 13: Configuration of instruments installed on the Multisensor Core Logger at UiT; direction in which the core is moved by the core pusher (right) through the instrument array is indicated by red arrows.

**Gamma Ray Attenuation Bulk Density**

Determination of the sediments' wet bulk density was done by measuring the attenuation of gamma rays through the core. The decay of a 10-milligram Caesium-137 capsule (source) produces a steady stream of photons (gamma ray), which is collimated and directed through the core in the cross-core direction before encountering a scintillation counter on the wall of the instrument opposite the source. Along their way through the core, many of these photons collide inelastically with the electrons of the core sediments and pore fluid (Compton scattering). Collisions result in a transfer of energy from the incident photons to the encountered electrons. The ratio of photons attenuated during transmission through the core is related to the density of electrons in the core material. The scintillation counter is capable of distinguishing diminished-energy photons from those which retain their initial energy (0.662 MeV); this ratio is used in the calculation of the sediment’s wet bulk density, \( \rho \), by the equation below, where \( \mu \) = the Compton attenuation coefficient, \( d \) = the sediment thickness, \( I_0 \) = the intensity of the gamma source and \( I \) = the measured intensity through the sample.

\[
\rho = \frac{1}{\mu d} \ln \frac{I_0}{I}
\]

Because the instrument measures water-saturated sediments contained within a plastic liner it is necessary to adjust for attenuation through the core liner and pore fluids. This is done by comparison to measurements of a core liner with known contents – namely water.
and a known amount of aluminum that is variable downcore. By measuring this tube periodically it is possible to calibrate the core measurements; it has the added benefit of controlling for the decay of the cesium source over time.

The size (diameter) of the collimator used to bombard the core with gamma radiation is user-controlled; measurements presented in this thesis were performed using the 5 mm (diameter) collimator.

**P-Wave Amplitude and Velocity**

P-waves (primary waves) are compressional (pressure) mechanical waves such as the sound waves used in the chirp and multibeam surveys. P-wave velocity is affected by many variables such as the temperature, density and rigidity of the material through which the p-wave travels. The multi-sensor core logger is able to measure the velocity of p-waves through the core section by passing the section through two transducers, one of which produces a short p-wave pulse similar to the projectors of the geophysical equipment described earlier while the other receives the signal in much the same way as the hydrophones. Using the precise measurements of the core diameter, the time it takes for the signal to travel from the projector to the receiver is used to calculate the p-wave velocity. In order to minimize the amount of air the transducers measure through, both transducers are spring-mounted to lie flush against the core liner and the liners are continuously wetted to improve the acoustic coupling between the core liner and the transducers. Poor coupling – either between the core liner and the transducers or between the sediments and the core liner – is indicated by weak p-wave amplitude; low p-wave amplitude can be taken as an indication that the p-wave velocity measurement is unreliable.

**X-Radiography**

X-radiographs allow for visualization of material interiors on the basis of density differences, which, in sediments, are largely controlled by lithology, grain size, and compaction and nature of the pore fluids. Digital radiographs of all four cores were taken in spring 2014 with the GEOTEK MSCL-XCT using a source voltage of 120 kV and 350 µA
Materials and Methods

current. Images were taken every 20mm and stitched together using MSCL’s software; they are presented in such a way that the highest-density intervals appear white.

Color Photography

The color photographs presented in this thesis were taken in November 2014, five months after the opening of the cores, using the line-scan camera mounted on UiT’s XRF scanner. These photos were taken with an exposure time of 10 ms and an aperture of f/11. Prior to photographing, the surfaces of the cores were cleaned (smoothed) using a plastic card and the surfaces were allowed to dry as uneven surfaces can give the impression that stratification or sedimentary structures exist where they do not. Wet surfaces photograph poorly as water reflects light from the camera.

Creation of Lithological Logs

Immediately after opening the gravity cores, their colors were described as color changes have been reported to begin to occur as soon as the cores are exposed to air (Hansen 2014). The surfaces of the cores were described and logged with attention paid to color, sedimentary structures, consistency/compaction, clast-matrix relationship, sorting, and the nature of the contacts between the beds and units approximately three months after opening. Colors reported in the lithological logs are those observed on the split core surfaces approximately three months after opening, after having been cleaned with a plastic card. The Munsell color scale codes are presented in Table 3.

The division of the sediment cores into units was done on the basis of these initial observations; units reflect visible changes in dominant sediment grain size, the presence or absence of stratification, and color. Observations of clasts on the x-radiographs were combined with this information to generate the logs presented in the Results chapter. Nongenetic lithofacies codes have been assigned according to the scheme presented in Table 4.
### Materials and Methods

#### Munsell Color Code

<table>
<thead>
<tr>
<th>Munsell Color Code</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>5Y5/2</td>
<td>Olive grey</td>
</tr>
<tr>
<td>5B5/1</td>
<td>Bluish grey</td>
</tr>
<tr>
<td>N6/</td>
<td>Grey</td>
</tr>
<tr>
<td>2.5Y4/1</td>
<td>Dark grey</td>
</tr>
<tr>
<td>2.5Y6/1</td>
<td>Greyish brown</td>
</tr>
<tr>
<td>10YR/2</td>
<td>Light olive brown</td>
</tr>
<tr>
<td>2.5Y5/4</td>
<td></td>
</tr>
</tbody>
</table>

*Table 3: Munsell color codes assigned during logging*

#### Facies Code

<table>
<thead>
<tr>
<th>Facies Code</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fl</td>
<td>Fines, Stratified</td>
</tr>
<tr>
<td>Fm</td>
<td>Fines, Massive</td>
</tr>
<tr>
<td>Dmm</td>
<td>Diamict, matrix-supported and massive</td>
</tr>
<tr>
<td>Dml</td>
<td>Diamict, matrix-supported and stratified</td>
</tr>
</tbody>
</table>

*Table 4: Lithofacies codes assigned during logging*
Element Geochemistry

X-Ray Fluorescence Scanning

The element geochemistry of the sediment cores was determined using the Avaatech XRF core scanner at the Department of Geology, UiT. This instrument uses an x-ray source and the principles of energy dispersive x-ray fluorescence to rapidly and non-destructively determine the elemental composition of split sediment cores. The configuration of the components that compose the scanner is shown as Figure 14. Much like the MSCL, the XRF scanner provides near-continuous measurements at user-defined down-core and cross-core intervals. In this case, all four cores were scanned at one-centimeter down-core and 12 mm cross-core resolutions. Additionally, selected stratified intervals were further scanned at a downcore resolution of two millimeters using a cross-core slit size of five millimeters.

In x-ray fluorescence spectrometry, radiation is directed at the split core surface. Once the beam encounters the core surface, the x-ray photons collide with the atoms that compose the sediments. For every element, electrons in each shell have a specific energy at which they are able to overcome their electrostatic attraction to the nucleus, allowing them to be expelled (ionization energy). By exciting the inner electrons such that they vacate their orbitals, an electron from a higher orbital can fall to occupy the void thereby stabilizing the structure. As the outer-shell electrons drop from higher-energy positions to the lower-energy inner shells the excess energy is released. This ‘characteristic energy’ is dependent on the falling electron’s original orbital, the destination orbital and the composition of the x-ray source anode (Rhodium, in the case of the source used herein). The characteristic energy is indicative of the element encountered by the x-ray beam and therefore the content of the core. The energies themselves are identified by a silicon drift detector that turns the x-rays into a voltage signal which is in turn processed by a digital pulse processor. Ultimately this processed signal is interpreted using WinAxil X-Ray Analysis Software.
Figure 14: Components of the XRF core scanner

Within the source, electrons emitted from a heated filament cathode are accelerated through a vacuum chamber to a Rhodium anode. Upon collision with the Rhodium target the electrons produce both x-rays and heat; the x-rays travel out of the vacuum through a beryllium window as a beam while the heat is dissipated using circulating water. The amount of electrons produced at the cathode is a function of the current at the filament (how hot the cathode is heated). In Avaatech XRF core scanner, the maximum filament current is 2.4A. The degree to which the electrons are accelerated is proportional to the electric potential tension (voltage) between the cathode and the anode. The voltage across the vacuum is controlled by the user and can be adjusted to excite elements of interest (minimum voltage to induce radiation generation: 10kV; maximum operating voltage: 50kV).
The secondary radiation produced during the collision of the x-ray beam with the core sediments is measured through a helium-filled chamber directly in contact with the core surface, save for a four-micron Ultralene® film. Secondary radiation, especially that produced by light elements, is easily absorbed by high-friction materials such as air and water (Richter et al 2006; Tjallingii et al 2007). Measuring through helium as opposed to air reduces the amount of friction to which the radiation is exposed, thereby preserving the sensor’s ability to detect all elements, however it is not possible to remove all air and water from the surface due to the uneven nature of split sediment cores. Deleterious effects of measuring through water and air caused by the inherently uneven nature of sediment core surfaces are termed matrix effects; they can be minimized by presenting geochemical data as element ratios instead of simply as counts (Tjallingii et al. 2007; Weltje & Tjallingii 2008). An attempt to minimize secondary radiation absorption by water and air at the sediment surface was made by ensuring the cores were at room temperature before applying the film, in order to prevent condensation from forming on its underside. The Ultralene® film prevents the contamination of the sensor as it moves down the core, however, especially while measuring coarse sediments, small amounts of air and water can become trapped between the core surface and the foil. A roller was used to remove air bubbles and further ensure the best possible contact between the foil and the sediment surface.

Measurements on each section were done in multiple runs, each designed to measure a distinct range of elements by selecting an appropriate voltage (see above) and using metal filters to block bands of secondary radiation produced by elements lighter than the range targeted for investigation (noise). Application of these filters allows the operator to obtain more accurate measurements of the relative concentrations of heavier elements, as radiation from the lighter elements does not reach the sensor. The runs were configured as follows: elements from Mg to Co were measured using a 10kV electric potential and a 1000µA beam current; Ni to Mo using a 30kV electric potential and a 2000µA beam current with a thick lead filter; Mo to U using a 50kV electric potential and a 2000µA beam current with a copper filter.
Presentation and analysis of geochemical data

Although element geochemistry data is obtained in the form of numbers of counts (further normalized by measurement time and area), the data is presented as element ratios (ratio of element counts to each other or to a sum) to reduce the effects of the absorption of secondary radiation due to roughness and water content (matrix effects, described in Tjallingii et al. 2007), variations in beam intensity, and changes in sediment compaction and density.

In sediment cores, element ratios have been widely used to determine the relative contributions of terrigenous and pelagic sedimentation to the deposits of deep sea basins; periods of increased terrigenous input can be recognized by higher ratios of common siliciclastic rock-forming minerals (e.g., Al, Ti, and Fe) relative to those which make up biogenic sediments (e.g., Ca and Sr; Sr commonly substitutes for Ca as noted in St-Onge et al 2007). The Ti:Ca proxy has limited utility in fjords, where sedimentation rates are very high and the biogenic, authogenic contribution is largely drowned out by terrigenous input.
Materials and Methods

**Grain Size Analysis**

The grain size distribution throughout the four gravity cores has been determined via a combination of sieving and the use of a Beckman Coulter Laser Diffraction Particle Size Analyzer (LS 13 320). Sieving was necessary because the sediments contained within the gravity cores – as noted on the x-radiographs and during visual logging – exceed the maximum size of those measureable based on the applied settings of the Beckman Coulter Laser Diffraction Particle Size Analyzer (2 mm; chosen as it is the maximum size of sand). Sampling resolution is not constant throughout the cores but rather was determined after examining the x-radiographs and visual logs and carried out in such a manner that did not cross-cut layer boundaries. In this thesis, sediments are classified according to the scheme put forth by Shepard (1954)

**Beckman Coulter Laser Diffraction Particle Size Analyzer**

The Beckman Coulter Laser Diffraction Particle Size Analyzer determines grain size distributions by transmitting a laser beam through a sample of sediments suspended in water. The sediments scatter the light based on their optical properties and dimensions; the distribution of the scattered light is recorded by a set of sensors mounted on the chamber’s opposite wall. This distribution (“composite scattering pattern”) is then deconvoluted to determine the relative abundances of particles of different sizes using either the Fraunhofer or Mie theory of light scattering, where the scattering angle is assumed to have been created by a perfectly spherical particle. For this reason, the grain size distribution is more precisely a distribution of the relative volumes of particles with like spherical equivalents. In this thesis, the optical model NGUreal.rf780d was used, where the suffix ‘rf780’ indicates that a 780 nm laser was used and the suffix d indicates that the PIDS analyzer was enabled. This model uses the Mie

<table>
<thead>
<tr>
<th>Fluid Real Refractive Index</th>
<th>Sample Real Refractive Index</th>
<th>Sample Imaginary Refractive Index</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.333</td>
<td>1.6</td>
<td>0.1</td>
</tr>
</tbody>
</table>

*Table 5: Parameters of the optical model NGUreal.rf780d*
Materials and Methods

theory of light scattering, which more accurately represents small particle sizes but requires the user to provide information on the refractive indices of the fluid and sample. Those used in the NGUreal.rf780d model are given in Table 5.

Because the light must reach the sensors on the chambers opposite wall, the opacity of the solution cannot be more than 14%. Due to the small size of the measurement chamber, it is only possible to measure very small amounts of material. The machine is designed to dilute samples by draining a percentage of the sample and then adding more water to the system. Each iteration of the dilution process, however, increases the risk that the sample that is measured is not representative of the sample initially inserted into the instrument. Unfortunately, in some samples, the machine diluted the sample upwards of twenty times; there is presently no way to quantify how representative the remaining (measured) sample is after the dilution process. Each sample was measured twice, for sixty seconds each time, and the results presented in the following chapters are averages of the two runs.

On average, the cores were sampled approximately every six centimeters for analysis. Selected laminated intervals in the innermost two cores were sampled more densely. Due to the small amount of material needed for this analysis it was possible to sample in cubes as small as 6 mm to a side. Prior to measurement, the samples were treated with acetic acid, for one day, and hydrogen peroxide, for five days. This was done in order to remove material deposited by processes independent of flow competence and capacity and with the potential to bias the grain size distribution (e.g. foraminfera, organic matter, in situ sediments). Immediately before measurement, each sample was centrifuged twice for four minutes each time. In order to break apart clay aggregates, formed during the centrifuging process or previously, two drops of Calgon were added to each sample and each sample was placed in an ultrasonic bath for approximately seven minutes. In two of the gravity cores, HH12-956 and HH12-954, the individual samples were dried and weighed after treatment with acid and peroxide in order to quantify the mass lost as a percentage of the samples’ dry weight.

The use of laser diffraction particle size analyzers for the analysis of sediments was initially hindered by the inability of the apparatus to detect clay-sized particles (McCave et al 1995), as sub-micron sized particles produce scattering patterns that become smoother and more
difficult to measure by the sensor array as particles become smaller. The development of the Polarization Intensity Differential Scattering (PIDS) detector has effectively expanded the range of grain sizes quantifiable by laser diffraction particle size analyzers such that they can now be used to quantify the size distribution of all fines. PIDS uses a second set of light sources of shorter wavelength (250, 600 and 900 nm) to induce an oscillating dipole moment in the sediment particle, causing the particle to emit light in all directions except in the direction of the oscillation. This can be controlled by polarizing the light sources using a set of integrated filters. Six photodiode sensors then record the intensity of the scattered light; the difference in scattering intensity under vertical and horizontal polarizing conditions for each wavelength is used to determine the size of small particles.

**Paleocurrent Strength from sortable silt mean grain size**

The mean grain size of so-called ‘sortable’ silt grains (s3) is often used as a proxy for the strength of bottom currents (Hass 2002; McCave et al 1995; Sternal et al 2014). Although grains as small as 4 μm are generally included in the silt grain size fraction, many grains this small can reasonably be expected to have been transported as aggregates. The sortable silt fraction is therefore often defined as being those silts larger than 10μm (10-63μm; McCave et al 1995; Hass 2002). These aggregates are assumed to have been broken apart by the addition of calgon and the use of the ultrasonic bath during the treatment process prior to grain size analysis. The mean grain size within the fraction 10-63 μm was determined by calculating a weighted average of the volume percentages reported by the corresponding channels. The percentage of fines greater than 10 μm was determined by summing volumes detected by those channels and dividing by the sum of the volumes detected by all <63 μm channels (total fines).

The validity of near-bottom current strength estimations derived from s3 measurements is contingent on the fulfillment of certain conditions. Namely, that transport and deposition are solely controlled by one, unidirectional current (Hass 2002; McCave et al 1995; Sternal et al 2014). Because of this, the use of sortable silt mean grain size as a proxy for near-bottom current speed has largely been confined to deep sea sediments. Because they often contain sediments deposited via a variety of processes and coming from multiple sources, fjords are not generally seen as ideal environments in which to study bottom current
Materials and Methods

strength variations via sortable silt. In Sermilik Fjord (southeast Greenland), Andresen et al (2014) used $S_S$ to reconstruct the North Atlantic Oscillation-driven inflow of shelf waters into the fjord basin.

In fjord settings, the influence of ice rafted debris contaminates the silt fraction, complicating the use of $S_S$ (Hass 2002). Assuming that sediments in the sand size fraction are too massive to be transported by bottom currents but that sand and silt are incorporated into icebergs at equal rates, Hass (2002) developed a method to statistically separate the iceberg-derived silt from the total silt. After fitting a regression equation to the plot of weight percent sand versus $S_S$ in phi units Hass used that equation to determine how $S_S$ would have varied absent a variable bottom current ($S_S_{pot}$). By subtracting $S_S_{pot}$ from the $S_S$ determined by the particle size analyzer, Hass determined the sortable silt mean grain size solely as a function of current strength variation (deviation from $S_S_{pot}$; $\Delta S_S$). I have applied this method to the sediment cores from St. Jonsfjorden, using regression equations calculated using the Analysis TookPak for Microsoft Office Excel. In determining the (geometric) mean sortable silt size, Gradistat software was used (Blott & Pye 2010).
Geochronology: Radiometric Dating

The predictable and measurable decay rates of naturally occurring radioactive isotopes have allowed them to be developed into reliable geologic dating tools. Radioactive isotopes decay at different rates, and their half-lives determine the timescales on which they are useful: ages are determined by calculating the ratio of parent to daughter isotopes present in a sample; after approximately five half-lives these numbers are often too low to reliably quantify.

Radiocarbon Dating: The Carbon Cycle

Carbon-14, with a half-life of 5,730 years, is the only naturally occurring radioactive isotope of carbon and makes up less than one percent of carbon on Earth. Produced via the continual collision of neutrons and Nitrogen-14 atoms in the atmosphere, Carbon-14 atoms readily bond with oxygen to form carbon dioxide molecules. These enter the global carbon cycle and travel through the same pathways as carbon dioxide molecules formed from stable carbon isotopes. Carbon-14 decays to Nitrogen-14 (stable) via β- decay (one nuclear proton is transformed into one nuclear neutron while one electron is ejected; Bowman 1990).

Organisms incorporate carbon from the environments in which they live into their tissues. The uptake of Carbon-14 by living organisms is assumed to be roughly in equilibrium with the Carbon-14 present in the atmosphere at the location of uptake (Lowe and Walker 1997). Upon death, the creation of new tissue ceases and the body can be regarded as a closed system; due to the short lifespans of plants and animals relative to the half-life of Carbon-14, the ratio of Carbon-14 to Nitrogen-14 or Carbon-13 can be used along with the decay rate of Carbon-14 to quantify the amount of time since the death of the organism.

In the marine environment, radiocarbon dating faces unique challenges owing to fractionation and residence times. Carbon enters the hydrosphere via dissolution at the sea surface and as dissolved organic carbon from the terrestrial environment. The ratio of Carbon-14 in the marine environment is not the same as in the atmosphere because ocean waters preferentially absorb Carbon-14 at the sea surface. Furthermore, mixing in the oceans is not instantaneous and organisms living at the sediment-water interface can
reasonably be expected to incorporate older carbon into their tissues than those living at the sea surface; this is termed ‘marine reservoir effect’ and is commonly taken to be approximately 400 years; this principle has been the foundation for ocean ventilation studies (Stuiver et al 1983).

**Radiocarbon Dating: Dated Material**

No whole, in situ macrofossils were located in any of the four gravity cores. For this reason, benthic foraminifera were chosen for radiocarbon dating.

Monospecific samples of the benthic foraminifera *Nonionellina labradorica* and *Cibicides lobatulus* were used in the AMS-radiocarbon dating of sediments in gravity cores HH12-954, HH12-955 and HH12-956. These specimens were picked from the 125-500 micron grain size fraction of freeze-dried samples (vertical resolution: 1 cm) sieved at the UiT in January 2015.

*N. labradorica* is a shallow infaunal species, meaning that it lives in the uppermost centimeters of the water column and feeds on sediments presumably older than those at the surface (Corrliss 1991). Therefore, these radiocarbon ages are expected to overestimate the ages of the intervals dated using *N. labradorica*. In Hornsund, a fjord roughly 200 km southeast of St. Jonsfjorden, *N. labradorica* specimens were observed living as deep as 9 cm beneath the sediment-water interface (Zająckowski et al 2010).

In contrast, *Cibicides lobatulus* is an epifaunal species, living at the sediment-water interface, feeding directly on material suspended in the lowermost portion of the water column. While the carbon in these tests can reasonably be expected to accurately reflect benthic conditions during the lifespan of the foraminifera the microhabitat of the *C. lobatulus* also makes them more likely to be redistributed via current action. Because of this, the largest (presumably most massive and least likely to be redeposited by bottom currents) *Cibicides lobatulus* specimens possible were chosen for dating.

At three stratigraphic levels, enough foraminifera were present that it was possible to submit both monospecific samples of *C. lobatulus* and *N. labradorica* for dating. These dates will be compared and their validity will be discussed later in this thesis. Of the intervals investigated, gravity core HH12-954 contained only one sample with enough carbonate
Materials and Methods

material to be radiocarbon dated – three large foraminifera tests which were submitted as one sample. All samples were rinsed with distilled water prior to being submitted to the Ion Beam Physics Laboratory at the Swiss Federal Institute of Technology (ETH) Zurich for dating.

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth, cm</th>
<th>Species</th>
<th>Weight, g</th>
</tr>
</thead>
<tbody>
<tr>
<td>HH12-956-GC</td>
<td>75.5-76.5</td>
<td>Nonionellina labradorica</td>
<td>0.0023</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Cibicides lobatulus</td>
<td>0.0026</td>
</tr>
<tr>
<td>HH12-956-GC</td>
<td>174.5-175.5</td>
<td>Nonionellina labradorica</td>
<td>0.0015</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Cibicides lobatulus</td>
<td>0.0030</td>
</tr>
<tr>
<td>HH12-956-GC</td>
<td>234-235</td>
<td>Nonionellina labradorica</td>
<td>0.0028</td>
</tr>
<tr>
<td>HH12-955-GC</td>
<td>86-87</td>
<td>Nonionellina labradorica</td>
<td>0.0020</td>
</tr>
<tr>
<td>HH12-955-GC</td>
<td>186-187</td>
<td>Nonionellina labradorica</td>
<td>0.0020</td>
</tr>
<tr>
<td>HH12-955-GC</td>
<td>246-247</td>
<td>Nonionellina labradorica</td>
<td>0.0019</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Cibicides lobatulus</td>
<td>0.0007</td>
</tr>
<tr>
<td>HH12-954-GC</td>
<td>267-268</td>
<td>Combined sample of N. labradorica &amp; C. lobatulus</td>
<td>0.0005</td>
</tr>
</tbody>
</table>

Table 6: Depths and weights of samples submitted for radiocarbon dating

Radiocarbon Dating: Procedure

Radiocarbon dates presented herein were calculated at the ETH Ion Beam Physics Laboratory in April 2015 by Dr Lukas Wacker. The measurements were performed using a compact Accelerator Mass Spectrometry (AMS) system developed at ETH for use with very small samples. This method is new and continually being improved upon. The details of a
similar system, also developed at ETH, are presented in Bard et al (2015); this method is an outgrowth of the method presented in Wacker et al (2013).

With the exception of the sample from gravity core HH12-954, which was too small, all samples were leached with 100 µl 0.02M HCl in order to remove authigenic carbonates and surface contamination. Carbon from this leached fraction was measured independently, and measurements of the leached fraction were compared to the main fraction to determine the reliability of the obtained dates (Wacker, personal communication).

**Radiocarbon Dating: Calibration**

In the atmosphere, large fluctuations in Carbon-14 fluctuations are expected over a variety of timescales; because of this, there is considerable difficulty in converting radiocarbon years to calendar years (Fairbanks et al 2005). In this thesis, dates have been calibrated using the software Calib Rev 7.0.4, using the calibration dataset Marine04 (Stuiver et al). This calibration takes into account the changing rate of $^{14}$C production as well as a time-dependent global reservoir age of approximately 400 years; a local difference of 105 ± 24 has been added as suggested by Mangerud et al (2006). Calibrated dates presented herein are rounded to the nearest decade as suggested by the authors of the CALIB program for dates with standard deviations (one sigma ranges) greater than 50 years.

**Lead-210 Dating: The Lead Cycle**

The decay of radioactive isotopes of Uranium and Thorium to stable isotopes of Lead in the Earth’s crust accounts for much of the natural radioactivity on our planet (Robbins 1978). The decay chain of Uranium-238 passes through thirteen intermediate daughter isotopes before eventually forming the stable lead granddaughter isotope (end product) Lead-206 (Robbins 1978). Many of the intermediate daughter isotopes have short half-lives, on the orders of seconds to hours (Robbins 1978). Radon-222, however, has a relatively long half-life (3.8 days) (Robbins 1978). As an inert gas it can diffuse out from the crust, where it decays through a series of heavy metals which attach to natural aerosols – in this case small particles of solids or liquids suspended in a gas, such as water vapor (fog, steam), dust or smoke – and enter the atmosphere (Robbins 1978). These metal decay products have half-lives of less than 30 minutes except for Lead-210 (22 years), which decays to Bismuth-210
(5 days) and further to Polonium-210 (138 days) before ultimately becoming stable Lead-206 (Robbins 1978).

The ratio of Lead-210 to the daughter isotope Polonium-210 can be used as a geochronological tool and has been used to study sedimentation rates since the early 1970s (Krishnaswami et al 1971 & Koide et al 1972 in Robbins 1978). However, lead in recently deposited sediments comes from two sources: atmospheric deposition at the sediment surface ('unsupported lead' or 'excess lead') and the in-situ decay of its parent isotope, Radium-226 ('supported lead'). Only the unsupported lead should be used in the determination of sedimentation rates. In the deepest parts of the cores, only freshly-produced supported lead is present (no unsupported lead remains) (Robbins & Edington 1975). In order to quantify the unsupported lead activity in the uppermost (younger; dateable) portions of the core, the average supported lead activity in the lower core was subtracted from the total amount of lead activity in the upper core, as per Szczuciński et al (2009), first suggested by Robbins & Edington (1975).

**Lead-210 Dating: The Constant Initial Concentration Model**

The concentration of Lead-210 (activity per unit area) at the sediment-water interface is dependent on both the unsupported lead flux (activity per unit area per unit time) as well as the sedimentation rate (unit sediment depth per unit area per unit time). In order to turn activity measurements into sedimentation rates, and eventually dates, it is necessary to assume that either the activity deposited at the sediment-water interface is either constant per unit time (Constant Flux; activity is variable per unit volume as sedimentation rates change; Constant Rate of Supply) or constant per unit dry sediment volume (sedimentation rates are still variable but the amount of activity per unit volume varies similarly; Constant Initial Concentration) (Cohen 2003). The Constant Initial Concentration model is used in this thesis and is explained below.

The Constant Initial Concentration model established by Robbins and Edington (1975) assumes a constant (linear) sedimentation rate throughout the interval on which unsupported lead is present and a constant flux of excess (unsupported) lead to the sediment-water surface; taking into account the sediment porosity, the density of the solid
sediment and the compaction of the sediments with depth, Robbins and Edington suggest three equations that together describe the profile of lead through a sediment column. These three equations are given below:

\[ A(z) = \frac{P}{\rho_s R(1-\phi_0)} \exp \left[ -\left( \frac{\lambda}{R_0} \right) f(z) \right] + A' \]  \hspace{1cm} (1)

\[ f(z) = \frac{1}{(1-\phi_0)} \left[ z(1 - \phi') + \frac{\phi(z) - \phi_0}{\beta} \right] \]  \hspace{1cm} (2)

\[ A\Delta z_j = \frac{1}{\Delta z} \int_{z_j}^{z_{j+1}} A(z')dz' \]  \hspace{1cm} (3)

where \( P \) is the flux of Lead-210 at the sediment water interface, \( z \) is depth, \( \lambda \) is the radioactive decay constant for Lead-210, \( R \) is the sedimentation rate in centimeters per year (\( R_0 \) being at the sediment-water interface), \( A' \) is the activity of the supported lead as calculated earlier from the older sediments, \( z \) is depth, \( \phi \) is porosity, \( \rho_s \) is the density of the solid phase of the sediment, and \( \beta \) is sediment density.

Nineteen samples from the two inner cores were submitted for dating via Lead-210 to the Gamma Lab at the Institute of Geology, Adam Mickiewicz University (Poznań, Poland) and the dates were obtained by Dr. Witold Szczuciński. These samples were taken from two-centimeter slices representing approximately half the volume of the split core working half on the investigated intervals. Samples were submitted from the surface (0-2cm downcore) and from the intervals 10-12cm, 20-22 cm, 50-52cm downcore and approximately every half meter thereafter. Large clasts (greater than sand-sized) cannot be submitted for dating; because of this, sampling clast-rich diamict intervals was avoided.

The samples were dried at 60C overnight, crushed (powdered) using an electric mortar and sealed in plastic bags at UiT prior to being shipped to Poznań for analysis. The mean weight (dry) of the samples was 25.6 grams.
Materials and Methods
Results – Acoustic Data

Within this chapter, structures identified on the seafloor are described first, in the Bathymetry section. Their internal architectures are described subsequently, in the section titled Seismic Profiles. Interpretations of the fjords landforms and stratigraphy rely on both descriptions from the bathymetry and seismic profiles sections. The stratigraphy in the immediate vicinity of the core locations is described and interpreted together with the lithostratigraphic results presented in the Discussion chapter. Locations and extents of the datasets are presented in the Materials and Methods chapter as Figure 10.

Bathymetry

The large-scale bathymetry of St Jonsfjorden, which can be seen in Figure 15, is composed of smooth-floored basins, bisected by ridges and plateaus oriented at shallow angles (subparallel) to the fjord axis. In the outer fjord, shallow banks (shoals) dominate the bathymetry. Two shoals, Farmgrunnen (formerly known as Megrunnen) and St Jonsgrunnen, are known from bathymetric charts constructed by Norges Svalbard- og Ishavs-undelsøkelser from individual depth soundings made in the first half of the 20th century (npolar.no). At 7 km from present-day terminus of Osbornebreen a broad, shallow ridge bisects the width of the fjord; the landward extensions of this ridge have previously been mapped and interpreted as a terminal moraine dating to the Little Ice Age (Preisner 1988 in Król et al 2010).

In the basins, streamlined, elongate landforms can be observed; throughput the fjord these have variable length-to-width ratios. In the outer fjord their heights are fairly uniform (approximately 5 m) while in the inner fjord subglacial streamlined landforms are between one and thirteen meters high. Regardless of their location, these forms are oriented along the fjord’s local axis. Furthermore, they often appear in association with, articulated to, transverse ridges. These transverse ridges infrequently span the entire width of the fjord.
Acoustic Results: Multibeam

Figure 15: The large-scale bathymetry of St Jonsfjorden.

A note on units: steepness is defined, for the purpose of this thesis, as the absolute value of the slope and is presented as a directionless percentage of change. Slopes of lines that increase from left to right are traditionally (conventionally) termed positive; this convention is less useful in the analysis of three-dimensional shapes. The use of absolute values avoids having to define one direction as positive. Steepnesses in this thesis are estimated using the profiling tool in Fledermaus. Along linear features, steepness has been measured approximately perpendicular to the feature’s strike and is usually presented as a maximum value observed through the examination of multiple profiles.
Acoustic Results: Multibeam

The Inner Fjord

The portion of the fjord head mapped in the multibeam survey is between 50 and 100 meters deep, where gently sloping (steepness less than four) but smooth-floored basins (66 to 94 meters below sea level) separate areas of positive relief (figure 13) oriented in roughly the same direction as the shoals described later in this chapter. These high-relief areas can rise as high as 30 meters from the surrounding basins; within them, landforms are often so closely spaced that it is not possible to determine if they are distinct, individual features or multiple, overlapping features that create composite forms. Because of this, any description of individual forms involves at least a modest degree of interpretation.

Five major zones of positive relief can be identified; these are referred to as zones A (innermost) to E (outermost) and are shown in Figure 16. Of these five, only the outermost (zone E) exhibits a clearly defined ridge crest. This ridge is arcuate, and the trace of this ridge crest is approximately one kilometer in plan view although the seafloor is elevated across the width of the survey area. The ridge crest is most pronounced at the northern extent of the dataset and becomes broader and less pronounced in the southern half of the dataset such that it becomes unrecognizable as a ridge. The remaining four major zones of positive relief are each between 350 and 600 meters wide and often most narrow towards the midline of the fjord. Zones C and D exhibit multiple local topographic maxima separated by rough saddles. Zone B is not continuous on the width of the fjord: a 125-200m wide smooth-floored basin divides this zone into a northern and southern region (marked in stippled line in Figure 16-II). Zone A occupies the northeastern corner of the dataset. While it is notably shallower than its surroundings, the limited representation of the feature in the dataset makes it difficult to describe with confidence. Zone A does, however, appear to have at least a modest crest (6 meters prominence).

Elongate, tapered forms oriented parallel to the fjord axis are clearly visible throughout the inner fjord. These fjord-parallel ridges exist on a spectrum where one end member is bullet-shaped; with broader, steeper and higher snouts facing the fjord head and gently sloping pointed tails tapering in the direction of the fjord mouth. Their length-to-width (elongation) ratios are 5:1; they can be as wide as 170 m and as prominent as 13 meters above the surrounding seafloor. Ridges at this end of the spectrum are between 500 and
Acoustic Results: Multibeam

Figure 16: Landforms in the inner fjord. In subfigure II zones of high relief are identified and labeled A - E. In subfigure III the crests of fjord-parallel ridges are marked with green lines; thickness of these green lines is a reflection of the length-to-width ratio of the feature, with the broadest ridges relative to their lengths marked with the thickest lines. Transverse ridge crests are also shown in subfigure III, as black lines.
1000 meters long. The thinnest longitudinal ridges appear nearly linear and do not taper out-fjord. Their length-to-width ratios can exceed 8:1, they are rarely higher than two meters and are from 350 to 800 meters in length. Fjord-parallel ridges frequently begin or terminate at zones of high relief, contributing to the complex geometry of the high relief zones.

Transverse ridges less than two meters high and twenty meters wide are present on many of the larger transverse and fjord-parallel ridges. These are confined to the fjords shallower areas but are also present to a lesser degree in the deeper portions of the basins as well. The traces of these ridges are sub-linear. The ridges themselves can be up to 350 meters in length (cross-fjord), but are, on average, less than 150 meters long. They rarely occur alone, with fairly evenly spaced groups of up to ten ridges arranged nearly parallel to each other occurring most commonly. The distance between ridges in groups of this type is normally between 40 and 50 meters, with some ridges spaced as frequently as 20 meters.
Acoustic Results: Multibeam

**Broad, Shallow Ridge**

A broad, shallow ridge, shown in Figure 17, bisects the fjord between Konowfjellet and Piriepynten. The ridge itself is over 3 km in the along-fjord direction and rises from 90 meters below sea level in the east and 110 meters below sea level in the west to reach a maximum height of 33 meters below sea level at its crest. The sides slope gently, with steepness never exceeding 5.

While the ridge flanks are generally smooth, anastomosing, sinuous ridges up to two meters high and thirty meters long can be seen superimposed on the crest and shallowest portions of the ridge’s western flanks (Figure 17-A). Hemispheric depressions up to 65 meters in diameter and four meters in relief are present on both slopes of the ridge and occur most frequently beneath approximately 44 meters water depth (Figure 17-D). Similar depressions and raised features can be found throughout the fjord. Positive hemispheric features are also found on the ridge crest and shallower areas of the flanks (Figure 17-B and -C); these have similar dimensions to the depressions mentioned above.

Due to deeper water depths west of the ridge, the western flank of the ridge is longer than the eastern. On this flank, at the northern and southern margins of the bathymetric survey, it is possible locally to see curvilinear traces where the seafloor rises sharply away from the midline of the fjord. The bathymetric survey generally terminates 500-600 meters from the coastline, but can be over 1km from the coastline in areas where shallow bays adjoin the fjord. Gaffelbreen terminates in one such bay on the northern shore of the fjord, just west of the western ridge flank; the bathymetric survey stops 1.6 m from its terminus. Opposite the fjord, the glacier Løvliebreen terminates on land 1.5km from the edge of the bathymetric survey. On the northern fjord wall, in front of Gaffelbreen, elevation changes are modest, on the order of 4 meters change over 110-160 meters and 5 meters change over 170 meters. On the southern fjord, in front of Løvliebreen, steeper slopes can be seen - changes of 15 meters over 80 meters, 10 meters over 150 meters, and 6 meters over 90 meters.
Figure 17: The broad, shallow ridge that bisects the inner fjord between Konowfjellet and Piriepynten.
The Outer Fjord

In the outer twelve kilometers of the fjord, from Gjertsenodden to the fjord head just outside Ankerneset-Müllerneset, the fjord bathymetry is characterized by shoals, ridges and cliffs which punctuate a generally smooth-floored and deep basins (see Figure 15). The northern margin of the bathymetric survey is noticeably deeper than the area off the fjord’s southern shore. Many of the fjord’s transverse ridges are not continuous across the width of the fjord and these features are concentrated along the southern third of the survey.

In the 5 kilometers east of the fjord mouth, a basin in excess of 100m deep occupies the area north of the shoals confined to the southern portion of the survey area. In front of Ankerbreen this basin reaches 160 meters deep. The basin is between 300 and 400 meters wide. Because the 150-meter deep basin is bordered by the 140-meter deep plateau, and this plateau abuts the 40-meter cliff and attached 105-meter plateau, this section of the fjord has previously been described as having a step-like appearance (Forwick et al 2014).

Streamlined and Transverse Ridges

West of – and perhaps extending south from – Gjertsenodden is a ridge 180-650 meters wide and 30 meters high. This ridge is visible in Figure 17 and does not bisect the entire fjord. A local peak at its southern limit, approximately at the midline of the fjord, rises approximately ten meters from a saddle on the ridge’s crest (Feature I). Twelve hundred meters west of this ridge lies a second, larger and irregularly shaped ridge with a steeper western slope and more gently sloping eastern slope. This ridge rises 40-50 meters from the seafloor, shown in Figure 19 (Feature II). East of Bullbreen, a northwest-southeast oriented ridge extends from the southern shore of the fjord (Feature III).

Streamlined, elongate linear features, similar to those found in the inner fjord, extend westward from all three ridges. Like those described from the inner fjord, they appear to begin at larger bathymetric features. These longitudinal ridges are up to 1 km long, 3 meters high and 60 meters across and are oriented parallel with the axis of this section of the fjord. A profile, A-A’, crossing twelve such ridges is shown in Figure 19. Ridges originating at Feature II are generally narrower and longer than those originating from the Gjertsenodden ridge (I).
Extensions of the thinnest fjord-parallel ridges originating from Feature II continue beyond a 15-meter cliff, shown in Figure 18 (plan view) and Figure 19 (as seen from the seafloor, looking northeast at the cliff wall). Unlike in other areas of the fjord, beyond this cliff the ridges do not appear to pinch out so much as they terminate in broad lobes. The cliff itself drops 15 meters over about 70 meters in the north and central parts of the fjord (profiles B-B’ and C-C’, respectively).

Feature III is broad and approximately 30 meters high. Bullet-shaped landforms elongate tails from this ridge westward from a broad, 30-meter tall ridge. These streamlined forms follow the same orientation as those described from other ridges west of Gjertsenodden although they have lower elongation ratios.

Figure 18: Streamlined subglacial bedforms in the outer fjord, showing fjord-parallel ridges extending from features mentioned in the section Streamlined and Transverse Ridges
Figure 19: Ridges oriented along the axis of the fjord extending from Feature II and continuing over the cliff. Two profiles over the cliff are presented as B-B’ and C-C’
Acoustic Results: Multibeam

**Shoals**

In the multibeam survey, gaps in the data from the outer fjord correspond to shallow areas known from navigational charts as Farmgrunnen and St Jonsgrunnen and are visible in Figure 21. These two shoals may be one continuous ridge that is widest in the northwest and tapers toward the southeast. It is at least 1.8 km wide at its widest point while most of the tail is roughly 1.1 km wide. On the interval that was mapped, the majority of the ridge crest lies approximately 42 meters below the sea surface with some areas as low as 60 and some as high as 32 meters below the sea surface. The ridge is steep-sided, with steepnesses as high as 17 and 12, respectively. At Ankerneset, the shoal is separated from the fjord wall by a depression nearly 700m wide and over 100m deep. A smaller raised form is present 500m northeast of St. Jonsgrunnen and may be an independent shoal. This shoal is shallower than 55 meters and clearly visible on the singlebeam dataset published in Król et al (2010).

The southeasternmost portion of the St. Jonsgrunnen-Farmgrunnen ridge crest is not smooth, as the sides are. This irregular seafloor could be several short, non-articulated sinuous ridges or several circular-to-oblong depressions, in which case they are mostly less than 40 centimeters in relief and 10 meters in diameter.

**The Fjord Mouth**

Beyond the banks (shoals) lie two smooth-floored basins up to 163 and 172 meters deep separated by a lip approximately 145 meters deep (this sinuous ridge extends from the western edge of the shoal and is described above; shown in Figure 20). Both basins house numerous circular depressions 30 to 100 meters in diameter and 50 centimeters to 5 meters deep. Two such depressions can also be seen on the lip that separates the basins. Indentations less than 30 meters across can be seen on the walls surrounding the southern, shallower basin, including on the southern flank of the shoal and the southern fjord wall. Beyond the southern basin the seafloor drops suddenly to a depth of greater than 200 meters over a distance of less than 700 meters.
Acoustic Results: Multibeam

Figure 20: The outer fjord, including the shoals Farmgrunnen (northwest) and St Jonsfrunnen (southeast), the two basins of the fjord mouth and the curvilinear features in Forlandsundet (pink; depicted in profile A-B) and the fjord mouth (red; see profile A-B).

Forlandsundet

The portion of Forlandsundet mapped as the ship was in transit to St Jonsfjord from the Central Fram Strait is 600 to 800 meters wide (widest in the south, where water depths as deep as 280 meters). In the 17 kilometers south of the fjord mouth (shown as Figure 21; approximately from the fjord mouth to Eidembukta) significantly shallower areas are conspicuous. The water depth in the northernmost 17 kilometers of the Forlandsundet portion of the multibeam survey rarely exceeds 200 meters, however from 8 to 17 kilometers from the fjord mouth water depths are as shallow as 165 meters. This 35-meter high smooth-sided feature appears to be part of something larger; three distinct lobes are represented on the multibeam survey two of which appear to be terminating toward the
Acoustic Results: Multibeam

west and northwest, one of which continues further into the sound. The northermost lobe is twice as steep (steepness up to 3.5) on the southwestern flank than on the northwestern flank (steepness less than 1.5). The southernmost lobe appears to be roughly symmetrical, with steep sides like the southwestern facing slope of the northern lobe. The middle lobe has an intermediate slope and is clearly shown to terminate in water depths of approximately 200 meters. Circular depressions up to 1.5 meters deep and 20 meters across are found both on and around this feature. They appear nearly identical to those found elsewhere throughout St Jonsfjorden.

Two additional features rising approximately 40 meters from the seafloor (from 210 to 170 meters water depth) are present in the 8 kilometers between the fjord mouth (described above as where the water depth increases suddenly from 160 to 220 meters) and the shoal’s inception (shown in Figure 21). The southern of the two is oriented northnorthwest-southsoutheast and is approximately 420 meters wide and slightly curvilinear. Only a 400-meter by 600-meter portion of the northern feature is visible in the survey area; this feature

Figure 21: Landforms in Forlandsundet directly south of the mouth of St Jonsfjorden
Acoustic Results: Multibeam

appears to extend to the northwest. Both features are steep sided (steepness between 3 and 6).

In this area between the fjord mouth and the shoal and surrounding the two 40-meter features, ridges similar to the fjord-parallel ridges observed in and described from the inner fjord are present. Due to the limited width of the area of Forlandsundet that was mapped, it is not possible to say how long these features are. However, they are up to 140 meters wide and three meters tall. These are oriented roughly north-south. Much like the linear features of the fjord head and those of the middle fjord, these ridges begin or end at (on) larger features (the 40-meter high ridges); in Figure 20 these curvilinear features are marked in rink and a representative profile through three such ridges is given as B-B'. They are up to 100 meters wide and 2-4 meters high.
Seismic Profiles
Throughout much of the fjord the seafloor reflection is underlain by a strongly stratified seismoacoustic unit up to 0.02 seconds thick (two-way travel time), which generally increases in thickness in the outfjord direction and is absent on steep slopes, where the seafloor horizon is synonymous with the upper boundary of the acoustic basement. Where it does not overlie the basement directly, this uppermost, strongly stratified seismostratigraphic unit often overlies a package of similar thickness where reflections are of a similar high amplitude but are much less continuous, bordering on chaotic. Within this unit, and between this and the overlying unit, acoustically transparent lenses are common, especially on slopes. These lenses can be up to 0.006 seconds thick and pinch out in both lateral directions. One such prominent lens on the western flank of the broad shallow ridge measures up to 932 meters in the long-fjord direction.

In various locations throughout the fjord clusters of parabolic (convex up) artefacts are observable. The locations of these artefact clusters are roughly coincident with the locations of the groups of 2m-high transverse ridges described above. The lower boundaries of these artefacts are often at the seafloor horizon, however, they can also be associated with lower, high amplitude reflections.
Acoustic Results: Chirp
The Inner Fjord

In the inner fjord, five major zones of positive relief (described in the bathymetry section) are identifiable in at least one of the seismic lines (Figure 22). The uppermost (strongly stratified) seismostratigraphic unit is approximately 0.012-0.014 seconds thick (all thickness in this section are presented as two-way travel times). It is visible in all areas where the seafloor appears smooth on the swath bathymetry; in highly rugose areas of the seafloor this unit may be obscured by distortion related to this rugosity or it may be absent.

Between these ridges, basins house weakly-stratified-to-acoustically-transparent units of variable thickness, sometimes divided by one or more continuous, high amplitude reflectors; these inter-ridge units can locally exceed 0.04 seconds in thickness and are capped by the same, laterally-continuous strongly stratified seismostratigraphic unit found throughout the fjord.

Beneath the uppermost seismostratigraphic unit, an acoustically transparent unit underlies the smooth-bottomed basin which separates zones A and C. In seismic line 4 this unit overlies a feature that may be the southward extension of zone B.

In addition to the five zones of positive relief described in the Bathymetry section, an additional prominent ridge can be identified on the seismic profiles, at the base of the broad shallow ridge’s eastern flank. This ridge is exposed at the surface in Line 1 and appears buried under the uppermost strongly stratified unit in Line 4.

Figure 22: pages 70 and 71 show five inlines along the length of the inner fjord, the location of each is shown in the inset on page 71.
Acoustic Results: Chirp
**Broad Shallow Ridge**

At the ridge crest, stratified low amplitude reflections conformably overlie the acoustic basement (marked in blue on Figure 23). The distance from the acoustic basement to the sediment-water interface rarely exceeds 0.050 seconds. The seafloor horizon, marked in red, is irregular at the ridge crest, as earlier described in the multibeam dataset. The acoustic basement is undulatory. Positive features under 0.0005 seconds twtt may be rims around the circular depressions described earlier, or may be independent positive features.

![Figure 23: Seismic line 8 as it crosses the broad, shallow ridge between Konowfjellet and Piriepynten. Acoustic basement is marked in blue; seafloor is marked in red. Location of the extent of the seismic line is shown as an inset in the bottom left of the figure.](image)
Shoals

Chirp line 2 bisects the southernmost section of the ridge while line 6 bisects the shallower, middle section. Both lines show that the ridge is almost entirely acoustically transparent, save for an overlying package of poorly organized, chaotic reflections less than 0.005 seconds thick (two way travel time). This is absent on the ridge’s western flank in seismic line 6, where the top of the acoustic basement is also the seafloor. The thin package of chaotic reflections is interpreted as a thin cover of unconsolidated sediments. Weak internal reflectors are present on the lower, gentler slopes, where the sediment package is thickest. At its thinnest, the sediment cover shown in Line 2 is thinner than 0.002 s (twtt), where the seafloor is shown as red and the acoustic basement is marked in blue.

Figure 24: Seismic line 2 as it passes through shoal in the outer fjord. Acoustic Basement is marked in Green. Seafloor is marked in Red. Location of the line is shown as an inset in the bottom left of the figure.
Interpretation of Acoustic Results

Interpretation

As the chirp sub-bottom profiler cannot penetrate especially dense material, the acoustic basement in St Jonsfjorden is likely either bedrock or glacially overridden (consolidated) sediments (Forwick & Vorren 2010). All reflections and acoustically transparent seismoacoustic units overlying this basement are therefore assumed to be unconsolidated sediments. If fast-moving ice had flowed through St Jonsfjorden surrounding the Last Glacial Maximum, sediments previously existing in the fjord were likely transported out of the fjord (compare with Elverhøi 1995 and Forwick & Vorren 2010). Sediments overlying the basement are therefore interpreted as being emplaced during or after the Late Weichselian glaciation.

Shoals

Because of the limited extent of the multibeam dataset presented in this study, it is difficult to resolve the shapes of the shoals, or even to confirm that they are two independent shoals, as they are presented in Norwegian Polar Service maps. Król et al (2010) presents the results of a single-beam bathymetric survey performed by the Norwegian Hydrographic Service; this dataset was briefly described in the introduction (physiography section) of this thesis.

These shoals have previously been interpreted as being related to tectonic movements along the West Spitsbergen Fold-Thrust Belt by Forwick et al (2014). This interpretation is supported by the apparent agreement between the strikes of many West Spitsbergen Fold-Thrust faults south of the fjord (Bergh et al 2003) and the orientation of the St. Jonsgrunnen-Farmgrunnen ridge (see Figure 25). The acoustically transparent core of the ridge (acoustic basement) is interpreted as being a bedrock ridge overlain by a thin cover of unconsolidated sediments.

Thin-to-nonexistent sediment packages on the upper, steepest slopes may indicate that these slopes are too steep for unconsolidated sediments to accumulate. Internal reflectors within the lower, thicker sediment packages may be related to the downwasting of sediments from these steep slopes.
Beyond the shoals and in Forlandsundet, Król et al (2010) note no ‘evidence of glacial relief.’ In the multibeam data presented in this thesis, however, streamlined landforms are identified exiting the fjord, west of the shoals; these are interpreted as being of a glacial origin on the basis of their marked similarity to landforms within St Jonsfjorden.

Figure 25: Faults in the area surrounding St Jonsfjorden (mapped from data published at npolar.no, 2014)

Pockmarks and Rocks

Circular depressions on the shoal crest, as seen in the multibeam data, may be pockmarks related to fluid escape from the underlying sedimentary rock. Pockmarks are common in Spitsbergen fjords and are interpreted as being the surface expressions of fluid moving upward through the sediments and entering the water column (Forwick et al 2009). These pockmarks likely represent the movement of water through the sediments, as opposed to gas (Morten Often, personal communication).

Seemingly positive features noted on the Chirp seafloor horizon may be rims around these pockmarks, or may be rocks, possibly of an ice rafted origin. Furthermore, the possibility that these features are artefacts – clusters of bad picks overlooked in processesing – cannot be ruled out. These features are less than 40 cm high, the measured apparent height of linear artefact traces in the inner fjord.
**Interpretation of Acoustic Results**

**Bedrock-controlled end moraine**

On the basis of single-bean echosounder data, the broad, shallow ridge between Konowfjellet and Piriepynten has been interpreted (by Król et al 2010) as the subsea extension of moraines found on either side of the fjord. These subaerial moraines were previously interpreted by Preisner (1988) as delimiting the LIA maximum extent of the Osbornebreen-Konowbreen Glacier Complex (Preisner 1988 in Król et al 2010). The moraine complex at Piriepynten (southern landward extension) is described by Evans and Rea (2005) as comprising glaciotectonized muds.

**Subglacial landform assemblages**

That the orientations of areas of high relief in the inner fjord are roughly congruent with the orientations of faults mapped terrestrially in the catchment of inner St Jonsfjorden suggests that these features are at least partially bedrock-controlled.

The landform assemblage in the head of St Jonsfjorden is similar to those found in other Svalbard fjord heads, especially those through which glaciers have repeatedly advanced and retreated (e.g., Flink et al 2015; Ingolfsson et al 2011; Ottesen & Dowdeswell 2006; Plassen et al 2004) including thick and high transverse and fjord-parallel ridges – herein interpreted as morainal banks and streamlined subglacial bedforms – overlain by thin and low sinuous transverse ridges, interpreted as annual retreat moraines. Streamlined subglacial bedforms in the inner fjord are likely reworked sediments deposited, possibly as morainal banks – during and after the last glacial. This type of reworking has been previously suggested for Tempelfjorden (Flink et al 2015) to be associated with surging glaciers but is herein suggested to be related to advance during the Little Ice Age (see Discussion – Glacial History of St Jonsfjorden).

Glacial landforms in the outer fjord resemble those of the inner fjord. Transverse ridges may be bedrock or sedimentary but are suggested to be fixed bedrock; as ice moved over these ridges the protuberances may have initiated low pressure cavities behind which attracted soft sediments, creating the streamlined forms which are conspicuously articulated with transverse ridges (Benn & Evans 2010; Hart 1997).
Figure 26: above: x-radiographs, color photographs and lithologic logs of the four cores. X-radiograph is shown as a positive image for HH12-953 (densest areas white) while x-radiographs of the other cores are shown as negative images, in which the densest areas appear darkest, except for in the top section of gravity core HH12-953, which is shown as a positive image. Below: locations of the four gravity cores.
Gravity Core HH12-953

Gravity core HH12-953 was taken from an even patch of seafloor 82 meters deep, located west of zone five described in the acoustic results section (see location map given in Figure 26). The core is 346 centimeters long and comprised entirely of clayey silt with included lithified clasts (see log, x-radiographs, color photographs and grain size distribution in Figure 27 and additional grain size information in Figure 28). Two distinct muddy units – one finely and strongly stratified (953-3) and one weakly stratified (953-2) – overlie a dense diamict (953-1). The contact between the upper two layers occurs gradually between 218 and 228 cm. The lowermost unit is a clast-rich but matrix-supported diamict with large clasts. The grain size distributions of all three units, excluding clasts greater than 2 mm, are shown in Figure 28.

This core contains no bioturbation or visible fossils. Sediments in this core become more compacted downcore, showing steadily increasing density and decreasing fractional porosity, especially in units 953-1 and 953-2. This is shown in the log presented in Figure 27. Geochemical data, presented as Figure 30, shows disagreement between data obtained in spring 2014 and autumn 2014.

Description of Units

Unit 953-1: 346 – 308 cm

Unit 953-1 is a matrix-supported diamict with a clayey silt matrix. The clast content of this lower unit increases markedly beneath 290 cm, beneath 308 cm downcore it is locally clast supported. The lowermost 38 centimeters of the core are comprised of a dense (2.07 grams per cubic centimeter average), firm and clast-rich diamict. The matrix of this diamict is neither sticky nor plastic. Two distinct clast pavements exist at 316 and 330 centimeters downcore.

Figure 27 (next page): Physical properties of gravity core HH12-953
HH12-953-GC-MF

- 953-1
- 953-2
- 953-3

Lithologic Results

Volume % Sand

Volume % Clay

Volume % Silt

Clay

Sand

Silty Clay

Sand-Silt-Clay

Clayey Sand

Silty Silt

Sandy Silt

Sandy Sand

Volume % Silt

0 25 50 75 100

0 25 50 75 100

sandy horizons
Figure 28 (opposite): Grain size distribution of 131 samples from gravity core HH12-953, obtained via Beckman Coulter Laser Particle Size Analyzer. Sandy horizons in unit 953-3 are indicated by arrows.

Unit 953-2: 308 – 228 cm

In the lower mud unit, dense dark clayey silt contains lithified clasts as weak pavements, clusters, and scattered, individual clasts. The clast content of this unit as identified in the x-radiographs is not as high as the diamict it overlies, but is significantly higher than that of unit 953-3. This unit is weakly stratified and has an average density of 2.03 grams per cubic centimeter. Staining is present on the surface of the core, which may indicate decaying organic matter. The density of the matrix decreases in the up-core direction from the lower unit boundary at 308 cm as the consistence becomes less firm. Units 953-2 and 953-1 have high magnetic susceptibilities, which peak at 262.6 cm downcore (144.3 x10^5 SI; see Figure 27), a clast rich interval (gravel pavement). Because the entire unit is weakly stratified, the upper boundary of this unit is difficult to determine. On the interval 190 – 228 cm bands of sediments up to 8 cm thick resemble the sediments of 953-2 and the overlying stratified mud unit 953-3. These bands are visible on both the split core surface and x-radiographs. This unit likely represents a gradual transition from 953-2 depositional conditions to 953-3 depositional conditions. Somewhat arbitrarily, these transitional sediments are assigned to unit 953-3, described below.

A diagonal mark on the color photographs between 270 and 280 cm downcore is a gouge created when the knife dragged a 4 cm clast upcore from 280 to 270 cm during the opening of the cores.

Unit 953-3: 228 – 0 cm

In the upper unit, stratification is between one millimeter and three centimeters thick. The layers are greyish brown, bluish grey, reddish brown, and grey, with infrequent dark grey sandy horizons up to 2 cm thick. Boundaries between the laminations are generally sharp as opposed to gradational. Many of the laminations are variable in thickness and some pinch out visibly.

The sandy horizons can have gradational but irregular boundaries and veins of the underlying/overlying mud can be seen to cut into the sandy horizons (see example
presented as Figure 29). Eight such sandy horizons can be identified on the core surface; two – those at 64-66mm and 428-431 mm – are indicated with arrows on the ternary diagram presented as Figure 28.

![Image](image.png)

**Figure 29: Sandy horizon at approximately 60-61 cm downcore [HH12-953-GC unit 3]**

Dark, lithified clasts up to 6 mm across are present throughout the laminated unit both as weak pavements and as scattered individual clasts. The unit has an average density of 1.80 g/cc.

The transition from the upper to the lower unit is not sudden, and occurs over the interval from 191 to 228 centimeters down core. Beneath 190 cm, stratification becomes stronger and magnetic susceptibility begins to increase at a faster rate. Between 218 and 228 there are four thin (1-4 cm) layers, which alternate between the laminated clayey silts of the upper unit (953-3) and the denser, massive clayey silts of the lower unit, 953-2.

Throughout unit 953-3, the June geochemical data shows peaks in Silicon, Potassium and Aluminum corresponding to troughs in the Iron trend. This pattern is not visible in the November data. Furthermore, Silicon and Aluminum vary together with a distinct sawtooth pattern. Geochemical data is presented in Figure 30.
Figure 30: Element geochemistry of the gravity core HH12-953 displayed as ratios of individual elements to the sum of the most common elements (top) and as ratios of individual elements to each other (bottom). Grey line (infilled) indicates measurements taken in June 2014 (10 mm downcore resolution) and data is shown as a three-point running mean. Red line indicates measurements taken in November 2014 (2 mm downcore resolution) and data is shown as an eleven-point running mean. Sand percentages are also shown on the plot of geochemical ratios, as it simplifies comparisons between grain size and geochemistry. In both halves of the figure lithologic units are indicated by the background color, where pink indicates unit 953-1, purple indicates 953-2 and green indicates 953-3. Silica=Si. Aluminum=Al. Iron=Fe. Calcium=Ca. Potassium=K. Titanium=Ti and Zircon=Zr. Sand percent is shown as a volume percent; this data comes from the grain size analysis done using the Beckman Coulter Laser Particle Size Analyzer described earlier.
Chronology

Ten samples have been submitted for lead dating at depths of 0-2 cm, 10-12 cm, 20-22 cm, 50-52 cm, 100-102 cm, 150-152 cm, 200-202 cm, 250-252 cm, 300-302 cm and 345-347 cm. Only the top five samples were analyzed, and the measured activities are presented in Table 7 and in Figure 32. Age estimates, shown in Figure 31 represent sedimentation at one centimeter per year, the minimum sedimentation rate proposed by Witold Szczuciński (personal communications).

The first occurrence of cesium in the atmosphere is generally taken to have been in the early 1950s and therefore the lack of activity in the 100-102 cm sample can be interpreted as an indication that these sediments were deposited prior to this point (Witold Szczuciński, personal communication). Excess $^{210}$Pb decreases steadily through at least the top hundred centimeters but is absent in the 150-152 cm sample. In samples with low initial surface
activity (less than 200Bq per kilogram) the detectable limit for excess $^{210}$Pb is suggested by Appleby (2001) to be at sediments deposited approximately 130 years prior to sampling. The presence of excess $^{210}$Pb in the sample 100-102 cm downcore indicates that these sediments were deposited in the 130 years prior to sampling (post-1882).

![Graphs of $^{210}$Pb and $^{137}$Cs activities](image)

**Figure 32:** Measured $^{210}$Pb (left) and $^{137}$Cs (right) activities in gravity core HH12-953, with error bars. This figure was originally created by Witold Szczuciński and later modified for clarity in print form.

<table>
<thead>
<tr>
<th>sample depth [cm]</th>
<th>lab No.</th>
<th>$^{137}$Cs Bq/g</th>
<th>uncertainty</th>
<th>$^{210}$Pb Bq/g</th>
<th>uncertainty</th>
<th>ex $^{210}$Pb Bq/g</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-2</td>
<td>IG0979</td>
<td>0.00225</td>
<td>0.00096</td>
<td>0.07912</td>
<td>0.01831</td>
<td>0.03948</td>
</tr>
<tr>
<td>10-12</td>
<td>IG0980</td>
<td>0.00255</td>
<td>0.00117</td>
<td>0.07658</td>
<td>0.01761</td>
<td>0.03695</td>
</tr>
<tr>
<td>20-22</td>
<td>IG0981</td>
<td>0.00172</td>
<td>0.00090</td>
<td>0.06815</td>
<td>0.01763</td>
<td>0.02851</td>
</tr>
<tr>
<td>50-52</td>
<td>IG0982</td>
<td>0.00463</td>
<td>0.00118</td>
<td>0.06710</td>
<td>0.01606</td>
<td>0.02746</td>
</tr>
<tr>
<td>100-102</td>
<td>IG0983</td>
<td>0</td>
<td>0.05760</td>
<td>0.01750</td>
<td>0.01796</td>
<td></td>
</tr>
</tbody>
</table>

**Table 7 (Above):** Measured activities of $^{210}$Pb and $^{137}$Cs and determined excess (unsupported) $^{210}$Pb activity determined by subtracting the amount of supported lead activity – that determined to have been produced by the in situ decay of $^{226}$Ra – from the measured activity in the sample. Witold Szczuciński performed these measurements and created this table.
Gravity Core HH12-954

Gravity core HH12-954 was taken just east of the crest of the broad shallow ridge at a water depth of 37 meters (location: Figure 26). The core is composed largely of thinly bedded fine sediments (clayey silt) with lithified clasts throughout (Figure 33, Figure 34). These clasts are up to ten centimeters and, throughout most of the core, appear to be distributed randomly, however distinct horizons of sand and gravel are not uncommon and there are some intervals upon which clasts are conspicuously absent. The thin beds and laminae are often mildly sub-parallel and both upper and lower contacts are most often gradational. This core contains no bioturbation or visible fossils.

Figure 33 (above): Ternary diagram showing the grain size distribution of 63 samples from the gravity core HH12-954. Samples from the matrices of the diamicts in unit 954-2 are marked with arrows.

Figure 34 (opposite): Physical properties and logs of gravity core HH12-954
Lithologic Results

HH12-GC-MF

Grain Size Distribution
(Cumulative % by volume)

Wet Bulk Density
(ρw[m3/cm3])

Acoustic Impedance
([m/s, ρg/cm3])

Fractional Porosity

Volume Specific Magnetic Susceptibility
([SI x 10^-5])

Primary Wave Velocity [m/s]

954-3

954-2

954-1
**Description of Units**

**Unit 954-1: 356 – 272 cm**

Unit 954-1 is a stiff, coarse and clast rich interval existing from 272 cm downcore to the core's bottom at 356 cm. The matrix of 954-1 is considerably sandier and stiffer than that of the overlying units (Figure 33). Furthermore, it is structureless (Figure 34). Clast content increases in the downcore direction, as does the general size of the clasts. Many of these clasts are light colored, sub-round, and smaller than 6 mm. Most, however are dark grey and larger than 8 mm.

Horizontal, evenly spaced lines present on the x-radiographs in the core's fourth section are artefacts.

**Unit 954-2: 272 – 211 cm**

Above 272 cm, weakly stratified light olive brown (2.5Y5/4) clayey silts are separated by two significant diamict intervals, approximately 211-224 and 250-260 cm downcore, both of which can be seen sloping downward to the left in the color photos and x-radiographs. Lenses of brownish-yellow clay are common within the lamina, and are often found in clusters. A distinctive trend of two-three broad peaks can be seen in the K:Ti, Zr:Rb & Ca:Sr element ratio data, those these peaks are not in phase with each other (Figure 35).

The diamict intervals are massive and contain lithified clasts up to 7 cm supported by a stiff clayey silt matrix that is, on average, coarser than the clayey silt of unit 954-3 but less coarse than that of unit 954-1. In the grain size data, only the diamict matrix was measured in the laser particle size analyser. The matrix of the lower diamict interval (sample from 255-256 cm downcore) is over 90% sand by volume while the matrix of the upper diamict interval (sample from 221-222 cm) is approximately 20% sand (see Figure 33, diamict matrices are indicated with arrows).

In addition to the lithified clasts that occur in the diamict units, solitary lithified clasts are distributed throughout the stratified sections, however, they are notably absent from the uppermost eighteen centimeters of this unit. Within the stratified intervals two anomalous beds are immediately visible: a 4 mm bed of light olive brown clay at ten centimeters downcore and an 8 mm bed of light reddish brown clay at 48 cm downcore.
**Unit 954-3: 211 – 0 cm**

The uppermost 211 centimeters of gravity core HH12-954 contain stratified clayey silt with a variable clast content. Stratification is visible both at the core surface and in the digital x-radiographs. This unit is visually very similar to the uppermost unit of gravity core HH12-953.

Grayish-brown silt (10YR5/2) dominates the uppermost ten centimeters of the core, and includes within it a 6-millimeter horizon of bluish grey sand (Munsell Color 5B6/1) at a depth of 11-17 mm. Lithified clasts are scattered throughout these top ten centimeters and a weak pavement exists at ten centimeters down core.

Beneath the pavement, and continuing down core to 211 centimeters, the clayey silts are laminated-to-thinly bedded (less than three centimeters). Like those of Unit 953-3, these thin beds are greyish-brown (2.5Y5/1), grey (2.5Y6/1) and bluish grey and have indistinct boundaries. Weak pavements of lithified clasts are present at 55, 63, and 85 centimeters down core while dispersed lithified clasts begin beneath 81 cm down core. A distinct cluster of lithified clasts also appears (on x-radiographs; Figure 34) to be present beneath the pavement at 63 centimeters.

Between 100 and 131 centimeters down core, dispersed lithified clasts are absent. Beneath 131 cm lithified clasts less than five millimeters along their longest axis, dark grey, and platy are scattered throughout the sediments. Two anomalously large lithified clasts are found at 50 cm down core, one being sub-round and oblong, 9.5 cm oriented in the cross-core direction and 3.5 cm in the downcore direction, and the other being a sub-round subsphere of diameter 1.8 cm. In addition to lithified clasts, between 54 and 66 centimeters down core stiff, dark grey unlithified lenses of sediments can be observed. These soft sediment clasts contain coarser grains than those they are surrounded by, predominantly fine sand but also distributed very coarse sand grains. Two additional soft sediment clasts are located at 76 and 77 centimeters down core.

Figure 35 (next two pages): Element geochemical (XRF) data from the gravity core HH12-954 presented as a ratio of certain elements to the sum of the six most common elements (top) and as element ratios between two measured elements (bottom). Red line indicates measurements performed in November 2014 while the measurements performed in June are plotted in grey.
**Chronology**

Ten samples have been submitted for lead dating at the following depth intervals: 0-2 cm; 10-12 cm; 20-22 cm; 50-52 cm; 100-102 cm; 150-152 cm; 200-202 cm; 250-252 cm; 327-331.5 cm and 350-352 cm. Activities measured in the top four samples are presented in Table 9 and plotted in Figure 37; estimated sedimentation rates are shown alongside Figure 36.

The presence of excess lead in the uppermost three samples indicates that these sediments were deposited within approximately the past century. The unsteady decrease of lead activity with depth in the top 20 cm of the core was suggested by Witold Szczuciński (personal communication) to represent higher sedimentation rates on the interval 10-20 cm downcore than in the top ten centimeters; a sedimentation rate of 0.2-0.4 centimeters per year is suggested for the upper portion of the core (Witold Szczuciński, personal communication). The presence of $^{137}$Cs activity at 10-12 cm downcore but not...

*Figure 36: Lead decay derived (linear) sedimentation rates and estimated minimum ages of sediments in gravity core HH12-954. The radiocarbon date obtained from three benthic foraminifera is also shown, although these organisms can reasonably be expected to have been redeposited and therefore no sedimentation rate has been calculated from this age.*
at 20-22 cm downcore indicates that the introduction of Cesium to the atmosphere, in 1952, occurred somewhere on this interval. This is consistent with the average sedimentation rate estimated for the upper portion of the core. In Figure 36, dates inferred from the measured samples are shown, in brown, as a range, where the younger age is calculated using the higher end of the proposed sedimentation rate range and the older age is calculated using the low end of the proposed sedimentation rate range. Beneath 50 cm, very rough, rounded estimates of the depositional age are shown in grey; these estimates assume that the upper core sedimentation rate range is valid along the entire uppermost unit, 954-3. These age estimations were determined using the sedimentation rate range midpoint, 0.3 centimeters per year, and are presented alongside the number of years one would need to add or subtract to obtain an age estimate based on the maximum or minimum sedimentation rate proposed, respectively.

One sample, containing three benthic foraminiferal tests, was submitted for AMS $^{14}$C dating. The details of this date are presented in Table 8:

Because of its small size, this sample was not leached using HCl before measurement.

<table>
<thead>
<tr>
<th>Depth (cm) downcore</th>
<th>$^{14}$C year</th>
<th>Laboratory number</th>
<th>Material</th>
<th>$\delta^{13}$C</th>
<th>Age BP ($\sigma_1$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>267-268 cm</td>
<td>2505 ± 85</td>
<td>ETH 60275.1.1</td>
<td>0.0005 grams</td>
<td>19 per mil</td>
<td>2030 ± 110</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>N. labradorica &amp; C. lobatulus</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 37: Measured $^{210}\text{Pb}$ (left) and $^{137}\text{Cs}$ (right) activities in gravity core HH12-954, with error bars. This figure was originally created by Witold Szczuciński and later modified for clarity in print form.

<table>
<thead>
<tr>
<th>sample depth [cm]</th>
<th>lab No.</th>
<th>$^{137}\text{Cs}$ Bq/g</th>
<th>uncertainty Bq/g</th>
<th>$^{210}\text{Pb}$ Bq/g</th>
<th>uncertainty Bq/g</th>
<th>ex $^{210}\text{Pb}$ Bq/g</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-2</td>
<td>IG0984</td>
<td>0.00492</td>
<td>0.00149</td>
<td>0.06712</td>
<td>0.01880</td>
<td>0.02748</td>
</tr>
<tr>
<td>10-12</td>
<td>IG0985</td>
<td>0.00364</td>
<td>0.00113</td>
<td>0.05331</td>
<td>0.01462</td>
<td>0.01368</td>
</tr>
<tr>
<td>20-22</td>
<td>IG0986</td>
<td>0</td>
<td></td>
<td>0.05434</td>
<td>0.01778</td>
<td>0.01470</td>
</tr>
<tr>
<td>50-52</td>
<td>IG0987</td>
<td>0</td>
<td></td>
<td>0.03964</td>
<td>0.01182</td>
<td>0</td>
</tr>
</tbody>
</table>

Table 9: Measured activities of $^{210}\text{Pb}$ and $^{137}\text{Cs}$ and determined excess (unsupported) $^{210}\text{Pb}$ activity determined by subtracting the amount of supported lead activity – that determined to have been produced by the in situ decay of $^{226}\text{Ra}$ – from the measured activity in the sample. This table was created by Witold Szczuciński.
Gravity Core HH12-955

Description of Sediments

*HH12-955: 254 – 0 cm*

Gravity core HH12-955 is considered here as one unit. All included sediments are clayey silt (Figure 38) and the grain size distribution remains nearly constant on the 255 centimeters of the core’s length. Weak changes in sediment color (grey and brown mud) on the centimeter scale are not significant enough to warrant classification as individual units and no substantial changes in the density of the clayey silt matrix can be seen in the x-radiographs (Figure 39).

Lithified clasts are visible in the x-radiographs on the entire length of the core, though their concentration is variable. Two prominent clast-supported pavements, from 81-86 cm downcore and from 179-186 cm downcore, mark changes in the general trend of clast concentration. Beneath 186 cm clast concentration is highest, on the interval 86-179 clasts are rare, and above 81 cm clasts concentration is intermediate and clasts are organized as weak bands approximately 5-8 cm thick. Foraminifera were dated from beneath these clast-supported pavements, yielding $^{14}$C ages of $1743 \pm 85$ and $1316 \pm 85$, respectively; the details of these dates are presented in Table 10.

In the geochemical data (shown as Figure 40) local maxima in iron and zirconium may be related to local minima in calcium and potassium.
Figure 38 (above): Ternary diagram showing the grain size distribution of 87 samples from gravity core HH12-955 analyzed using the Beckman Coulter Laser Particle Size Analyzer

Figure 39 (opposite): Physical properties and logs of gravity core HH12-955

Figure 40 pages 100 & 101): Element geochemical (XRF) data from the gravity core HH12-955 presented as a ratio of certain elements to the sum of the six most common elements (pg100) and as element ratios between two measured elements (pg 101).
Chronology

Four monospecific samples of benthic foraminifera taken from three 1-cm slices of gravity core HH12-955 were submitted for AMS $^{14}$C dating; details of these dates are presented as Table 10. Samples from 86-87 cm and 186-187 cm are taken from directly beneath the two most prominent clast pavements in this core. Two unique species ($C$. Lobatulus and $N$. labradorica) were submitted from the interval 246-247 cm in order to compare and evaluate the resultant dates.

<table>
<thead>
<tr>
<th>Depth (cm) downcore</th>
<th>$^{14}$C year</th>
<th>Laboratory number</th>
<th>Sample Weight</th>
<th>$\delta^{13}$C</th>
<th>Age BP ($\sigma_1$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>86-87 cm</td>
<td>1743 ± 85</td>
<td>ETH 60271.1.1</td>
<td>0.002 g</td>
<td>5.2 per mil</td>
<td>1200 ± 65</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$N$. labradorica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>186-187 cm</td>
<td>1316 ± 85</td>
<td>ETH 60272.1.1</td>
<td>0.002 g</td>
<td>3.5 per mil</td>
<td>750 ± 70</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$N$. labradorica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>246-247 cm</td>
<td>2373 ± 85</td>
<td>ETH 60273.1.1</td>
<td>0.0019 g</td>
<td>3.1 per mil</td>
<td>1880 ± 85</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$N$. labradorica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>246-247 cm</td>
<td>2564 ± 85</td>
<td>ETH 60274.1.1</td>
<td>0.0007 g</td>
<td>9.7 per mil</td>
<td>3000 ±100</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$C$. lobatulus</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 10: Material submitted for radiocarbon dating from gravity core HH12-955

Obtained radiocarbon dates and inferred sedimentation rates are shown in Figure 41. Because sediments from 186-187 cm dated so much younger than those from 86-87 cm, sedimentation rates for deposition under three possible accumulation models are presented to the right of the core log.
Figure 41: Radiocarbon dates and inferred sedimentation rates in the gravity core HH12-955. Rates in majenta exclude the date from 186-187 cm while rates in light pink exclude the date from 86-87 cm; both are based on dates from _N. labradorica_ samples. The rate shown in blue was calculated using the date obtained from the _C. lobatulus_ sample at 246-247 cm assuming that sediments at the top of the core were deposited in 2012. All sedimentation rates assume linear sedimentation.
**Gravity Core HH12-956**

The outermost gravity core, HH12-956, comprises four units. Distributed single lithified clasts are present throughout the core, though they are variable in size and composition. Upon opening this core, the surface was grey with black mottles. The black coloration largely disappeared after the first day. However, when the core was sampled for radiocarbon dating material it was noted that the interior of the core (greater than 2-3 centimeters deep) remained black. A photograph of the surface of HH12-956, taken immediately after opening, is shown as Figure 42.

The element ratios recorded in this core (see Figure 45) show changing geochemistry from unit to unit. Notably, within the stratified silt units, high Fe, Ti and Zr in the lower stratified silt unit (956-2) relative to the upper stratified silt unit (956-4) leads to much higher Fe:Ca and Zr:Rb ratios in 956-2 relative to 956-4 and much a higher K:Ti throughout 956-4 than on 956-2.

![Figure 42: mottled surface of the gravity core HH12-956 immediately after opening. Dashed lines indicate lines produced by the plastic card during the cleaning of the core surface. The scale bar to the left indicates centimeters downcore.](image-url)
Description of Units

Unit 956-1: 351 – 318 cm downcore

Sediments at the bottom of gravity core 956 are coarse and dense diamict with a silt matrix (see grain size distribution, ternary diagram Figure 44). This unit has high calcium and low potassium content relative to the other major elements detected in the core (). The bottommost 17 cm (334-351 cm downcore) are the densest and coarsest, containing abundant lithified clasts. The remainder of this unit (uppermost 16 cm, from 334 to 318 cm) is distinctly layered. Each band of sediment (individually up to 7 cm thick) closely resembles either the sediments of the lower half of unit 956-1 or the sediments of overlying unit, 956-2, which are lighter, stratified, less dense and lack lithified clasts. Attempts to remove foraminifera from these lighter, stratified units were unsuccessful as no foraminifera were located on either the interval 329-330 or the interval 333-334 cm downcore. This upper half of unit 956-1 is therefore suggested to represent a transition between the depositional conditions of the lower part of 956-1 and that of 956-2.

Unit 956-2: 318 – 233 cm

Stratified clayey silts in unit 956-2 largely lack lithified clasts. The layers are olive grey (5Y5/2), bluish grey (5B5/1), grey (N6/) in color. These color variations are only visible on the surface of the split core surface and in x-radiographs; physical properties measurements and element geochemical data do not show stratification. In sampling for radiocarbon dating material, it was observed that sediments deeper than approximately 2 centimeters are uniformly black. Furthermore, the sediments increase markedly in water content and stickiness from the base to the top of the unit. The top of the unit is defined by a sharp and uneven contact, beneath which one monospecific sample of the benthic foraminiferal species Nonionellina labradorica taken for radiocarbon dating (234-235 cm). This sample yielded an age of 11,345 ± 102 14C years BP.

Figure 43 (opposite): Physical properties and logs of gravity core HH12-956
Figure 44: Ternary diagram showing the grain size distribution of 82 samples from gravity core HH12-956

**Unit 956-3: 233 - 170/172 cm**

Unit 956-3 is sandier and more red than the underlying sediments of unit 956-2 and of 956-4, above. Furthermore, unlike these other units, it appears structureless, although a slight fining upward trend is visible in the grain size distribution data (see log Figure 43).

At 233 cm downcore, the sharp and uneven base of unit 956-3 is marked by distinct peaks in sand content, magnetic susceptibility, and density. These trends decrease gradually in the upcore direction, ultimately stabilizing at approximately 172 cm downcore. In this unit, the ratio of potassium to titanium increases in the upcore direction from being low but relatively stable in units 956-1 and -2 to being high and relatively stable in unit 956-4.

From the sharp basal boundary to the upper gradational limit, sediments in this unit are noticeably sandier and redder than sediments elsewhere in the core. Distributed individual
lithified clasts and white flecks are visible on the split core surface. These white flecks were later determined to be shell fragments and foraminifera; no whole shells were located in any of the four cores.

The upper boundary of this unit is not easily identifiable by examination of the core surface, however, x-radiographs show a diamict interval at 165-172 cm downcore. The base of this diamict can be seen to slope from left to right on the interval 170-172 cm. This diamict is assigned to unit 956-4 and the unit boundary is placed at the lower limit of the diamict, 170-172 cm downcore. Immediately beneath this diamict, from the interval 174.5-175.5 cm downcore, monospecific samples of the benthic foraminiferal species *N. labradorica* and *C. lobatulus* were submitted for radiocarbon dating to determine the maximum age of the diamict’s emplacement. These samples yielded ages of 3,985 ± 65 14C years BP and 4,158 ± 64 14C years BP respectively (see Table 11).

*Unit 956-4: 170/172 – 0 cm*

The uppermost 170 cm of gravity core HH12-956 are dark grey clay [2.5Y4/1] with very weak stratification visible on the core surface but not in the x-radiographs. Relative to the other units, 956-4 is higher in calcium and lower in iron relative to the other major elements of the core *(Figure 45).*

Immediately overlying the diamict at this unit’s base, very few lithified clasts are present and sand contents are consistently less than 5% by volume. From 76 to 66 cm downcore a massive, matrix-supported diamict corresponds to a local peak in the ratio of iron to other major elements in the core. In the x-radiograph, the clasts that compose this diamict appear up to 4 cm. Above this diamict interval clast frequency increases markedly, however, the top ten centimeters of the core are conspicuously devoid of clasts. At 50-52 centimeters downcore a brown nodule surrounded by orange staining was uncovered. Along with the mottled appearance of the freshly opened core surface *(see Error! Reference source not found.)* I suggest that this is an indication of bioturbation.

*Figure 45 (pages 110 & 111):* Element geochemical (XRF) data from the gravity core HH12-956 presented as ratios of certain elements to the sum of the six most common elements (pg 110) and as element ratios between two measured elements (pg 111). Sand percentage (volume) is also shown on page 111.
Chronology

Five monospecific samples of benthic foraminifera from three 1-centimeter slices were submitted for AMS $^{14}$C dating, two of which are double samples, where two species were picked from the same interval. Details of the dated material and obtained date are presented in Table 11 while sedimentation rates derived from these dates are presented alongside the log in Figure 46.

<table>
<thead>
<tr>
<th>Depth (cm) downcore</th>
<th>$^{14}$C year</th>
<th>Laboratory number</th>
<th>Sample weight and species</th>
<th>Age BP ($\sigma_1$) (Calibrated)</th>
</tr>
</thead>
<tbody>
<tr>
<td>75.5-76.5</td>
<td>1738 ± 60</td>
<td>ETH 60266.1.1</td>
<td>0.0023 g</td>
<td>1200 ± 65 N. labradorica</td>
</tr>
<tr>
<td>75.5-76.5</td>
<td>2229 ± 60</td>
<td>ETH 60267.1.1</td>
<td>0.0026 g</td>
<td>1710 ± 85 C. lobatulus</td>
</tr>
<tr>
<td>174.5-175.5</td>
<td>3985 ± 65</td>
<td>ETH 60268.1.1</td>
<td>0.0015 g</td>
<td>3850 ± 100 N. labradorica</td>
</tr>
<tr>
<td>174.5-175.5</td>
<td>4158 ± 64</td>
<td>ETH 60269.1.1</td>
<td>0.0030 g</td>
<td>4070 ± 105 C. lobatulus</td>
</tr>
<tr>
<td>234-235</td>
<td>11345 ± 102</td>
<td>ETH 60270.1.1</td>
<td>0.0028 g</td>
<td>12720 ± 95 N. labradorica</td>
</tr>
</tbody>
</table>

Table 11: $^{14}$C ages are unrounded and uncalibrated while calibrated ages ($\sigma_1$) are rounded to the nearest ten years, as suggested by the authors of the Calib program.
Figure 46: Radiocarbon dates and inferred sedimentation rates in the gravity core HH12-956. Rates in blue indicate those calculated from *C. lobatulus* dates while those in pink indicate those calculated from *N. labradorica*. Dates in grey were calculated using the average of the two species. All sedimentation rates are calculated assuming linear sedimentation and the top of the core is assumed to have been deposited in 2012 AD.


Discussion

Sources of Error and Uncertainty
The precision and utility of results, especially grain size analyses, element geochemical qualifications and radiocarbon dates, presented within this thesis are affected by a variety of factors; a short explanation of their limitations and is given below.

Grain Size Analysis
Sediment grain size distributions as measured by the Beckman Coulter Laser Particle Size Analyzer may not accurately represent the sediments sampled, as there is no way of knowing if the sample maintains its distribution after multiple dilutions. Samples in this thesis were diluted up to 22 times each, with only six samples being diluted once or not at all. Sand was noted visually in many samples that did not register sand in the LS outputs. Furthermore, very coarse sand was noted to remain at the bottom of the sample input chamber after the rinsing process on multiple occasions. This may indicate that the current used may not have been sufficient to support very coarse sand, a prerequisite for measuring suspended samples.

The process by which sediment samples were prepared for analysis is not above reproach. Hydrochloric acid was used to remove foraminifera but may also have dissolved carbonate and metacarbonate silt and sand sized grains that should have been included in the analyses. The percent mass loss removed by HCl was used in identifying depths to sample to obtain foraminifera specimens for radiocarbon dating, however, no correlation was found between the amount of mass removed and the concentration of microfossils observed while picking foraminifera. Many samples identified as having medium-high mass loss percentages ultimately yielded fewer foraminiferal specimens than those samples that underwent low rates of attrition. It is important to note that only the grain size fraction greater than 125 micrometers was investigated; it is possible that high concentrations of foraminifera exist in the grain size fraction smaller than 125 microns accounting for the material dissolved by the HCl.
Radiocarbon Dating
Given the extreme age differences observed between ages obtained from *N. labradorica* and *C. lobatulus* specimens taken from the same sediment sample age models that combine the radiocarbon dates from both species are suspect. This age difference decreases with depth, from 510 cal years at 75.5-76.5 cm to 120 cal years at 174.5-175.5 cm, in gravity core HH12-956 (Figure 46) suggesting that the effects of compaction with depth decrease the apparent age difference. However, given that only two pairs of samples were submitted, this claim is feeble. Indeed, the greatest difference in calibrated age in a pair of depth-equivalent samples comes from the bottom of gravity core HH12-955 (1110 cal years, Figure 41) although no paired samples were submitted from above this interval for comparison.

Given the lifestyles of the two species, dates from epifaunal *C. lobatulus* are regarded as more reliable than those obtained from infaunal *N. labradorica* although it is not possible to say with any certainty that the *C. lobatulus* specimens are not redistributed by bottom currents without knowing the strength of such currents. Furthermore, *C. lobatulus* is known to inhabit high-energy environments (Hald & Korsun 1997; Klitgaard-Kristensen et al 2002 and references therein) and has been used (interpreted) as an indicator of strong bottom currents in the past (e.g., Holtegaard Nielsen et al 2010).

In gravity core HH12-955, foraminifera (*N. labradorica*) from 186-187 cm downcore returned radiocarbon dates younger than foraminifera (*N. labradorica*) from 86-87 cm downcore. It is not uncommon for older sediments to be reworked into younger deposits, making it easy to dismiss the older, stratigraphically higher age as invalid for this reason. The younger age, furthermore, fits the calibration curve-age model better (shown in Figure 47).

The rate obtained via exclusion of the uppermost date (lighter pink sedimentation rates reported in Figure 41) is therefore taken to be more reliable than the rate obtained discarding the young *N. labradorica* date at 186-187 cm (darker pink rate). However, all rates are approximately 2-4 times higher than the rate obtained using the single *C. lobatulus* date at 246-247 cm. This can suggest either that the submitted *C. lobatulus* sample incorporated older, redeposited specimens, that the depth to which *N. labradorica* can burrow causes it to significantly and consistently date too young, or both.
In gravity core HH12-956, the leached fraction of the sample submitted from 234-235 cm depth was very high (L. Wacker, personal communication), indicating possible contamination by younger carbonate material. Dating of additional samples deeper in the unit 956-2 is therefore suggested.

**X-Ray Fluorescence Scanning**

Core sections scanned twice, five months apart, yielded significantly different element ratios (see Figure 30 and Figure 35), especially in the light elements (e.g., silicon and aluminum) where trends can be seen locally to be reversed (former silicon and aluminum peaks now appear as troughs and vice versa; see Figure 30). No explanation for this is suggested beyond the matrix effects discussed in Tjallingii et al (2007) and Weltje & Tjallingii (2008) and possible contamination of the core surfaces despite diligent cleaning with a plastic card before both scannings.
Seismostratigraphy

Throughout much of the fjord, a thin package of stratified reflections exists at the top of the sediment column, directly underlying the seafloor reflection. In the inner fjord, this stratified unit directly overlies a shallow acoustic basement and likely represents sedimentation since retreat from the Little Ice Age Limit; this claim is supported from lead decay based sedimentation rates from two sediment gravity cores in the inner fjord (see Discussion: Glacial History of St Jonsfjorden).

In the outer fjord, beyond the shallow ridge running between Piriepynten and Konowfjellet, this stratified package is interpreted as glacimarine sedimentation deposited throughout the Holocene and is broadly correlative to the seismoacoustic units S5 (Early Holocene; weak stratification) and S6 (Mid to Late Holocene; distinct and continuous acoustic stratification) described by Forwick and Vorren (2011) based on observations from Isfjorden, 40 km south. In the outer fjord – and only in the outer fjord – S5 overlies thick stratified low amplitude locally discontinuous reflectors, which are herein correlated to the seismostratigraphic unit s4 of Forwick & Vorren (2011). As per Forwick & Vorren, the unit S4 is divided into subunits S4a and S4b, both of which comprise stratified low amplitude reflections which drape buried structures (the nature of this draping is indicated by a stippled line in Figure 48. Units S4, S5 and S6 are all interpreted as being deposited in a glacimarine environment.

Forwick & Vorren (2011) also report acoustically transparent lobes (termed s3 in Isfjorden) intercalated with their unit s4 on the outfjord facing slopes of high relief areas. Sediment wedges in outer St Jonsfjorden, however, have no preferred orientation, which is to say they face both outfjord in the direction of paleoice movement as well as infjord, in the direction of retreating ice. These transparent units are interpreted as mass transport deposits; lithostratigraphic unit 956-3 represents one such lobe (see Figure 52) and two stacked lobes are highlighted in Figure 48, in gold and green.
Figure 48: Seismic line 1, representative of the thick sedimentary packages of the outer fjord. Stippled lines indicate internal reflectors and show the semi-transparent drape of unit s4. Transparent lobes shown in green and yellow are interpreted as redeposited sediments from the steep slope pictured in the west of this profile. Both subfigures depict the same portion of Line 1, shown as an inset in the top subfigure.
**Correlation of sediment cores and acoustic data**

**Gravity core HH12-953**

Gravity core HH12-953 was taken from an even patch of seafloor 82 meters deep, located west of zone E described in the acoustic results section and shown again in Figure 49. An acoustically stratified package 0.04 seconds thick (two-way travel time) overlies both Zone E and a v-shaped transparent unit which separates this ridge from Zone F, located on the lower eastern flank of the broad shallow ridge which bisects the fjord between Konowfjellet and Piriepynten. The 0.04 seconds of stratified material are suggested to correlate with units 953-2 and 953-3, while the densest diamict of 953-1 most likely correlates to the underlying transparent seismostratigraphic unit. The stratification in unit 953-2 may contribute to the strong, high amplitude reflectors at the base of the acoustically stratified unit.
Figure 49: Location of gravity core HH12-953 relative to the fjord's bathymetry (above) and its penetration into the sediment column (below; indicated by yellow box). The acoustic basement beneath the transparent package is marked by a thin blue line between zones E and F. A thicker, turquoise line indicates the contact between the acoustically transparent and stratified sediments, interpreted as the contact between lithological units 953-3 and 953-2. The trace and extent of seismic profile 8 presented in the lower figure is indicated by a grey line the upper subfigure.
**Gravity core HH12-954**

Gravity core HH12-954 was taken just east of the crest of the broad shallow ridge at a water depth of 37 meters, as shown in Figure 50. West of the ridge crest a fairly well defined acoustically stratified package of reflections conformably overlies the acoustic basement. East of the ridge's apex, however, the material overlying the basement is thicker and more chaotic. Three distinct seismoacoustic packages can be identified above the acoustic basement: an uppermost weakly stratified unit and a lower unit with a more grainy and chaotic appearance and isolated sediment lobes that exist and pinch out between the two previously described seismostratigraphic units. Unit 954-1 may come from one such lobe, interpreted as having been transported from the inner fjord, while units 954-3 and -2 are interpreted as having been deposited in a glacimarine setting.
Figure 50: Location of gravity core HH12-954 relative to the fjord’s bathymetry (above) and its penetration (below) into the sediment column (indicated by yellow box). A white line on the upper subfigure indicates the trace and extent of seismic line 4 as depicted in the lower subfigure. At the core location the acoustic basement beneath the transparent package is marked by a thin blue line. Note that west of the ridge apex, a stratified acoustic unit overlies the basement directly.
Gravity core HH12-955

Gravity core HH12-955 was taken from the eastern side of a landform at a water depth of 88 meters. This core is correlated as representing entirely the uppermost, acoustically stratified package shown on the seismic profile in Figure 51 to increase in thickness to the right of the profile (in the downslope direction). All sediments contained in this core are expected to belong to the uppermost acoustically stratified package (above the green line in the seismic profile), which is expected to be over three meters thick based on the p-wave velocity as measured by the multi-sensor core logger. These sediments are younger than 3 ka cal BP (as per radiocarbon date ETH 60274.1.1, Table 10) and belong to seismostratigraphic unit S6. Beneath the penetration of the core, a weakly stratified low amplitude package overlies an acoustically transparent lobe emplaced atop a ridge suggested to be a bedrock feature, which initiated the streamlined subglacial bedforms seen to the west of the core site.
Figure 51: Location of gravity core HH12-955. All sediments in this core are interpreted as glacimarine sedimentation deposited as suspension rainout during the Holocene with a modest contribution from ice rafting.
**Gravity core HH12-956**

Gravity core HH12-956 was taken from a smooth patch of seafloor located northwest of the complex of shoals in the outer fjord as described in the acoustic results section. This core was recovered from a water depth of 111 meters, in the trench that separates the shoals from the northern fjord wall (location is given in Figure 52, top). As only one deep channel cuts through the fjord mouth, this trench is the most likely path for inflow of deeper water masses into the fjord from Forlandsundet and/or the shelf, should that occur.

In the chirp line (Figure 52, bottom), the uppermost .0012 seconds (twtt) of stratified high amplitude reflections correspond to the clast-rich upper interval of unit 956-4. The lower interval of 956-4 is represented by the underlying .0018 seconds – which is also stratified but of lower amplitude. An acoustically transparent wedge below this unit is correlated to unit 956-3. A high-amplitude reflection is suggested to correlate to the lower boundary of unit 956-3. The interval of weakly stratified sediments underlying unit 956-3 is interpreted to include both units 956-2 and 956-1 (indistinguishable in seismic line). Furthermore, because the EdgeTech 3300 is advertised as penetrating up to 6 meters in coarse sand and 80 meters in clay (EdgeTech 2015), this thick sediment package is expected to be fine.
Figure 52: Location of gravity core HH12-956 (upper) and its penetration into the sediments. Red line marks the seafloor horizon, also the top of seismostratigraphic unit 6. Yellow line marks the base of S5, also interpreted as the base of lithologic unit 956-4. Orange line marks the erosive base of a mass transport deposit (956-3), beneath which are lithologic units 956-2 and 956-1 and seismostratigraphic unit S4, interpreted as deposition in a glacial marine setting where the delivery of ice rafted sediments is high in 956-1 and low in 956-2.
Discussion

Chronology
Later in this chapter, radiocarbon dates obtained in this study are compared to earlier radiocarbon dates obtained by researchers who have previously investigated glacial episodes in St Jonsfjorden. These previous studies often present the ages of dated material as reservoir-corrected but uncalibrated years before present. In order to facilitate comparisons between sets of radiocarbon ages, and because of significant advances in radiocarbon calibration science, I have calibrated all samples using CALIB 7.1 (Stuiver et al 2014) and the marine calibration dataset. These dates are presented as Table 12. All dates are retrieved from *Mya truncata* shells or fragments of *Mya truncata* shells.
<table>
<thead>
<tr>
<th>Lab Number</th>
<th>$^{14}$C age BP</th>
<th>Reservoir correction</th>
<th>Location</th>
<th>Cal Age BP (CALIB)</th>
<th>Collection Elevation</th>
<th>Study</th>
</tr>
</thead>
<tbody>
<tr>
<td>DIC 3055</td>
<td>9,535 ± 90</td>
<td>-425 years</td>
<td>Sub-littoral sand Ankerbreen</td>
<td>9,110 ± 90</td>
<td>13.5 m</td>
<td>Forman (1989)</td>
</tr>
<tr>
<td>DIC 3056</td>
<td>8,265 +80/-90</td>
<td></td>
<td>Sub-littoral sand Piriepynten</td>
<td>7,840 ± 90</td>
<td>3.5 m</td>
<td></td>
</tr>
<tr>
<td>GU 8069</td>
<td>480 ± 60</td>
<td></td>
<td>Thrust silts west of Charlesbreen</td>
<td>Not calibrated because of young age</td>
<td>21 m</td>
<td></td>
</tr>
<tr>
<td>GU 8070</td>
<td>9,400 ± 70</td>
<td></td>
<td>Gravel bench, Gjertsenodden</td>
<td>8,960 ± 70</td>
<td>19.5 m</td>
<td></td>
</tr>
<tr>
<td>GU 8071</td>
<td>3,920 ± 50</td>
<td></td>
<td>Thrust silts west of Smalgangen</td>
<td>3,480 ± 50</td>
<td>13.5 m</td>
<td></td>
</tr>
<tr>
<td>GU 8072</td>
<td>9,690 ± 90</td>
<td></td>
<td>Foreset sands west of Bullbreen</td>
<td>9,250 ± 90</td>
<td>15.5 m</td>
<td>Evans &amp; Rea (2005)</td>
</tr>
<tr>
<td>GU 8073</td>
<td>10,040 ± 90</td>
<td>-440 years</td>
<td>Bottomset marine silts over bedrock west of Gjertsenodden</td>
<td>9,600 ± 90</td>
<td>13 m</td>
<td></td>
</tr>
<tr>
<td>GU 8074</td>
<td>9,910 ± 80</td>
<td></td>
<td>300 m in front of Bullbreen</td>
<td>9,470 ± 80</td>
<td>12 m</td>
<td></td>
</tr>
<tr>
<td>GU 8075</td>
<td>9,830 ± 80</td>
<td></td>
<td>Diamict surface west of Vestre Holmsletbreen</td>
<td>9,390 ± 80</td>
<td>19 m</td>
<td></td>
</tr>
<tr>
<td>GU 8076</td>
<td>9,120 ± 60</td>
<td></td>
<td>Diamict surface east of Bullbreen</td>
<td>8,680 ± 60</td>
<td>10 m</td>
<td></td>
</tr>
</tbody>
</table>

Table 12: Radiocarbon dates from previous studies, calibrated using CALIB 7.1 (Stuiver et al 2014)
Sediment sources and depositional processes

Suspension Fallout

The stratified clayey silts of units 956-4, 956-2, 954-3, 953-3 and the entirety of gravity core HH12-955 are all suggested to have been deposited primarily from suspension settling from meltwater plumes. The high density of seawater relative to that of the freshwater produced as glaciers melt is often enough to support meltwater plumes at the top of the water column. When plumes enter the water column above the seafloor the plume will sink to the seafloor or to a depth at which it is neutrally buoyant if the sediment concentration in the meltwater plume is high enough that the plume becomes sufficiently dense (Benn & Evans, 2010; Syvitski et al. 1987). Furthermore, even for very dense plumes, if the discharge of the stream is sufficiently high, the plume can continue to exist at the depth at which it enters the water column through the force of its momentum (Benn & Evans 2010). Suspension settling occurs when meltwater plumes mix with other water masses in the fjord, resulting in freshening of the ambient fjord waters and a loss of capacity by the plume, leading to the deposition of suspended sediment as it settles through the water column. The coarser, bedload-transported fraction of the meltwater-transported sediment is deposited in the immediate vicinity of the entrance point of the meltwater stream into the water column, often termed the efflux point. A study of modern sedimentation in Kongsfjorden determined that coarse sediments are almost exclusively deposited within four kilometers of the tidewater cliff (Elverhøi et al. 1983).

In the summer months, glacial meltwater and snowmelt deliver sediment-laden meltwater plumes to St Jonsfjorden both directly from tidewater glaciers and as proglacial streams; overflow can be seen on aerial photographs and satellite images taken most years between July and September. Figure 54 shows an overflow plume entering the fjord at its head (Figure 54-A) and traveling over nine kilometers west, in the outfjord direction. Although side glaciers contribute some sediment to the plume as it moves outfjord, the plume visibly becomes less concentrated as it travels further from the fjord head (compare subfigure B to subfigure A). This may suggest that, at least under modern conditions, sediments produced
at the fjord head account for a major fraction of those composing the meltwater plume and that as the plume moves into the outer fjord sediments settle out of the plume as it mixes with the surrounding, ambient water masses and loses momentum (an overview of this process is given in Benn & Evans 2010).

Stratified, fine sediments in fjords have previously been interpreted as cyclical changes in fjord hydrography and sediment supply either on seasonal or shorter timescales and Ó Cofaigh & Dowdeswell (2001) presents a review of processes resulting in laminated or stratified sediments in the glacimarine environment. Because no regular or cyclical color changes are observed in any of the stratified units it is unlikely that this stratification is due to annual cycles; furthermore, lead-decay derived sedimentation rates in units 953-3 and 954-3 are too low for the observed stratification to have been caused by diurnal, tidal cycles. The stratification in these clayey silts is suggested therefore to reflect changes in sediment source, possibly controlled by the distribution of the plume. As meltwater plumes sourced from different areas of the catchment enter the fjord they are distributed by winds, tides, bottom currents and the Coriolus effect (Ó Cofaigh & Dowdeswell, 2001). Changing conditions at the air-water interface, specially seasonal changes in wind patterns as suggested by Plassen et al (2004) to control color stratification of sediments in Tempelfjorden, and changing seasonal meltwater discharge and plume strength, can both contribute to shifting sediment distributions in the fjord.

Units 953-2 and 954-2 may comprise glacimarine sediments in which high ice rafted sediment delivery has partially muted the suspension fallout signal. This may indicate an environment that is more glacier-distal, where by the time the plume reaches this distal environment it has mixed with the ambient waters enough that little sediment remains suspended and the surface water is warm enough to melt passing icebergs. This scenario – clayey silts with high clast content – is described in Plassen et al (2004) from Yoldiabukta, where a high clast content in distal glacimarine sediments is attributed to high levels of sediment delivery from icebergs relative to suspension settling.

Stratified muds found in St Jonsfjorden largely lack sand, except for as thin, discrete horizons. As per Elverhøi et al (1980), it is unlikely that these sands are deposited as suspension rainout. In the glacier proximal environment, mass flow has been identified as
an emplacement mechanism for sandy horizons (Ó Cofaigh & Dowdeswell, 2001 and references therein). Figure 53 shows the movement of mass flows down slopes to emplace sandy horizons intercalated with stratified fine sediments in the proximal glacimarine environment. Because HH12-953 was taken from the out-fjord facing slope of a ridge, the sandy horizons therein are interpreted as mass transport deposits where the sandy material is sourced from higher on the ridge's flank.

![Figure 53: Deposition of sediments in glacial marine environment, not to scale. Arrows indicate suspension rainout. Figure was originally published in Ó Cofaigh & Dowdeswell, 2001](image)

**Sediment Sources**

The matrices of units 956-4, 954-3, 953-3 and the entirety of gravity core HH12-955 are all suggested to have their provenance in the post-Devonian sedimentary rocks of the inner fjord.
Unit 956-2 has a distinct color, texture and geochemical signature when compared to the other units suggested to be of glacimarine origin (dominated by suspension rainout). Clayey muds in this unit are suggested to be of a local source and comprise debris abraded from the Neoproterozoic basement rocks of the outer fjord. The clasts of unit 956-1 may represent IRD deposition coming from the inner fjord, based on similarities between the geochemistry of this unit and that of the cores taken from the inner fjord.
Figure 54: Meltwater plumes entering the fjord in front of Devikbreen, Paulbreen, Vegardbreen and Osbornebreen (A) travel east toward the fjord mouth. The sediment concentration of the plume at the surface is lower by the time it reaches Konowfjellet and Piriepynten (B); additional sediments can be seen entering the fjord in front of Gaffelbreen. Photo: Norwegian Polar Institute. Summer 2009.
Ice Rafting

Lithified clasts in the units 956-4, 956-2, 954-3, 953-3 and the entirety of gravity core HH12-955 greater than sand size are all interpreted as being ice rafted. The processes by which sediments are incorporated into ice are numerous and complex and an inexhaustive overview is shown below as Figure 55. Shorefast sea ice freezes in beach sediments, fine sediments suspended in the water column can become trapped in the fjord’s seasonal ice cover and icebergs incorporate both sediments abraded and plucked from the glacier bed as well as supraglacial sediments transported by wind or avalanches on the confining valley walls (Hambrey 1994; Vorren et al. 1983 in Ehlers 1983).

Sediments incorporated into ice are deposited primarily by melt-out – where the ice melts, resulting in the release of frozen-in sediments – and by dumping, where the ice becomes gravitationally unstable and overturns, releasing the sediments which had rested on the surface (Vorren et al. 1983 in Ehlers 1983). These methods of sediment release are shown in Figure 56. Mounds of sediments deposited via dumping can be up to two meters high (Benn & Evans 2010). In order for sediments to melt out of ice bodies the surrounding water must be sufficiently warm. For this reason, sediments deposited in glacier proximal environments where the surface water is very cold may not show many ice rafted clasts.
Furthermore, deposition from meltwater plumes is expected to be highest closest to the efflux point; this sedimentation can drown out ice rafting-derived sedimentation close to the termini of tidewater glaciers (e.g., masking of the IRD signal by suspension rainout in glacier proximal deposits from Tempelfjorden in Plassen et al 2004).

Figure 56: Illustration of modes of iceberg sedimentation, from Vorren et al 1983
Subglacial Bedforms

Within this thesis, all streamlined features oriented along the axis of the fjord are interpreted as subglacial features where fast moving ice has transported, deposited and reworked sediments while flowing toward the mouth of the fjord. They are taken to be indicative of fast ice flow in the out-fjord direction. On the basis of their singlebeam (NHS) bathymetric survey, Król et al (2010) have previously identified both glacial lineations and flutes in the outer fjord, and their interpretation of landforms in the outer fjord is presented alongside the multibeam data presented in this thesis as Figure 57.

Figure 57:

Top: Multibeam bathymetry showing subglacial bedforms in outer St Jonsfjorden.

Bottom: Singlebeam bathymetry with interpretations of subglacial bedforms in St Jonsfjorden reprinted from Król et al 2010.

FwB indicates flutings with initiating boulders, GL signifies glacial lineations.


Glacial History of St Jonsfjorden

Deglaciation

As a former inter-ice area with a complex record of emergent beaches, the configuration of the Svalbard-Barents Sea Ice Sheet in the vicinity of St Jonsfjorden has been the subject of considerable discussion over the past half century (e.g., Evans & Rea 2005; Forman 1989).

The presence of drumlins, glacial lineations and annual retreat moraines in the outer fjord (Figure 18, Figure 19 and Figure 20) supports the interpretation that grounded ice has at one point occupied the outer fjord. The extension of ice beyond the fjord mouth is indicated by north-south oriented glacial lineations in Forlandsundet, between St Jonsfjorden and Isfjorden. Furthermore, ice-contact sedimentary deposits such as terminal moraines or grounding line wedges are conspicuously lacking in the outer fjord, suggesting that ice was never perched here for an extended period of time. This implies rapid breakup of the ice as it retreated from Forlandsundet to the onset of the annual retreat moraines.

A radiocarbon age from foraminifera (N. labradorica) at the top of Unit 956-2, interpreted as glacimarine muds, indicates that the fjord mouth was free of grounded ice by 12,625 cal BP (this study). Independent carbon counts from carbonate material leached from this sample and the main fraction, however, give different activities. Higher activities in the leached fraction indicate that surface contamination may have caused this sample to date artificially young (L. Wacker, personal communication); the 12,720 ± 95 cal BP date is therefore taken as an absolute minimum age for deglaciation of the outer fjord as these sediments are likely older. Forman (1989) determined that glaciers in St Jonsfjorden had begun to retreat from their maximum positions at or before 10,500 cal BP, and had retreated significantly by 9,500 cal BP, exposing open water in the outer fjord (Mya truncata, laboratory code DIC 3055, from the north shore of the fjord west of Ankerbreen). The radiocarbon date at top of unit 956-2 (this study) extends the minimum age of deglaciation by at least two thousand years.

At the location of gravity core HH12-956, the sediment package extends 0.0047 seconds (twtt) beneath the base of the mass transport deposit. This corresponds to over 10.6
meters of sediment, using the average p-wave velocity of units 956-1 and 956-2, 1582 meters per second, as obtained by the multisensor core logger. If the sedimentation rate over the intervals 953-1 and 953-2 averages the same as 956-4, then the 10.6 meters of sediment should represent approximately 25.5 ka of sediment accumulation.

Forman (1989) notes glacimarine sediment thicknesses of 3-5 meters in emergent marine sections on the shores of the fjord. In the fjord’s interior these sediment packages are considerably thicker. If grounded, fast-moving ice flowing through the fjord removed all sediments predating the Last Glacial, as suggested by Elverhøi et al (1995) for Isfjorden, which also housed fast flowing grounded ice during the last glacial maximum (Elverhøi et al 1995; Forwick and Vorren 2010), then these sediments comprise glacimarine sediments that postdate the onset of deglaciation and potentially also, in the lowest portion of the sediment package, ice-contact sediments deposited at the end of the last glacial. Without a second date from unit 956-2 it is not possible to estimate a sedimentation rate for this interval.

The distinctive grey color of these sediments, in contrast with the dominantly brown and red sediments composing the other stratified mud units, may indicate that these sediments are derived, at least in part, from the pre-Devonian basement rocks of the outer fjord and delivered to the fjord basin by local valley glaciers such as Bullbreen and Ankerbreen. It is also possible that catchment boundaries were not the same 13,000 years ago and that this ice body transported sediments from points further east which now lie outside the catchment limits. Low amounts of ice-rafted debris can indicate that these sediments were deposited in an environment where high meltwater production and associated sediment rainout related to the rapid disintegration of the ice stream drowns out the IRD signal.

**The Early Holocene**

As discussed above, deglaciation of the fjord was underway by 12,625 cal BP. Shells (*Mya truncata*) collected from sublittoral sediments underlying beach gravels at Piriepynten (3.5 masl) by Forman (1989) yielded a radiocarbon date of 7,840 ± 90 cal BP (laboratory number DIC-3056). If these shells are in situ and lived in a littoral environment nearly
8,000 years ago this indicates that the area between Piriepynten and Konowfjellet was free of grounded ice by this time.

Gravity cores in the outer fjord lack sediments deposited in the Early Holocene. In gravity core HH12-956 these might have been removed during the mass transport event that emplaced unit 956-2, sometime between 12,720 ± 95 and 3850 ± 100 cal BP. The radiocarbon date obtained from the Nonionella labradorica sample above the mass transport deposit is used here as a maximum age for the mass transport deposit(s) only because it postdates the age acquired from the Cibicides lobatulus specimens (Table 10). In nearby fjords, warm oceanographic and climatic conditions with reduced ice rafting (Isfjorden, Rasmussen et al 2012; Billefjorden, Baeten et al 2010) and low sedimentation rates are typical of the early Holocene (e.g., Isfjorden, Forwick & Vorren 2009; Van Mijenfjorden, Hald et al 2004 and 2007).

Late Holocene (4 ka BP to present)

In the outer fjord, unit 956-1 is interpreted as distal glaciomarine sediments deposited in the past 4,000 years on the basis of the radiocarbon dates of 3850 ± 100 and 4070 ± 105 cal BP obtained from the base of this unit (Table 11). Between approximately 4 and 2 ka cal BP low clast content, as shown in the sediment log in Figure 46, is interpreted as suppressed IRD delivery in a cool climate following the Holocene Climate Optimum.

Gravity core HH12-955 represents glaciomarine sedimentation through the past 3 ka (Figure 41). These sediments resemble those of 956-4 in color and texture and were likely deposited under similar conditions. Sedimentation rates in HH12-955 are higher than those in HH12-956, taken together with the location of HH12-955 nearly 7 km in-fjord this likely indicates that these sediments were deposited in a more glacier proximal environment.

The Little Ice Age

On the basis of single-beam echosounder data, the broad shallow ridge between Piriepynten and Konowfjellet has been interpreted (by Król et al 2010) as the subsea extension of moraines found on either side of the fjord. These subaerial moraines were previously interpreted by Preisner (1988, in Król et al 2010) as delimiting the LIA
maximum extent of the Osbornebreen-Konowbreen Glacier Complex (Preisner 1988 in Król et al 2010). Bedrock knobs visible onshore at Piriepynten suggest that this topographic feature may be a bedrock ridge or riegel (Evans and Rea 2005). Furthermore, bedrock and structural maps of western Spitsbergen (Bergh et al 2003; npolar.no) suggest that one or many faults may cross the fjord at this location.

A lack of features associated with grounded ice (subglacial bedforms, terminal moraine, recessional moraines) visible on the seafloor between the ridge crest and Zone F may suggest that ice was only grounded at the sides of the fjord, on land and/or outside the limits of the multibeam survey. Between zones E and F recessional moraines are only found in the 250 meters nearest the northern margin of the dataset in water depths less than 80 meters (see Figure 16). However, in the chirp data (see Figure 22) distortion on the shallow portions of the ridge (indicated by arrows) banks may be buried indicators of ice grounding. This is not to say that the distortion is not caused by bedrock features or the fjords geometry; it very well might be.

The interpretation of the ridge as bedrock feature is supported by the shallow depth of the acoustic basement (rarely over 0.0025 seconds owtt, or approximately 4 m). The stratified nature of the reflection package on the lower flank of the ridge (Figure 58-B) could indicate that at least part of the package is glacimarine sedimentation meltout from the underside of floating ice.

The terrestrial moraine complex at Piriepynten (southern landward extension) is described by Evans and Rea (2005) as comprising glaciotectonized muds; In situ *Mya truncata* shells recovered from glaciotectonized muds just east of Piriepynten yield a radiocarbon date of 480 ± 60 uncalibrated radiocarbon years before present (GU8069); as these shells are transported from the inner fjord they are interpreted as the maximum age of the glacier complex’s advance to this point (Evans and Rea 2005). *Mya truncata* is an extant, infaunal species of marine bivalve mollusk (saltwater clam) which burrows several centimeters beneath the sediment-water interface, while extending an intake siphon to feed on material suspended in the lowermost portion of the water column. These clams can live in excess of 70 m water depth and are found in coastal regions throughout the north Atlantic.
Figure 58: Crest and shallow eastern flank of the broad shallow ridge between Konowfjellet and Piriepynten. Arrows indicate distortion suggested to be caused by buried former pinning points or bedrock features.

The average distance between the inner fjord’s annual recessional moraines is approximately 50 meters, indicating retreat of 50 meters per year. From historical
observations of the glacier front an annual retreat rate of 46 meters per year can be calculated over the past century. Assuming that ice had retreated at a rate of 50 meters per year from the ridge crest, four known historical positions of the Osbornebreen-Konowbreen ice front were used to calculate a very rough estimate of the onset of the glacier complex’s retreat from this point. Using this method, the onset of ice retreat is expected to have occurred between 1820 and 1860 AD. In contrast, the average lead decay derived sedimentation rate for the upper portion of this core, 0.3 cm per year places the boundary between units 954-2 and 954-3 at approximately 640 cal BP (Figure 36). If the upper seismoacoustic unit (see Figure 50; lithologic units 954-2 and -3) is taken to represent glacimarine deposition since retreat from the Little Ice Age maximum sedimentation would have had to have been up to six times higher during that retreat. This view is rejected because sedimentation in the extremely glacier-proximal glacimarine environment can reasonably be expected to be higher than after the glacier has retreated several kilometers, as discussed in Elverhøi (1983).

The Little Ice Age on Svalbard was asynchronous (Werner 1993), however is associated with an abrupt rise in temperature at the beginning of the 20th century (Svendsen & Mangerud 1997). A lack of landforms associated with grounded ice on the of annual retreat moraines between the ridge crest and zone E and Because photographs from St Jonsfjorden depict the 1909 ice margin as existing over 4 km east of this point (Figure 59)

An additional radiocarbon date of 3,480 ± 50 cal years BP (laboratory number GU8071), also from redeposited Mya truncata shells, was interpreted by Evans and Rea (2005) to have been deposited during this same Osbornebreen advance but on the north side of the fjord, just west of Smalgangen glacier. The presence of this older clam in the diamict implies that at least some portions of the inner fjord, east of Piriepynten, was ice free. A radiocarbon date of 2030 ± 110 cal BP, obtained from unit 954-1 (diamict east of the ridge crest) supports the hypothesis that the inner fjord was at least partially ice free prior to 2030 cal. years BP. Assuming continuous sedimentation at a constant rate of 0.2-0.4 cm/year, a rate derived from lead activity in the upper portion of this core, the unit 954-3 represents 540-1080 years of glacimarine sediment deposition, indicating that the redeposited forams were delivered after 2010 and before 480 cal BP. It is therefore
possible that the diamict unit 954-1 and the glacitectonized muds of Evans and Rea (2005) were deposited contemporaneously.

Figure 59: Recent extents of glaciers in the fjord head. 1909 limit comes from Norwegian Polar Institute archive photos from the 1909 and 1910 Isachsen Spitsbergen Expedition. 1986-1991 Osbornebreen and Konowbreen limits are from Kverndal 1991. 2009 limit is from Konig 2013. Background photograph was taken June 9, 2012 and is available from GoogleEarth, copyright 2015 DigitalGlobe. Radiocarbon dates from Evans & Rea 2005 (pink) and Forman 1989 (blue) have been calibrated using the most recent version of CALIB, except for the youngest age, which is outside the range on which CALIB is effective.

Sedimentation rates estimated from lead decay (Figure 36) indicate that the upper portion of gravity core HH12-953 was deposited at a rate greater than one centimeter per year.
Keeping this in mind, the sediments of 953-3 are suggested to represent glacimarine sedimentation since the retreat of the Osbornebreen glacier complex from its Little Ice Age maximum.

Only one surge of Osbornebreen is known to have occurred, from 1986 to 1989 (Kverndal 1991); the maximum extent of this surge is beyond the eastern limit of the bathymetric and seismic surveys. The subglacial features visible in the bathymetry of the inner fjord are suggested to have been deposited during the ice front’s advance to the bedrock-controlled ridge between Piriepynten and Konowfjellet. In the seismic lines, no evidence exists to suggest that any of the major ridges overlie each other. For this reason, all transverse ridges – both major ridges and annual recessional moraines – are suggested to have been deposited since the Little Ice Age retreat from the bedrock-controlled morainal bank between Piriepynten and Konowfjellet.
Conclusions

- Streamlined subglacial bedforms – drumlins, flutes and glacial lineations – in outer St. Jonsfjord and Forlandsundet confirm that fast flowing ice moved through St Jonsfjorden during the last glacial
- Grounded ice was absent from the outer fjord by 12,625 cal BP
- The inner fjord has been at least partially ice-free during much of the Holocene
- Morainal banks in the inner fjord (zones A – E) were likely deposited in the inner fjord during the last deglaciation and subsequently drumlinized during Holocene readvance(s)
- Ice advanced to the broad shallow ridge between Konowfjellet and Piriepynten at least once within the past thousand years, including during the Little Ice Age, although it was not necessarily grounded in the deeper parts of the fjord at this time
- The Osbornebreen glacier complex retreated from its Little Ice Age maximum at the end of the 19th century, forming annual retreat moraines in the fjord’s inner 4 km approximately 50 meters apart
- Sedimentation in St Jonsfjorden has occurred primarily via suspension settling since 4 ka cal BP, with a significant contribution from ice rafting
- Prior to 4 ka cal BP sediment delivery via icebergs and sea ice was likely enhanced, creating the thick sediment packages observed in the outer fjord
- Sediment reworking and mass wasting also contribute to sedimentation in St Jonsfjorden
- IRD at all stratigraphic levels confirms the presence of tidewater glaciers continuously since 4 ka cal BP
- Tidewater glaciers may have been present in the fjord during the entire postglacial period, including the Holocene Climate Optimum, but a gap in the sediment core record between 12720 ± 95 and 4070 ± 105 cal BP prevents this from being confirmed
- Increased IRD deposition since 1.7 ka cal BP in the outer fjord and since 1 ka cal BP in the middle fjord is suggested to be related to increased glacier activity in the fjord head and/or local valley glaciers
- Due to the great thickness of the underlying stratified seismoacoustic sediments in the outer fjord, grey, fine-grained sediments in unit 956-2 (the 70 cm beneath 12720 ± 95 cal BP) very little IRD are suggested to be deposited largely by suspension settling from local, (glacio-)fluvial sources at a time when iceberg production or transport was suppressed after the deglaciation of the fjord mouth in
addition to meltwater produced in association with the retreat of the Osbornebreen glacier complex

- Despite its widespread use in paleoenvironmental reconstructions, the shallow infaunal foraminiferal species *Nonionellina labradorica* may not be an ideal choice for radiocarbon dating because it consistently returns dates that are younger than the sediments from which it is recovered; in high sedimentation environments such as fjords this presents less of a problem than in low sedimentation environments such as further out on the continental shelf where these organisms can easily penetrate thousands of years into the sedimentary archives
References


EdgeTech 2015. 3300 Hullmount sub-bottom profiling system brochure. EdgeTech.com


Forman, S. 1989. Late Weichselian glaciation and deglaciation of Forlandsundet area, Western Spitsbergen, Svalbard. Boreas 18: 51-60


Hogan, K. A., Dowdeswell, J. A.; Noormets, R., Evans, J., & Ó Cofaigh, C. 2010 Evidence for full-glacial flow and retreat of the Late Weichselian Ice Sheet from the waters around Kong Karls Land, eastern Svalbard. Quaternary Science Reviews 29(25-26): 3563-3582


Jessen, S. P., Rasmussen, T. L., Nielsen, T., & Solheim, A. 2010. A new Late Weichselian and Holocene marine chronology for the western Svalbard slope 30,000-0 cal years BP. Quaternary Science Reviews 29: 1301-1312


Lønne, I & Lyså, A. 2005. Deglaciation dynamics following the Little Ice Age on Svalbard: Implications for shaping of landscapes at high latitudes. Geomorphology 72: 300-319


Rasmussen, T. L., Thomsen, E., Ślubowska, M. A., Jessen, S., Solheim, A., & Koç, N. 2007. Paleoceanographic evolution of the SW Svalbard margin (76°N) since 20,000 14C yr BP. Quaternary Research 67: 100-114


Stuiver, M., Reimer, P. J., & Reimer, R. W. 2014. CALIB 7.1. calib.qub.ac.uk


